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The detrital record of orogenesis: A review of approaches and techniques used in the Himalayan sedimentary basins

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Abstract

The sediment archive, of material eroded from an active tectonic region and stored in adjacent basins, can provide a valuable record of hinterland tectonism especially when information in the source region itself is obscured by later metamorphism or removed by tectonism or erosion. Using the sediment record to document tectonism is a well established approach, but more recently there has been a burgeoning of the number of isotopic techniques which can be applied to detrital material, in particular single-grain analyses. Thus the scope for application of detrital studies to a number of different tectonic problems has widened considerably. In this review, the example of sediments eroded from the Himalayan orogen and preserved in the suture zone basin, foreland basin, remnant ocean basins and deep sea fans is used to illustrate the approach. Techniques as diverse as petrography, heavy mineral, XRF and Sr–Nd studies; single grain dating by Ar–Ar, U–Pb and fission track methodologies; and single grain Sm–Nd and Pb isotopic analyses, are described. The paper documents how the sediment record can be used to determine the thermal and tectonic evolution of the orogen, constrain mechanisms of continental deformation, exhumation rates and palaeodrainage.

Keywords: detrital sediment record; provenance techniques; Himalaya; orogenesis; exhumation; erosion

1. Introduction

Study of the process and product of erosion of a mountain belt is critical to our understanding of the causes and consequences of orogenesis, for which the Himalaya is taken as a paradigm.

A number of mechanisms have been proposed to accommodate convergence during the India–Asia continental collision, for example lateral extrusion (Tapponnier et al., 1982), continental underthrusting (Zhao et al., 1993) and crustal thickening (Dewey et al., 1988). A knowledge of the amount of material that has been removed from the mountain belt by erosion through time, will allow the relative importance of these mechanisms to be assessed.

The traditional view that tectonism results in erosion (a "one way" cause and effect) has more recently been reconsidered with a greater understanding of the dynamic interaction between these two processes. Where erosion rates are high, erosion rather than

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tectonic exhumation may be the dominant mechanism by which unloading and cooling of a rock is achieved in orogenic settings (Burbank, 2002). An understanding of the timing and magnitude of erosion will allow constraint to models of its influence on orogenic structure (e.g., Willett, 1999) and exhumation, as proposed in the coupled tectonic-surface process models of Beaumont et al. (2001) and Zeitler et al. (2001a,b).

Throughout an orogen's protracted evolution, numerous processes and events have combined to result in the mountain belt seen today. However, information on much of this early history is obscured in the mountain belt itself, overprinted by later metamorphism or removed by tectonism or erosion. Hence, the sediment record of material eroded from the mountain belt can provide an invaluable archive of the early development of the mountain belt.

Study of the erosive product of orogenesis contributes towards an understanding of the consequences of continental deformation as well as the causes. It has been proposed that increased erosion due to formation of the Himalaya has resulted in the rise in global marine $^{87}Sr/^{86}Sr$ values since ca. 40 Ma (Richter et al., 1992) as well as Cenozoic global cooling (Raymo and Ruddiman, 1992). Testing these hypotheses requires detailed information on the erosive history of the mountain belt.

As ever more sophisticated analytical methods have been developed, the sedimentary record has been used to document orogenic evolution using an increasing number of approaches. In the Himalaya, the often excellent exposure of the rocks and the large numbers of magnetostratigraphically dated sections, combined with the proposed importance of the mountain belt to global climate and marine geochemistry, has resulted in a relatively large number of studies. The aim of this paper is to provide a comprehensive review of the contribution to our understanding of Himalayan tectonic evolution as determined from the sedimentary record. This paper is not intended to document the evolution of the sedimentary basins and drainage in detail, for which excellent papers are already in existence (e.g., Behrensmeyer and Tauxe, 1982; Willis, 1993a; Burbank et al., 1996; Zaleha, 1997; Nakayama and Ulak, 1999; Friend et al., 2001). It is the intention that this paper should be applicable beyond the realm of Himalayan geology and document methods pertinent to a number of orogenic settings.

2. The geology of the Himalaya

The Himalaya formed when the Tethys ocean, which had been subducting beneath Eurasia, closed, and India and Asia collided. Microplates and island arcs are sandwiched between the two continents along the line of collision. The eastern and western lateral margins of the India–Asia contact zone are transpressional boundaries, formed as India continued to move north and indent further into Asia. These transpressional margins are characterised by accretionary prisms (the Makran and the Indo–Burman ranges) and strike slip faults that form the boundaries between the continents (Fig. 1).

India–Asia collision is an ongoing process. The timing of initiation of collision is constrained by a number of datasets which give varying ages from ~70 to 38 Ma and younger (Jaeger et al., 1989; Beck et al., 1995 and references therein; Rowley, 1998; Yin and Harrison, 2000; Aitchison et al., 2002). The degree of diachroneity along strike is also disputed, with estimates varying from insignificant (e.g., Searle et al., 1997) to substantial (e.g., Rowley, 1998). The approaches used to constrain the timing of the start of collision are as diverse as documenting the age of faunal mixing (Jaeger et al., 1989) to dating the youngest subduction-related magmatism (Searle et al., 1987). It is this variety of techniques, each of which represent a different stage in an ongoing collisional process, as well as the definition of collision employed, that results in the range of ages. “Initial continental contact” is most commonly defined as the start of collision (Friend, 1998) but this is difficult to date because the original northern extent of Indian continental crust is unknown.

The most commonly quoted age of collision, 55–50 Ma, is taken from a variety of data including palaeomagnetic methods which show the Indian plate velocity to have decreased rapidly around this time (Patrait and Achache, 1984; Klootwijk et al., 1992), the time of cessation of marine facies and development of flexure-related unconformities in the suture zone and Tethys Himalaya (e.g., Garzanti et al., 1987; Searle et al., 1997), and P–T–t paths of eclogites from which it is

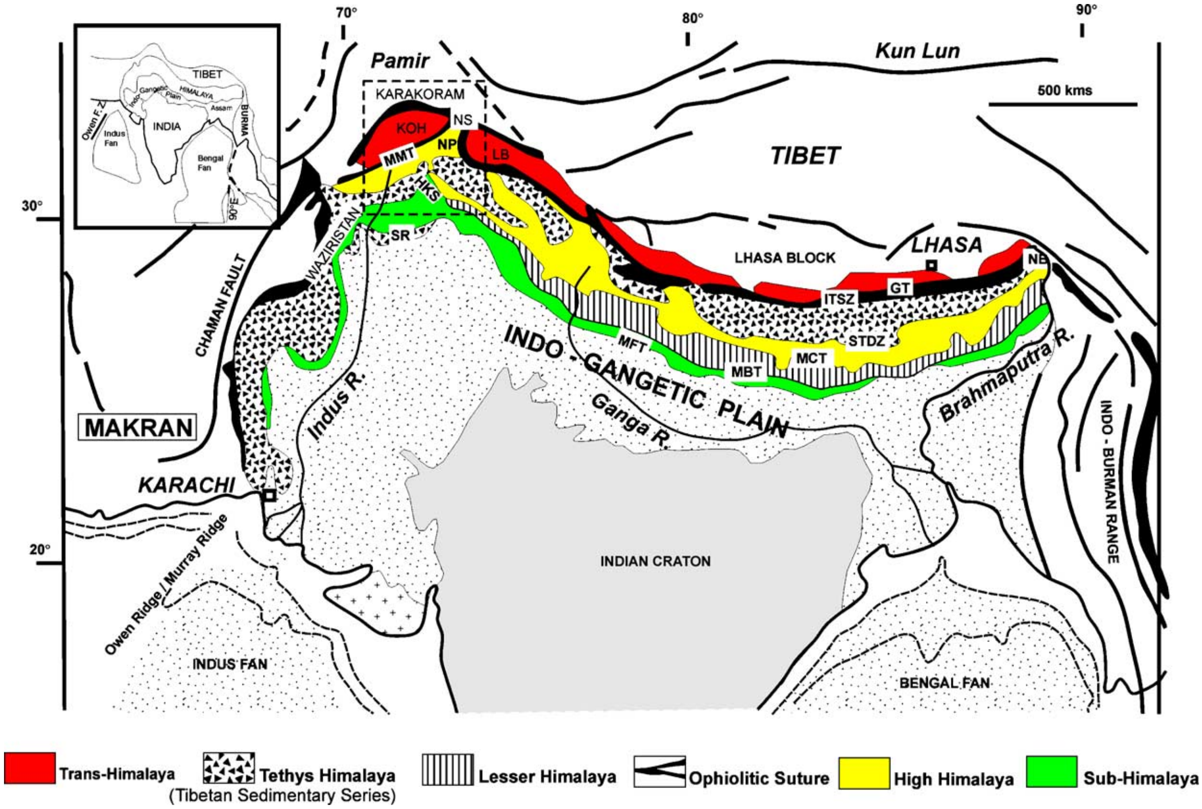


Fig. 1. Location map of the Himalayan region (additional annotations added to original map of Critelli and Garzanti, 1994). NB = Namche Barwa, GT = Gangdese thrust, HKS = Hazara–Kashmir Syntaxis, ITSZ = Indus Tsangpo Suture Zone, KOH = Kohistan Island arc, LB = Ladakh batholith, MBT = Main Boundary Thrust, MCT = Main Central Thrust, MFT = Main Frontal Thrust, MMT = Main Mantle Thrust, NP = Nanga Parbat, NS = Northern Suture, SR = Salt Range, STDZ = South Tibetan Detachment Zone. Inset: location of Himalayas in a wider geographical context. Boxed region (dashed line): shown in more detail in Fig. 2.

interpreted that initial collision in the subduction regime occurred at 55 Ma, followed by a switch to the collisional regime at 47 Ma (De Sigoyer et al., 2000). This age, and other estimates, where constrained by the sediment record, are discussed in Section 6.1.

The mountain belt consists of a series of southward-propagating thrust sheets, the consequence of crustal thickening which began soon after collision in the Eocene (Ratschbacher et al., 1994; Searle et al., 1997). From north to south the main thrusts are the Gangdese Thrust, Main Central Thrust (MCT), Main Boundary Thrust (MBT) and Main Frontal Thrust (MFT) (Fig. 1). Furthest north lies the Asian crust, which consists of a number of metamorphosed and sedimentary terrains including the Lhasa Block and Karakoram. The Trans-Himalaya represents the original Andean-type continental arc of the Asian active margin, when Tethyan oceanic crust was being subducted. The Indus Tsangpo suture zone (ITSZ), which separates the Indian and Asian crusts, contains sedimentary rocks, melange and ophiolitic material. The timing of ophiolite obduction onto the northern Indian passive margin is disputed, either Late Cretaceous obduction with Eocene rethrusting associated with collision (e.g., Searle et al., 1997), or Palaeocene–Early Eocene obduction (e.g., Garzanti et al., 1987; Gaetani and Garzanti, 1991).

South of the suture zone is the Tibetan Sedimentary Series (TSS), a Palaeozoic–Eocene sedimentary succession which was deposited on the northern passive margin of India (e.g., Gaetani and Garzanti, 1991). To the south, the metamorphosed Indian plate rocks of the

Higher Himalaya are separated by the normal-faulting South Tibetan Detachment Zone (STDZ) from the Tibetan Sedimentary Series above, and by the Main Central Thrust from the Lesser Himalaya below. Since collision inception, the Higher Himalaya have been subjected to Barrovian metamorphism. Burial and heating continued until at least 25–30 Ma, culminating in crustal melting and emplacement of leucogranites, aged mostly between 23–12 Ma (Hodges et al., 1996; Searle, 1996; Searle et al., 1997; Edwards and Harrison, 1997; Vance and Harris, 1999; Godin et al., 1999; Prince et al., 1999; Foster et al., 2000; Simpson et al., 2000). Some workers portray the metamorphism as an early Mid Eocene–Early Oligocene Eo–Himalayan Barrovian phase and the culmination as a later Neo-Himalayan phase (ca. 25 and 15 Ma). The culmination of metamorphism is co-incident with movement along the MCT and STDZ, the structures along which the High Himalaya exhumed, leading to rapid cooling. More recent Late Miocene–Pliocene reactivation of the MCT is also proposed but disputed (e.g., Harrison et al., 1997; Robinson et al., 2003). Two additional aspects of the metamorphosed Himalayan unit are of particular note: the syntaxial regions of Nanga Parbat to the west and Namche Barwa to the east, which show unusually young metamorphism (<10 Ma) and extremely rapid recent exhumation (Treloar et al., 1989; Zeitler et al., 1993; Burg et al., 1997; Ding et al., 2001), and the "inverted metamorphism" of the MCT Zone and Higher Himalaya (e.g., Le Fort, 1975; Searle and Rex, 1989).

South of the Higher Himalaya lies the Lesser Himalaya. This unit consists of unmetamorphosed or low-grade Indian crust material of mostly Precambrian to Palaeozoic age (Valdiya, 1980; Tewari, 1993; Frank et al., 1995; Hodges, 2000). A further subdivision of this unit into Inner and Outer Lesser Himalaya has been made on the basis of geochemical differences (Ahmad et al., 2000; see Section 4.6.1). South of the Lesser Himalaya lie the Sub-Himalaya foreland basin sedimentary rocks (Section 3.2), separated from the Lesser Himalaya by the MBT, active at ca. 10 Ma (e.g., Meigs et al., 1995).

A slightly different configuration exists in Pakistan (Fig. 2). Here the suture zone diverges around the Kohistan Island Arc (Khan et al., 1993). The Northern Suture represents the line of collision between Kohistan and Asia, which occurred prior to India–Asia

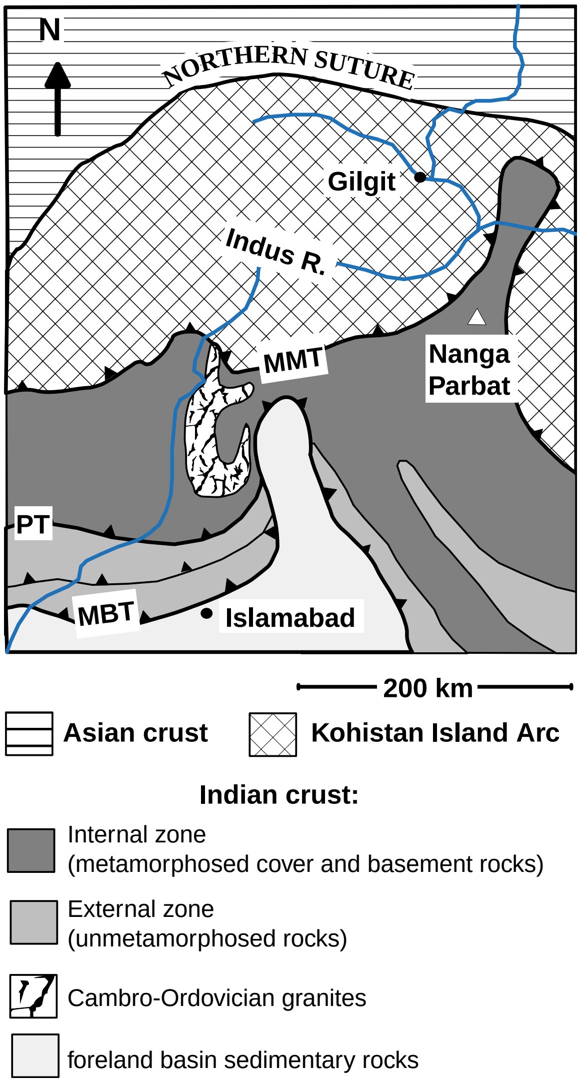


Fig. 2. Map showing the geology of the boxed region (dashed line) in Fig. 1 in more detail. MMT = Main Mantle Thrust, PT = Panjal Thrust, MBT = Main Boundary Thrust.

collision. The Main Mantle Thrust represents the line of subsequent collision between the combined arc plus Asian plate with the Indian plate (Treloar et al., 1989). The Indian crust can be subdivided into the internal metamorphosed zone and the external unmetamorphosed zone (Treloar et al., 1991). The metamorphosed zone consists of thrust imbricated basement and cover sequences, with only the cover affected by Himalayan metamorphism. Mineral ages may indicate earlier metamorphism compared to the east, continuing to Early Miocene times (Treloar and Rex, 1990; Smith et al., 1994; Chamberlain and Zeitler, 1996; Schneider et al., 1999; Foster et al., 2002). The metamorphosed zone is separated from the unmetamorphosed zone below by the Panjal Thrust, and the Kohistan arc above by the Main Mantle Thrust (MMT). The MMT may also have acted as an extensional fault in the Miocene or earlier (Burg et al., 1996; Argles and Edwards, 2002), resulting in rapid cooling and tectonic exhumation of the Indian crust metamorphic rocks.

The total convergence across the entire orogen, (including the Himalaya but also further north, e.g., Tibet), since the start of collision is estimated between 1400 and 3215 km (Molnar and Tapponnier, 1975; Dewey et al., 1989; Lepichon et al., 1992; Hodges, 2000; Yin and Harrison, 2000; Guillot et al., 2003). Proposed mechanisms by which this convergence has been accommodated include: distributed shortening followed by lithospheric delamination (Dewey et al., 1988); crustal underthrusting of Indian crust beneath Eurasia (Zhao et al., 1993); and lateral extrusion whereby material is pushed out towards the east along major transform faults (Tapponnier et al., 1982). These models differ in the timing and rates of major events. For example, lithospheric delamination late in the orogenic process would have resulted in relatively recent rapid uplift of Tibet. Yet there is a lack of consensus on the timing of uplift and the development of significant topography of the Himalayan–Tibet region. Suggested timing of uplift ranges from pre-collisional (Murphy et al., 1997) to Late Miocene–Early Pliocene (Kirby et al., 2002). This diversity in timing no doubt reflects, to some extent, the different areas studied, since regions of Tibet may well have uplifted at different times (e.g., Tapponnier et al., 2001). This paper demonstrates how the sediment record can document such events.

3. The sedimentary repositories of the Himalaya; (Table 1)

Sediments eroded from the mountain belt have been deposited in the marine depositional environments of remnant ocean basins and deep sea fans, and in sedimentary basins on land of which the major depocentres are the foreland basin and suture zone (Fig. 3).

3.1. The suture zone and Tethys Himalaya

The suture zone marks the contact between the continents of India and Asia. Continental collision resulted in the evolution of the marine forearc to an intermontane setting. Most information on the suture zone molasse has been gathered from its western end, in Ladakh, with more recent studies encompassing the eastern region in south Tibet.

Stratigraphy in the suture zone (Table 1) is complicated by a variety of local names and a paucity of age data (e.g., Van Haver, 1984; Garzanti and Van Haver, 1988; Searle et al., 1990; Sinclair and Jaffey, 2001; Clift et al., 2001a; Aitchison et al., 2002). In NW India, the Chogdo Formation is the first continental sedimentary formation that lies stratigraphically above a wholly marine forearc succession. Its facies are alluvial and its age is constrained by the Maastrichtian rocks below and Nummulitic limestones above, dated by O. Green (cited in Sinclair and Jaffey, 2001, and Clift et al., 2001a) at 54.9 Ma. Above these limestones, the rocks are wholly continental and age data are sparse. The Nurla Formation overlies the Nummulitic limestone and is overlain by the Choksti Conglomerate/Formation. Above lies the Nimu Formation. The enigmatic Hemis Conglomerate has been variously regarded as lying above (Sinclair and Jaffey, 2001) or correlated with (Clift et al., 2001a) the Choksti Conglomerate, or as stratigraphically equivalent to or below the Nurla Formation (Van Haver, 1984). Since the Hemis Conglomerate is tentatively dated on the basis of fossil evidence at Upper Eocene–Lower Oligocene, it is unfortunate that its stratigraphic position is so unclearly defined. The only other age constraints are provided from detrital apatite fission track data. Interpreted unreset ages of 35.7 Ma and reset ages of 14.3 Ma in the Nimu Formation provide maximum and minimum constraints for this unit, whilst reset ages of 12 Ma in apatites in the Choksti Conglomerate provide a minimum age for these host sediments (Sinclair and Jaffey, 2001; Clift et al., 2002a). Interpreted facies are alluvial for the Nurla Formation and Choksti conglomerate, and lacustrine for the Nimu Formation (Garzanti and Van Haver, 1988; Sinclair and Jaffey, 2001).

In the east, in southern Tibet, the conglomerates of the Kailas, Qiuwa, Dazhuqu and Luobosa Forma-

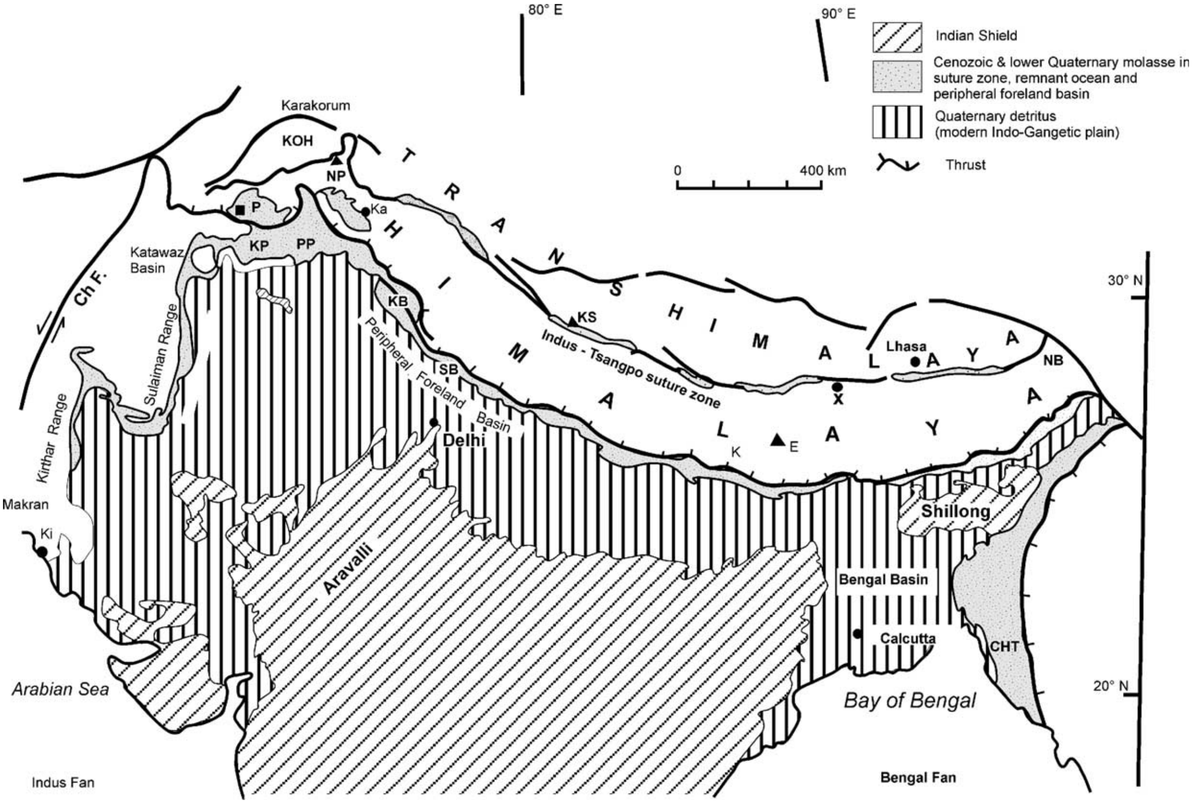


Fig. 3. Main sedimentary repositories of detritus shed from the Himalaya as described in text. Basins of the Indus–Tsangpo suture zone, peripheral foreland basin (including Kangra Sub-basin (KB), Subathu Sub-basin (SB), Kohat Plateau (KP), Potwar Plateau (PP)), Bengal Basin, Chittagong Hill Tracts (CHT) Katawaz remnant ocean basin, Sulaiman and Kirthar Ranges, Makran accretionary prism, Indus and Bengal Fans. Other abbreviations: E — Everest, K — Kathmandu, Ka — Kashmir Basin, KOH — Kohistan arc, X — Xigase, Ki — Karachi, Ks — Kailash, NP — Nanga Parbat, P — Peshawar. Annotations added to base map from Le Fort (1996).

tions are grouped together into the Gangrinboche conglomerates by Aitchison et al. (2002). These conglomerates are of alluvial and braid-plain facies (Yin et al., 1999; Wang et al., 2000), unconformably overlie rocks of the Lhasa terrain from Kailas to Namche Barwa, and are dated at Lower Miocene on the basis of structural, fossil and indirect geochronological evidence (Harrison et al., 1993; Aitchison et al., 2002). Fossils are of Oligo–Miocene age and depositional contacts occur with underlying plutons that had completed their exhumation by 23 Ma. Minimum ages are constrained by cross-cutting dykes dated at 18 Ma, and thrusts on which movement is bracketed between 18–10 Ma. With the current level of stratigraphic information, it would be imprudent to attempt correlation with the conglomerates further west at this stage. Older than the Gangrinboche conglomerates are the Liuqu conglomerates, poorly dated at Palaeogene age (Davis et al., 2002). The conglomerates lie on overturned Mid Aptian rocks. The age of structural inversion is tentatively placed at 70 Ma. Felsic dykes, dated at 18 Ma, cross-cut thrusts which disrupt the conglomerate, thereby placing a minimum age on these sediments. Plant fossils are found in the conglomerates that have affinity with Palaeogene floras from outside Tibet. These conglomerates are of subaqueous and terrestrial facies. Depositional contacts have been observed with the underlying Indian plate, ophiolitic terranes and subduction complexes.

South of the suture zone, post-collisional sedimentary rocks are preserved in a few places in the Tethys Himalaya. Whilst the Tethys Himalaya cannot be considered a major sedimentary repository, brief mention

Table 1
Stratigraphic chart of major Himalayan sediment repositories

Age	Suture zone		Distal and Remnant ocean basins			
	Zanskar	S. Tibet	Katawaz basin	Makran accretionary prism	Sulaiman and Kirthar ranges	Bengal basin
L. Pliocene						Dupi Tila Fm
E. Pliocene				Makran Gp		Tipam Fm
L. Miocene					Siwalik Gp	Boka Bil Fm
M. Miocene						Bhuban Fm
E. Miocene	Nimu Fm	Gangrinboche congl	Khojak Fm			
L. Oligocene					Vihowa	
E. Oligocene				Panjgur Fm	Chitarwata	Barail Fm
L. Eocene						
M. Eocene	Choksti congl					Kopili Fm
E. Eocene	Nurla Fm	Liuqu congl	Nisai Fm		Kirthar Fm	Sylhet Lst
L. Paleocene	Nummulitic Lst	(not in contact with			Gazij Fm	Cherra Fm
E. Paleocene	Chogdo Fm	Gangrinboche)				

Age	Peripheral foreland basin				Deep sea fans	
	Pakistan	India (Subathu sub-basin)	India (Kangra sub-basin)	Nepal	Indus Fan	Bengal Fan
L. Pliocene	Siwalik Gp: Soan, Dhok Pathan, Nagri & Chinji Fms	Siwalik Gp	Siwalik Gp: Lower, Middle & Upper	Siwalik Gp: Lower, Middle & Upper	Indus Fan marine sediments	Bengal Fan marine sediments
E. Pliocene						
L. Miocene						
M. Miocene	Kamlial Fm	Kasauli Fm	Dharamsala Fm	Dumre Fm		
E. Miocene		Dagshai Fm				
L. Oligocene	Murree Fm					
E. Oligocene						
L. Eocene						
M. Eocene						
E. Eocene	Patala / Gazij Fm	Subathu Fm		Bhainskati Fm		
L. Paleocene						
E. Paleocene						

Diagonal line indicates a period of time represented by an unconformity. Dashed line implies maximum depositional age for that formation (determined from detrital mineral ages). Sources referenced in text.

is given here to rocks of the Chulung La Formation in NW India (Critelli and Garzanti, 1994), and the Pengqu (Wang et al., 2002) or Shenkeza and Youxia Formations (Zhu, 2003; Zhu et al., 2005) in southern Tibet, since they are often used to constrain the time of collision. The Chulung La Formation consists of red beds of deltaic facies. Its age is usually given at Lower Eocene (Gaetani and Garzanti, 1991). However, since this age assignment is based only upon its stratigraphic position unconformably above the Dibling Limestone and its assumed but unproven correlation with the Kong Slates of that age (Garzanti et al., 1987, 1996), this age can only be viewed as a robust maximum. In southern Tibet, the Qumiba section has received considerable attention since identification of fossils as young as Priabonian. According to Wang et al. (2002), the sandstones and mudstones of the Pengqu Formation sits conformably above the Upper Danian to Lower Lutetian Zhepure Shan limestones. They divide the Pengqu Formation into a lower green coloured Enba member and an upper red coloured Zhaguo member, both of marine facies. Wang et al. biostratigraphically date the Pengqu Formation at upper Lower Lutetian to upper Priabonian (~47–34 Ma) and consider the transition between members to be transitional. By contrast, Zhu (2003) and Zhu et al. (2005), who call this section the

Shenkeza section, record fossils consistent with P8 (50.5 Ma) in the green facies. They also note the presence of an unconformity between the green and red beds, and consider the red bed facies to be continental, the marine fossils within it as reworked, and thus the red beds may be considerably younger than the age ascribed by Wang et al. In view of this, they consider the red and green beds to be different formations, naming the marine green beds as the Youxia Formation, and the overlying continental red beds as the Shenkeza Formation.

3.2. The peripheral foreland basin

Peripheral foreland basins form in collisional belts in response to flexural loading of the lithosphere by the adjacent thrust belt (Beaumont, 1981; Jordan, 1981). The Himalayan peripheral foreland basin stretches east–west along the length of the orogen, flanking its southern margin from eastern India to Pakistan. In contrast to the main arc of the Himalaya, the axis of the foreland rotates at its eastern and western transpressional margins, such that it is approximately N–S and parallel to the bounding ranges. The north–south-trending basins are discussed in Section 3.3.

Along-strike over the entirety of the east–west-trending peripheral basin, rocks are broadly uniform. Palaeocene–Eocene marine facies are separated from Oligo–Miocene and younger alluvial strata by a basin-wide unconformity. The Palaeogene alluvial rocks go by a number of regional names, whilst the overlying Neogene deposits are uniformally termed the Siwalik Group. The Siwalik Group has been lithostratigraphically divided into the Lower, Middle and Upper Siwalik sub-groups, broadly identified as representing significantly mudstone facies, predominantly sandstone facies, and substantial input of conglomeratic facies, respectively, although the first conglomeratic facies are found in the Middle Siwalik sub-group.

The majority of the foreland basin rocks are located in the Sub Himalaya, but Palaeogene strata also occur in the Lesser Himalaya in some regions, with significant outcrop in Nepal.

3.2.1. Pakistan

The East–West trending peripheral foreland basin consists of marine facies of Palaeocene–Eocene age, with a plethora of local names (e.g., Shah, 1977; Pivnik and Wells, 1996). These marine facies are unconformably overlain by continental facies of the Murree Formation. The Murree Formation encompasses the Balakot Formation in the Hazara–Kashmir Syntaxis to the north, and extends south into the Kohat and Potwar plateaus and east into India (e.g., Pinfold, 1918; Wadia, 1928; Gill, 1951a; Bossart and Ottiger, 1989; Singh and Singh, 1995; Najman et al., 2002a). The Balakot Formation is dated by Ar–Ar ages of detrital micas at younger than 37 Ma (Najman et al., 2001). The younger southern Murree Formation is considered to be Lower Miocene (Fatmi, 1973; Abbasi and Friend, 1989). This is consistent with their maximum depositional age determined from Ar–Ar ages of detrital white micas (taken from Murree Formation rocks at Murree Hill Station) which show a peak of ages at 22 Ma (Najman and Pringle, unpublished data). Facies are interpreted as alluvial and/or tidal (Bossart and Ottiger, 1989; Singh and Singh, 1995; Najman et al., 2002a). The Murree Formation is conformably overlain by the Kamlial Formation, magnetostratigraphically dated from 18–14 Ma (Johnson et al., 1985), which is conformably overlain by the Siwalik Group. The Siwalik Group is sub-divided into the magnetostratigraphically dated Chinji Formation (12–14 at base to 9–11 Ma at top — the range at the boundaries is the result of slightly diachronous units in Pakistan), Nagri Formation (9–11 to 7–8 Ma), Dhok Pathan Formation (7–8 to 4–7 Ma) and Soan Formation (<4–7 Ma) (Burbank et al., 1996 and references therein). The Chinji Formation broadly corresponds to the Lower Siwalik sub-group, the Nagri and Dhok Pathan Formations to the Middle Siwalik sub-group, and the Soan Formation to the Upper Siwalik sub-group.

Facies variation and a dearth of literature preclude an accurate summary of the Murree and Kamlial Formation sedimentary rocks. Detailed description of the Kamlial Formation in the Chinji village region, Potwar Plateau area shows a sandstone-rich facies, with multistoried sandstone bodies up to ca. 58 m thick indicating deposition by a large river (Stix, 1982; Hutt, 1996). Regional facies variations are significant but generally the Murree Formation appears to have a recognisably higher mudstone to sandstone ratio (ca. >50%) compared to Kamlial Formation rocks at Chinji village region. Sandstone units are also considerably thinner (<10 m) compared to Kamlial Formation strata at Chinji village, but such quantitative records only exist for the older northern Balakot Formation in the Hazara–Kashmir Syntaxis (Bossart and Ottiger, 1989; Najman et al., 2002a), and for the Murree Formation to the east at Jammu (Singh and Singh, 1995).

Above the Kamlial Formation there is a return to more mudstone dominated facies in the Chinji Formation. Chinji Formation channel sandstone beds make up about 30% of the Chinji strata, and storeys are separated by ca. 25 m of overbank fines (Willis, 1993a). The Nagri Formation heralds a dramatic return to sandstone dominated facies, with increased sandstone thickness, sediment accumulation rate, palaeochannel size and palaeodischarge. Palaeodischarge is estimated at more than double that for the underlying Chinji Formation (Willis, 1993a; Zaleha, 1997). By contrast, the Nagri–Dhok Pathan transition is characterised by a decrease in all these parameters, except for sediment accumulation rate that continues to increase. The first conglomeratic facies are found within the Middle Siwalik sub-group (Cotter, 1933; Gill, 1951b). Facies of the Soan Formation are extremely variable, with a significant proportion of conglomerate, reflecting its proximal position.

3.2.2. India

To the east in the Subathu sub-basin (Raiverman et al., 1983), the oldest foreland basin deposits consist of the shallow marine Subathu Formation facies which are predominantly green, but with minor red facies. These rocks are unconformably overlain by the continental, predominantly alluvial, red beds of the Dagshai Formation, and younger Kasauli Formation and Siwalik Group (Bhatia, 1982). Further west in the Kangra sub-basin, the alluvial strata that underlie the Siwalik Group are called the Dharamsala Formation (Chaudhri, 1975).

The Subathu Formation has been dated at upper Palaeocene–lower Mid Eocene on the basis of Nummulites and Assilines (Mathur, 1978; Batra, 1989). The Dharamsala Formation is sub-divided into Lower and Upper Members that are magnetostratigraphically dated between 21–17 and 17–13 Ma, respectively, although the base of the succession is not seen (White et al., 2001). Further east, tectonism precludes magnetostratigraphic dating of the Dagshai and Kasauli Formations. Fission track dating of detrital zircons and Ar–Ar dating of detrital micas constrains the base of the Dagshai Formation at younger than 30 Ma and the Kasauli Formation at younger than 28 Ma at two localities, and younger than 22 Ma at a third (Najman et al., 1997, 2004). The overlying Siwalik Group is divided into the Lower, Middle and Upper Siwalik sub-groups, as described above, with the boundaries falling at ca. 11.5 and 7 Ma (Meigs et al., 1995; Brozovic and Burbank, 2000).

The oldest continental facies, the Dagshai Formation, contains a significant proportion of fine-grained material, in contrast to the overlying Kasauli Formation that consists predominantly of sandstone facies. The broadly correlative pre-Siwalik rocks of the Dharamsala Formation show two coarsening-up megacycles corresponding to the Lower and Upper Members. Each cycle begins with subordinate sandstone proportions compared to mudstone and siltstone, and relatively thin sandstone beds. Further up-section, sandstone dominated facies with multistorey sandstone beds on average 5–30 m thick characterise the succession (White et al., 2001; Najman et al., 2004). The overlying Siwalik Group broadly follows the facies described above for the Lower, Middle and Upper Siwalik sub-groups of Pakistan (Brozovic and Burbank, 2000). The mudstone-dominated Lower Siwalik sub-group displays sandstones <5 m thick, rarely multistoried to thicknesses of 30–40 m, whereas the sandstone-dominated Middle Siwalik sub-group displays numerous multistoried sandstones up to 45 m thick, but mostly between 10–20 m. As in Pakistan, this boundary also marks the initiation of increased sediment accumulation. The Middle Siwalik sub-group shows the first evidence of conglomeratic facies at 10 Ma in some sections (Meigs et al., 1995), prior to the main Upper Siwalik sub-group "Boulder conglomerate" depositional stage.

3.2.3. Nepal

The oldest foreland basin strata, which overlie Cretaceous–Palaeocene passive margin facies, are the shallow marine Bhainskati Formation, dated at Lower to Mid Eocene (Sakai, 1983). The overlying continental Dumri Formation has been dated at younger than 30 Ma on the basis of detrital zircon fission track data (Najman et al., 2005), younger than 20 Ma on the basis of Ar–Ar ages of detrital white micas (DeCelles et al., 2001), and magnetostratigraphically from 21–16 Ma

(T.P. Ojha, unpublished data, cited in DeCelles et al., 2001 and Robinson et al., 2001). Siwalik sedimentation dates from 16 Ma (Gautam and Fujiwara, 2000), with the Lower–Middle Siwalik boundary taken between 9–11 Ma, and the Middle–Upper boundary at <4.5 Ma (e.g., Quade et al., 1995; DeCelles et al., 1998b; Gautam and Rosler, 1999; Nakayama and Ulak, 1999; Ojha et al., 2000). Dumri Formation facies have sandstone channels 10–40 m thick (DeCelles et al., 1998b). Siwalik facies variation broadly follows that found throughout the foreland basin. The Lower Siwalik sub-group is mudstone-dominated, the Middle Siwalik sub-group is sandstone-dominated with greater channel thicknesses, and conglomerates are characteristic of the Upper Siwalik sub-group (e.g., DeCelles et al., 1998b; Burbank et al., 1996). Sediment accumulation rates increase at 11 Ma (Meigs et al., 1995; Burbank et al., 1996).

3.3. The distal, remnant ocean, and deep marine basins

Due to irregularities in crustal margins, suturing of continents can result in diachronous development of the fold–thrust belt whilst remnant ocean basins remain (Graham et al., 1975; Ingersoll et al., 1995). Clastic sediments eroded from uplifting areas are transported transversely to peripheral foreland basins and axially to remnant ocean basins. Hence a molasse–delta–submarine fan facies continuum exists. In the case of the India–Eurasia collision, material was transported to the west by the palaeo-Indus River. Localities where the deposits of the palaeo-Indus may be found include the Western Fold Belt (Sulaiman and Kirthar Ranges) the Katawaz basin, the Indus Basin (e.g., Sulaiman, Kirthar, Pab regions), the Gulf of Oman (Makran region) and Indus Fan. To the east, the Ganges and Brahmaputra rivers deposited sediments into the Bengal Basin and Bengal Fan.

3.3.1. The western margin

The Tertiary units of the Western Fold Belt (e.g., Sulaiman, Kirthar, Pab provinces) go by a variety of regional names (e.g., see Hemphill and Kidwai, 1973; Shah, 1977), and correlation between regions, and dating is at a reconnaissance level. In the Sulaiman and Kirthar regions, Early to Mid Eocene rocks include the Ghazij and Kirthar marine shelf facies. These are overlain by marine facies of the Oligocene Nari Formation and Miocene Gaj Formation (better developed to the south). To the west in the Pab Province, the Eocene marine strata of the Kude Formation are overlain by the turbiditic sandstones of the Kohan Jhal formation (Smewing et al., 2002). In the Sulaiman ranges, the Chitarwata Formation has been correlated with both the Nari and the Gaj Formations (see discussion in Downing et al., 1993 and references therein), and its age has been considered Early Miocene (Friedman et al., 1992) or Oligocene (Welcomme et al., 2001). Its facies have been interpreted as fluvial (Waheed and Wells, 1990), deltaic (Welcomme et al., 2001) and coastal; estuarine, strandplain and tidal flat/tidal channel (Downing et al., 1993).The overlying fluvial Vihowa Formation is correlated with the Kamlial Formation, possibly slightly older on the basis of mammalian fauna, at Early–Mid Miocene age (Friedman et al., 1992), above which, Siwalik-like fluvial facies persist to present day.

The Katawaz remnant ocean basin (Qayyum et al., 1996) consists of Eocene–Oligocene deltaic and submarine fan turbidites of the Khojak Formation which overlies the shallow marine Early Eocene Nisai Formation. In the Makran the Oligocene Panjgur Formation was deposited as deep sea turbidites prior to accretion into the Makran accretionary complex (Critelli et al., 1990).

During the Palaeogene, the locus of Indus Fan deposition was further west than the present day, but some fan material was deposited in the current region of the Indus Fan (Clift et al., 2000, 2001b). Initiation of sedimentation on the Indus Fan at its current location is not easily determined because the thickness of sediment precludes drilling to the base of the fan. At distal site DSDP Site 221 the oldest fan sediments are of Upper Oligocene age (Whitmarsh et al., 1974), but sedimentation undoubtedly began earlier in more proximal areas. Two approaches have been used to date fan initiation in the more proximal regions; seismic studies and drilling on the Owen Ridge (DSDP Site 224) where later uplift saved early fan sediments from deep burial. Seismic data show a tilted Palaeogene sequence which predates Early Miocene uplift of the Murray Ridge (Clift et al., 2000) and correlation of multichannel seismic data with well data indicates that 35% of fan sediments predate the Early Miocene (Clift et al., 2001b). Significant Oligocene–Miocene deposits have also been identified by seismic studies in the offshore Indus Basin by Daley and Alam (2002). Turbidites on the Owen Ridge, interpreted as palaeo-Indus Fan deposits (Clift et al., 2001b), are of Mid Eocene age. However, provenance is disputed, with some workers suggesting an Arabian margin source (Jipa and Kidd, 1974; Mallik, 1974) — see later discussion in Section 5.3.1.

The modern Indus Fan underwent its first period of rapid sediment accumulation in the deep ocean region in the Early Miocene (Davies et al., 1995). Seismic and well data have been used to document the sediment accumulation patterns through time, with sometimes contrasting results. Increased sediment accumulations have been proposed variously during the Mid Miocene, Late Miocene and Mid Pliocene, with the Late Miocene also proposed as the time of reduced sediment supply to the fan (Rea, 1992; Metivier et al., 1999; Clift and Gaedicke, 2002). Today, the Indus Fan, although dwarfed by its giant neighbour the Bengal Fan, is one of the most major sediment bodies in the ocean basins, at volumes of $\sim 5 \times 10^6$ km^3.

3.3.2. The eastern margin

The Bengal Basin deltaic complex and deep sea Bengal Fan form a foreland basin—remnant ocean basin pair into which the Ganges and Brahmaputra rivers debouch. The Bengal Basin covers ca. 144 000 km^2 onshore and 63 000 km^2 offshore, containing sediment derived predominantly from the Himalaya, and subordinately the Indo–Burman ranges to the east and Indian shield to the west. Tertiary strata are preserved in a Shelf region in the NW, a central basin, and an eastern fold–thrust belt called the Chittagong Hill Tracts. In a study of a western region of the Bengal Basin, the timing of initiation of major deltaic sedimentation is imprecisely bracketed between 50 and 30 Ma, based on the use of seismic stratigraphy to map sequence boundaries. Lindsay et al. (1991) mapped a 49.5 Ma sequence boundary above which the large scale prograding clastic units of the major delta began. However, the authors note that there was a delay before rapid sedimentation commenced above this sequence boundary. A minimum age for this event is provided by the overlying 30 Ma sequence boundary which defines the top of this basal prograding clastic wedge of the transitional delta. Whether these earliest sediments are Himalayan or Indian-craton derived is a matter of debate (see Section 5.3.2). The first unambiguous evidence of Himalayan input is found in the Miocene (e.g., Johnson and Nur Alam, 1991; Uddin and Lundberg, 1998a).

The traditional Tertiary lithostratigraphic units of the Bengal Basin are termed the Sylhet, Kopili, Barail, Bhuban, Boka Bil, Tipam and Dupi Tila Formations. Reimann (1993) points out that confusion over the age of some of these formations has arisen from inaccurate correlation with rocks in Assam (Evans, 1932) coupled with a lack of good biostratigraphic control. The marine Sylhet Limestone and Kopili Shale are of Eocene age. Deltaic conditions persist throughout deposition of the Barail Formation and Surma Group (Bhuban and Boka Bil Formations). The Barail Formation is considered to be of Oligocene age, extending into the Miocene (Reimann, 1993), and the Surma Group of Miocene age extending into the Pliocene (Banerji, 1984; Johnson and Nur Alam, 1991). The top of the Surma Group-the Upper Marine Shale — is dated by Worm et al. (1998) at 3.5 Ma. The Tipam Formation is of Pliocene age (Reimann, 1993) and overlain by the Dupi Tila Formation. Facies are fluvial in the north and marine in the south. All rocks are predominantly arenaceous apart from the Eocene beds which are mixed carbonate and sandstone material.

Because the current dates are often poorly constrained and rely on tentative correlations over long distances and between different facies, some workers suggest that a significant revision or abandonment of the traditional lithostratigraphic units is required (e.g., Partington et al., 2002; Gani and Alam, 2003). Partington et al. (2002, 2005) (see also Cairn Energy Internal Report, 2000) correlate new biostratigraphic, well, seismic and fieldwork data with the existing lithostratigraphy, to define three regionally correlatable unconformity-bound Megasequences. Megasequence 1 correlates with the Bhuban and Boka Bil Formations, Megasequence 2 consists of the Tipam Formation and extends into the Dupi Tila Formation, whilst Megasequence 3 then continues to present day. The boundary between Megasequences 1 and 2 lies within calcareous Nannoplankton zone NN16–NN15 of Martini (1971), late Early Pliocene, whilst the boundary between MS2 and MS3 lies within NN19–20, the Early Pleistocene.

The Bengal Fan has only been drilled to depths corresponding to the Lower Miocene. Estimates of the age range between 17 and 22 Ma. Microfossil evidence indicates that the maximum age of the drilled sediment is 18 Ma (Gartner, 1990). However, this age requires a significantly higher sedimentation rate between the base of the drilled hole and the overlying dated biostratigraphic horizon compared to the accumulation rate above this level. Because of this, and the fact that microfossil preservation is poor in this succession due to deposition below the CCD, Galy et al. (1996) consider that biostratigraphic age estimates may be too young. Extrapolation of sedimentation rates from the overlying interval would date the base of the drilled section at 22 Ma, a conclusion also reached by Curray (1994).

Determination of the time of initiation of Bengal Fan sedimentation is poorly constrained. At the distal ODP Sites, Cochran (1990) suggested that the increase in pelagic horizons and carbonate content in the lowest drilled sediments, accompanied by decreasing grain-size, could indicate that the base of the fan was approached at the extent of the drilled interval. However Cochran also noted that a Palaeocene–Mid Eocene hiatus, recognised by an unconformity on seismic data, separates pre-fan from fan sediments (Curray and Moore, 1971; Moore et al., 1974) and could be interpreted as implying that fan sedimentation began after this time, although this boundary was likely time-transgressive. The earlier age for fan initiation at this distal site is supported by Curray (1994) who notes that there is no seismic record attributable to initiation of fan deposition in the Early Miocene, and that extrapolation of sedimentation rates to the interpreted base of the fan at 2 km depth suggests an approximate Mid Oligocene age for initiation. This is in accord with Davies et al. (1995) who record the start of sediment accumulation in the Bengal Fan during the Oligocene. Furthermore, both Cochran (1990) and Curray (1994) note that fan sedimentation in the more proximal northern Bay of Bengal would have begun earlier, with Curray suggesting the Early Eocene on the basis of the start of sedimentation above the regional unconformity.

4. Techniques

In this section, the background to the methodologies used on the above-described sediment record to document Himalayan evolution are discussed. Sedimentological and provenance techniques can be employed to constrain mechanisms of accommodation of strain, and as an indicator of tectonism in the hinterland. Isotopic dating of detrital minerals provides information on hinterland metamorphism, cooling and exhumation rates, as well as contributing to provenance discrimination.

4.1. Sedimentology and stratigraphy

This section includes information that can be gained on orogenic evolution and palaeo-drainage from sedimentological and stratigraphic studies. These studies include temporal and spatial facies variation of basin sediments, palaeocurrent data and sediment accumulation rates, determination of sediment volume and age, and the occurrence of unconformities. Information on orogenic development can be recognised in the sediment record by building up a picture based on any number of the above datasets, often combined with provenance data, (discussed in Sections 4.2–4.6).

Estimation of the amount of material eroded from the Himalaya and preserved in adjacent sedimentary basins has been attempted by workers including Curray (1994), Johnson (1994), Einsele et al. (1996) and Rowley (1995). The degree of difference in the results is due to the significant uncertainties associated with the calculations. Some sediment will be unaccounted for (e.g., that lost by subduction), the width and thus volume of the original foreland basins is not precisely known (i.e., the extent of continuation of sediment beneath thrust sheets or alluvium is uncertain) and the degree of mixing with non-Himalayan derived sources that may have occurred in some repositories (e.g., the Bengal Fan) has not been accurately ascertained. Some possible repositories have not been included since it is currently not known if the sediments within them are Himalayan-derived (e.g., some sediment areas in Assam and Nagaland). Similar degrees of uncertainty arise in estimations of the areal extent of the Himalaya pre thrusting, and in the estimated volumes of restored sections of the mountain belt where crustal thickness and amount of shortening is not well constrained.

Johnson (1994) undertook the most wide ranging study in terms of spatial and temporal coverage of the sedimentary repositories, and in his considerations of the significance of the results. Removal of crustal material by erosion is one mechanism by which the continued convergence of India and Asia has been accommodated. Lateral extrusion (Tapponnier et al., 1982), underthrusting (Zhao et al., 1993) and crustal thickening (Dewey et al., 1988) are others (Section 2). Comparison of the total volume of sediment eroded from the orogen, compared to the total amount of shortening, permits an estimation of the extent to which other mechanisms must be invoked to accommodate strain. Johnson (1994) calculated the total sediment volume eroded from the Himalayan orogen; that preserved in the major sedimentary repositories of the foreland basin, Ganges–Brahmaputra Delta, Indus and Bengal Fans, plus that carried in solution, to be 1.9×10^7 km^3. This figure is adjusted for compaction. By comparison with the area of the Himalaya, he was able to consider the vertical and areal extent of source area that is required to account for that sediment volume, thereby drawing conclusions on provenance and palaeodrainage (see Section 5.3.2). By volume balancing using restored sections of the Himalaya (taking values for length of the restored section between 750–1250 km and crustal thicknesses of 25–38 km), he also drew conclusions on the degree to which accommodation of convergence can be accounted for by erosion, and consequently that proportion that must be taken up by other mechanisms (see Section 6.2.3).

The sediment record is ideally suited to determining the timing of continental collision (Butler, 1995). Collisional influence has been recognised in the sediment record by a number of indicators, further discussed in Section 6.1, including an increase in subsidence expressed by a change in depositional character as well as provenance changes (e.g., Rowley, 1996), the age of the oldest strata that overlap both continents, (e.g., Clift et al., 2002a; Beck et al., 1995), and the cessation of marine facies (e.g., Searle et al., 1997). Furthermore, spatial variation in these factors record the degree of diachroneity of collision, as proposed in the Himalaya by e.g., Rowley (1996, 1998), Uddin and Lundberg (1998a) and Najman and Garzanti (2000).

Burbank and Raynolds (1988) illustrated the sedimentary responses that one might expect to see resulting from hinterland thrusting. These responses include increased sediment accumulation rates, palaeocurrent changes as drainage is diverted, facies variation including upward coarsening of sediments basinward of the thrust and ponding of sediments hinterlandward of the thrust, the development of unconformities, and changes in provenance. However, interpretation of orogenic tectonism from the timing of the sedimentary response is not simple, with progradation of coarse-grained facies attributed both to periods of uplift or quiescence. In the syn-tectonic scenario, increased sediment input occurs concurrently with tectonism whereas subsidence lags behind, and thus evidence of progradation of the sediment wedge can be taken as the timing of tectonism (e.g., Burbank et al., 1988). In the post-tectonic scenario, subsidence follows on quickly from tectonism but sedimentation lags behind (e.g., Heller et al., 1989). In this case, it is the fine-grained sections of the sediment record that record periods of tectonic uplift, when coarser-grained sediments are concentrated in more proximal regions.

As well as reflecting thrusting, as described above, unconformities on the distal cratonic side of the foreland basin can be the result of a number of events. These include passage of the peripheral forebulge cratonwards as it passes through the region of the back-bulge depozone (DeCelles and Giles, 1996); migration of the peripheral forebulge back towards the hinterland due to redistribution or rejuvenation of the load or relaxation of a visco-elastic lithosphere (e.g., Quinlan and Beaumont, 1984; Sinclair et al., 1991); and rapid uplift consequent to slab break-off due to the removal of a large load and cessation of subduction-related dynamic subsidence. A major Late Eocene–Oligocene unconformity separates marine from continental facies in the Himalayan foreland basin, as documented in Section 3.2, and the significance of this is discussed in Section 6.2.3.

In the Himalaya, sedimentological changes have been used to constrain the timing of movement of the MBT and more southerly thrusts, (Burbank and Raynolds, 1988; Meigs et al., 1995) as described in Sections 5.1.1 and 5.1.2. In addition, mapping of drainage evolution, by facies, palaeocurrent and provenance (e.g., Friend et al., 1999) can help to discriminate between tectonic vs. erosionally driven mountain uplift (Burbank, 1992) and aggradation vs. progradation in the foreland basin.

4.2. Petrography and dense minerals

4.2.1. Sandstone facies

Data from detrital modes, achieved by point-counting sandstones, often provide the first step in provenance determination. This method provides quantitative information on contributing source region lithologies for sandstones of mixed provenance, by identification of constituent minerals and lithic fragments. Traditional parameters (e.g., proportion of quartz, feldspar, lithic fragments — the latter also subdivided into sedimentary, igneous and metamorphic sub-groups) are used to construct ternary diagrams within which distinct provenance fields: recycled orogen, magmatic arc and cratonic block, are located (Dickinson and Suczek, 1979; Dickinson, 1985). More recently, an extended spectrum of key indices for lithic grains has been used, incorporating distinction according to composition (e.g., metapelite, metafelsite, metabasite) and metamorphic grade of metasediments (e.g., slate, phyllite, schist, gneiss), to allow further provenance discrimination (e.g., White et al., 2002; Najman et al., 2003a; Garzanti and Vezzoli, 2003). Dense mineral studies can bring an extra dimension to provenance discrimination. The occurrence of specific metamorphic index minerals gives information on the metamorphic grade of the source region, the proportion of ultrastables allows interpretation of the degree of sediment recycling, and a high proportion of a mineral distinctive to one region provides information on drainage evolution and exhumation history of that source.

It should be remembered that other factors apart from source region can affect petrographic parameters: weathering, recycling, diagenesis and grain-size. Grain-size influence is minimised using the Gazzi–Dickinson approach, in which the datum point for a coarse-grained fragment is assigned to the constituent mineral under the cross-hair (e.g., Ingersoll et al., 1984; Zuffa, 1985). In Himalayan strata deposited in the western regions of the foreland basin, the wealth of labile volcanic and metamorphic lithic fragments indicates that petrography has not been drastically affected by weathering or diagenesis (e.g., Najman and Garzanti, 2000; Najman et al., 2003a). There has, however, been some alteration of these unstable grains, and there is evidence of extensive carbonate replacement and some pressure solution, although framework grain dissolution is minor. The low proportion of dense minerals suggests intrastratal solution of unstable mafic material, whilst the high proportions of ultrastables in some sandstones points towards sediment recycling. In sediments deposited further east in the foreland basin of Nepal and in the Bengal Basin, the diagenetic imprint is greater.

In the Himalaya, the application of petrographic and dense mineral techniques to the sediment record has been used to constrain the timing of collision (Section 6.1.4), tectonics, exhumation and drainage history of the orogen (Section 6.2).

4.2.2. Conglomeratic facies

Where conglomerate clasts are present, these provide an easier means to link sediment with source compared to provenance studies in sand grade material where the source material is often broken down into its constituent grains. In the Himalayan foreland basin, from Pakistan to Nepal, provenance determination from clast identification has constrained the timing of exhumation of the clast source region. This has provided information on the timing of movement on the MBT, frontal thrusting and flexure-related basement faulting (Baker et al., 1988; Burbank and Beck, 1989; Meigs et al., 1995; DeCelles et al., 1998b; Brozovic and Burbank, 2000, see Sections 5.1.1–5.1.3). In the suture zone, provenance of conglomeratic facies has aided understanding of drainage development and tectonics of the region, as well as in correlation and stratigraphy (Clift et al., 2001a; Sinclair and Jaffey, 2001; Aitchison et al., 2002, see Section 5.2).

4.3. Whole rock geochemistry

The geochemical composition of clastic sediments is often dominantly influenced by source rock composition. Major and trace elements have therefore been successfully used as an indication of provenance, provided the elements used are those with a low solubility during weathering and a low residence time in seawater (e.g., Wronkiewicz and Condie, 1987; Cullers et al., 1988). Whole rock geochemistry is especially useful to add accuracy to detrital modal analyses, where less resistant grains may have been preferentially broken down into matrix. X-ray fluorescence, ICP-AES and ICP-MS are commonly used, usually carried out on fine-grained sediments that preserve a source signature most accurately, being better mixed and more homogenous than coarser grained fractions. In the Himalaya, chrome and nickel concentrations have been used to detect mafic input (Garzanti and Van Haver, 1988, see Section 5.2; Najman and Garzanti, 2000, see Section 5.1.2), of which there is often scant evidence in petrographic studies due to preferential breakdown of mafic nesosilicates and inosilicates. Ni and Cr are good indicators of mafic input because they substitute for Mg and Fe in the early (mafic) phases of fractional crystallization; nickel is primarily found in olivine and chrome in spinel and, to a lesser extent, diopside and augite.

In interpreting such whole rock geochemical signatures in terms of provenance, due regard should be paid to other potentially influencing factors, for example weathering, diagenesis and metamorphism; (Wronkiewicz and Condie, 1987; Condie and Wronkiewicz, 1990; McLennan and Taylor, 1991). Chrome and nickel are considered immobile during the processes of metamorphism and hydrothermal activity (Condie and Wronkiewicz, 1990; Rollinson, 1993). Chrome is susceptible to the effects of weathering and sedimentation but Condie and Wronkiewicz (1990) demonstrated its successful use as an indicator of provenance. The effects of adsorption of metals from sea water on to clay minerals has not been fully evaluated. Therefore some of the variation between values from marine and continental facies, as described in Section 5.1.2, could potentially be explained by environment of deposition. However, a study of muds from the Amazon River delta does not show significant major or trace element variation as a function of distance seaward of the river mouth (Kronberg et al., 1986; Wronkiewicz and Condie, 1987). In addition, the degree of variation shown in the study in Section 5.1.2 is an order of magnitude greater than might be expected from seawater adsorption effects.

Trace elements and also major elements have also been used from Indian foreland basin (Najman, 1995), and Tethyan rocks in southern Tibet (Zhu, 2003; Zhu et al., 2005, see Section 5.2), to discriminate between derivation from different tectonic environments. For example, Roser and Korsch (1986), use a K_2O/Na_2O vs. SiO_2 discrimination diagram to identify sediments derived from passive margin, active margin and arc fields, and Bhatia and Crook (1986) used Th–Sc–La and Th–Sc–Ze/10 ternary diagrams to discriminate between different fields of sediments deposited in settings of oceanic island arc, continental island arc, active continental margin and passive continental margin. Whilst rocks from the Tethyan Himalaya, southern Tibet (Youxia Formation) plot in the active margin field, consistent with their petrography, rocks from the Indian foreland basin plot in the passive margin field. This is consistent with the definition of passive margin which includes sedimentary basins on trailing continental margins supplied from collisional orogens, e.g., the Bengal Fan (Bhatia, 1983; Bhatia and Crook, 1986).

4.4. Mineral geochemistry

This involves determination of the composition of individual minerals because particular compositions can be tied to different generic types of source area. Analyses are most commonly undertaken using electron microprobe techniques.

4.4.1. Spinel geochemistry

Spinel composition is largely a function of magma and source composition. The variation in its geochemistry with different mafic–ultramafic rock types and geodynamic settings (Barnes and Roeder, 2001) has resulted in its common use as a provenance indicator.

Dick and Bullen (1984) used Cr# (Cr/(Cr+Al)) to differentiate between spinels derived from different geotectonic environments. They concluded that spinels from abyssal peridotites and basalts of Mid Ocean Ridge setting have Cr#'s of less than 0.6. In contrast, spinels in rocks of arc-related settings, continental layered intrusives and oceanic plateau basalts have Cr# above 0.6. Spinels from Alpine-type (ophiolitic) peridotites and associated volcanics display a complete range of spinel compositions and can be divided into: Type I ophiolites, those with spinels which share the same composition as spinels from Mid Ocean Ridge rock (Cr#<0.6), and therefore likely represent sections of ocean lithosphere formed in this tectonic setting; Type III ophiolites, where spinel composition has a Cr# of >0.6 and therefore largely falls outside the compositional field of Mid Ocean Ridge type spinels. A sub-volcanic arc provenance is inferred for these rocks; Type II ophiolitic rocks have spinel compositions which span the full range of compositions of Type I and III rocks. Type II peridotites and volcanics are inferred to represent composite origins involving complex multi-stage melting histories.

These may be found in tectonic settings where, for example, a young volcanic arc was constructed on older oceanic crust, or sections across the transition from arc to ocean lithosphere. Kamenetsky et al. (2001) used wt.% TiO_2 vs. Al_2O_3 discriminant plot to differentiate between arc, Mid Ocean Ridge Basalts (MORB), Oceanic Island Basalts (OIB) and Large Igneous Provinces (LIP). TiO_2 is higher in volcanic rocks compared to residual mantle peridotites, and also high in spinels from LIP and IOB settings.

Power et al. (2000) called into question the use of chrome spinel as a provenance indicator, suggesting that the discrimination diagrams were not based on representative data. Later, Barnes and Roeder (2001) plotted data from >26,000 spinel analyses and thus characterised the composition of spinels from different tectonic regimes and rock types more comprehensively.

Detrital spinel composition has been used in the Himalaya to discriminate between suture zone–arc provenance and input from the Indian craton (Najman and Garzanti, 2000; Zhu, 2003; Zhu et al., 2005), showing that erosion from the orogen was occurring by Eocene times (see Sections 5.1.2 and 5.2).

4.4.2. Amphibole geochemistry

Using the electron microprobe and the ion microprobe, Lee et al. (2003) determined the major and trace element composition of detrital amphiboles eroded from different tectonic units of the Western Himalaya, in order to investigate the potential of this mineral for provenance studies. The common occurrence, and complex and varied composition of amphibole is a critical aspect for its use in provenance studies. Its propensity for diagenetic alteration is a drawback, but use of the ion microprobe allows unaltered areas of the grain to be analysed. Lee et al. found the trace element ratios Nb/Zr and Ba/Y the most discriminating between amphiboles from different tectonic units. Although overlap between the fields of different tectonic units was found, their data showed that there was sufficient discrimination to allow exclusion of certain provenances for specific grains. Using detrital grains collected from rivers eroding only one tectonic unit allows a better average of grains to be sampled from a wider area, than if amphiboles from a bedrock sample were used. Analyses from a number of rivers per tectonic unit would tighten up the provenance fields further.

4.5. Isotopic dating of detrital minerals

Single grain techniques are particularly important when gaining information from the sediment record since whole rock approaches can only provide an average value when sources are mixed. For a given isotopic system, each mineral has a different closure temperature, below which age information is retained. Hence, isotopic ages can provide information on a range of temperature conditions from the time after crystallisation of a grain to its lower temperature stages of cooling, dependent on the isotopic system and mineral being studied.

The most common use of detrital mineral age data lies in their contribution to provenance discrimination. Different source areas may be identified if they have been subjected to contrasting metamorphic histories, reflected in the age of their constituent mineral populations. Other applications include using detrital minerals to provide a maximum depositional age to the host sediments where no other dating method is available, and determination of host rock and sediment exhumation and cooling histories.

4.5.1. Ar–Ar dating of detrital minerals

Minerals commonly dated by this method include hornblende, muscovite, biotite and feldspar. In studies of sediments eroded from the Himalaya, white mica is the most commonly dated mineral and feldspar has also been used. Ar–Ar white mica ages record the time of cooling through their closure temperature of ca. 350 °C (350 °C; Jager, 1967; 350–420 °C; McDougall and Harrison, 1999; von Blanckenburg et al., 1989). The closure temperature for feldspar is more variable, with a range of >300 to ~125 °C (Foland, 1974; Harrison and McDougall, 1982).

4.5.1.1. Constraining maximum sediment depositional age. If during erosion, deposition and burial a detrital mineral has not been subjected to conditions likely to cause alteration, or temperatures sufficient to result in post-depositional resetting, its age records the time of cooling through its closure temperature in the source region. In these circumstances, it can thus be used to provide a maximum depositional age for the host sediment, since sediment depositional age must be younger than detrital mineral cooling age. White micas are ideal for this purpose, having a closure temperature high enough that minerals are not easily reset but low enough to provide a close constraint on the depositional age of the sediment. In Himalayan sedimentary rocks, several tests have been employed on the mica grains to detect alteration, and on the host rock to estimate if burial temperatures could have been sufficient to cause resetting (Najman et al., 1997, 2001). Mica grains have been subjected to step-heating experiments where flat plateaus are indicative of minimal alteration, and electron microprobe traverses to obtain compositional profiles, where alkali loss at the grain margins indicates alteration. Petrographic study of the non-detrital minerals in the host rock and the degree of crystallinity of the <2 micron fraction of authigenic illite can indicate the post-depositional pressure–temperature conditions to which the rock and its constituent minerals have been subjected. The results show that in the regions studied to date, Ar–Ar ages of detrital micas in the Himalayan foreland basin record the timing of cooling in the source region. They can therefore be used to provide maximum ages for the initiation of continental alluvial sedimentation throughout the basin from Nepal to India and Pakistan (Najman et al., 1997, 2001; DeCelles et al., 2001, see Section 3.2). This provides constraint to the duration of the foreland basin regional unconformity (Section 3.2*),* and hence to the timing, diachroneity and mechanisms of collision and orogenic processes (see Sections 6.1.2 and Section 6.2.3).

Harrison et al. (1993) employed the Ar–Ar technique to constrain the depositional age of the suture zone Kailas conglomerate. They used Ar–Ar analyses of feldspar in a clast derived from the Gangdese volcanics to model the thermal history of the grains by diffusion domain theory. A period of rapid cooling in the source region is interpreted as exhumation of the region prior to its erosion and incorporation into the conglomerate. Hence the maximum age of deposition of the conglomerate is determined (Section 3.1).

4.5.1.2. Constraining exhumation. Lag times are defined as the difference between a detrital mineral age, (which reflects the time since cooling through closure temperature at depth in the source region), and its host sediment depositional age, (determined magnetostratigraphically or biostratigraphically). Lag times provide information on hinterland exhumation rates. A short lag time indicates rapid exhumation in the source region during that period. As the number of such studies using coeval Himalayan foreland basin rocks along strike increase, a spatial and temporal picture of orogenic exhumation will be built up (e.g., White et al., 2002; Najman et al., 2002b, 2003a; Szulc, 2005; Szulc et al., submitted for publication). Interspersed periods of long lag time must be interpreted with more caution since temporal variations may be the result of drainage diversion or sediment storage rather than hinterland exhumation rate changes.

From lag time information, the exhumational state of the orogen can be deduced. The sediment record of an orogen in exhumational steady-state will show peak detrital mineral ages which young with time, and have constant lag times (Garver et al., 1999). Quantification of exhumation rates determined from lag times, and investigation of lag time response to changing exhumation rates, is possible from simple thermal modelling which utilises an appropriate thermal structure for the crust, allowing for advection modification of the geotherm (e.g., Garver et al., 1999; White, 2001; Najman et al., 2003a,b).

In the Himalaya, this approach has been used in the foreland basin of Pakistan and India using Ar–Ar dating of detrital micas (White et al., 2002; Najman et al., 2002b, 2003a) and fission-track dating of detrital zircons (Cerveny et al., 1988) (see Sections 5.1.1 and 5.1.2). In Nepal (Section 5.1.3) Ar–Ar dating of detrital feldspar has been employed, a mineral more susceptible to alteration and post-depositional resetting due to its lower closure temperature and more porous structure (Harrison et al., 1993). Szulc et al. (submitted for publication) utilised the Ar–Ar technique on the Siwalik rocks in Nepal using micas. A similar study has also been undertaken on the Bengal Fan sediments, using detrital feldspars and micas (Copeland and Harrison, 1990; Section 5.3.2). However, in this study, mineral ages were in a number of instances younger than the sediment depositional age, which points towards mineral alteration or age resetting, ascribed, in this instance, to hydrocarbon alteration.

4.5.1.3. Constraining provenance. Micas are a good indicator of source provenance in the Himalaya where the lithotectonic units can be characterised, to some extent, by different aged mica populations. In the central Himalaya, published mica ages from the presently exposed Lesser Himalaya are sparse but typically Proterozoic: ages of ~1200–1600 and 800–900

Ma have been reported (Copeland et al., 1991; Frank et al., 1995; Najman, 1995). By contrast, micas of the metamorphosed Higher Himalaya have Himalayan ages (i.e., <55 Ma). Steck (2003) records that micas <700 Ma have been documented in the protolith Higher Himalaya as, for example, exposed in NW India (Thakur, 1998). I am unaware of any published detrital mica ages from the Tethyan Himalaya lithotectonic zone. To the west, in Pakistan, division of lithotectonic zones into Higher and Lesser Himalaya is not so clearly defined (see Section 2). Indian crust cover rocks have been affected by Himalayan metamorphism and thus contain mica ages reset by this event, whilst Indian crust basement contains micas of Precambrian and Palaeozoic age (Treloar et al., 1989).

In conjunction with other provenance techniques, White et al. (2002) used Ar–Ar ages of detrital micas to determine the provenance of foreland basin sediments in NW India (Section 5.1.2). Najman et al. (2002a, 2003a) used similar methods to document the provenance of foreland basin rocks in Pakistan (Section 5.1.1).

4.5.2. U–Th–Pb dating of detrital minerals

Zircon and monazite are commonly dated by the U–Th–Pb method, using isotope dilution thermal ionization multicollector mass spectrometry, the ion microprobe and the more recently developed laser ablation plasma ionization multicollector mass spectrometry (LA-PIMMS). The speed and adequate precision of the LA-PIMMS technique is well suited to analysis of detrital mineral populations, and spot analyses allow recovery of polyphase histories from grains which have grown in multiple episodes (e.g., White et al., 2001). Monazite and zircon have high closure temperatures (>750 °C, Spear and Parrish, 1996) and thus their U–Th–Pb ages record the time of crystallisation from igneous melts. However, in the Higher Himalayan Tertiary leucogranites, inheritance is ubiquitous in zircons (Hodges et al., 1996; DeCelles et al., 1998b) and therefore monazite, which is less affected, is the mineral of choice to date crystallisation of these intrusions. Monazite is also found in metamorphic rocks and its U–Th–Pb age records the time of prograde metamorphism to 520–550 °C (Smith and Barreiro, 1990) as well as decompression (Foster et al., 2000).

In Nepal, the different tectonic zones of the Himalaya contain zircons with characteristic U–Pb ages: Lesser Himalayan zircons are generally older than 1700 Ma (ca. 1867–2657 Ma), Higher Himalayan aged zircons are characteristically less than 1700 Ma (ca. 967–1710 Ma), and Cambro–Ordovian granites found in the Lesser Himalaya, Higher Himalaya and Tethyan Himalaya likely contain zircons of that age (Parrish and Hodges, 1996; DeCelles et al., 2000). Taking advantage of the zircon ages characteristic to each tectonic zone in the Nepal Himalaya, DeCelles et al. (1998a,b) used U–Pb ages of detrital zircons from the sediment record to establish provenance changes and hence the timing of thrusting and erosion of source region units (see Section 5.1.3). In addition, DeCelles et al. (2004) placed constraints on the degree of diachroneity of collision by identifying the first input of Himalayan-derived zircons to the foreland basin from their U–Pb ages (Section 5.1.3).

In contrast to zircon, the majority of published U–Th–Pb ages of metamorphic and magmatic monazite from the Himalayan orogen give Tertiary ages (Harrison et al., 1997; Searle et al., 1999a,b; Simpson et al., 2000). On the assumption that monazite in unmetamorphosed Higher Himalaya, Lesser Himalaya and Tethyan Himalaya rocks will have similar ages to that of zircons in the same units, detrital monazites can be used as an indicator of provenance. Such an approach was used by White et al. (2001) on the Dharamsala Formation foreland basin strata in India, where the detrital monazites were also used to date the timing of prograde garnet grade metamorphism in the source region and the subsequent exhumation history of these metamorphosed rocks to the surface (see Section 5.1.2).

4.5.3. Fission track dating of detrital minerals

Fission tracks result from fission of ^{238}U, which causes linear zones of radiation damage in the crystal lattice of the host mineral. Apatite and zircon are the most commonly used minerals for this technique. Dating by the fission track method follows the same principle as other isotopic systems. The fission tracks start to accumulate as the mineral cools through its partial annealing zone (PAZ) which for apatite lies in the temperature range of ~60–110 °C (Laslett et al., 1987) and for zircon is ~200–350 °C (Tagami et al., 1998). The number of fission tracks in a mineral is used to determine the time since it passed through its partial annealing zone, i.e., the mineral's "age", whilst the track length distribution provides information on how quickly it passed through this zone. Thus, with a quantitative understanding of track lengths, this method can be used to construct the cooling history of the mineral as well as providing an age. Such a cooling model is well established for apatite (Laslett et al., 1987; Ketcham et al., 1999), but is still in the developmental stages for zircon (Yamada et al., 1995; Tagami et al., 1998).

The temperature of resetting through the PAZ of apatite is typical of that encountered in sedimentary basins and hence in detrital studies apatite is often used to constrain basin evolution. In the Himalaya, it has been used to constrain the timing of thrusting within the foreland basin (Najman et al., 2004 see Section 5.1.2) and, in conjunction with illite crystallinity methods, to constrain the timing of maximum burial and cessation of sedimentation in the suture zone basin (Sinclair and Jaffey, 2001 — see Section 3.1). Where burial temperatures in the basin have not been sufficient to reset apatite fission tracks, the data can be used to document hinterland exhumation rates by comparison of detrital mineral age with sediment depositional age. This approach, determination of lag times, uses the same rationale as described in Section 4.5.1. It has been demonstrated with apatites in the Bengal Fan by Corrigan and Crowley (1990), (Section 5.3.2).

Temperatures attained by burial in the sedimentary basin are generally insufficient to reset zircon fission track characteristics and thus detrital zircons preserve a record of the low temperature cooling history of the hinterland. Detrital zircon ages have been used in the Himalayan foreland basin to provide information on the cooling and exhumation history of the orogen by comparison of detrital mineral age with host sediment depositional age as described in Section 4.5.1 (Cerveny et al., 1988, Section 5.1.1). Detrital zircon fission track ages have also been used to provide maximum age constraints for the foreland basin sediments (Najman et al., 2004, Section 3.2.2) and to identify the first arrival of syn-orogenic grains to the foreland basin thus constraining the age and diachroneity of collision (Najman et al., 2005, Section 5.1.3).

4.6. Provenance discrimination by isotopic fingerprinting

Different source terrains may be characterised by their isotopic signatures, which are a function of the age and composition of the rocks units. The characteristic signature can be recorded in whole rock or individual minerals of the source region. In detrital studies, single grain techniques have greater potential in provenance discrimination when mixed contributions from more than one lithotectonic source exists. This is because whole rock sediment analyses can provide only an average value, and subordinate source contributions may be obscured. Two techniques used in the Himalaya are discussed below. Discrimination between lithotectonic zones on the basis of their Rb–Sr and Sm–Nd characteristics is primarily of importance in differentiating between units of the Indian crust. By contrast, the use of Pb isotopic characteristics of feldspar is of more relevance to discrimination between Asian, arc and Indian lithologies. It is an approach only recently applied to the Himalaya, and still in a more developmental state in this region.

4.6.1. Isotopic fingerprinting by Sm–Nd and Rb–Sr techniques

4.6.1.1. Detrital sediment whole rock studies. The Sm–Nd system (e.g., Depaolo, 1981) is not easily reset by metamorphism or radically altered by the processes of erosion or sedimentation and thus it is an ideal technique for provenance use. The rare-earth elements are relatively insoluble and fractionation by sedimentary sorting or grain size variation generally only has a limited impact (Taylor and McLennan, 1985). However, within the framework of the fine grain-sized fraction analysed by preference (on the assumption that it represents the best average composition of the source rock eroded) recent studies show that grain-size does influence the results to some extent (Galy et al., 1996; Najman and Bickle, unpublished data). However, the magnitude of variation is less than the difference between the Himalayan lithotectonic zone sources. This should be borne in mind when direct comparisons are made between analyses of fine-grained sediments of different grain size. In contrast, Sr is both soluble and shows large differences between minerals that have different Rb/Sr ratios. Nevertheless, the large difference in $^{87}Sr/^{86}Sr$ between the Himalayan source terrains appear to be reflected in the characteristic $^{87}Sr/^{86}Sr$ of their detrital sediments (e.g., Galy et al., 1999).

The Sr and Nd characteristics of the various Himalayan lithotectonic units have been documented by a number of workers, (e.g., Vidal et al., 1982; Deniel et al., 1986, 1987; Scharer et al., 1990; France-Lanord et al., 1993; Petterson et al., 1993; Parrish and Hodges, 1996; Harris et al., 1998; Whittington et al., 1999; Ahmad et al., 2000; Robinson et al., 2001). These data show that the Sr and Nd isotopic compositions of the main Himalayan metasedimentary units (the Lesser and Higher Himalayan Series) are distinctly different. A recent assertion that tectonostratigraphic units were not isotopically distinct (Myrow et al., 2003) ignores thrusting which has superimposed units from initially disparate tectonic units (Ahmad et al., 2000). Values below are quoted in T_{DM} (model age with respect to the depleted mantle reservoir) and ε_{Nd} (a measure of the difference in $^{143}Nd/^{144}Nd$ from CHUR — CHondritic Uniform Reservoir). The Lesser Himalaya has T_{DM} ages of 1939 to 2600 Ma and ε_{Nd} range -16 to -25, average > -20, whereas rocks of the Higher Himalaya have T_{DM} of 1361 to 2200 Ma and ε_{Nd} range -19 to -5, average ca. -15. The $^{87}Sr/^{86}Sr$ values of the Higher Himalaya are ~0.73 to 0.82. By contrast, Lesser Himalayan rocks have a wider range between ~0.73 up to ~5 with a mean of 0.98. Metacarbonates give particularly high values (English et al., 2000; Bickle et al., 2001; Quade et al., 2003). Ahmad et al. (2000) recognised some critical sub-divisions of the Higher and Lesser Himalaya in the Garhwal Himalaya: The metamorphosed rocks of the Main Central Thrust zone are isotopically identical to the Inner Lesser Himalaya, and low-grade rocks in the Outer Lesser Himalayan region have isotopic compositions identical to the Higher Himalaya. The Tibetan Sedimentary Series has a similar Sm–Nd isotopic composition to the Higher Himalaya and T_{DM} 1142–2166 Ma, but can be distinguished by its $^{87}Sr/^{86}Sr$ values (~0.705–0.730) which are lower than those of the Higher Himalaya. Comparatively few samples are available from the diverse Indus Tsangpo Suture Zone. However their relatively low $^{87}Sr/^{86}Sr$ ratios (~0.710) and range of mantle-like ε_{Nd} values (-7 to $+13$) are consistent with the abundance of ophiolite in the unit and the Mesozoic calc-alkaline rocks north of the Tibetan Zone and sediments derived from them. It should be remembered that a Sm–Nd model age calculated for sedimentary or metasedimentary rocks may not in itself represent a specific event in geologic history but a weighted average of the range of ages at which the sources were extracted from the mantle.

The characteristic signatures of the Himalayan lithotectonic source units have been used to characterise provenance in the Bengal and Indus Fans, suture and foreland basins (France-Lanord et al., 1993; Najman et al., 2000; Clift et al., 2001a,b; Huyghe et al., 2001; Robinson et al., 2001; see Sections 5.1–5.3). In particular, the younger foreland basin deposits document a change in provenance which can be related to exhumation of the Lesser Himalaya (Section 5.1.3). This change is not documented in the Bengal Fan sediment succession however, which illustrates the value of more proximal small-basin scale studies where subordinate detritus may be less diluted. If direct comparisons of Sm–Nd values of detrital material is to be made between the sedimentary basins, it should be noted that different workers have used different sub-fractions within the fine-grained material (whole rock and <2 μm) and, as described above, this could influence the results.

4.6.1.2. Detrital mineral Sm–Nd characteristics. To overcome the limitations inherent in using whole rock detrital material for provenance determination from mixed sources (as outlined above), White (2001) extended the Sm–Nd isotopic fingerprinting approach to use on single grain detrital monazites from the Himalayan foreland basin (Section 5.1.2), with the assumption that they record the Sm–Nd characteristics of the lithotectonic source region. ε_{Nd} values at $t=0$ (present day) can be calculated and comparison with source region lithotectonic lithologies allows provenance of individual minerals to be assigned. Since U–Th–Pb ages can be determined on the same grain (Section 4.5.2), combined provenance information from U–Th–Pb age dating and Sm–Nd isotopic fingerprinting provides a powerful approach, and ε_{Nd} values at time of crystallisation and model T_{DM} ages can be calculated.

4.6.2. Isotopic fingerprinting using Pb isotopes in K-feldspar

Clift et al. (2000) analysed the Pb isotopic character of detrital K-feldspars in Himalayan-derived sediments. They used the ion microprobe rather than conventional mass spectrometry for analysis, which has the advantage that unaltered grain cores can be analysed. K-feldspars are good provenance indicators because they are common in continental settings and typically contain enough Pb to allow measurement of the isotopic ratio using modern microbeam mass spectrometry.

The technique exploits the variation in feldspar composition between the Asian rock units of the Lhasa Block, Karakoram and Trans-Himalaya, Kohistan and the Indian crust when plotted as $^{207}Pb/^{204}Pb$ against $^{206}Pb/^{204}Pb$. This variation reflects differences in source rock composition and the age of melt extraction from the mantle. Although analytical uncertainties for ^{204}Pb are higher than for other Pb isotopes, use of this isotope is crucial because it provides significant separation between the different sources considered. This technique is at a developmental stage in the Himalaya where the Pb isotopic characteristics of the source areas are currently incompletely characterised: early detrital provenance interpretations were based on comparison with source areas defined by a modest number of grains, whole rock analyses only, or analyses from a long distance along strike in the central Himalaya (Clift et al., 2001a). More recent work is addressing this issue (Clift et al., 2002b). These authors note that increased data is already resulting in modification of source fields, and resulting partial overlap suggests that absolute discrimination between sources may be impossible even if truly comprehensive source region compositions were defined. Nevertheless, as more data become available this technique will continue to be refined. At its current level Clift et al. use this technique to discriminate between Indian and non-Indian crust sources for detrital feldspars of the Indus Fan and suture zone deposits (see Sections 5.2 and 5.3.1).

5. The sediment archive as a record of orogenic evolution: regional case studies

5.1. The Himalayan peripheral foreland basin

5.1.1. Pakistan foreland basin

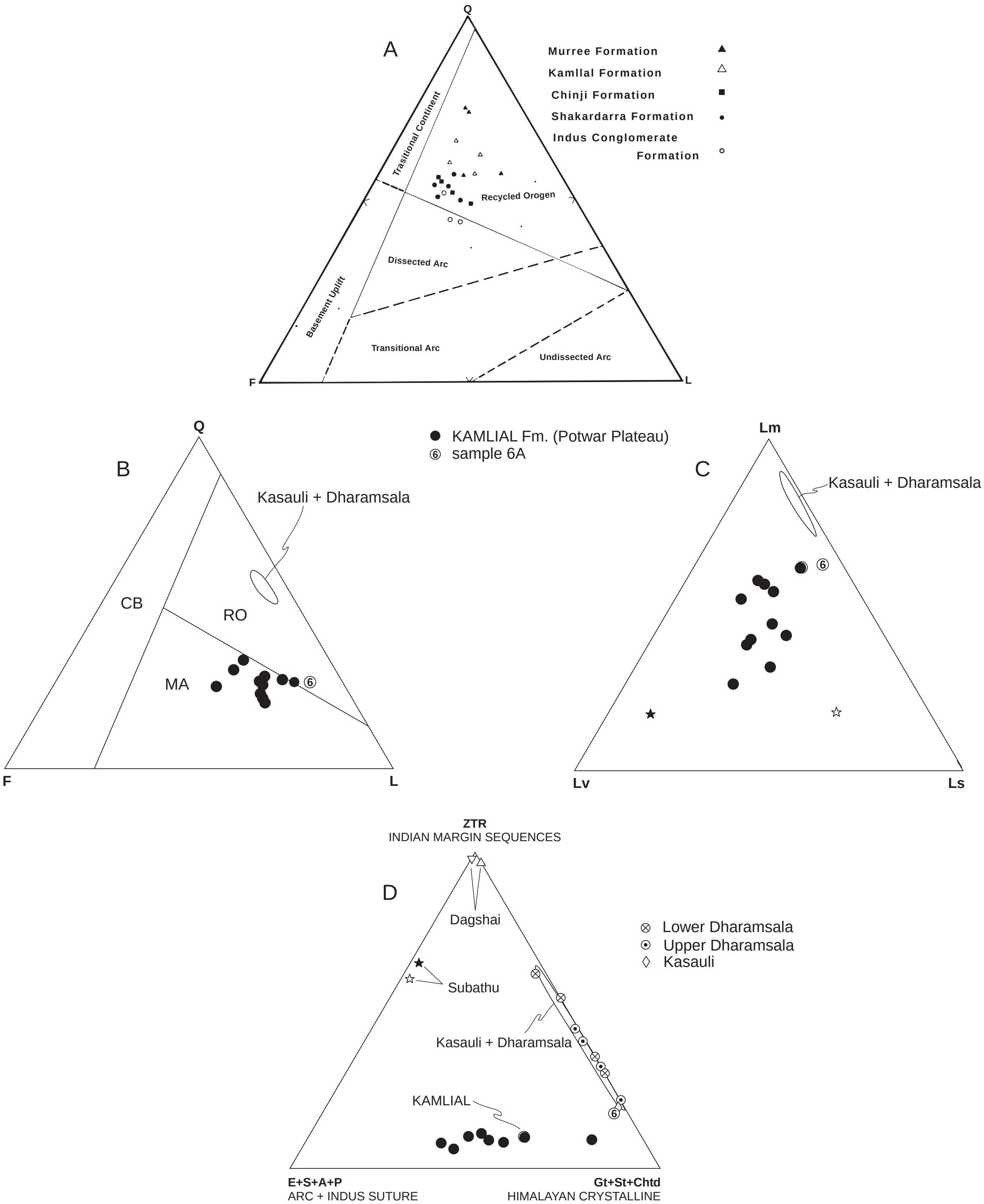
As described in Section 3.2.1, the peripheral foreland basin in Pakistan consists of Palaeocene–Eocene marine facies with a number of local names, overlain unconformably by the Oligo–Miocene Murree Formation, the overlying Kamlial Formation dated 18–14 Ma, and younger Siwalik Group, subdivided into the Chinji, Nagri, Dhok Pathan and Soan Formations.

5.1.1.1. Data

5.1.1.1.1. Palaeocene–Eocene marine facies. In the Kohat region, shallow marine passive margin facies change to deep water Tarkhobi shales of the Gazij Group in the Late Palaeocene. This is followed by shallowing up to continental facies of the uppermost Lower Eocene Mami Khel Group before marine conditions resume until the Eocene–Oligocene hiatus (Pivnik and Wells, 1996). A sedimentary source dominates, with some metamorphic input. Co-eval marine strata of the Patala Formation (P6, 53–55 Ma) to the east in the Kazara–Kashmir Syntaxis contain pure quartz arenites in the lower part and record an influx of felsitic volcanics, phyllite, chert and serpentine schists in the middle and upper parts (Critelli and Garzanti, 1994).

5.1.1.1.2. Murree Formation. Above the regional unconformity, the provenance of the continental Murree Formation heralds the first major influx of metamorphic material (Fig. 4A). The rocks plot in the recycled orogen provenance, lithic fragments are predominantly sedimentary or of low-grade metamorphic origin, and feldspar is rare (Abbasi and Friend, 1989; Critelli and Garzanti, 1994; Singh, 1996). The 37 Ma peak of detrital mica ages in the older northern Murree Formation (Balakot Formation) attests to influx of material metamorphosed during the Himalayan orogeny (Section 3.2.1, Najman et al., 2001, 2002a). Low-grade metamorphic lithic fragments are consistent with derivation from the metamorphosed Himalaya, whilst volcanic material shows derivation from both arc and suture zone sources (Critelli and Garzanti, 1994). Cr-spinel composition, indicating detrital input from the suture zone/arc, is similar to that of spinels found in the Subathu Formation, Indian foreland basin (Najman and Garzanti, 2000).

5.1.1.1.3. Kamlial Formation. A petrographic change occurs at the boundary between the Murree Formation and the Kamlial Formation. There is increased contribution from a metamorphic source in the Kohat Plateau (Abbasi and Friend, 1989) and a drastic change to substantial input from arc-derived material in the Chinji village region, Potwar Plateau (Hutt, 1996; Najman et al., 2003a), (Fig. 4B–D). Ar/Ar dating of detrital white micas shows lag times which decrease from 5–7 My during the period 18–17.7 Ma, to <1 My after 17.4 Ma to the end of the analysed interval at 14 Ma, indicating rapid exhuma-


Q
A
Murree Formation
Kamlal Formation
Chinji Formation
Shakardarra Formation
Indus Conglomerate Formation
Trasitional Continent
Recycled Orogen
Dissected Arc
Basement Uplift
Transitional Arc
Undissected Arc
F
L
B
Q
Kasauli + Dharamsala
CB
RO
MA
F
L
KAMLIAL Fm. (Potwar Plateau)
⑥ sample 6A
C
Lm
Kasauli + Dharamsala
⑥
Lv
Ls
ZTR
INDIAN MARGIN SEQUENCES
D
Dagshai
Lower Dharamsala
Upper Dharamsala
Kasauli
Subathu
Kasauli + Dharamsala
KAMLIAL
⑥
E+S+A+P
ARC + INDUS SUTURE
Gt+St+Chtd
HIMALAYAN CRYSTALLINE

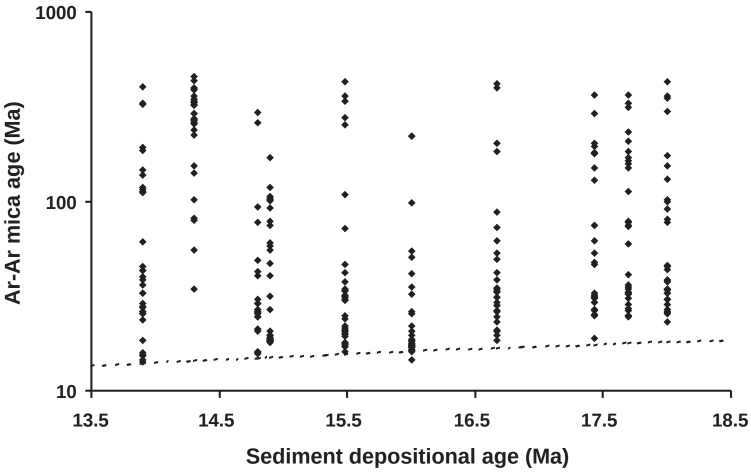


Fig. 5. Sediment depositional age (determined by magnetostratigraphy; Johnson et al., 1985) vs. Ar–Ar ages of detrital white micas from the Kamlial foreland basin sedimentary rocks, Chitta Parwala section, Potwar Plateau, Pakistan. Dashed line = line of zero lag time, where mica age equals sediment depositional age and rapid exhumation is indicated. The data show a period of rapid exhumation in the source region between ca. 17–13 Ma. From Najman et al. (2003a).

tion of the source region during this period (Fig. 5). This result is not dissimilar to the work of Cerveny et al. (1988) who obtained a 4 My lag time from a single basal Kamlial sample using fission track dating on detrital zircons. Simple thermal modelling of the mica data indicate that the rapid exhumation of this source began at ca. 20 Ma (Najman et al., 2003a, Section 4.5.1). Modal mineral abundances show that the micaceous source was subordinate to the main arc source and the co-occurrence of garnet and staurolite since 18 Ma indicates its grade. Kyanite/andalusite first appears at 16 Ma.

5.1.1.1.4. Siwalik Group. After the return to more mudstone-rich facies in the Chinji Formation (Section 3.2.1), the abrupt return to sandy facies at the start of the Middle Siwalik sub-group is coeval with increased sediment accumulation rates at 11 Ma (Meigs et al., 1995; Burbank et al., 1996). The first conglomerates are also found within the Middle Siwalik rocks, and are of metamorphic and sedimentary provenance (Cotter, 1933; Gill, 1951b). Upper Siwalik (Soan) conglomerates contain clasts which commonly have distinctive local provenance, for example the Talchir granite clasts of Salt Range provenance found in the Siwalik deposits since 5 Ma (Burbank and Beck, 1989).

Siwalik Group rocks predominantly plot in the recycled orogen petrographic provenance field (e.g., Fig. 4A). Their source is predominantly medium to high grade metamorphic, with only a subordinate contribution from sedimentary, arc and ophiolitic sources (Abbasi and Friend, 1989; Critelli and Ingersoll, 1994; Garzanti et al., 1996; Pivnik and Wells, 1996). Distinctive arc-derived blue-green hornblende

Fig. 4. Petrographic (Fig. 4A,B,C) and heavy mineral data (Fig. 4D) for the foreland basin sediments of Pakistan, and comparison with co-eval Indian foreland basin rocks along strike. Fig. 4A: evolution to decreasing quartzose content through time, from Murree to Indus Fm, in the Kohat Plateau region, from Abbasi and Friend (1989). Chinji Formation (oldest), Shakardarra Formation and Indus Conglomerate Formation (youngest) are subunits of the Siwalik Formation. Fig. 4B and C (from Najman et al., 2003a) show the composition of the Kamlial Formation at Chitta Parwala section in Potwar Plateau, with much greater influence of arc-derived material compared to the co-eval rocks in Kohat region (as above), and in India (Kasauli and Dharamsala Formation field shown; Najman and Garzanti, 2000; White et al., 2002). Open star symbol represents green Subathu facies, filled star symbol represents red Subathu facies in Indian foreland basin (Section 5.1.2). Sample 6A is anomalous in terms of all provenance indicators. Fig. 4D: heavy mineral studies show the much greater input of arc-derived material in the Kamlial Formation of Chitta Parwala, Potwar Plateau, compared to rocks of the Indian foreland basin. The figure also shows the evolution of provenance in Indian foreland basin deposits through time. Q = total quartzose grains, F = total feldspar, L = total lithic fragments, Lm = metamorphic lithic fragments, Ls = sedimentary lithic fragments, Lv = volcanic and subvolcanic lithic fragments, ZTR = zircon, tourmaline and rutile, Gt = garnet, St = staurolite, Chtd = chloritoids, E = epidotes, S = spinels, A = amphiboles, P = pyroxenes.

makes up a substantial proportion of the heavy mineral assemblage from 11 Ma (Johnson et al., 1985; Cerveny et al., 1989). Zircons are ubiquitous throughout the succession. This abundance allowed Cerveny et al. (1988) to document rapid exhumation throughout the period 18–0 Ma from lag times of 1–5 My calculated from detrital zircon fission track data. However, since the sampling was sparse throughout the period (4 Siwalik samples in one sedimentary succession, 3 samples in another, plus one Kamlial sample), this can only act as a first order approximation. Variation in exhumation rates may well exist on a more refined scale, as already demonstrated for the Kamlial Formation by Najman et al. (2003a,b) (see above).

5.1.1.2. Interpretations. Subsidence followed by regression to continental embryonic foreland basin facies, plus the appearance of volcanic and serpentinite schist lithic fragments in the Upper Palaeocene to Lower Eocene marine strata point towards ophiolite obduction, flexural loading, and initiation of collision at this time. Collision was between India and either Eurasia or microcontinents/intra-oceanic arcs (Pivnik and Wells, 1996; Garzanti et al., 1996; Davis et al., 2002). The quartz-arenites of the oldest rocks are interpreted as the response to erosion from an uplifted peripheral forebulge (Critelli and Garzanti, 1994).

The major change from sedimentary to metamorphic provenance occurring between the Eocene marine rocks and the Oligo–Miocene Murree rocks is taken to represent the first substantial input from the rising Himalayan thrust belt. The paucity of feldspar and predominance of lower grade metamorphic detritus indicates only limited dissection and dominant contribution from supracrustals at this time (Critelli and Garzanti, 1994). Detrital minerals with ages <55 Ma indicate erosion from sources affected by the Himalayan metamorphism. The peak of mica ages at 37 Ma allows calculation of cooling rates from peak metamorphism of 500–600 °C at ca. 50 Ma (Chamberlain and Zeitler, 1996) at an averaged rate of 20 °C/Ma during this 50 to 37 Ma period (Najman et al., 2002a). Using younger ages of peak metamorphism and higher peak metamorphic temperatures (e.g., Foster et al., 2002), results in cooling rates between 35 and 75 °C/Ma. Redating of the Balakot Formation at <37 Ma, from its previously held Early Eocene age, removes existing evidence of early erosional exhumation of the orogen (Treloar, 1999). Tectonic exhumation may have been achieved by extensional movement on the MMT (Section 2), and the peak of ages at 37 Ma may provide constraints on the timing of such movement (Najman et al., 2002a). The mix of Himalayan and pre-Himalayan aged micas indicates that exhumation of both basement and cover sequences in the internal metamorphosed zone of the orogen (see Section 2) occurred early during orogenic evolution. Continued input from the arc/suture zone is reflected in the presence of detrital Cr-spinels and in petrography.

Najman et al. (2003a,b) interpreted the provenance and sedimentological changes at the Murree–Kamlial Formation boundary at Chinji village, Potwar Plateau (Section 3.2.1) as reflecting a major change in palaeodrainage. They suggest a transition from drainage to the foreland by fluvial systems which predominantly drain the Indian crust thrust-stack, only cutting back into the arc to a limited extent, to drainage by a large river with a catchment area extending further north, draining significant regions of the arc (Fig. 6). This they interpret as likely to be the first time the route of the palaeo-Indus drained from the suture zone, through the arc, and into the foreland, consistent with a catchment area encompassing the interpreted NPHM (Nanga Parbat Harramosh Massif) or Asian source regions for the rapidly exhumed micas. If an NPHM source for the rapidly exhuming micas is accepted, subsequent decrease in arc-derived detritus in the Siwalik Formation deposits could be interpreted as the result of progressive breaching of the arc carapace passively uplifted above the rapidly exhuming Indian crust.

First routing of the palaeo-Indus through the arc to the foreland basin at 18 Ma is inconsistent with the previously held belief that this occurred at 11 Ma, based on the increased fluvial discharge at this time and the presumed first major erosion from the arc based on the first appearance of arc-derived hornblende (Cerveny et al., 1989). However, since the recent dataset of Najman et al. (2003a,b) shows major arc erosion at 18 Ma, the later appearance of hornblende may be explained by either (1) gradual erosion through deeper levels of the arc carapace or (2) diversion to the region of a river that was already

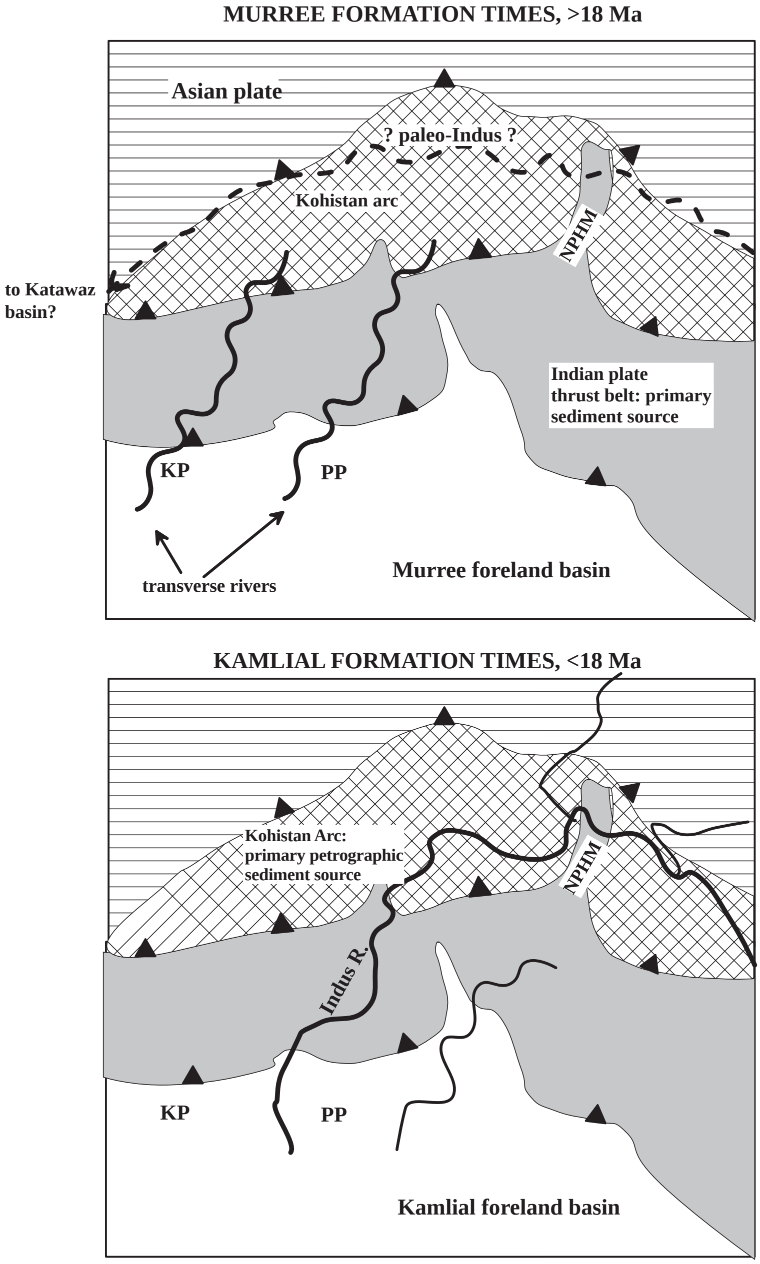


Fig. 6. Interpreted change in fluvial drainage, with first input of the palaeo-Indus to the foreland basin at 18 Ma, the Murree–Kamlial Formation boundary (from Najman et al., 2003a). KP = Kohat Plateau, PP = Potwar Plateau, NPHM = Nanga Parbat Harramosh Massif.

carrying hornblende. Progressive erosion through the arc is consistent with the change from a higher proportion of shallow level arc volcanic detritus in the Kamlial Formation, to significant input of hornblende and pleochroic orthopyroxene in the Siwalik Group, probably associated with amphibolite-granulite terrain in the roots of the arc. Possible river diversion, plus the concurrent increased palaeodischarge and sediment accumulation rates at 11 Ma, suggests active tectonic deformation due to a

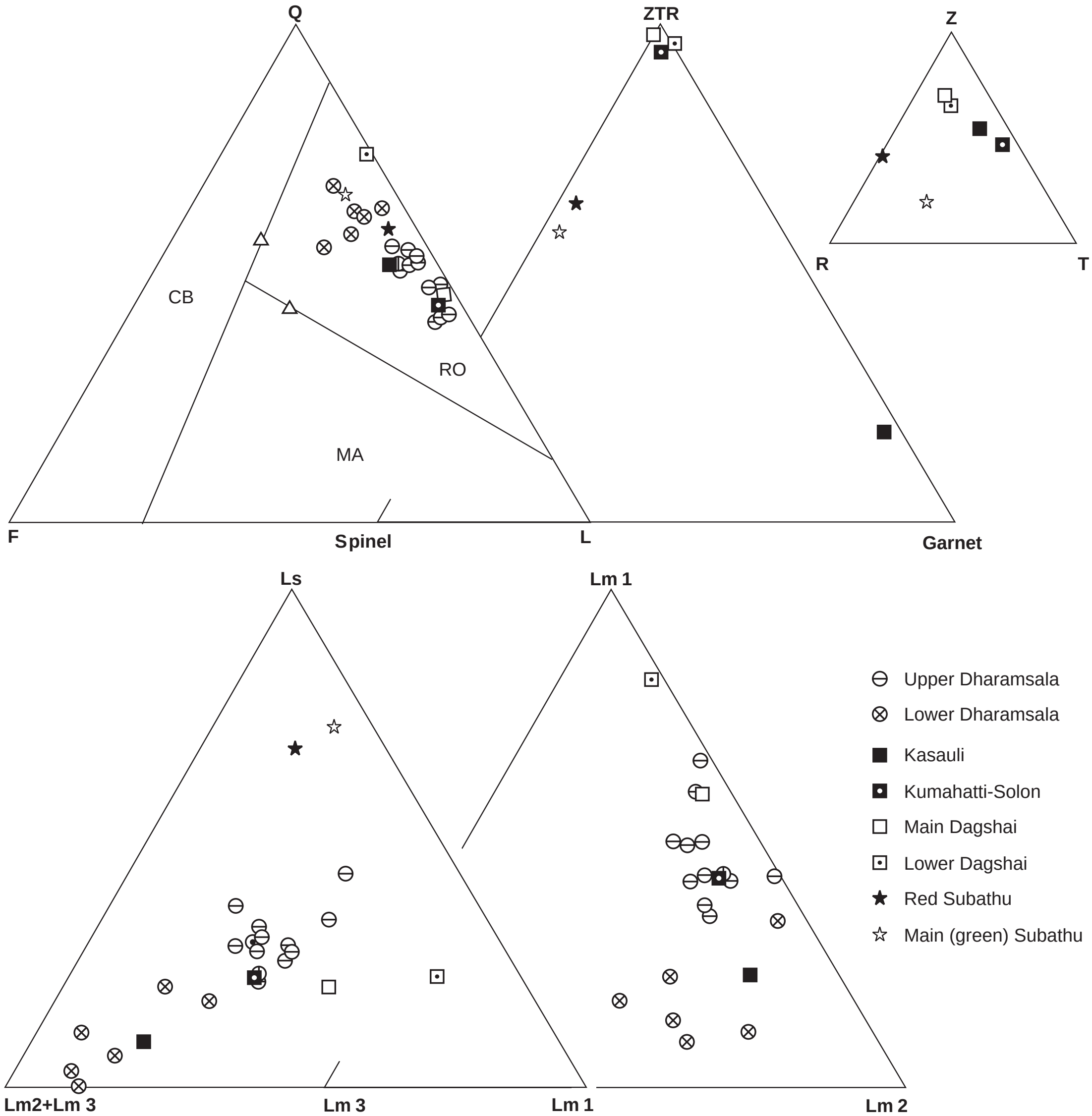


Fig. 7. Petrographic and heavy mineral composition of the foreland basin sediments of India. Data from Najman and Garzanti (2000) and White et al. (2002). All data points shown for Dharamsala Formation, average value shown for Subathu, Dagshai and Kasauli Formations. Note evolution of erosion to deeper metamorphic levels shown by petrographic data from the Subathu Formation, through Dagshai Formation to Kasauli and Lower Dharamsala Formations. Then there is a distinct and abrupt change in petrography in the Upper Dharamsala Formation, where metamorphic grade of detritus decreases compared to older rocks. Q = total quartz grains, F = total feldspar, L = total lithic fragments, Ls = sedimentary lithics, $Lm_{1,2,3}$ = very-low-grade, low-grade and medium-grade metamorphic lithic fragments, respectively. CB = continental block provenance field, RO = recycled orogen field, MA = magmatic arc field (Dickinson, 1985). Progressive evolution is also mirrored in the heavy mineral proportions, with first appearance of garnet, staurolite and chloritoid in Kasauli Formation (Subathu Sub-basin) and Dharamsala Formation (Kangra sub-basin) (see also Fig. 4D). Kumahatti–Solon unit is an informal name assigned by Najman and Garzanti (2000) to the transitional facies as the Dagshai–Kasauli boundary.

significant new thrust load shifting its locus towards the foreland (Willis, 1993b; Burbank et al., 1996). Increased accumulation rates are seen throughout the foreland basin (Section 3.2), interpreted as a response to exhumation of the Lesser Himalaya and eventually its southernmost thrust, the MBT (see Section 6.2.2).

Transfer of slip to the Main Frontal Thrust, called the Salt Range Thrust in Pakistan, occurred at 5–6 Ma. The evidence for the timing of thrusting, based on first appearance of distinctive Salt Range-derived conglomerate clasts, has been called into question by Baker et al. (1988) who consider these clasts may be derived from a basement fault scarp. However, thrusting at 5–6 Ma is also supported by fluvial facies and palaeocurrent changes as well as the development of unconformities in the basin record, and differential amounts of rotation across these unconformities (Burbank and Beck, 1989; Burbank et al., 1996).

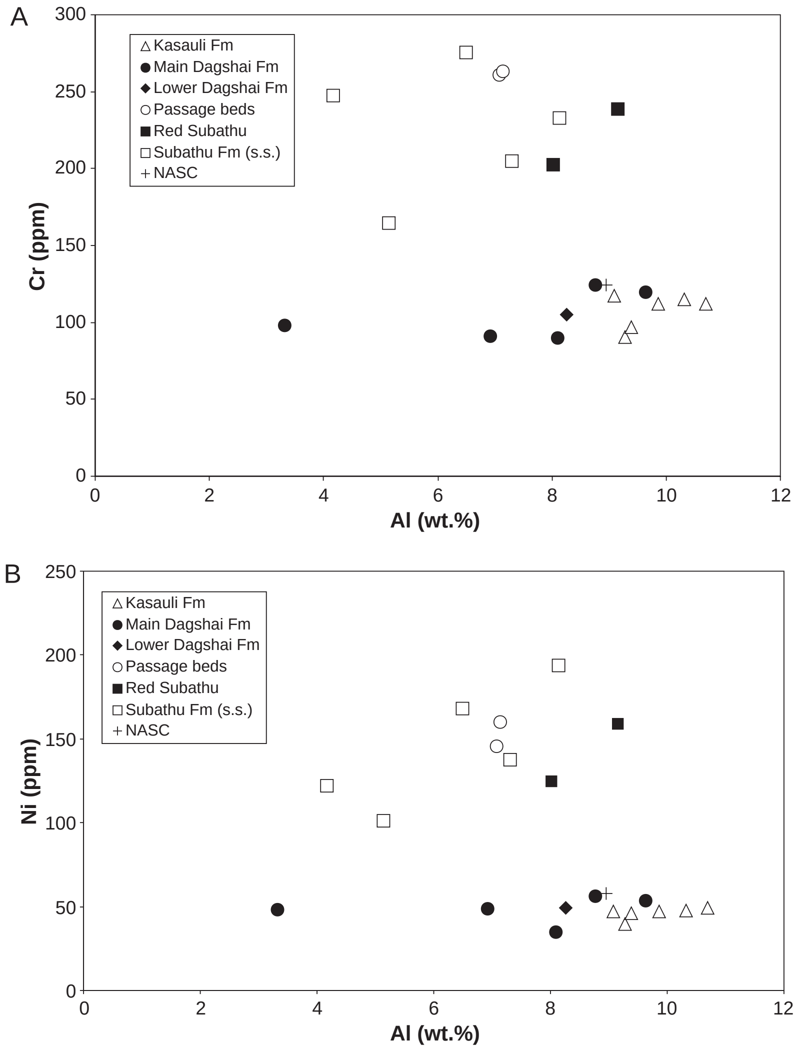


Fig. 8. Composition of foreland basin mudstones, Subathu sub-basin, India, determined by XRF. High Cr (Fig. 8A) and Ni (Fig. 8B) contents are indicative of contribution from a mafic igneous source. Data is plotted again Al to test for the effects of dilution. NASC = North American Shale Composite, average composition for passive margin sediments. Subathu s.s. (sensu stricto) refers to main green Subathu Formation facies. From Najman and Garzanti (2000).

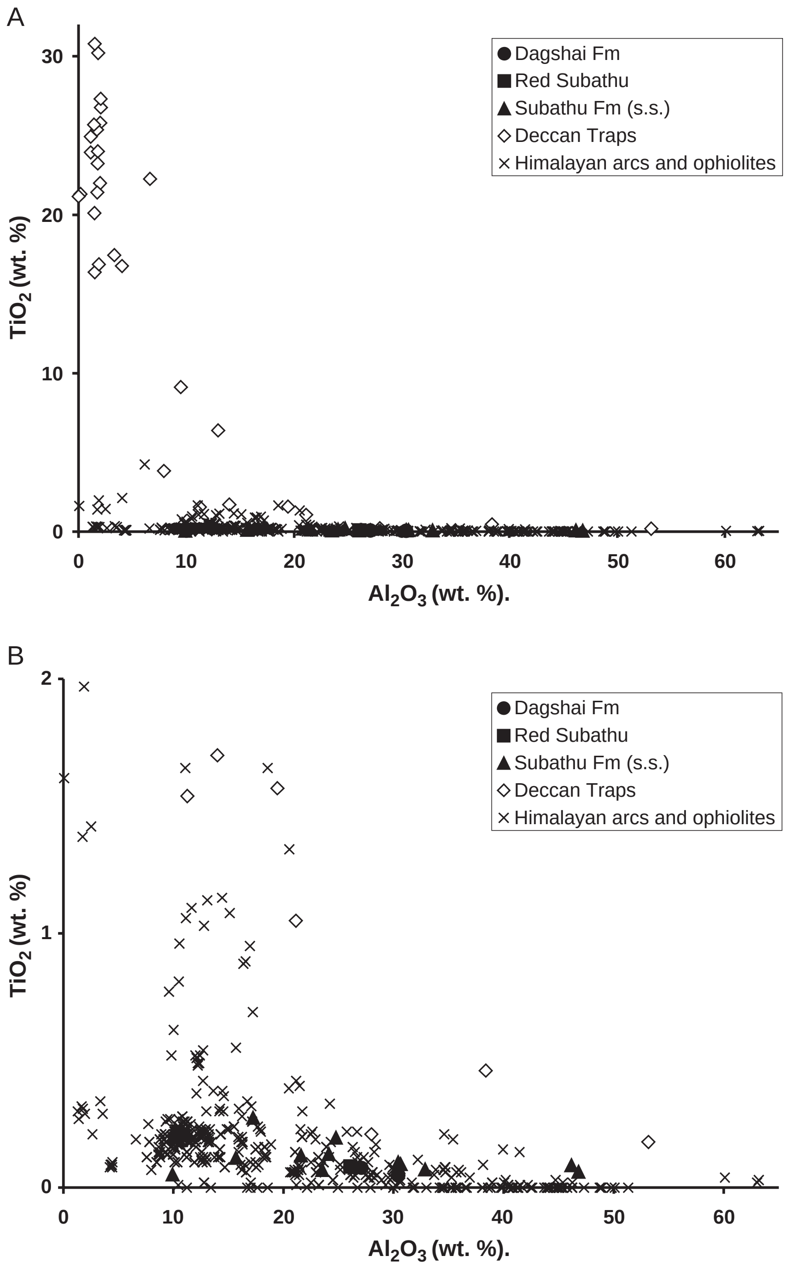


Fig. 9. Graph A, and the lower part of its Y axis expanded in Graph B, show composition of spinels from the Eocene Subathu Formation foreland basin rocks, Subathu sub-basin, India (Najman and Garzanti, 2000), plotted for comparison with spinels from Himalayan ophiolite and arc rocks and Deccan Trap spinels. Data taken from spinel database of Barnes and Roeder (2001) and references therein. Analyses with wt% totals less than 98% or greater than 102% are excluded. Typically, spinels from continental flood basalts, such as the Deccan Traps, have higher TiO_2 content compared to spinels from ophiolite and arc rocks, as illustrated in the example above. Subathu Formation spinels resemble spinels found in Himalayan ophiolite and arc rocks, and are unlike those found in the Deccan Traps.

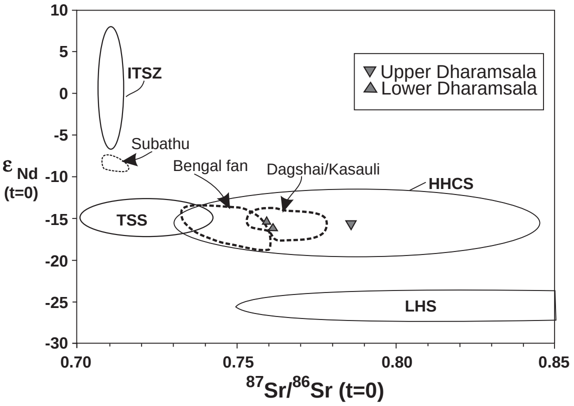


Fig. 10. ε_{Nd} and $^{87}Sr/^{86}Sr$ for source regions (ITSZ = Indus Tsangpo suture zone, TSS = Tibetan Sedimentary Series/Tethys Himalaya, HHCS = Higher Himalaya, LHS = Lesser Himalaya) and foreland basin deposits of India. Data from White et al. (2002) and Najman et al. (2000). Also shown is field for Bengal Fan, from France-Lanord et al. (1993). Source region references given in text.

5.1.2. Indian foreland basin

As described in Section 3.2.2, Fig. 3, the regions studied are the Subathu sub-basin, consisting of the Eocene marine Subathu facies, unconformably overlain by the Oligo–Miocene alluvial Dagshai and Kasauli Formations, and the Kangra sub-basin consisting of the Miocene Dharamsala Formation (Raiverman et al., 1983). Miocene–Recent alluvial Siwalik deposits are found in both sub-basins.

5.1.2.1. Data

5.1.2.1.1. Subathu Fm (Subathu sub-basin, Eocene aged). Petrographic parameters show a predominantly sedimentary provenance with subordinate felsite and less common serpentine schist input in the main (green) Subathu Formation. Red Subathu facies contain a high proportion of volcanic, mostly felsitic, detritus (Figs. 4 and 7). High proportions of Ni and Cr in mudstones (Fig. 8) are also indicative of a mafic contribution. Heavy mineral analyses show the presence of spinel (Fig. 7). Spinel geochemistry indicates that the source region was ophiolitic/arc derived, rather than from continental flood basalts from the Indian craton (Section 4.4.1, Fig. 9) (see also Najman and Garzanti, 2000). Although the use of spinel geochemistry has been called into question by Power et al. (2000), in this case, the similarity between the spinels of the Subathu Formation and those of the Himalayan ophiolites and arc and the dissimilarity with documented spinel compositions from the Deccan Traps, suggests that the interpreted provenance is correct. Both the Kohistan Island Arc and the southern Ladakh ophiolites (e.g., Spontang) contain spinels which span the range of Cr# and TiO_2 values exhibited by the detrital spinels in the Subathu Formation (e.g., Jan and Windley, 1990; Jan et al., 1992, 1993; Arif and Jan, 1993; Maheo et al., 2004) whilst TiO_2 values in spinels from the Deccan Traps are higher. Sr–Nd data

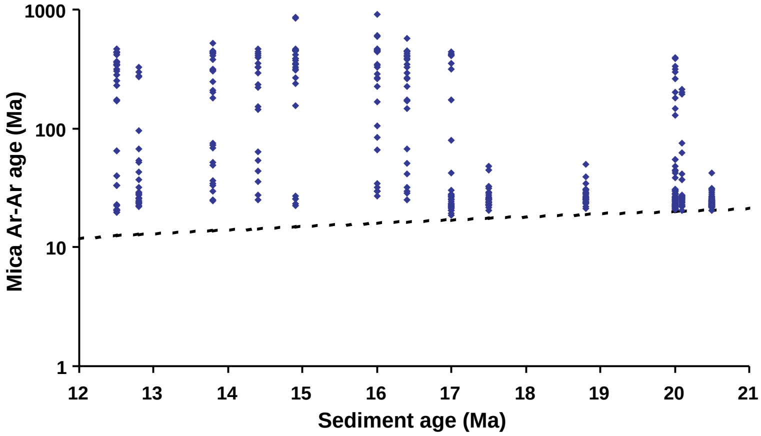


Fig. 11. Sediment depositional age (determined by magnetostratigraphy) vs. Ar–Ar ages of detrital white micas from the Dharamsala Formation foreland basin sandstones, India. Dashed line = line of zero lag time, where mica age equals sediment depositional age and rapid exhumation is indicated. The data show an increase in lag times at 17 Ma, accompanied by an increase in pre-Himalayan aged grains. This is interpreted as the time when the apex of erosion shifted from the metamorphosed Higher Himalaya to the footwall of the MCT (see Fig. 13). From White et al. (2002).

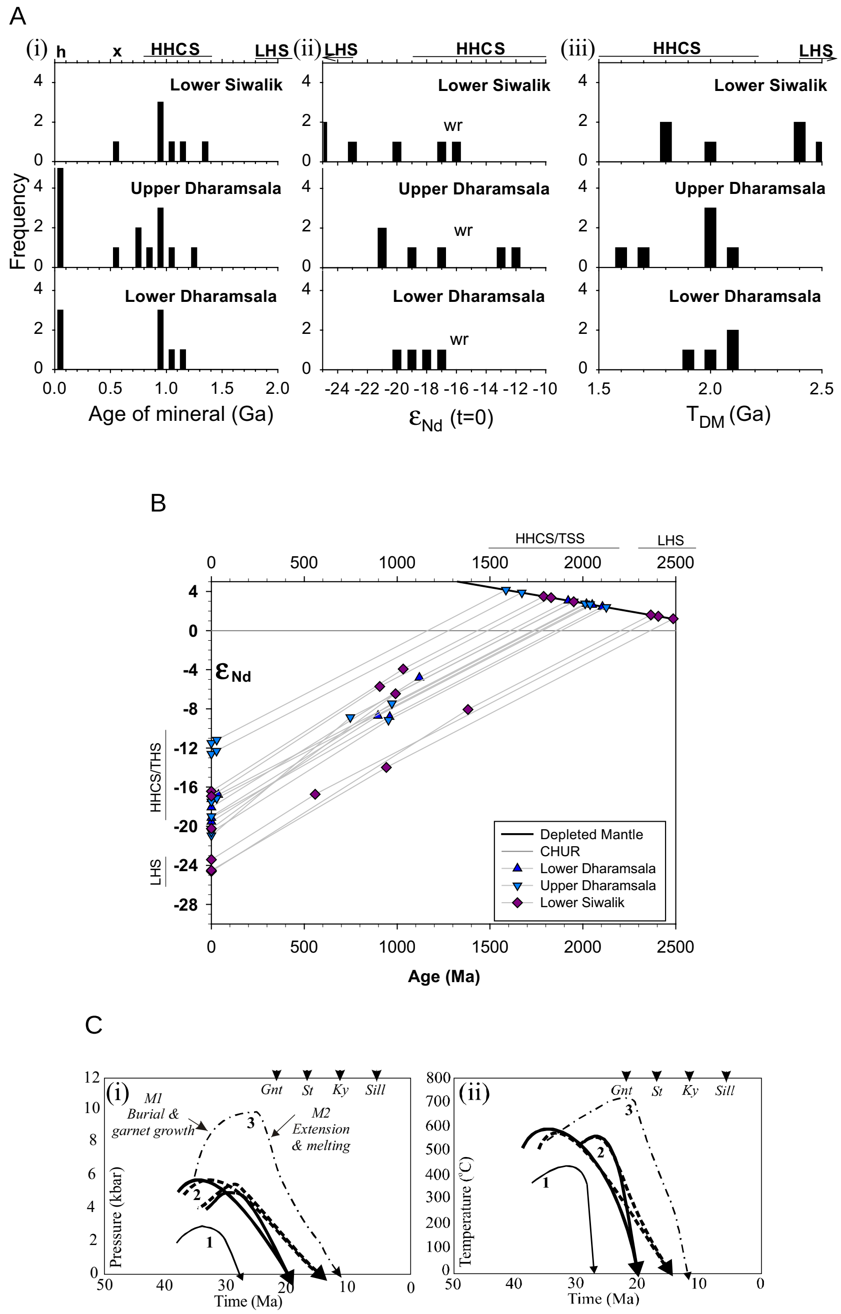
(Section 4.6.1*)* support a northern provenance, with mixed contribution from the Tibetan Sedimentary Series of the Indian passive margin and the Indus Tsangpo Suture Zone (Fig. 10), (Najman et al., 2000).

5.1.2.1.2. Dagshai and Kasauli Formations (Subathu sub-basin, Oligo–Miocene aged). In the Dagshai Formation, petrography shows that lithic fragments are predominantly of very low grade to low grade metamorphic type (Fig. 7). Mafic input decreases compared to the underlying Subathu Formation, as evidenced by whole rock geochemistry (Fig. 8) and only exceedingly rare occurrences of mafic minerals such as spinel. "Himalayan aged" micas and zircons (Section 3.2.2); (Najman et al., 1997, 2004), and High Himalayan ε_{Nd} signatures (Fig. 10, Najman et al., 2000) all attest to High Himalayan provenance. Characteristics of the Kasauli Formation are similar to those of the underlying Dagshai Formation (Figs. 4d, 7, 8 and 10) but petrography and heavy mineral data indicate erosion to deeper levels of the Higher Himalayan metamorphic orogen. There is an increase in metamorphic grade of lithic fragments, and the first appearance of garnet, of almandine type, typical of derivation from regionally metamorphosed schists (Najman and Garzanti, 2000). Detrital apatite fission track ages of 5 Ma, interpreted as reset post-deposition, record cooling in the basin at that time (Najman et al., 2004).

5.1.2.1.3. Dharamsala Formation (Kangra sub-basin, 20–13 Ma). Micas (Fig. 11) and monazite ages (Fig. 12A–i), Sr–Nd whole rock and monazite compositions (Figs. 10, 12a-ii and b), petrography (Fig. 7) and heavy minerals (Fig. 4d) document erosion of the garnet–staurolite grade Higher Himalaya during deposition of the Lower Dharamsala Member, from 20–17 Ma (White, 2001; White et al., 2001, 2002). Monazite ages (Section 4.5.2) indicate that the source was metamorphosed between 37–28 Ma. Short mica lag times (<3 Ma) indicate that this source was then exhumed rapidly. Simple thermal modelling demonstrates that the observed distribution of ages can be simulated with high rates of exhumation, up to 5 mm/yr prior to 20 Ma. At 17 Ma, deposition of the Upper Dharamsala Member began, accompanied by an abrupt change in petrographic characteristics and mica age populations. A major influx of micas with pre-Himalayan ages, and lithic fragments of dominantly sedimentary and very low to low grade metamorphic character (Fig. 7), indicate an additional new source region. Consistent with the Higher Himalayan-like signatures of the monazite T_{DM} and ε_{Nd} data (Fig. 12A-ii and iii), and whole rock Sr–Nd data (Fig. 10), such a source should be low grade material of Higher Himalayan isotopic character. White et al. (2002) interpret the source to be unmetamorphosed/low grade Higher-Himalayan protolith, likely the Haimanta Formation (Thakur, 1998; Steck, 2003) (Fig. 13). Similar rocks defined as the Outer Lesser Himalaya by Ahmad et al. (2000) (Section 4.6.1) would also be compatible. The lack of detrital mica grains

Fig. 12. A: Histograms showing distributions of: (i) concordant U–Pb, U–Th–Pb ages of detrital monazite and zircon from the Dharamsala and Lower Siwalik Formations. For comparison the range of U–Pb ages of detrital zircons from metasediments of the High Himalayan Crystalline Series (HHCS), and Lesser Himalayan Series (LHS) are shown on the upper x-axis (Parrish and Hodges, 1996; DeCelles et al., 2000), h is the time of Himalayan collision and x the time of intrusion of Cambro–Ordovician orthogneisses and granites. Note that zircons are recording protolith ages, monazite ages are recording both protolith ages and ages resulting from Himalayan metamorphism. (ii) ε_{Nd} (t=0) of detrital monazite (also analysed by U–Pb methods). For comparison the ε_{Nd} (0) of whole rock samples from the same detrital sedimentary units are annotated by "wr", and the range of whole rock ε_{Nd} (0) of samples from the HHCS and LHS are displayed on the upper x-axis (from Ahmad et al., 2000). (iii)T_{DM} of detrital monazite (calculated as discussed in White, 2001). The range of T_{DM} of the Himalayan units are displayed for comparison on the upper x-axis (from Ahmad et al., 2000). Note that whereas the detrital ages would suggest a High Himalayan/Tibetan Sedimentary Series source, the ε_{Nd} (0) and T_{DM} of three grains from the Lower Siwalik sub-group have a similar isotopic composition to the Lesser Himalaya. From White et al. (2001) and White (2001). B: ε_{Nd} vs. time (Ma) plot, showing results of calculation of ε_{Nd} at time=0 (symbols lie on the y-axis), ε_{Nd} at time of monazite crystallisation (as calculated from U–Pb ages on the same grains), and the probable modal age (T_{DM}). For comparison, the ranges of ε_{Nd}(0) and T_{DM} for samples of the High Himalayan Crystalline Series (HHCS), the Tibetan sedimentary series (TSS) and Lesser Himalayan Series (LHS) are shown on the x (ε_{Nd}(0)) and y (T_{DM}) axes. From White (2001). C: Pressure–time (i) and Temperature–time (ii) paths. From White et al. (2001). 1. Trajectory (thin line) for rocks of sub-garnet grade, which were eroded by Dagshai Formation times (based upon mica ages, Najman et al., 1997). 2. Trajectories (thick solid line — Lower Dharamsala, thick dash line — Upper Dharamsala) from garnet–staurolite grade metamorphism (crystallisation of monazite) to deposition in Dharamsala Formation. 3. Trajectory (dash dot line) for presently exposed High Himalayan kyanite-grade pelites from Zanskar (Vance and Harris, 1999; Searle et al., 1999a,b). The time of first appearance in the detrital record of metamorphic index minerals is shown.

older than 1000 Ma is not consistent with Inner Lesser Himalaya contribution. However contribution from the Inner Lesser Himalaya cannot be ruled out since the Ar–Ar mica ages of Lesser Himalayan rocks are currently poorly characterized (Section 4.5.1) and not abundant in the rocks, and the Sr–Nd whole rock technique may not be sufficiently sensitive to pick up such an influx, when diluted by High Himalayan material. However, there is no specific need to invoke this provenance source.



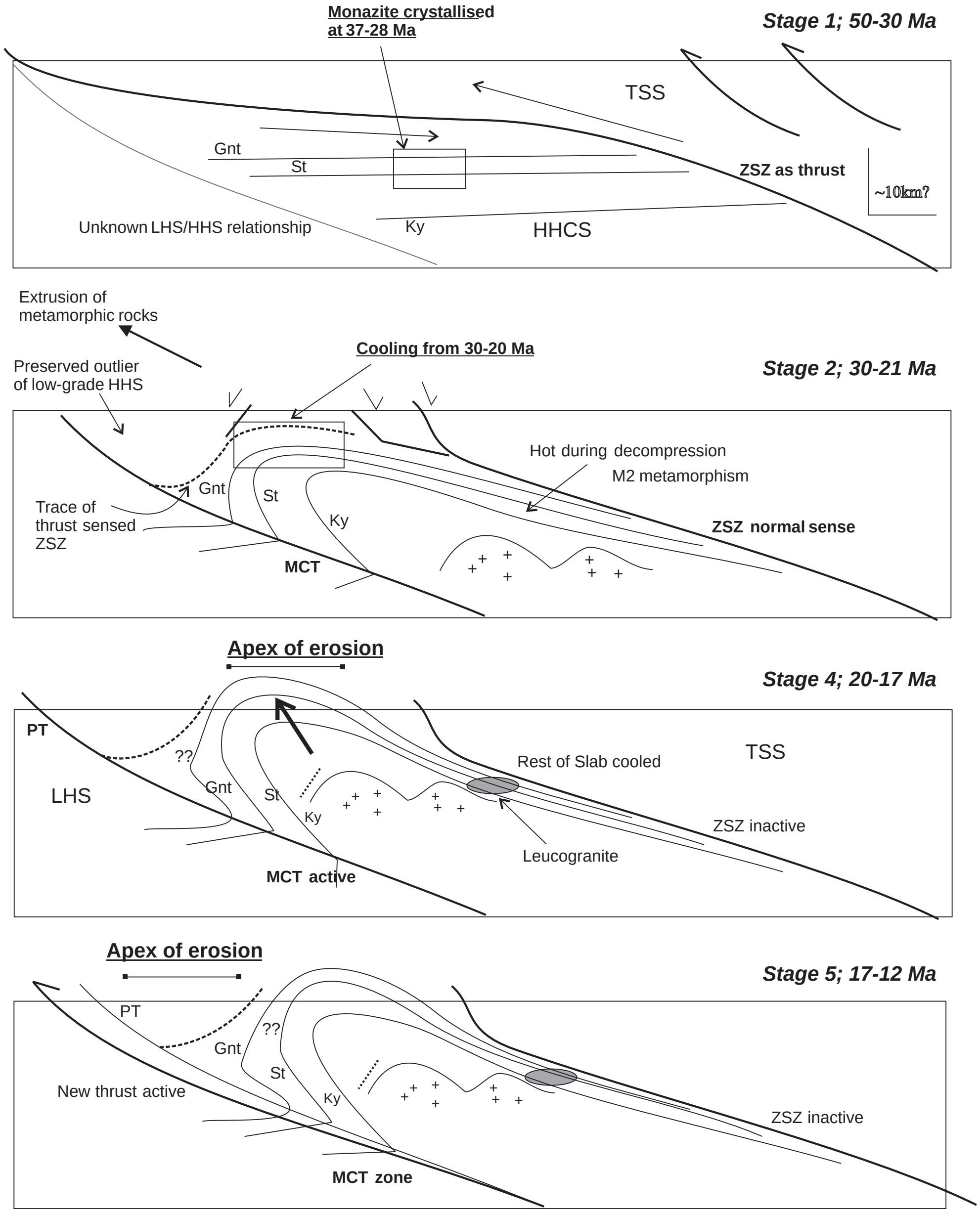


Fig. 13. Schematic model for the evolution of the High Himalayan Slab in NW India. Contributions from the detrital record, as documented in White (2001), White et al. (2001, 2002), are shown as bold and underlined. TSS: Tethyan Sedimentary Series, HHCS: High Himalayan Crystalline Series, LHS: Lesser Himalayan Series, ZSZ: Zanskar Shear Zone (part of south Tibetan Detachment Zone), MCT: Main Central Thrust, PT: Panjal Thrust. Isograds: gnt = garnet, st = staurolite, ky = kyanite, ++ = sillimanite. Diagram taken from White (2001), compiling her interpretation and previous work of others as referenced therein.

Continued occurrence of garnet and staurolite detritus, Himalayan-aged monazites, a small fraction of the mica population with Himalayan ages, and whole rock Sr–Nd values of Higher Himalayan signature, indicate continued contribution from the metamorphosed Higher Himalaya. Higher Himalayan exhumation rate was reduced however (lag times 5–7 My, exhumation rate 1 mm/yr) compared to that occurring during Lower Dharamsala Formation times.

5.1.2.1.4. Siwalik Group (Kangra sub-basin, <13 Ma). Siwalik Group rocks show the first appearance of kyanite at 13 Ma, followed by sillimanite at 8 Ma (Najman et al., 2003a,b). Himalayan-aged micas are common. Youngest mica ages remain constant throughout the dated section (13 to 4 Ma), and thus lag times increase up section (Najman et al., 2002b). These data confirm that significant contributions from the now more slowly exhuming Higher Himalaya continued during Siwalik Formation times. At 10 Ma there is an abrupt facies change, from sandstone to conglomeratic facies of Lesser Himalayan provenance as determined by ε_{Nd} signatures (Meigs et al., 1995; Najman et al., 2002b). This is approximately coeval with the timing of increased accumulation/subsidence rate at 11 Ma (Meigs et al., 1995; Burbank et al., 1996).

5.1.2.2. Interpretations. Subathu Formation provenance clearly shows that these rocks are not passive margin facies. A foreland basin had developed by this time into which northerly-derived igneous detritus was being deposited (Najman and Garzanti, 2000). Both the ophiolitic rocks south of the Indus–Tsangpo Suture, and the Kohistan–Ladakh arc and Trans Himalaya north of the suture provide suitable sources for the igneous material. However, the detrital igneous material in the Subathu Formation is predominantly felsic, indicating most probable derivation from north of the suture since only very minor amounts of andesitic material are associated with the predominantly mafic–ultramafic ophiolites to the south (Pedersen et al., 2001; Maheo et al., 2004).

The occurrence in the foreland basin of material derived from north of the suture zone indicates that India–Asia collision had occurred by this time.

Derivation of igneous material from ophiolites south of the suture zone would also constrain timing of collision whether the ophiolites are considered to have been obducted synchronous with continental collision in the Eocene, or prior to continental collision in the Cretaceous (see Section 2). Even if obduction was in the Cretaceous, such detritus could not simply represent continued erosion of the obducted ophiolite prior to continental collision because the Palaeocene, more northern, Dibling limestone shows no evidence of ophiolitic input. Ophiolitic detritus into the foreland basin would either be associated with the ophiolite's rethrusting in the Eocene, associated with collision, or with initial obduction of the ophiolite in the Eocene, again associated with collision.

The dramatic shift to low grade metamorphic provenance, with the elimination of the igneous source in Dagshai Formation times indicates the development of a thrust barrier composed of upper crustal levels of the Higher Himalaya by the close of the Oligocene. Development of this thrust barrier resulted in burial and Barrovian metamorphism in the Higher Himalaya. U–Th–Pb ages of 37–28 Ma on detrital monazites from the Dharamsala Formation confirm their origin from the same prograde phase of the metamorphic cycle as reported by Sm–Nd and U–Th–Pb ages between 45–25 Ma from garnets and monazites in the Higher Himalaya (Prince et al., 1999; Vance and Harris, 1999; Foster et al., 2000). The products of this initial phase of burial were then exhumed to the surface by 20 Ma, at cooling rates between 60–40 °C/Ma. These monazite ages, coupled with petrographic and mica age constraints, allow a portion of the P–T–t path of the exhumed rock to be constrained (Fig. 12c, White et al., 2001). This is consistent with simple one dimensional models of over-thickened crust in which less deeply buried rocks are heated less and exposed earlier (England and Richardson, 1977; England and Thompson, 1984).

The slowdown in exhumation rates of the Higher Himalaya, as determined from lag time calculations, is consistent with cessation of movement along the bounding structures of the MCT and STDZ (Hubbard and Harrison, 1989; Hodges et al., 1996; Dezes et al., 1999; Walker et al., 1999). This change in exhumation rate, accompanied by the distinct provenance change to lower metamorphic grade, signifies a major reorganisation of the orogenic wedge geometry in the NW Himalaya at 17 Ma, with propagation of thrusting into the footwall of the MCT envisioned at this time (Fig. 13; White, 2001; White et al., 2002). Continued denudation of the Higher Himalaya to increasing metamorphic levels during this period of slow exhumation is indicated by the progressive appearance of metamorphic index minerals of increasing grade.

Exhumational steady state, as defined by Willett and Brandon (2002) can be recognized in the sedimentary record as periods when lag times remain constant through time (Garver et al., 1999). The lag times throughout the period from 20–6 Ma indicate that the region of Higher Himalaya eroded into the catchment area studied was not exhuming in steady state during this time (Najman et al., 2002b).

At 10 Ma, the increased subsidence and major facies change to conglomeratic units, with Lesser Himalayan clasts, suggested to Meigs et al. (1995) and Brozovic and Burbank (2000) that the MBT was active by this time. This is consistent with the timing of further southward propagation of thrusting into the basin prior to 5 Ma, as evidenced by reset detrital apatite fission track ages (Najman et al., 2004).

5.1.3. Nepal foreland basin

As outlined in Section 3.2.3, Eocene–Recent foreland basin rocks in Nepal overlie the Cretaceous–Palaeocene Lesser Himalayan rocks of the Amile Formation. The foreland basin strata consist of the Lower-Mid Eocene marine Bhainskati Formation, overlain unconformably by alluvial deposits of the Miocene to Recent Dumri Formation and Siwalik Group.

5.1.3.1. Data

5.1.3.1.1. Bhainskati Formation (Lower to mid Eocene). Although petrographic parameters are similar to the underlying Amile Formation passive margin facies, i.e., pure quartz arenites with no indication of derivation from the Himalaya (Fig. 14), U–Pb and fission track ages of detrital zircons differ from those of the Amile Formation. Fission track data show that zircons in the Bhainskati Formation contain Himalayan-aged (syn-collisional) populations (Fig. 15) (Najman et al., 2005). U–Pb data show that whilst the vast majority of Amile Formation detrital zircons have ages in the range ca. 1800–2500 Ma, consistent with derivation from the Indian craton, Bhainskati Formation grains show a significant component of younger zircons in the range ca. 500–1000 Ma (Fig. 16) (DeCelles et al., 2004). Although these younger grains resemble those from source lithologies of the Higher Himalaya, DeCelles et al. consider provenance from the Tibetan Sedimentary Series, itself derived from the Higher Himalaya, as most likely since the Higher Himalaya was not yet exhumed to the surface at this stage (Section 4.5.2). The less negative ε_{Nd} values of the Bhainskati Formation compared to that

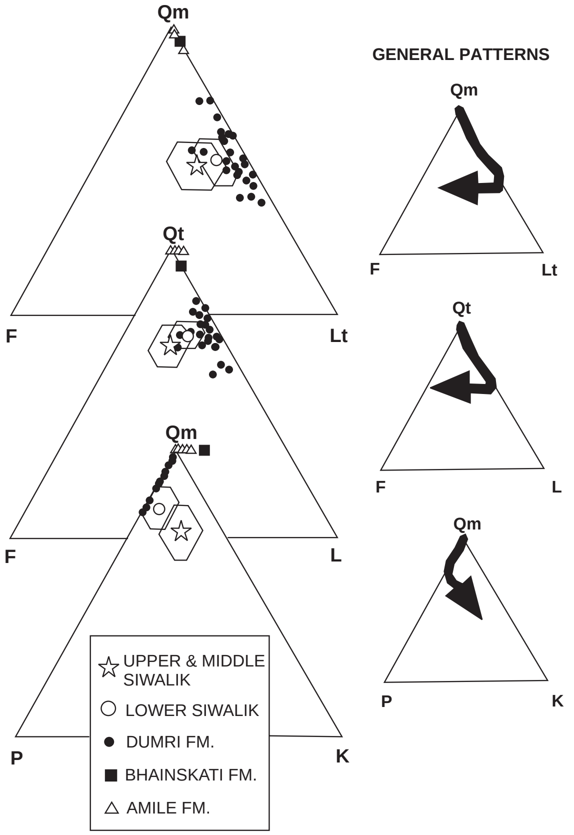


Fig. 14. Petrographic compositions, showing evolution from passive margin facies (Amile Formation) to foreland basin strata (Bhainskati, Dumri and Siwalik Formations) in Nepal. From DeCelles et al. (1998a). Qm = monocrystalline quartz, F = total feldspar, Lt = total lithic fragments, Qt = total quartzose grains, L = non-quartzose lithic fragments, P = plagioclase, k = k-spar.

expected from a Higher Himalayan provenance (Fig. 17) may also be the result of erosion from the Tethyan succession (Robinson et al., 2001) or input from the suture zone (Section 4.6.1).

5.1.3.1.2. Dumri Formation (21–16 Ma). Metasedimentary lithic clasts in the Dumri Formation present the first evidence of derivation from the rising metamorphosed Himalayan fold thrust belt (Fig. 14). A U–Pb detrital zircon population with ages similar to that of the Bhainskati Formation and Siwalik Group confirm provenance from the Tibetan Sedimentary Series/Higher Himalaya (DeCelles et al., 1998a,b), (Fig 16). The presence of plagioclase suggests erosion from Higher Himalayan crystallines, ε_{Nd} values match

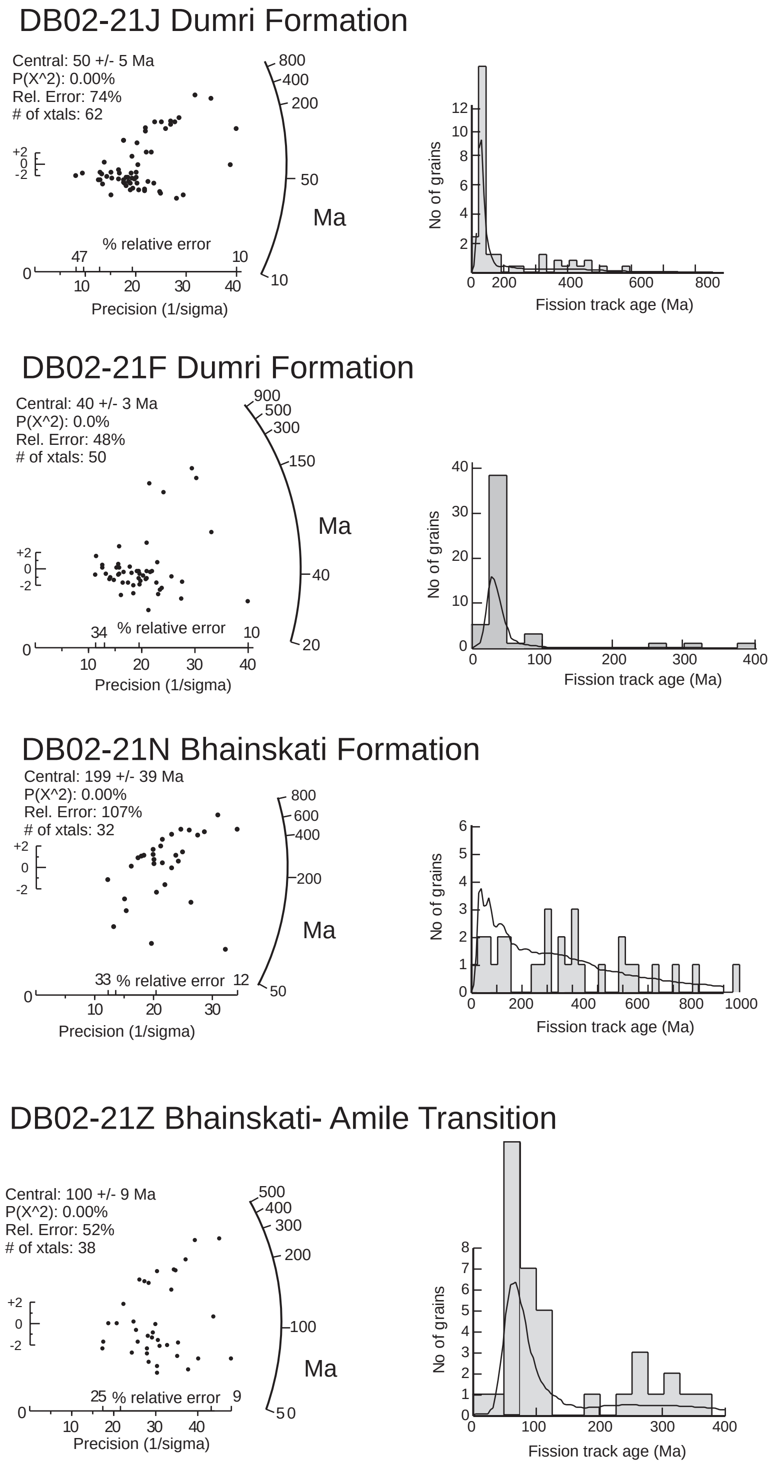


Fig. 15. Radial plots (Galbraith, 1988) and histograms showing distribution of detrital zircon fission track ages, from foreland basin strata, Nepal. Radial axes in Ma. From Najman et al. (2005).

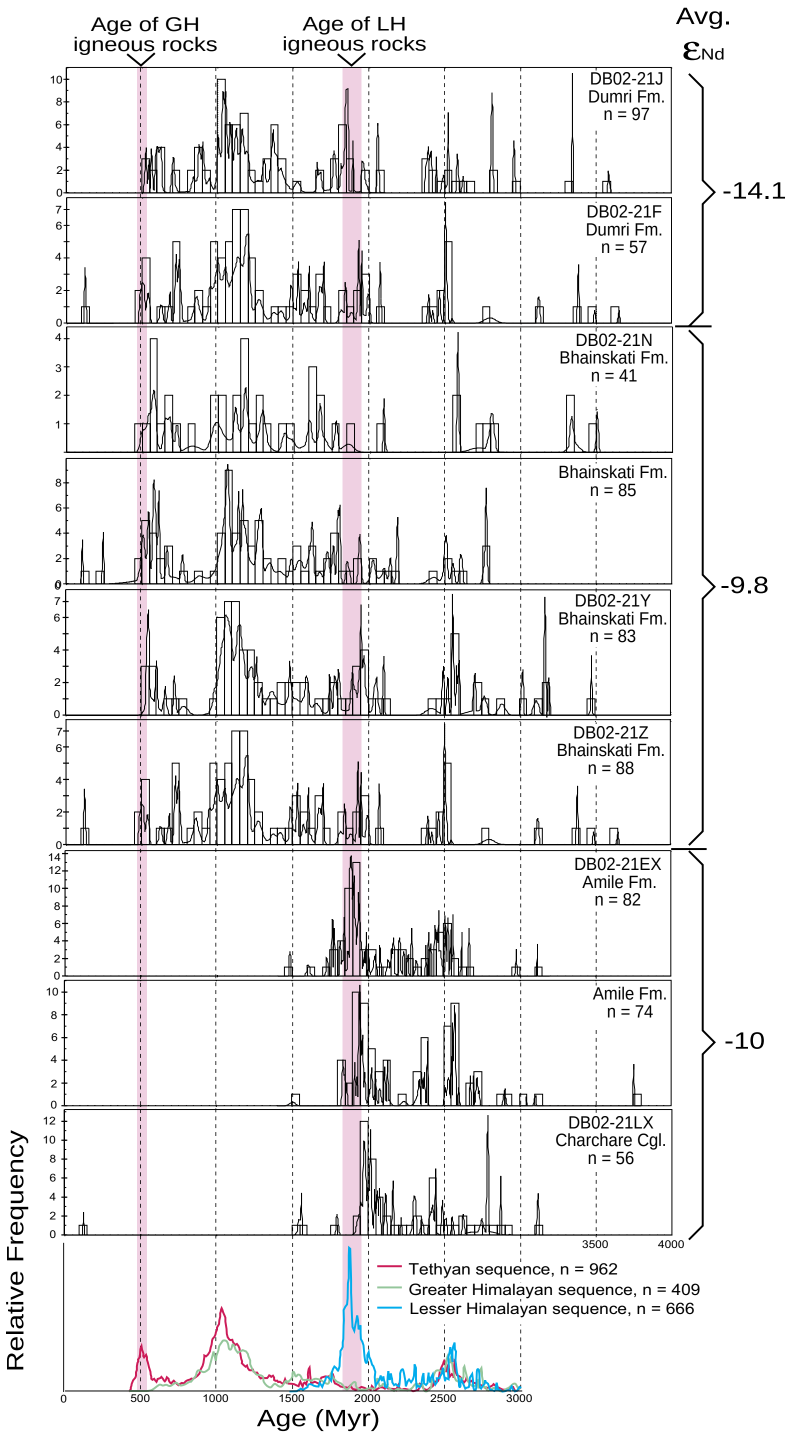


Fig. 16. Histograms showing distributions of concordant U–Pb ages of detrital zircon grains from Amile passive margin rocks, and Bhainskati and Dumri foreland basin rocks, Nepal. Also shown (base) are analyses of zircons from bedrock samples in the potential source regions of the Himalayan lithotectonic zones. From DeCelles et al. (2004).

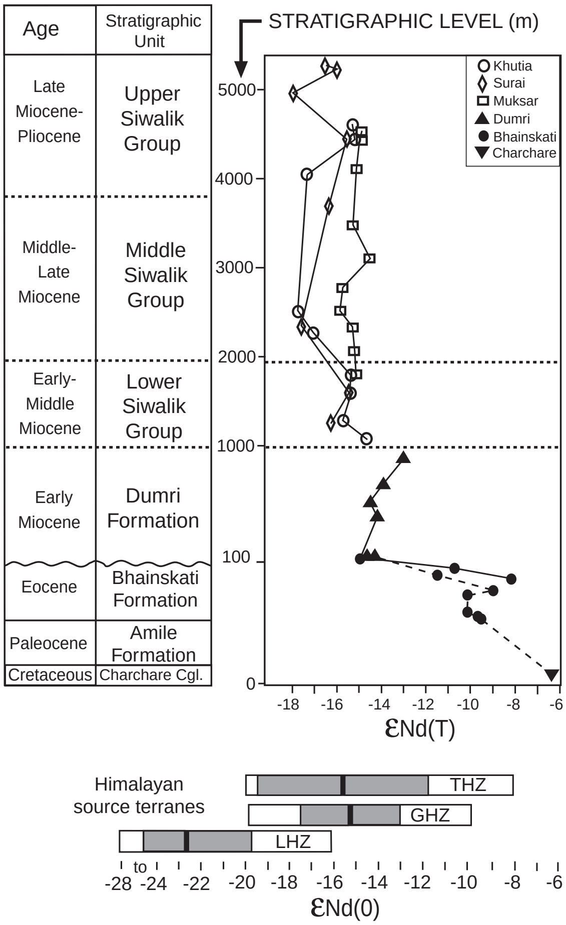


Fig. 17. ε_{Nd} values of foreland basin rocks of the Bhainskati, Dumri and Siwalik Formations, Nepal, and underlying passive margin (Chachare Formation). Khutia = Siwalik section in western Nepal, Surai = Siwalik section in central Nepal, Muksar Khola = Siwalik section in eastern Nepal. Also shown are source region values (THZ/TSS = Tibetan Sedimentary Series, GHZ/HH = Greater Himalayan zone [Higher Himalaya], LHZ = Lesser Himalayan Zone). From DeCelles et al. (2004).

the signature values of the Higher Himalaya (Robinson et al., 2001), (Fig. 17) and detrital mica Ar–Ar ages and detrital zircon fission track ages (Section 3.2.3) confirm Higher Himalayan input.

5.1.3.1.3. Siwalik Group: (15 Ma–Recent). Continued Higher Himalayan contribution is significant throughout this period. This is evidenced by a detrital zircon population with U–Pb ages which match those of a Higher Himalayan source (DeCelles et al., 1998b), Sm–Nd data (Huyghe et al., 2001; Robinson et al., 2001) (Fig. 17), Neogene Ar–Ar mica ages (Szulc, 2005; Szulc et al., submitted for publication) and petrographic data such as the continued occurrence of plagioclase and first appearance of high grade metamorphic minerals (DeCelles et al., 1998b). Garnet appears in the Lower Siwalik subgroup, with staurolite, kyanite and sillimanite appearing later in the same sub-group in the sections studied by DeCelles et al. (1998b) but not until the Middle Siwalik subgroup in the sections studied by Szulc et al. (submitted for publication). Increasing contribution from the Lesser Himalaya and the passively uplifted overlying Crystalline Thrust Sheets (Dhadeldura Thrust sheet — Fig. 18) is discerned in western Nepal from ca 11 Ma (DeCelles et al., 1998b; Huyghe et al., 2001; Robinson et al., 2001; Szulc, 2005; Szulc et al., submitted for publication). Lesser-Himalayan derived carbonate clasts and crys-

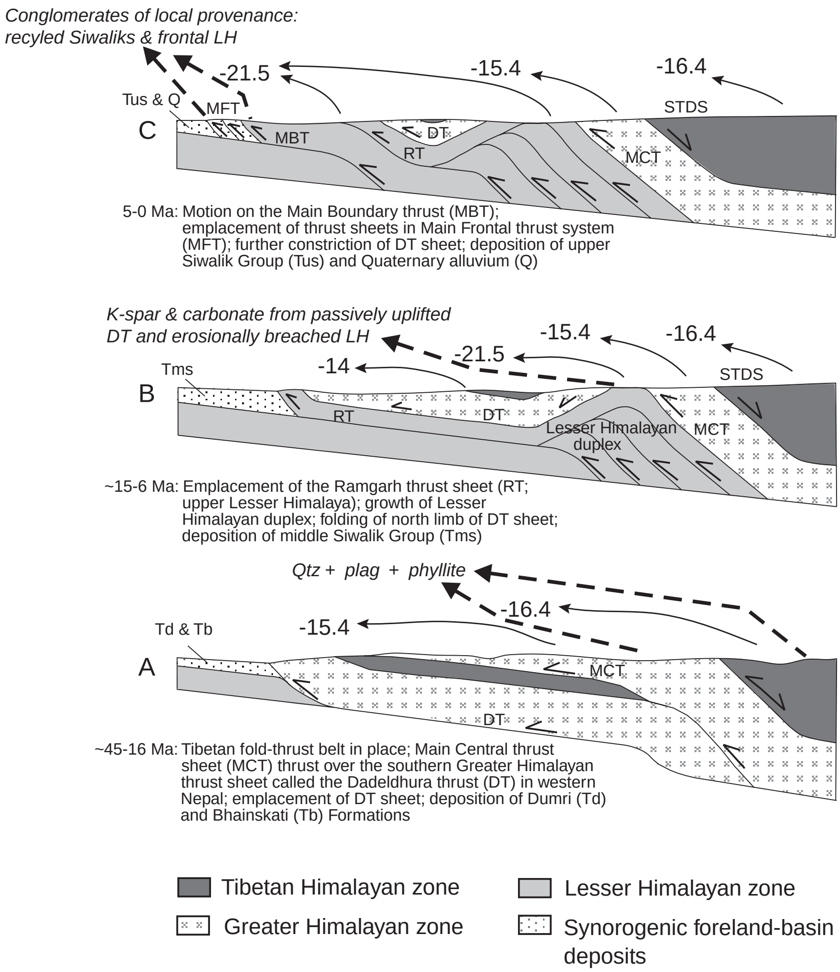


Fig. 18. Schematic reconstruction of the evolution of the Himalayan fold–thrust belt in western Nepal (from Robinson et al., 2001). Solid arrows and negative numbers are ε_{Nd} values of the tectonostratigraphic terrains that contributed detritus to the sediment record (as shown in Fig. 17). Dashed arrows and italic text are key new petrographic compositions of the foreland basin deposits at times shown, and their interpreted provenance (from DeCelles et al., 1998b — their Fig. 15).

talline-thrust-sheet derived K-spar increases throughout Siwalik times, along with the proportion of Lesser-Himalayan sourced detrital zircons with U–Pb ages >2 Ga (DeCelles et al., 1998b). ε_{Nd} values decrease at 10–11 Ma which is also indicative of significant contribution from the Lesser Himalaya at this time. Timing is approximately coeval with coarsening grain size at a microscopic (Huyghe et al., 2001) and macroscopic scale (Nakayama and Ulak, 1999) and with an increased sediment accumulation rate (Burbank et al., 1996). By contrast, there is as yet no evidence for a similar trend of increased Lesser Himalayan contribution at this time in eastern Nepal (Robinson et al., 2001). Finally, proximal facies are indicated by the appearance of Lesser Himalayan conglomerate clasts at ~4–5 Ma.

Szulc (2005) and Szulc et al. (submitted for publication) used lag time data (Section 4.5.1) determined from detrital micas in the Siwalik rocks aged from 16 Ma to investigate exhumation rates of the Higher Himalaya. At no time were zero aged lag times encountered. The shortest lag time (3.6 My) occurred at 16 Ma, and there were no lag times shorter than 5 My after 10 Ma. By contrast, shorter lag times were documented during the period 8–2 Ma from a different Siwalik sedimentary section, by Harrison et al. (1993)

who used detrital K-feldspar. They interpreted lag times of <3 My throughout this period as indicative of rapid exhumation of a part of the hinterland during this time, whilst the paucity of Palaeogene grains compared to abundant grains younger than 21 Ma indicates that the source region was not subject to rapid unroofing during the Palaeogene. Ar–Ar ages of micas from modern day sediments have minimum lag times of 4 Ma (Brewer et al., 2000), with Bertrand et al. (2001) recording some mica ages as young as 1.3+/−1.1 Ma.

5.1.3.2. Interpretations (Fig. 18). The U–Pb and fission track ages of detrital zircons, which provide evidence of Himalayan erosion in the Eocene, bring the sedimentary evidence into agreement with the determined age of crustal thickening, burial and metamorphism in the mountain belt (see Section 2). Denudation of a right way up metamorphosed High Himalayan succession was occurring by 20 Ma. Because no lag times studies have been conducted on the Dumre Formation, the period of rapid exhumation documented from strata with very short lag times between 20–17 Ma in India, has not been documented in the detrital record of Nepal. Lag times of less than 5 My persist in the sediment record in West Nepal until 10 Ma however, indicating continued exhumation of the Higher Himalaya, albeit at slower rates. In central Nepal, periods of relatively rapid exhumation of the Higher Himalaya continue to present day.

Provenance data from western Nepal is interpreted as showing significant input of Lesser Himalayan material by ca. 12 Ma. This constrains the timing of movement on the Lesser Himalayan thrust system. DeCelles et al. (1998b) use their data to produce the following tectonic scenario (Fig. 18): erosional breaching of the Lesser Himalayan duplex from beneath the crystalline (Dhadeldura) thrust sheets occurred at ca. 11 Ma. Movement along the major Lesser Himalayan Ramgarh thrust is constrained as prior to this time since it forms the roof thrust to the duplex, and maximum age constraint is provided by its cross-cutting relationship to the Dumri Formation (DeCelles et al., 2001). The appearance of Lesser Himalayan conglomerate clasts in the sedimentary succession at ~4–5 Ma indicates proximal facies, suggesting the movement of the more southerly MBT thrust at this time. The lack of a similar trend to increasing Lesser Himalayan contribution in eastern Nepal, indicates less erosional unroofing in this region (Robinson et al., 2001). This is consistent with the outcrop pattern, although much less research has been undertaken in this area.

5.2. The suture zone and Tethyan Himalaya

As outlined in Section 3.1, Palaeogene and Neogene marine and continental strata are preserved in the suture zone in Ladakh in the west, whilst in the east, post-collisional sedimentary rocks are confined to Upper Oligocene–Lower Miocene Gangrinboche conglomerates. The Tethys Himalaya is not considered a major sediment repository, but where sediments are critical to concepts of Himalayan evolution, they have briefly been included here.

5.2.1. Data

The Palaeocene Chogdo Formation is characterised by alluvial and lacustrine facies and NE-directed palaeocurrents (Searle et al., 1990; Clift et al., 2001a). Conglomerate clasts include platform sediments, radiolarian chert, ophiolitic and granitic material, indicative of provenance from the Spongtang ophiolite to the south and the Trans-Himalaya to the north. ε_{Nd} values are relatively high (Fig. 19), similar to those of the Trans-Himalaya, and Pb isotope compositions of feldspars also fall within the Asian plate fields (Fig. 20). The overlying Nurla Formation contains a higher proportion of granitoid material from the intrusive core of the magmatic arc, and much higher concentrations of the trace elements V, Cr, Co and Ni typically associated with ultramafic rocks (Garzanti and Van Haver, 1988). The petrography of the Nurla and the overlying Nimu Formation fall into the magmatic arc field on the QFL diagram (Fig. 21), with conglomerate clast compositions including metasediments (pelite, micrite and chert), granodiorite, peridotite, and acid lavas and tuffs derived from the Ladakh batholith (Sinclair and Jaffey, 2001). ε_{Nd} values show a shift towards more negative values compared to the Chogdo Formation (Fig. 19), indicating additional input from the Lhasa block, Karakoram or Indian crust, as well as the dominant Trans-Himalayan source. Detrital K-spars predominantly plot in the Asian crust fields of the Pb–Pb discriminant plots, with a subsidiary

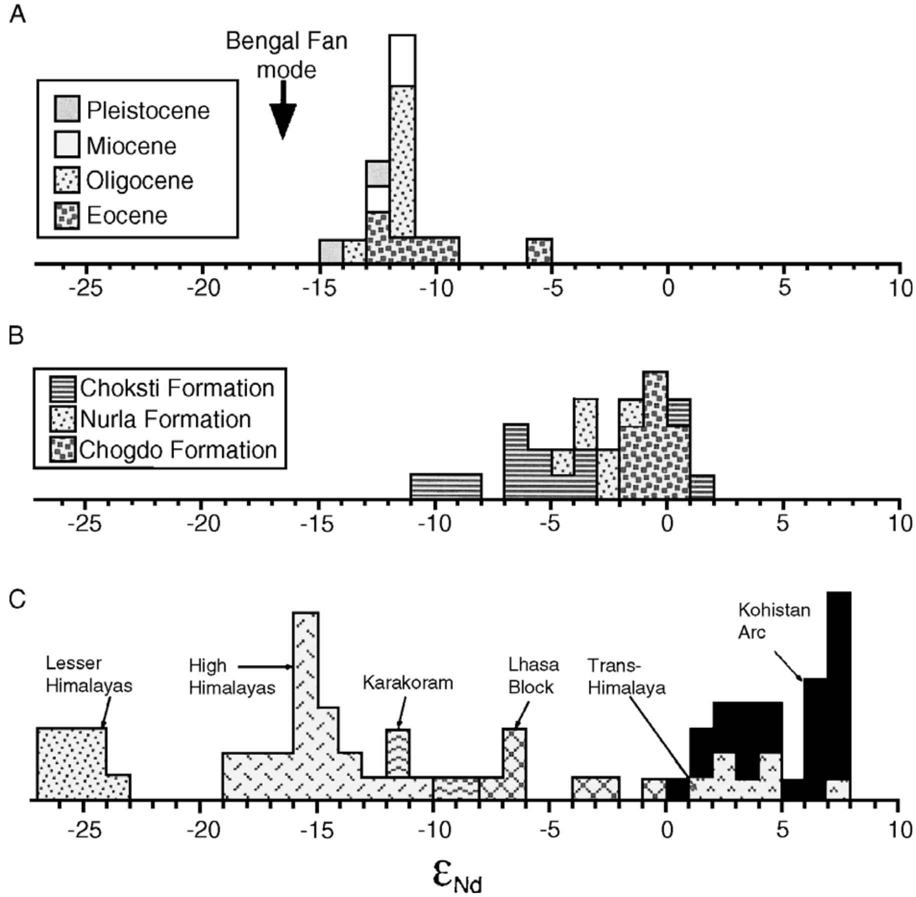


Fig. 19. ε_{Nd} values for detrital sediments of the Indus Fan (A), (mode of Bengal Fan analyses shown for comparison); Indus Molasse of the suture zone (B); and potential source terrains (C), which can be located on Fig. 1. From Clift et al. (2002a,b).

component of grains with more radiogenic signatures compared to the underlying Chogdo Formation (Fig. 20). Palaeoenvironments are alluvial and lacustrine, including turbiditic facies, and palaeocurrents show north and south directed transverse-, and east to west axial, components of flow (Garzanti and Van Haver, 1988; Searle et al., 1990; Sinclair and Jaffey, 2001; Clift et al., 2002a).

To the east, in southern Tibet, the Lower Miocene Gangrinboche conglomerates show clasts derived from both Asian (e.g., Lhasa block Gangdese granites and porphyry volcanics) and Indian (e.g., quartzite) sources (Aitchison et al., 2002). The Palaeogene Liuqu conglomerates (Davis et al., 2002) do not contain any material derived from north of the suture zone.

South of the suture zone in the Tethys Himalaya in southern Tibet, the Upper Palaeocene Jidula Formation shows no evidence of orogenic provenance in petrography, whole-rock or spinel geochemistry. By contrast, the Lutetian aged Youxia Formation contains volcanic and arc-derived lithics. Geochemistry of Cr-spinels indicates that they were derived from arc and ophiolitic rocks (Section 4.4.1) Whole rock major and trace element geochemistry confirms input from an arc setting (see Section 4.3) (Zhu, 2003; Zhu et al., 2005). Further west, in the Tethys Himalaya of NW India to the south, Critelli and Garzanti (1994) report arc and ophiolitic detritus, with subordinate sedimentary input, in the Lower Eocene or younger Chulung La Formation. These rocks plot in the magmatic arc field of the QFL ternary diagram (Section 4.2). Cr-

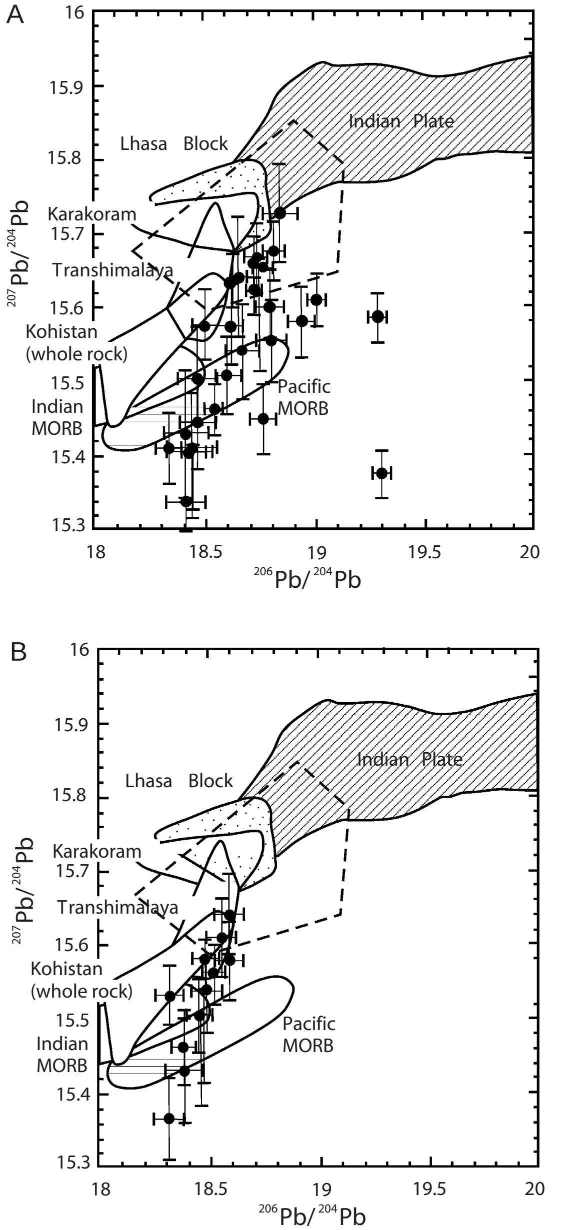


Fig. 20. Pb isotope discrimination diagram showing affinity of detrital K-spar values of the Chokst i (A) and Chogdo (B) Formations with source regions in or north of the suture zone, rather than Indian plate provenance (Clift et al., 2001a). Also, shown by dashed line is the revised field of the Indian plate source region, (from Clift et al., 2002b, their Fig. 9) which was redrawn as more data for all regions became available. With these revised fields, it can be seen that a subordinate contribution from Indian crustal source cannot be ruled out.

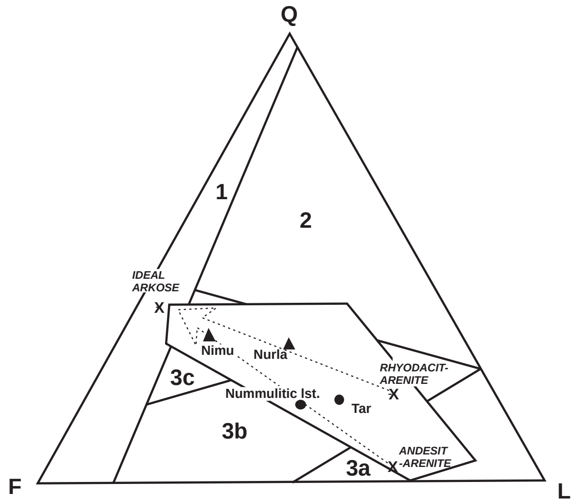


Fig. 21. Petrographic compositions of Indus suture zone rocks, showing evolution from undissected to dissected magmatic arc fields through time due to post-collisional active uplift of the calc-alkali batholiths of the Asian margin (from Garzanti and Van Haver, 1988). Sedimentary succession youngs from pre-collisional Tar Formation, to Nummulitics, and post-collisional Nurla and overlying Nimu Formation. 1 = continental block field, 2 = recycled orogen province, 3a,b,c = undissected, transitional and dissected arc provenance (Dickinson, 1985). Q = quartz, F = feldspar, L = lithic fragments. X marks positions of ideal arkose, rhyodacite–arenite and andesite–arenite.

spinel composition is similar to that of spinels found in the Youxia, Subathu and Murree Formations (Sections 5.1.1. and 5.1.2).

5.2.2. Interpretations

5.2.2.1. The timing of collision. The age of the first sedimentary succession to overlap both Asian and Indian continents provides a minimum constraint to the timing of collision. In the western part of the Indus suture zone, the pre −54.9 Ma Chogdo Formation is mapped as such by Clift et al. (2002a), but this is not in accord with mapping by Searle et al. (1990) who do not consider an Asian crust–Chogdo Formation depositional contact to be in evidence. Nevertheless, the presence of conglomerate clasts derived from both the south and north in these continental strata (Clift et al., 2001a) provides a strong supporting argument that collision occurred prior to 54.9 Ma. The 54.9 Ma aged Nummulitic limestones provide constraint to the timing of collision as defined by the cessation of marine environments (Searle et al., 1987). In the east, poorly dated Palaeogene conglomerates show no evidence of Asian provenance (Davis et al., 2002). They thus provide a maximum age constraint to India–Asia collision in the east. The first evidence of combined provenance from the Indian and Asian crust is recorded in the Lower Miocene Gangrinboche conglomerates, which thus provide a minimum age to collision. There are no known sedimentary rocks of intervening age in this region which can be used to provide tighter constraint.

South of the suture zone, in the Tethys Himalaya, Zhu (2003) interprets the change from craton provenance in the Jidula formation, to input of arc and ophiolitic material by 50.5 Ma (P8) in the Youxia Formation, as the first harbinger of orogenic foreland basin sedimentation derived from the arc and suture zone. Whilst similar petrography exists in the Chulung La Formation to the west, and similar interpretations have been made regarding the timing of collision from these data (Rowley, 1996), the lack of robust dating of these rocks (Section 3.1) precludes more accurate assessment of collisional age.

5.2.2.2. Evolution of drainage. An understanding of the drainage history in the Indus suture zone is important because development of the Indus River will have had significant effects on the denudation of the orogen as well as providing a possible feedback mechanism to tectonics and exhumation (see Section 6.2.2). As discussed in that section, the route of the palaeo-Indus, the upper reaches of which today flow west along the suture zone, is controversial. One school of thought suggests that the Palaeogene Indus Molasse represents the deposits of the palaeo-Indus river, and hence the river has been in existence along that route since early in the evolution of the mountain belt. By contrast, another school of thought considers that the Indus Molasse represents the deposits of internally drained basins, and that the Indus River first flowed west along the suture zone subsequent to their deposition. Cited evidence for both hypotheses involves provenance, palaeocurrent and facies data.

Based on palaeocurrent and provenance data, Clift et al. (2002a) interpret the Chogdo Formation as having been deposited in an internally drained basin. The first appearance of axially west-directed palaeocurrents and the shift in ε_{Nd} values in the overlying Nurla and Choksti Formations, interpreted by Clift et al. (2001a) as the result of erosion from the Lhasa block to the east, led those workers to conclude that these rocks were deposited by an axially draining river, the palaeo-Indus. Clift et al. consider that the shift in ε_{Nd} values could not be the result of contribution from the Karakoram to the north (see Fig. 19) due to palaeocurrent considerations, or from the Indian crust to the south on the basis that (1) the Pb compositions of detrital feldspars are not consistent with an Indian crust source (Fig. 20) and (2) the detritus lacks muscovite and garnet which they consider are typical of Indian crust Higher Himalayan derived sediments. As such, a Lhasa block source is strong supporting evidence for long-distance through-flow of drainage, i.e., a palaeo-Indus at this time. However, as discussed by Clift et al. (2002a), a second set of palaeocurrent indicators show transverse N–S flow, and this would be consistent with a local northern Karakoram source. Should be Clift (2002). Moreover, the new expanded Indian crust field of Pb K-feldspar compositions (dashed line box on Fig. 20, Clift et al., 2001a; their Fig. 6) (see Section 4.6.2) could permit the influx of more radiogenic detrital K-feldspars to be attributed to the Indian crust, although one might expect to also find evidence of highly radiogenic feldspars if the Indian crust is invoked as the additional source. The absence of muscovite and garnet does not preclude an Indian crust contribution since it has been shown that at this early stage of orogeny, muscovite and garnet were not being eroded from the upper crustal levels of the Higher Himalaya (Najman and Garzanti, 2000; Najman et al., 2004).

In contrast, the palaeocurrent dataset of Sinclair and Jaffey (2001) does not record any evidence of axial flow, although such evidence does exist in other areas, as outlined above. They consider that their palaeocurrent data, combined with the variable facies (lacustrine and alluvial), and limited range of conglomerate clast compositions predominantly of local origin, do not provide sufficient support to the theory of a long-distance through-flowing palaeo-Indus during the Palaeogene. Clearly, both schools of thought represent "work in progress" and these hypotheses will only be definitively resolved by the incorporation of large numbers of datasets which comprehensively cover the large area in question, both spatially and temporally.

5.2.2.3. Uplift of Southern Tibet. The timings and mechanisms of uplift of southern Tibet are still in dispute, with high elevation of the Lhasa block postulated as having been in evidence since before-, early-or not until late in the collisional process (e.g., Harrison et al., 1992; Murphy et al., 1997; Kirby et al., 2002). In the eastern part of the suture, adjacent to the Lhasa Block, the lack of any known sediment record of Eocene–Oligocene age led Aitchison et al. (2002) to consider a duration of sedimentary quiescence during this time, which suggests that the Lhasa block was not a topographic high contributing material to the basin during that period. Upper Oligocene–Lower Miocene conglomerate facies resting unconformably on the Lhasa block and initially with a high proportion of Lhasa terrain clasts documents the development of topographic relief on the Lhasa block at this time.

In the west, the interpretation of Clift et al. (2001a) of first contribution of Lhasa Block detritus to the imprecisely dated ca. Upper Eocene–Oligocene Nurla and Choksti Formation, led them to tentatively support those models which advocate early uplift of the Lhasa block, subsequent to collision. If a Lhasa Block source is accepted (see discussion above), its lack of contribution to the Palaeocene Chogdo Formation could be due to the change in sediment transport direction, as suggested by the palaeocurrent data, as discussed above, rather than lack of Lhasa block topography in the Palaeocene.

5.3. The distal, remnant ocean, and deep sea basins

5.3.1. The western, Indus system

5.3.1.1. Data

5.3.1.1.1. Katawaz remnant ocean basin. The deltaic and turbiditic facies of the Palaeogene Khojak Formation contain S and SW directed palaeocurrent indicators. This information, plus the large amount of sediment shed into the basin, led Qayyum et al. (1996) to conclude that the source was the rising orogen of the Himalaya. Petrographic data (Qayyum et al., 2001) shows that the Khojak Formation plots in the recycled orogen field (Fig. 22). In terms of total lithic fragments, sedimentary clasts are dominant, followed by subordinate low grade metamorphic clasts, the proportion of which increased in the Oligocene, and rare volcanic lithics. Coupled with the low feldspar content, this suggests that a magmatic arc was not a major source to the Khojak Formation, or that these labile materials were preferentially eliminated during transportation or diagenesis.

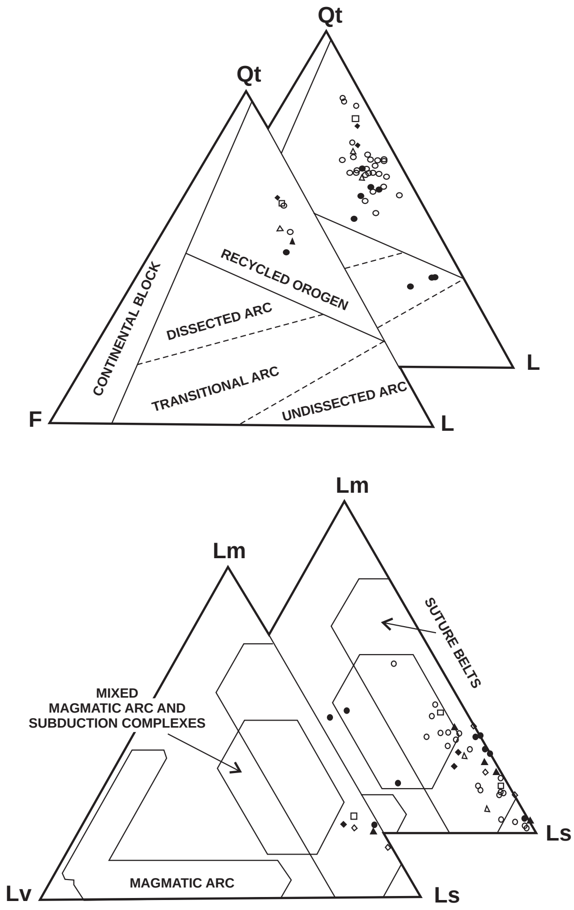


Fig. 22. Petrographic compositions of the Khojak Formation, Katawaz remnant ocean basin, Pakistan. Variation in symbols represents different sedimentary sections (Qayyum et al., 2001). Provenance fields are from Dickinson (1985). Qt = total quartz, F = total feldspar, L = lithic fragments, Ls = sedimentary lithic fragments, Lm = metamorphic lithic fragments, Lv = volcanic lithic fragments. Triangles in foreground represent mean of data from each section, triangles in background represent the range of data.

5.3.1.1.2. Makran. Sedimentary rocks of the Makran accretionary complex plot in the recycled orogen field, interpreted as Himalayan-derived (Critelli et al., 1990). Low to medium-grade metamorphic fragments dominate the proportion of total lithic fragments. Input of volcanic material, most likely derived from andesitic volcanic centres in northern Makran, peaks at ca. 16 Ma, after which time it decreases.

5.3.1.1.3. The Western Fold Belt (Sulaiman and Kirthar Ranges). The Eocene Gazij Group shows the first evidence of collision between India and either a microplate or the Asian landmass. There is a change in direction of palaeocurents, from westward off the craton in the Cretaceous, to eastward in the Palaeocene–Eocene, and a provenance which is predominantly sedimentary, but with a subordinate igneous input including Cr-spinels and basic volcanic extrusives indicative of ophiolite derivation (Kassi, 1986; Waheed and Wells, 1990; Warwick et al., 1998). The Late Eocene–Oligocene turbidites of the Kohan Jhal Formation are also considered to be the result of collision and are interpreted as Himalayan-derived (Smewing et al., 2002) although no detailed provenance work has been undertaken. Waheed and Wells (1990) took palaeocurrent measurements from the Chitarwata Formation aged Oligocene or Miocene (Section 3.3.1) and overlying Siwalik Formations and noted that data indicated drainage by a trunk river headed south, with local derivation from the uplifting Sulaiman ranges (denoted by eastward palaeocurrents) only found at a late stage.

5.3.1.1.4. Indus Fan. Early work in the Arabian Sea was largely confined to DSDP drill sites. Here it was only possible to penetrate into the oldest deposits of the Indus Fan at very distal sites (Arabian Sea DSDP Site 221 where the oldest fan sediments are upper Oligocene) or sites where subsequent uplift has allowed preservation of older Arabian sea sediments at relatively shallow levels (e.g., DSDP Site 224 on the Owen Ridge, where samples as old as Eocene were recovered) (DSDP Shipboard Scientific Party, 1974). Consequently, much work has been focussed on the more accessible younger parts of the fan, and comparatively little is known of its early evolution, although both Clift et al. (2001b) and Daley and Alam (2002) use seismic data to show significant Oligocene deposits.

Sediment budgets through time have been estimated using the two approaches of drill site and seismic data, sometimes with contrasting results. In the Early Miocene, high accumulation rates are first recorded in the deep offshore (Davies et al., 1995).The works of Clift and Gaedicke (2002) and Rea (1992) are similar in their documentation of significantly increased clastic influx in the Mid Miocene, approximately coincident with the construction of major channel and levee complexes, but in disagreement with that of Metivier et al. (1999) who record relatively low and constant rates until the Pliocene. The data of Clift and Gaedicke (2002) and Davies et al. (1995) record reduced supply to the fan since Late Miocene, in contrast to the major increase in flux recorded by Rea (1992), then and in the Mid Pliocene (cf Metivier et al., 1999). Clift and Gaedicke (2002) suggest these discrepancies might be the result of sediment budget calculations determined from drill sites which are small in number, often located in less than optimal positions to sample the largest sediment volumes, and not representative of the total system. However, Rea (1992) considering the Indus and Bengal Fans, judges that single sites may be representative of the system since a number of sites show similar sediment patterns and clastic flux records.

Provenance controls are taken from petrographic, Sm–Nd whole rock (Fig. 19), and detrital K-feldspar Pb isotopic composition data (Fig. 23). Taking Eocene sediments on the Owen ridge to be northerly Himalayan-derived palaeo-Indus Fan deposits, (Kidd and Davies, 1978), Clift et al. (2001b) compared the Pb isotopic compositions of detrital feldspars with those of Himalayan Indian and Asian sources to show that even by mid Eocene times, sources north of the Indian crust were contributing material to the fan. Subsequent additional analyses of K-feldspar grains in the source regions redefined and expanded some of the source fields, in particular that of the Indian crust (Clift et al., 2002b) (see Fig. 20). However, the basic premise, that at least some of the Eocene grains plot outside of the expanded Indian Plate field, remains intact (Fig. 23). Evidence for significant contribution from north of the Indian plate is augmented by ε_{Nd} values which are higher (less negative) than the typical Higher Himalayan signature (Clift et al., 2001a,b) — Fig. 19. The view that these Eocene Owen Ridge sediments are northerly-sourced palaeo-Indus Fan deposits, is not ubiquitously upheld. DSDP Shipboard Scientific party (1974), Jipa and Kidd (1974) and Mallik (1974) consider that the sediment distribution patterns, petrography, and heavy mineral data, favour a western source, most likely the varied assemblage of metamorphic and igneous rocks of Arabia. However, K-feldspar derived from the Arabian margin would likely

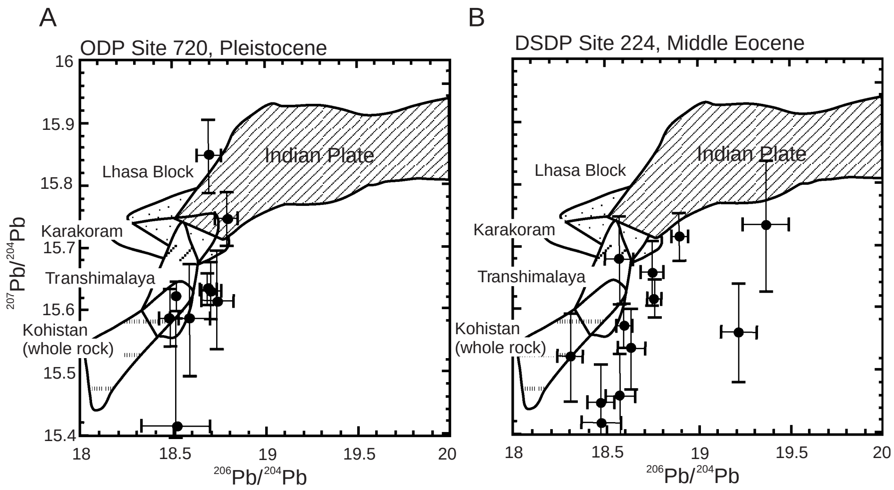


Fig. 23. Pb isotope discrimination diagram showing the affinity of detrital Pleistocene (A) and Eocene (B) K-feldspars in the Indus Fan ODP Site 720, with values characteristic of units within and north of the Indus suture, rather than the Indian plate. From Clift et al. (2001b).

be from Pan-African basement (Garzanti et al., 2003). Such feldspars would be highly radiogenic, rather than the unradiogenic feldspars derived from the arc.

5.3.1.2. Interpretations. If a northerly Himalayan source for Eocene sediments on the Owen Ridge is accepted, the presence of material derived from north of the Indian crust in these sands indicates that collision had occurred by this time (Clift et al., 2001b). This would agree with the suggestion that the Late Eocene–Oligocene Kohan Jhal turbidites of the Pab province of the Western Fold Belt are Himalayan-derived. Only sedimentary and low grade metamorphic source terrains were exposed to erosion during the Palaeogene, with erosion to deeper levels of the orogen indicated by rising proportions of low grade metamorphic fragments in the Oligocene (Qayyum et al., 2001). Qayyum et al. (1997) consider that the Early Miocene initiation of rapid sediment accumulation in the deep offshore reflects the diversion of the palaeo-Indus river from drainage into the Katawaz basin to its present course due to uplift of the Waziristan–Baluchistan ranges — this, and other data pertaining to Indus River palaeo-drainage, is discussed in Section 6.2.2). Clift and Gaedicke (2002), and Clift (2002) interpret their documented Mid Miocene increase in sediment accumulation as the result of tectonically and climatically enhanced exhumation and erosion of the Asian source areas of the Indus River drainage basin since 20 Ma, coupled with diversion of the fan from the west due to uplift of the Murray Ridge. They note that the subsequent Late Miocene drop in flux rates broadly correlates with that determined by Burbank et al. (1993) in the Bengal Fan and foreland basin, which has been related to reduced rates of tectonism and climatic change at that time. Whilst Clift et al. (2001b) use their evidence of the occurrence of Palaeogene sediments to strengthen the argument that early accommodation of strain in the orogen involved horizontal compression and the development of topographic uplift rather than dominance of lateral extrusion, Metivier et al. (1999) used their contrasting dataset (encompassing a large number of Asian basins which acted as Himalayan–Tibet sediment sinks) to suggest the opposite.

5.3.2. The eastern Bengal System

As described in Section 3.3.2, the Ganges and Brahmaputra rivers drain into the Palaeogene–Recent Bengal Basin deltaic complex and offshore Bengal Fan.

5.3.2.1. Data

5.3.2.1.1. Bengal Basin. Relatively few isotopic or geochemical provenance studies have been carried out on the sedimentary rocks of the Bengal Basin, and provenance interpretations are currently heavily reliant on petrographic and heavy mineral information. It should be born in mind however that, as pointed out by Reimann (1993), the different regions of the Basin (NW Shelf, Bengal Foredeep, Chittagong Hill Tracts) vary in their facies and potential provenance and thus datasets which amalgamate analyses from more than one region will obscure regional detail.

During the Eocene (Sylhet and Kopili Formations), sedimentation was extremely quartzose, plotting in the "craton interior" field of the QtFL diagram, and lithic grains are overwhelmingly of sedimentary character (Uddin and Lundberg, 1998a) (Fig. 24). The proportion of quartz decreases in the Barail Formation, but the rocks are still very quartz-rich. Lithic fragments continue to be dominated by sedimentary lithologies, but some low-grade metamorphic detritus is present (Johnson and Nur Alam, 1991). Rocks plot in the "craton interior" field on the QtFL diagram, and in the "quartzose recycled orogen" field in the QmFLt plot. Uddin and Lundberg (1998b) note that heavy minerals in the Eo–Oligocene (undifferentiated between Kopili and Barail Formations), are of low abundance and low textural maturity. They include tourmaline, garnet, rutile and zircon ascribed to igneous and metamorphic sources. The Surma Group heralds a major change in sedimentary petrography. Sandstones plot in the "recycled orogen" field, and show an abrupt increase in feldspar and lithic fragments at this time (Bhuban Formation) which continues upsection. Himalayan-aged micas are also found (Rahman and Faupl, 2003). The trend to increasing proportion of K-feldspar to plagioclase, increased metamorphic grade of lithic fragments and metamorphic index minerals (garnet increases greatly in the Bhuban Formation, when staurolite and kyanite also become more common. Kyanite increases significantly in the Tipam Formation, and sillimanite, present since the Bhuban formation, increases significantly in the Dupi Tila Formation) (Uddin and Lundberg, 1998b).

5.3.2.1.2. Bengal Fan. The initiation of Bengal Fan sedimentation, an important constraint to the timing of uplift and erosion of the orogen is poorly constrained. Estimates range from Eocene to latest Oligocene (see Section 3.3.2). In spite of these uncertainties, there is little doubt that a major increase in sediment delivery to the fan, and its very rapid progradation, took place in the Early Miocene. A period of accelerated deposition is recorded from around 11 Ma (Rea, 1992). This is followed by a period of decreased sedimentation rates and grain-size, accompanied by a shift in clay mineralogy from illite–chlorite dominated to smectite–kaolinite dominated, during the period 7 to ca. <1 Ma, when rapid sedimentation resumed (France-Lanord et al., 1993).

Throughout the drilled interval of ca. 18 Ma to present, whole rock ε_{Nd} values and $^{87}Sr/^{86}Sr$ values (Figs. 10 and 19) (France-Lanord et al., 1993; Galy et al., 1996) and Neogene aged muscovites and feldspars (Fig. 25) (Copeland and Harrison, 1990) indicate dominant contribution from the Higher Himalaya. The degree of contribution from non-Himalayan sources, e.g., the Indian subcontinent, is disputed (e.g., Bouquillon et al., 1990; Brass and Raman, 1990; Amano and Taira, 1992; France-Lanord et al., 1993), but is agreed to be relatively minor.

Exhumation of the Higher Himalaya to increasingly deeper crustal levels is indicated by the progressively higher grade of metamorphic minerals upsection (Amano and Taira, 1992). Qualitative assessment of exhumation rates during the time interval studied, using the determination of lag times (see Section 4.5.1), has been done with varying degrees of accuracy using fission track ages of detrital apatites (Corrigan and Crowley, 1990), Rb–Sr ages on detrital biotites (Galy et al., 1996), and Ar–Ar ages of detrital feldspar and muscovites (Copeland and Harrison, 1990). The detrital apatite fission track ages produce lag times between 0–10 million years, although each sample is picked from sediment core over a range of 70–120 m, and thus precision is reduced. These data are corroborated by lag times between 4.5 and 14 million years determined from Rb–Sr detrital biotite ages, although there are large uncertainties associated with the Rb–Sr technique in the detrital context and ages are only averages of bulk concentrates rather than individual grains. Ar–Ar age data on detrital feldspars and micas benefit from single grain analyses and show that throughout the drilled interval there are mineral ages which record zero lag time, implying extremely rapid exhumation of some part of the orogen throughout this interval (Fig. 25). However, it must be born in mind that the presence of negative aged micas (younger than depositional age, interpreted as due to hydrocarbon

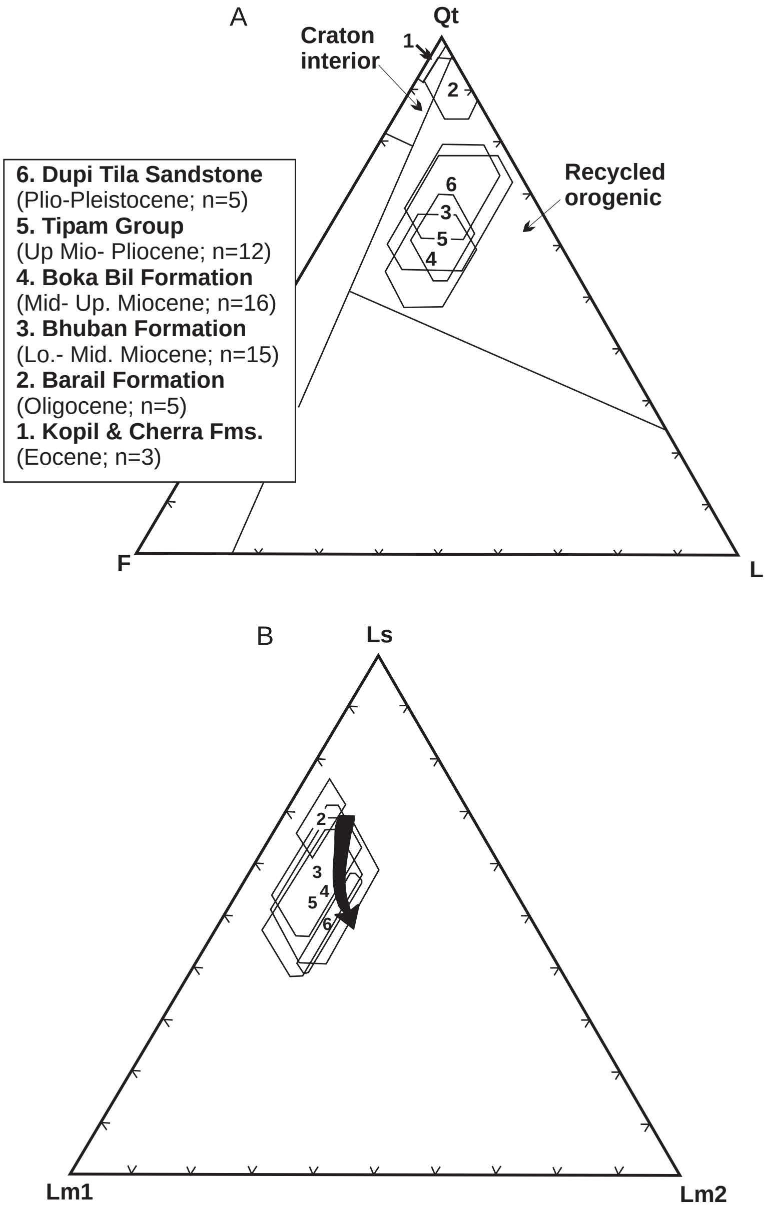


Fig. 24. petrographic compositions of rocks of the Bengal Basin, showing evolution to less quartzose composition (A) and to increasing proportion of metamorphic lithic fragments (B) through time. From Uddin and Lundberg (1998a). Eocene aged Kopili and Cherra Formations not shown on B due to insufficient numbers of lithic counts to plot. Qt = total quartzose grains, F = total feldspar grains, L = total lithic grains, Ls = sedimentary lithic grains, Lm_1 = very low to low grade metamorphic lithic fragments, Lm_2 = low to intermediate grade metamorphic lithic fragments.

interpreted by Copeland and Harrison, 1990) reduces the reliability of this data. In addition, the precision of this dataset suffers due to the minerals from each sample having been picked from a 70–120 m interval of core.

5.3.2.2. Interpretations. Interpreted provenance source varies with region. In the eastern part of the Bengal Basin, the thick deltaic sequence is interpreted as deposits of the palaeo-Brahmaputra and Indo-Burman draining rivers, whilst the later transition to

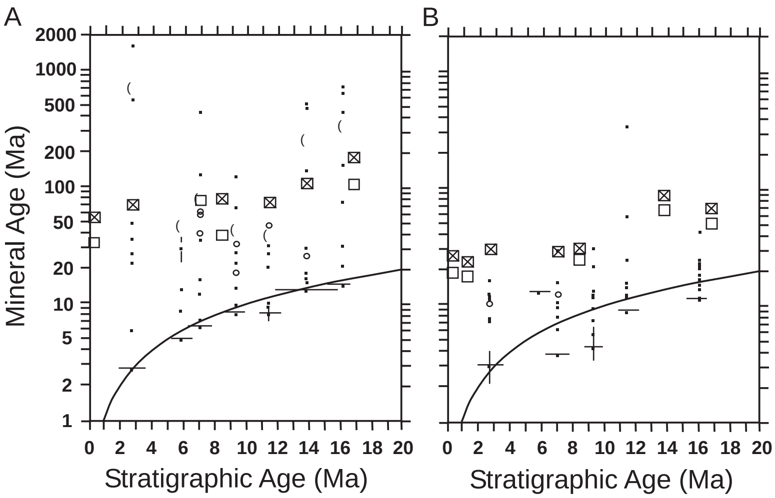


Fig. 25. Ar–Ar K-feldspar (A) and muscovite (B) detrital mineral ages vs. stratigraphic age from Bengal Fan sediments from ODP Site 116. Circles represent minimum age on age spectrum or total fusion age where only one step was performed for single-crystal analyses. Open squares represent minimum ages on age spectra from multigrain separates. Squares with X are total fusion ages of multigrain separates. Bracket symbols represent weighted average of single-crystal analyses from one stratigraphic interval. Line shows zero lag time, where sediment age is equal to mineral age. From Copeland and Harrison (1990).

alluvial facies in the North–East may reflect either derivation from the newly uplifted Shillong plateau to the north (Johnson and Nur Alam, 1991), or southward advance of the palaeo-Brahmaputra facies (Partington et al., 2005). In the North–West region (NW Shelf), the only very recent transgression of the Rangpur Saddle (which separates the Himalayan foreland basin from the Bengal Basin) leaves open the provenance of the earlier deltaic sediments as cratonic or Himalayan derived and raises questions as to the former route of the palaeo-Ganges (Johnson and Nur Alam, 1991; Lindsay et al., 1991; Alam et al., 2003; Uddin and Lundberg, 2004; Partington et al., 2005).

Opinions as to when the first Himalayan-derived detritus was delivered to the Bengal system, and the implications of this, are sharply divided. All parties are in agreement that by the time of deposition of the Bengal Basin Bhuban Formation in Miocene times, and to the extent of the drilled interval to ca. 18 Ma in the Bengal Fan, there is clear evidence of an orogenic signal. This is based on data from petrography, ε_{Nd} and $^{87}Sr/^{86}Sr$ isotopic signal, heavy minerals and mica ages. Progressive unroofing of the source area to deeper structural levels occurred through time. The predominantly Higher Himalayan signature in the Bengal Fan, coupled with the short lag times, indicates that at 18 Ma the gross structure of the Himalaya was already broadly similar to that of today (France-Lanord et al., 1993). Since the Higher Himalaya are not of granulite grade, which would be required if a region had been experiencing continually rapid exhumation for 17 million years, Copeland and Harrison (1990) suggest that short periods of rapid exhumation must be spatially and temporally distributed throughout the orogen. Rea (1992) suggests that pulses of tectonism may be recorded by the variations in Miocene–Recent sediment accumulation rates in the basin, since they do not correspond to changes in sea level, or climate as indicated by the $\delta^{18}O$ record. By contrast, the changes in clay mineralogy and sedimentology at 7 Ma may be ascribed to climate change, with longer storage and more intense weathering since this time (France-Lanord and Derry, 1994). Decreasing sediment accumulation rates may also be explained by a change in climate resulting in greater slope stability due to increased vegetation cover, or to decreasing Himalayan glaciation or tectonism (Burbank et al., 1993). Earlier work ascribed such a climatic change to monsoonal intensification, co-incident with the timing of monsoon-related upwelling in the Arabian sea (Kroon et al., 1991). However, more recent work indicates that monsoon inception occurred significantly earlier (see Section 6.3).

Whilst Miocene rocks are clearly orogen-derived, provenance of the pre-Miocene rocks of the Bengal Basin is debated. Johnson and Nur Alam (1991) consider that rocks of the Oligocene Barail Formation are the product of incipient Himalayan uplift. By contrast, Uddin and Lundberg (1998a) argue that the quartzose composition of the Eocene and Oligocene rocks do not show evidence of a significant contribution from any obvious orogenic source and suggest they are more likely to have been derived from passive margin sediments of the Indian craton, a theory also upheld by Banerji (1984). Whilst Uddin and Lundberg (1998a) consider that it is possible that these rock compositions could have been produced by derivation from a proto-Himalayan source accompanied by extreme weathering, the great distance to the orogen, compared to the relative proximity of the craton, suggests to them that a cratonic source as the more simple interpretation. However, it should also be noted that heavy mineral compositions do show evidence of a metamorphic source, and their low abundance is indicative of extreme weathering, although their textural immaturity suggests relatively close proximity to the source (Uddin and Lundberg, 1998a). Clearly, more work is needed, using a variety of techniques, to identify with certainty the presence of Himalayan-derived material in the Palaeogene record of the Bengal Basin.

By contrast, a different approach to determination of palaeo-source areas is used by Johnson (1994). Johnson calculated the sediment volumes in the major Himalayan repositories (of which the Bengal Fan is by an order of magnitude the largest), and compared it with the area of the Himalaya (see Section 4.1). Were the sediments entirely derived from the Himalaya, a 35-km vertical section would need to have been eroded, which is unlikely. He therefore concluded that the Karakoram and Tibet must have contributed significant quantities of sediment. This contribution was likely to have been greatest during early stages of evolution, prior to significant Himalayan uplift. The High Himalayan dominated provenance from the Bengal Fan from ca. 20 Ma would also support the contention of an early contribution from more northerly sources. Errors associated with sediment volume and Himalayan area calculations, plus input from northern sources to the smaller sediment repositories (e.g., the Indus Fan) preclude more accurate quantification of the amount of northern contribution to the Bengal Fan.

Whenever Himalayan-derived sediments did first reach the Bengal Basin–Fan system, the significance is also not clear cut. Uddin and Lundberg (1998a) consider that whilst the first obvious signs of Himalayan-derived sediments to the basin in the Early Miocene may represent the timing of initiation of uplift and exhumation in this part of the orogen, it more likely represents the migration of the sediment system to this area at this time. Older Himalayan-derived sediments of the Bengal system may be preserved in Assam or have been subducted beneath the Indo-Burman ranges.

6. The sediment record of Himalayan evolution: an overall perspective

This section seeks to summarise those aspects of Himalayan evolution which are best investigated by integrating sediment record data from more than one region. The regional sedimentological data and interpretations summarised in Section 5 are brought together and combined to provide a broader picture of the wider-scale evolution of the mountain belt.

6.1. The timing and diachroneity of collision

As discussed in Section 2, collision has been variously dated between ~70 and 38 Ma or younger (Jaeger et al., 1989; Klootwijk et al., 1992; Searle et al., 1997; Rowley, 1998; Yin and Harrison, 2000; Aitchison et al., 2002). This age range can be explained to a large extent by the variety of techniques used, each of which is recording one aspect of a spectrum of events which occur during collision. Here, the review is focussed on those techniques which utilise the sediment record, with referral to other lines of evidence where appropriate. Rowley (1996) provided a comprehensive review of stratigraphic constraints to the timing of collision, into which much new data can now be incorporated (Fig. 26).

The sediment record has been used in a number of ways to constrain collision. Dating the youngest marine rocks (Searle et al., 1997); evidence of subsidence resulting from loading (Rowley, 1998); the earliest strata to overlap both Indian and Asian plates (e.g., Clift et al., 2002a), to contain both Indian and Asian detritus (e.g., Clift et al., 2001a); or evidence of Himalayan erosion into a previously Indian passive margin setting (e.g., Critelli and Garzanti, 1994; Najman and Garzanti, 2000), have all been used to constrain the timing of collision. Variations in such data along strike have also led to ongoing discussion on the degree or otherwise of diachroneity of collision (Rowley, 1996, 1998; Searle et al., 1997; Wang et al., 2002). In the following sections, each type of dataset is summarised, and this is followed by an overall assessment of the extent to which arguments for diachroneity are supported by existing data. Where possible, ages in Ma or P zones (planktonic foraminiferal zonal scheme) are specified. Ages based on P zones quoted in the older literature are recalibrated here with the more recent timescale of Berggren et al. (1995).

6.1.1. Stratigraphic constraints to the timing of collision (Fig. 26A)

In Waziristan, northwest Pakistan, Beck et al. (1995) mapped marine shelf strata of P5 and P6 age (55 Ma) unconformably overlying the Kahi melange which is thrust over the Indian passive margin. The Kahi melange is an accretionary prism and trench complex which is interpreted to be part of the accretionary complexes, arcs and microplates which made up the southern margin of Asia. Thus, according to Beck et al., collision, as defined by thrusting of this melange over the passive margin, occurred between 66 Ma (the youngest sedimentary rocks preserved in the mélange) and 55 Ma (the age of the overlying unconformity).This interpretation is however disputed by Rowley (1996) who considers the event to be an intra-oceanic obduction event rather than contact of the Asian continent with India.

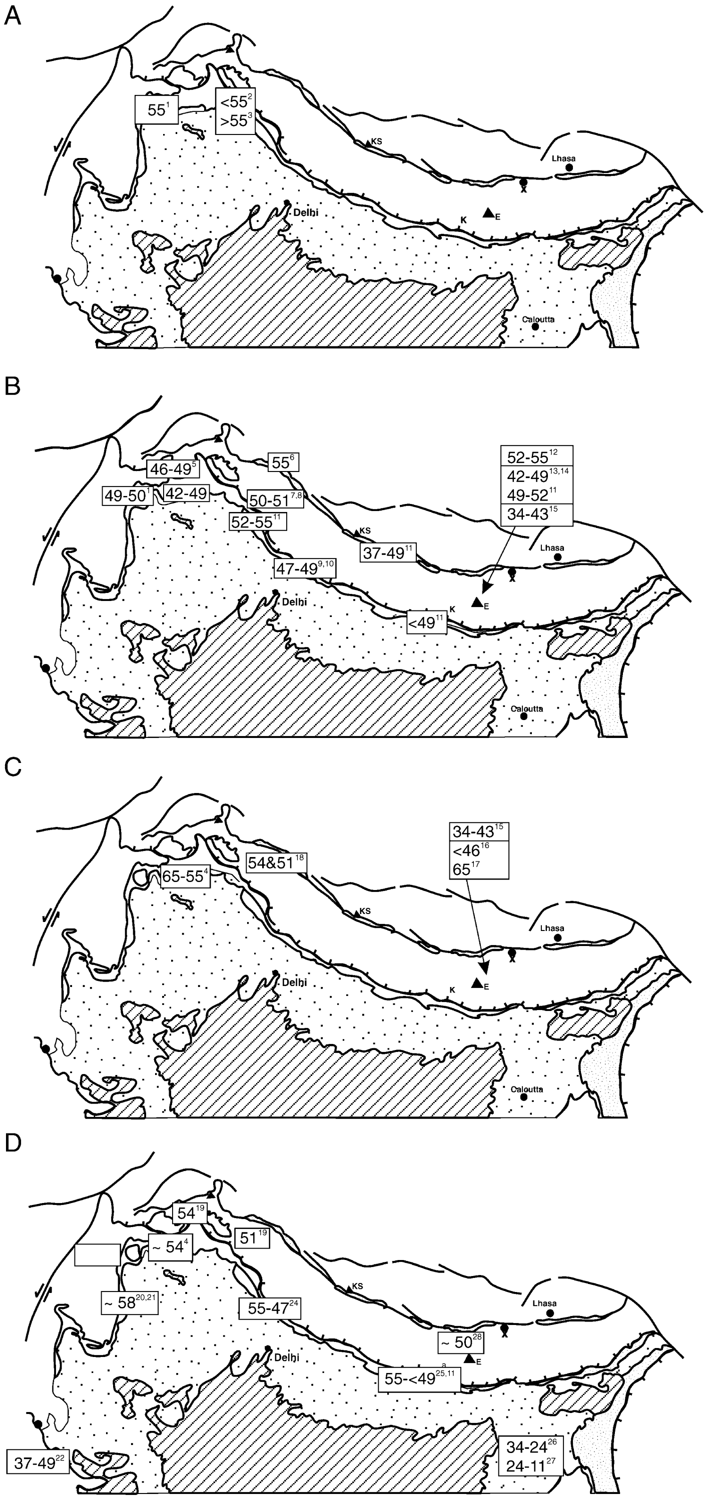
In the suture zone in Zanskar, NW India, Searle et al. (1990) map the <55 Ma aged Choksti conglomerate as the first overlap of the suture zone by strata which show derivation from both sides of the suture. More recent work by Clift et al. (2002a) identifies the older, >55 Ma aged Chogdo Formation in such a position, and with similar interpreted provenance (see Section 5.2).

6.1.2. Cessation of marine facies (Fig. 26B)

Cessation of marine facies provides a minimum age to collision since, as pointed out by Friend (1998), marine sedimentation on continental crust is common. Summarised below are data of youngest known marine rocks from across the Himalayan belt, including suture zone, Tethyan shelf and foreland basin locations. As described in Section 3.2, where accurate dating has been possible, it has been consistently shown throughout the foreland basin that the marine to continental facies transition is represented by a major unconformity. Thus, younger marine sediments may well have been deposited and have since been eroded from the succession.

Furthest west, in Waziristan, Pakistan, Beck et al. (1995) provide biostratigraphic data indicating marine conditions until P9 which corresponds to an age 49–50 Ma, Ypresian–Lutetian boundary. However, Beck et al. consider that collision occurred significantly earlier than this (see Section 6.1.1), and that these are marine facies that persisted post-collision. South of the Higher Himalaya, in a foreland basin setting, Pivnik and Wells (1996) date the youngest marine facies at Middle Eocene (42–49 Ma) in the Kohat Plateau. Further east, in a similar structural position, the youngest marine rocks in the Hazara–Kashmir Syntaxis, Pakistan, are early Lutetian age, (ca. upper P8–P10, 46–50 Ma) (Bossart and Ottiger, 1989), now

Fig. 26. Constraints to the timing of India–Eurasia collision provided by; stratigraphic information (i.e., oldest strata to overlap both continents, (A); cessation of marine facies (B); evidence of load induced flexure (C); and provenance information (i.e., first evidence of northern derived material onto the Indian margin, D). Boxed numbers are ages in Ma. A range of values, i.e., *xx–yy* indicates that dating is no more accurate than to a stage, converted to Ma using the geological timescale of Berggren et al. (1995). > and < symbols relate to the age of the sediment or event, (i.e., only a maximum/minimum age is documented), not to the age of the consequently inferred collision. Two sets of numbers in one box represents a dispute in the age of the Formation or event, or in its application — see text for details. [1]Beck et al., 1995; [2]Searle et al., 1990; [3]Clift et al., 2004; [4]Pivnik and Wells, 1996; [5]Bossart and Ottiger, 1989; [6]O. Green, unpublished data, cited in Sinclair and Jaffey, 2001; Clift et al., 2001a; [7]Baud et al., 1985; [8]Fuchs and Willems, 1990; [9]Mathur, 1978; [10]Batra, 1989; [11]Blondeau et al., 1986; [12]Willems and Zang, 1993a; [13]Willems and Zang, 1993b; [14]Willems et al., 1996; [15]Wang et al., 2002; [16]Rowley, 1998; [17]Liu and Einsele, 1994; [18]Garzanti et al., 1987; [19]Critelli and Garzanti, 1994; [20]Kassi, 1986; [21]Warwick et al., 1998; [22]Clift et al., 2001b; [23]Qayyum et al., 2001; [24]Najman and Garzanti, 2000; [25]DeCelles et al., 1998a; [26]Johnson and Nur Alam, 1991; [27]Uddin and Lundberg, 1998a; [28]Zhu, 2003.


A
55[1]
<55[2]
>55[3]
KS
Lhasa
Delhi
K
E
Calcutta
B
46-49[5]
55[6]
49-50[1]
42-49
50-51[7,8]
52-55[11]
37-49[11]
47-49[9,10]
<49[11]
52-55[12]
42-49[13,14]
49-52[11]
34-43[15]
KS
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Delhi
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65-55[4]
54&51[18]
34-43[15]
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54[19]
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~ 50[28]
55-<49[25,11]
37-49[22]
34-24[26]
24-11[27]
KS
Lhasa
Delhi
E

interthrust with non-marine red beds (Najman et al., 2002a).

In India, biostratigraphic data is available for the suture zone, Tethys shelf sequence and foreland basin south of the Higher Himalaya. The youngest recorded marine strata in the suture zone are the nummulitic limestones, which record a final brief marine incursion over older continental facies of the Chogdo Formation. They are dated at 55 Ma (O. Green, unpublished data, cited in Sinclair and Jaffey, 2001, Clift et al., 2001a). To the south, the youngest marine Tethyan rocks are the Kesi Formation, dated at P8, 50.5 Ma, (Garzanti et al., 1987; Gaetani and Garzanti, 1991) and younger Kong Formation, also dated at P8 (Baud et al., 1985; Fuchs and Willems, 1990). In the foreland basin, the youngest Subathu Formation rocks have been dated to Early Lutetian by Mathur (1978) and Batra (1989) in the more eastern region, but as no younger than Ilerdian (Early Eocene) further west by Blondeau et al. (1986).

Further east, marine facies in the suture zone region of Zhongba, Tibet persist until the mid Eocene (Blondeau et al., 1986). Data have also been collected from the Tethyan succession in the southern Tibet region near Everest. At the location of Gamba, the youngest marine rocks are dated as Ilerdian (Willems and Zang, 1993a), similar to the majority of ages further west, but in the Tingri region, located west of Gamba, the youngest marine rocks are younger: Lutetian marine facies are recorded in the Zhepure Shan section, Tingri, according to Willems and Zang (1993b) and Willems et al. (1996), although marine facies in this region are considered to extend only to late Ypresian by Blondeau et al. (1986). 4–5 km east, the Qumiba section, Tingri, contains marine facies as young as upper Lower Lutetian to Upper Priabonian (NP15–NP20, ~47–34 Ma) (Wang et al., 2002) which the authors consider reflects the regression of the epicontinental Tethys seaway that persisted for a substantial time after collision, defined as elimination of oceanic crust. Note, however, the query over that age control (Section 3.1). In the foreland basin to the south, marine foreland basin strata in the Tansen area are as young as Mid Eocene (Blondeau et al., 1986), although these authors note that differentiation between the fauna of Early and Mid Eocene age is very subtle.

It is important to note that onset of terrestrial facies in the sediment record, used as another measure of the extent of the collisional progress, need not be synonymous with cessation of marine facies. In the suture zone in Zanskar and the foreland basin in Kohat and further south in the Sulaiman–Kirthar ranges, terrestrial facies were deposited prior to the last marine incursion at similar times. In Zanskar, the terrestrial Chogdo Formation lies stratigraphically beneath the Nummulitic limestones, (Searle et al., 1990) dated at 55 Ma, whilst in Kohat, the terrestrial Mami Khel Formation is dated at upper Lower Eocene, and was deposited prior to resumption of marine conditions in the Mid Eocene as represented by the Kohat Formation (Pivnik and Wells, 1996). This is mirrored in the Sulaiman region where the upper part of the Upper Palaeocene to Mid Eocene Gazij Formation represents terrestrial facies before resumption of a shelf environment with the Mid Eocene Kirthar Formation (Warwick et al., 1998). Waheed and Wells (1990) suggest that this regression could be associated with collision between the Indian continent and a microplate, prior to main India–Eurasia collision.

Due to the unconformable nature of the marine–continental facies transition in the foreland basin (see Section 3.2), the age of the underlying marine facies can only be taken as a maximum time for onset of continental deposition. In the foreland basin, the first terrestrial deposits are dated at younger than 37 Ma in Pakistan (Balakot Formation, Najman et al., 2001) younger than 30 Ma in India (Najman et al., 1997) and at least 21 Ma in Nepal (T.P. Ojha, unpublished data, cited in DeCelles et al., 2001; Robinson et al., 2001). In the Tethyan Himalaya, (Chulung La Formation, NW India, and Shenkezar Formation, Southern Tibet) and suture zone, (Nurla Formation in NW India and Liuqu Conglomerates in south Tibet), the paucity of fossils and/or non-conformable or tectonic nature of the contact between continental facies and the underlying formations has precluded accurate dating of initiation of earliest continental sedimentation (Garzanti and Van Haver, 1988; Garzanti et al., 1987; 1996; Fuchs and Willems, 1990; Searle et al., 1990; Willems and Zang, 1993b; Najman et al., 2001; Davis et al., 2002; Zhu, 2003). The terrestrial Quiwu suture zone conglomerate, used as a constraint on the timing of collision by Searle et al. (1987), has recently been reassigned from the Eocene to the Lower Miocene by Aitchison et al. (2002).

6.1.3. Subsidence (Fig. 26C)

Rowley (1998) noted that collision should be marked in the sediment record of the Indian passive margin by initial rapid subsidence associated with thrust loading, followed by a change in provenance and coarsening of sediments. Rowley backstripped the stratigraphic section from 100–46 Ma at Zephure Shan, Tingri, South Tibet, and considered that it showed no evidence of accelerated subsidence. This, coupled with the lack of northerly-derived syn-orogenic detritus, led him to conclude that collision must post-date the youngest rocks, after 45.8 Ma. A deepening event and influx of orogenic detritus is recorded in the upper Lower Lutetian to upper Priabonian Pengqu Formation, located close by (Wang et al., 2002). However, Yin and Harrison (2000) noted that the sedimentary succession did show evidence of increased subsidence at ~70 Ma, as indicated by Willems et al. (1996) and thus in fact, this could be evidence of collision at this time. In addition, it should be noted that Liu and Einsele (1994), backstripping data from southern Tibet regions (including Tingri), representing northerly, middle and southerly locations on the Indian passive margin, concluded that collision *was* reflected in the subsidence record. All regions showed uplift, (although this was only relatively minor in the more southerly location), interpreted as the result of upwarping of the bending lithosphere. In addition, the authors suggest there was a period of relatively rapid subsidence in the Palaeocene in this southern area, which ceased by Eocene times.

Uplift rather than subsidence as a response to collision is also seen in the sediment record in Zanskar, India (Garzanti et al., 1987). Here, the onset of collision is reflected by shoaling of the outer shelf interpreted as the consequence of flexural uplift. This event is marked by unconformities dated at P6, 53–55 Ma and P8, 50.5 Ma, similar ages to the time range determined by Liu and Einsele (1994). Garzanti et al. (1987) suggest that the lack of an initial deepening stage during collision, typically characterised by flexural downwarping of the passive margin, could perhaps be explained by the particularly rigid and thick nature of the Indian plate. South, in the foreland basin, Pivnik and Wells (1996) consider the Palaeocene deepening event, recorded by the transition from shallow water passive margin facies of the Lockhart Limestone to the deep open marine Tarkhobi shales, to be the result of load induced subsidence (see Section 5.1.1).

6.1.4. Provenance information (Fig. 26D)

Provenance information can constrain the timing of collision by documenting 1) the first time when detritus from both Indian and Asian plates occur within the same sedimentary unit and 2) the first arrival of orogenic detritus onto the Indian passive margin.

In the western part of the suture zone, the Chogdo Formation is the oldest recorded unit to contain material from both north and south of the suture zone, constraining the collision as having occurred by the time of deposition of this unit (Clift et al., 2001a, Section 5.2). Further east the Lower Miocene Gangrinboche conglomerates show southern and northern derived provenance (Aitchison et al., 2002), which provides a minimum age to collision. The Palaeogene Liuqu conglomerate contains no evidence of mixed provenance (Davis et al., 2002), thereby providing a maximum age constraint.

A number of researchers working in the Tethyan region and foreland basin setting, have documented the time at which shallow marine shelf facies of carbonate or quartz arenite composition show the first evidence of igneous or metamorphic input. However the first signs of northern input can be extremely subtle (e.g., DeCelles et al., 1998a; Najman and Garzanti, 2000) and therefore very detailed provenance work is required. In addition, as cautioned by Friend (1998), it can be difficult on occasions to differentiate between the effects of the main India–Eurasia collision, and those resulting from collision between the Indian plate and microplates, intra-oceanic arcs or ophiolitic obduction events.

In Pakistan, in the Hazara–Kashmir Syntaxis area of the foreland basin, arc, ophiolitic and low grade metamorphic detritus is first observed in the upper part of the Patala Formation within zone P6 (53–55 Ma) (Critelli and Garzanti, 1994), (Section 5.1.1). To the southeast on the Kohat Plateau, the Gazij Shale, (equivalent to the Patala Formation), contains some metamorphic material (Pivnik and Wells, 1996).Further south in the Sulaiman and Kirthar ranges, the Gazij Formation contains evidence of ophiolitic input (Kassi, 1986; Warwick et al., 1998; Section 5.3.1). Influx of such detritus is interpreted as the result of collision between India and either Asia or a microplate or intra-oceanic arc. In the Indus Fan, detrital feldspars in mid Eocene sediments are interpreted as having been derived from the Asian plate (Section 5.3.1) (Clift et al., 2001b) and similar provenance is inferred for igneous and low grade metamorphic detritus in the Upper Eocene Katawaz basin sediments (Qayyum et al., 2001) (Section 5.3.1).

In India, the 50.5 Ma aged (P8) Kong Formation in the Tethyan region in Zanskar shows the first evidence of volcanic input (Critelli and Garzanti, 1994). In the foreland basin, the Subathu Formation of Lower to Mid Eocene age contains some very low grade metamorphic material and evidence of arc and ophiolitic input, which dominates in the "Red Subathu" Facies (Najman and Garzanti, 2000). This is interpreted as the erosional response to continental collision as discussed in Section 5.1.2. Eastward in Nepal, the evidence for orogenic input to the quartz arenites of the Lower to Middle Eocene Bhainskati Formation is based on U–Pb and fission-track ages of detrital zircons (Section 5.1.3).

As discussed in Section 5.3.2, the first input of Himalayan-derived sediments to the Bengal Basin is not clear cut. Whilst all workers agree that Miocene rocks are Himalayan-derived, there is dispute as to whether the Palaeogene strata are orogen-derived (Johnson and Nur Alam, 1991) or Indian craton sourced (Uddin and Lundberg, 1998a). Furthermore, those that propose the later first arrival of Himalayan sediments to the region consider that this may not reflect the first arrival of orogenic material into the region at this time since older orogen-derived sediments may have been subducted below the Indo-Burman ranges.

In the Tethyan region of southern Tibet, the Pengqu/Youxia Formation (age at base ~50–47 Ma) contains evidence of orogenic input whilst the Upper Palaeocene Jidula Formation is of non-orogenic provenance (Wang et al., 2002; Zhu, 2003). In the suture zone of southern Tibet, the Palaeogene Liuqu conglomerates lie in depositional contact with the Indian plate and ophiolitic and subduction complexes. The conglomerates contain evidence of igneous input, and these are interpreted as the result of intra-oceanic arc–India collision, prior to India–Asia continental collision (Davis et al., 2002).

6.1.5. Evidence of diachroneity

The proposed degree of diachroneity of collision waxes and wanes as new data, both that associated with the sediment record (Davies et al., 1995; Garzanti et al., 1996; Rowley, 1996, 1998; Searle et al., 1997; Uddin and Lundberg, 1998a; Najman and Garzanti, 2000; Wang et al., 2002) and from other lines of evidence, for example the age of metamorphism and subduction related magmatism (e.g., Searle, 1996; Hodges, 2000), continually modifies the picture.

Confining the discussion to those data which pertains to the sediment record, I believe that most indicators of collision do not as yet contain sufficient numbers of data points along and across strike of the orogen to allow broad interpretations to be made with certainty. In a region as large as the Himalaya, many more studies will need to be carried out along strike, before trends and anomalies can be clearly identified.

6.1.5.1. Stratigraphic constraints. Stratigraphic constraints provided by overlap of sediments over both Indian and Asian plates, are not documented east of Zanskar. Constraints from the time of onset of terrestrial facies suffers from comparison between sediments in different tectonic settings (suture zone vs. foreland basin), and imprecise dating.

6.1.5.2. Cessation of marine facies. In the foreland basin the youngest marine rocks are of similar age throughout: the Ilerdian to Mid Lutetian marine rocks interbedded with the Balakot Formation in Pakistan (Bossart and Ottiger, 1989; Najman et al., 2002a), and the Lower–Mid Eocene Subathu and Bhainskati Formations of India and Nepal (Mathur, 1978; Blondeau et al., 1986; Batra, 1989). Only two data points exist for the suture zone, 55 Ma in Zanskar (O. Green cited in Sinclair and Jaffey, 2001 and Clift et al., 2001a,b) and Mid Eocene in Zhongba, southern Tibet (Blondeau et al., 1986). Four data points are known for the Tethyan realm; 50 Ma in Zanskar and varying ages from locations close together in south Tibet (Blondeau et al., 1986; Willems and Zang, 1993a,b; Willems et al., 1996; Wang et al., 2002), some of this variation being explained by Willems et al. (1996) as due to the more basinal location of the younger sequences. An Ilerdian aged sequence occurs at Gamba, a disputed Ypresian or Lutetian age is given to the Zephure Shan section at Tingri and a Lutetian to Bartonian age at Qumiba, Tingri. Thus, the clearest indication of diachronous retreat of the Tethys sea is provided by data from the Qumiba section, although note the query pertaining to age dating of this section (Section 3.1).

6.1.5.3. Constraints from subsidence. Subsidence data is disputed. Rowley (1998) uses the lack of subsidence in the east as a strong argument for significant diachroneity when compared to the relatively well constrained collisional age, based on provenance criteria, in the west (Rowley, 1996). Yet the interpretation of this subsidence data is disputed (Yin and Harrison, 2000 — Section 6.1.3) and comparison between different types of collisional indicators introduce additional uncertainty since different indicators are recording different phases of an ongoing collisional process. Rowley's assertion of diachroneity is backed up by provenance comparison between the two regions. However, the revised age for the orogen-derived Balakot Formation and the uncertainty of the exact age of the Chulung La Formations in the west (Najman et al., 2002a) requires that the provenance comparison is made between the Kong and Patala Formations in the west, and eastern localities, as described below.

6.1.5.4. Constraints from provenance. Provenance information was used by Garzanti et al. (1987) to assert limited diachroneity based on the first evidence of volcanic material in the Tethyan realm during P6 in Pakistan and P8 in India. Volcanic detritus first found at ~50–47 Ma in southern Tibet (Section 6.1.4) is consistent with this trend. Najman and Garzanti (2000), studying the Subathu Formation in the foreland basin India, noted the diachronous arrival of ophiolitic and low grade metamorphic detritus to the foreland basin from Pakistan through to Nepal. Although age-dating of the Lower–Mid Eocene Subathu Formation is not sufficiently precise to determine relative first arrival of volcanic material compared to the P6 aged Patala Formation in Pakistan, the higher grade of metamorphic lithic fragments in the foreland basin in Pakistan (Balakot Formation) compared to the Subathu Formation in India was interpreted as the result of earlier or more intense collision in the west leading to earlier or deeper thrusting. However, with the reassignment of the Balakot Formation to a younger age, this comparison is no longer valid. The diachronous nature of initial input of igneous and metamorphic input eastward is still upheld with the quartz arenite composition of the Nepal Bhainskati Formation and rocks of the Bengal Basin. However, it is unclear whether this diachronous input can still be interpreted as the result of diachronous collision in view of Bhainskati Formation detrital zircons showing Himalayan U–Pb and fission track signatures (Section 5.1.3). It is more likely that the climatic influence on diagenesis can account for the variation along strike (Najman et al., 2005). In the suture zone, comparison between eastern and western localities is hampered by poor age control.

6.1.6. Summary

In conclusion, much evidence exists from the sediment record for an event at ca. 55–50 Ma. This is coeval with other indicators of change, for example a reported slow-down of the Indian plate, and the radiometric dates of eclogites (see Section 2). Yet what do these indicators represent? India–Asia collision is the most common interpretation (e.g., Klootwijk et al., 1992). Searle (2001) suggests that the 55 Ma eclogite dates represent the time of ophiolitic obduction prior to collision at ~50 Ma. De Sigoyer et al. (2000) consider it represents the time of initial India–Asia contact, after which continental subduction occurred, followed by continental collision at 47 Ma when full-thickness Indian crust arrived at the trench.

Whilst all the above represent details within the overall context of India–Asia collision at that time, Aitchison et al. (2002) consider that changes at 55–50 Ma, and subsequent initiation of compressional tectonics, relate to collision between an intra-oceanic arc and India in the Late Cretaceous–Early Palaeocene. In their model, India–Asia collision did not occur until Miocene time. Discussion relating to reinterpretation of non-sedimentological lines of evidence goes beyond the scope of this review, which concentrates on evidence from the sediment record. Their sedimentary provenance data from S. Tibet is consistent with collision at both 55–50 Ma or in the Miocene. The Palaeogene Liuqu conglomerate contains only Indian plate and arc detritus, consistent with an India-arc collision, whilst the Miocene Gangrinboche conglomerate contains evidence of Indian and Asian detritus. However, since the dating on the Palaeogene Liuqu conglomerate is poor (Section 3.1) and there is no sediment record between that and the Miocene Gangrinboche conglomerates, the timing of collision based on sediment provenance data is impossible to tie down more precisely from these rocks. Indeed, if the Liuqu conglomerate were to be Palaeocene, which seems likely since it is interpreted as the erosional bi-product of arc-continental collision in the Late Cretaceous–Early Paleoecene, we would expect that it would contain no Asian-derived detritus from a continental collision which occurred subsequently at 55–50 Ma. Thus, the available data: arc-continent collision in the Late Cretaceous–Early Palaeocene, erosion of arc and Indian plate material preserved in the Liuqu Conglomerates, and evidence of Asian and Indian detritus in the Miocene Gangrinboche conglomerate, is consistent with both the 55–50 Ma and proposed Miocene ages of collision. More difficult to reconcile is provenance data from the west where Eocene sedimentary rocks in the Suture Zone, foreland basin and Indus Fan contain detritus from north of the suture zone (Sections 5.1.2, 5.2–5.3.1).

6.2. Models of Himalayan evolution

Models which attempt to explain how, when and why the Himalayan orogen is evolving the way it is bring together disparate evidence, including numerical and analogue modelling, geophysical and geological data. In this section, the regional data outlined in Section 5 is summarised, to show the ways in which the sediment record can contribute to constrain these models.

6.2.1. Exhumation of the Himalaya

It is not possible to pinpoint the start of erosion of the metamorphosed Higher Himalaya, due to imprecise dating of the sediment record. In the foreland basin, although the first minor input of metamorphosed material is found in the Patala Formation, Pakistan, at 50–55 Ma (Section 5.1.1), the first major influx of material derived from the Higher Himalaya is found directly above the Late Eocene–Oligocene unconformity, dated at <37 Ma in Pakistan, <28 Ma but >20 Ma in India (Section 5.1.2) and <21 Ma in Nepal (Section 5.1.3). Offshore, where there is no hiatus in the sediment record, low-grade metamorphic detritus is found in Eocene–Oligocene times (Section 5.3), although in the east this input is minor until the Miocene. In both eastern and western areas offshore, attributing metamorphic detritus to a specific provenance is disputed between proto-Himalaya, Asian and Indian cratonic sources.

Lag time information from the foreland basin rocks from Pakistan through to Nepal suggests that there is both spatial and temporal variation in the exhumation rate of the Higher Himalaya (Section 5.1). Due to the limited number of sections studied to date, our current picture is far from complete. From the Kamlial section (18–14 Ma) studied in Pakistan, very short lag times, indicative of rapid exhumation, persisted from 17 until at least 14 Ma, which is the limit of the analysed succession. Indian foreland basin rocks show very short lag times, indicative of rapid exhumation of the Higher Himalaya, from the start of the dated section at 21, to 17 Ma. After this time, lag times increase, coincident with the time of propagation of thrusting into the footwall of the MCT. In Western Nepal, the progressive slowing down of exhumation of the Higher Himalaya is documented from lag times from at least 16 Ma in the Siwalik Group, which also record further southward thrust propagation to exhume the Lesser Himalaya at 12 Ma, co-incident with a further decrease in exhumation rate of the Higher Himalaya.

Exhumation of the Lesser Himalaya at this time is evidenced by provenance data in both Western Nepal (Section 5.1.3) and India (Section 5.1.2), and the basin wide acceleration in accumulation rates and upward coarsening facies at 11 Ma (Fig. 27), interpreted by Meigs et al. (1995) as the result of loading by, and significant attainment of, topography in the Lesser Himalayan thrust sheets. The data is consistent with apatite fission track data from the hanging wall of the MBT (Meigs et al., 1995) and more southerly thrusts (Najman et al., 2004), (Section 5.1.2). Note however that East Nepal does not record erosion of the Lesser Himalaya until later (Robinson et al., 2001).

Grujic et al., (1996, 2002), Beaumont et al. (2001, 2004) and Jamieson et al. (2004) explain

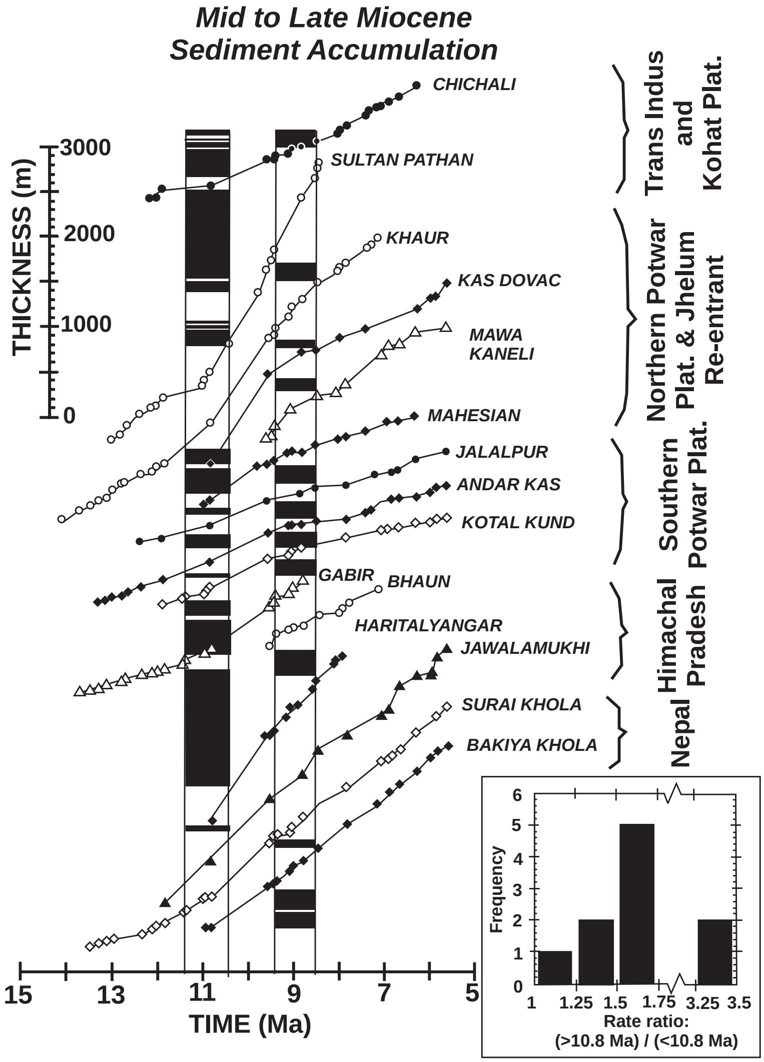

Fig. 27. Foreland basin sediment accumulation rates, showing basin-wide increase at 11 Ma (from Burbank et al., 1996, see also Meigs et al., 1995).

exhumation of the Higher Himalaya by extrusion of a mid to lower crustal channel coupled to focused surface denudation. In channel flow models, efficient erosion on the pro-flank of a Himalayan-type orogen can result in exhumation of a midcrustal channel along the erosion front. Reduction in denudation rate results in reduced exhumation and, at very low erosion rates, can lead to abandonment of the extrusion zone and detachment of the foreland to form a fold–thrust belt ahead of the channel. In the model runs to which the Himalaya is compared, orogenic erosion is not turned on until 30 Ma. This is broadly consistent with the sediment record as reviewed in this paper, and is necessary in order to build up sufficient thick hot crust to drive channel flow. High erosion rates and thus rapid exhumation is maintained from 24 to 14 Ma in the model, after which it declines progressively to the end of the run, again in keeping with the known constraints on Higher Himalayan exhumation as recorded in the sediment record. The timing of first exhumation of rocks of progressively increasing metamorphic grade is predicted by the model, and is broadly in agreement with the times of first appearance of metamorphic index minerals in the detrital record of the foreland basin in India and Nepal (Sections 5.1.2 and 5.1.3). The progressive slowing down of exhumation of the Higher Himalaya, documented in the sediment record since 17 Ma, with co-incident propagation of thrusting into the footwall of the MCT, and subsequently into the Lesser Himalaya at 12 Ma can be interpreted by the channel flow model as the result of decreasing rates of channel extrusion due to decreasing erosion since this time.

Initiation of Lesser Himalayan exhumation has been interpreted using critical taper theory (e.g., Dahlen and Suppe, 1988) as a time of frontal accretion as the over-critical thrust wedge returned towards steady state (Huyghe et al., 2001; Najman et al., 2002b). In the critical-taper wedge model, a foreland fold-and-thrustbelt growth revolves around maintenance of the critical taper angle, and thrust-front advance can only occur when the wedge is critically tapered. Thus, feedback between internal deformation, which acts to build taper, and frontal accretion, tectonic thinning and erosion, which act to decrease taper, dictates the kinematic history of the fold-and-thrust belt and can result in temporal oscillations between frontal accretion and internal deformation (e.g., DeCelles and Mitra, 1995; Meigs and Burbank, 1997; Hilley and Strecker, 2004; Whipple and Meade, 2004). Decreasing the erosion rate of the Higher Himalaya, as proposed in the channel flow model, may increase taper and thus facilitate conditions whereby an orogenic wedge could propagate forward by frontal accretion. Perhaps the switch from rapid exhumation of the Higher Himalaya to southward propagation of thrusting below the MCT and into the Lesser Himalaya, as documented in the Dharamsala and Siwalik rocks, reflects a change from internal deformation of the orogenic wedge to frontal accretion at that time. More work is required to understand the coupling between thick and thin-skinned tectonic style.

6.2.2. Evolution of the Indus drainage system and exhumation of the western Syntaxis

This paper is not intended to encompass a comprehensive account of the evolution of the Himalayan drainage network, and extensive reviews are found in, e.g., Burbank et al. (1996). However, data and arguments relating to the spatial and temporal evolution of the Indus river, from source to sink, have never been summarised as a whole, yet an understanding of its drainage development is important to constraints of models of orogenic evolution. As outlined in Section 5.3, Clift et al. (2001b) and Metivier et al. (1999) use Indus Fan accumulation rates to argue the relative importance of crustal thickening vs. lateral extrusion in the Palaeogene. Zeitler et al. (2001a,b) surmised that Namche Barwa and Nanga Parbat, the anomalously recently metamorphosed (<10 Ma) and actively rapidly exhuming massifs of the eastern and western syntaxes, respectively (see Section 2), owe their tectonic evolution to the orogen-scale rivers (the Yarlung and Indus) which crosscut them. In their tectonic aneurism model, they suggest that large-magnitude river incision focuses deformation of weak crust leading to lower crustal flow into the region, metamorphism and anatexis.

Today the Indus river flows west along the suture zone (its upper reaches), then makes a right angle bend to flow south across the arc and orogen transverse to tectonic strike, before debouching into the foreland basin, Indus Basin and continuing to final deposition in the Indus Fan (Figs. 1 and 2). Yet initiation of this drainage and its routing through time is controversial, and significant to our understanding and testing of orogenic models.

The first evidence of Indus drainage, in both its upper and lower reaches, is disputed. In the offshore Indus Basin and Fan, significant Palaeogene deposits of Oligocene age have been recognized in seismic and by drilling (Whitmarsh et al., 1974; Clift et al., 2001b; Daley and Alam, 2002 — Section 3.3.1). Kidd and Davies (1978), suggest that the first influx of terrigenous material to the northern Arabian sea in the Mid to Late Eocene is northerly derived and a result of the development of the Himalayan orogen. This is in agreement with the idea that Middle Eocene samples on the Owen ridge are palaeo-Indus Fan sediments (Clift et al., 2001b). The Middle Eocene Owen Ridge sediments are sourced from both igneous and metamorphic facies. Minerals present in the sediments include hornblende, pyroxene, monazite, garnet, sillimanite, kyanite and muscovite. Yet the rocks of the Katawaz basin and Indus suture zone, proposed Palaeogene palaeo-Indus river upstream depositional sites, (see Sections 5.2 and 5.3.1), provide no evidence that material of sufficient metamorphic grade was being eroded by that time in those catchment areas: either an Eocene onland fluvial sediment record of the Indus River with suitable metamorphic provenance is yet to be discovered, or the Eocene Owen Ridge sediments are of mixed provenance, with addition of material from areas other than the more northern reaches of the Indus River.

There seems little doubt that the Indus Fan was in existence in the Palaeogene, and thus a hinterland drainage-basin must have been in evidence by this time. Yet its route is controversial. One school of thought considers the Indus River to have been occupying a similar course to today since the time of collision (e.g., Friend, 1998). If this is the case, suitable alluvial rocks which might represent the palaeo-Indus during the Palaeogene should be present in its downstream reaches. In the Pakistan foreland basin, there are no suitable facies of Eocene age (Section 5.1.1). Whilst Downing et al. (1993) consider that there is no evidence of a palaeo-Indus until the Mid Miocene in the Sulaiman ranges (Vihowa Formation), Welcomme et al. (2001) consider that the older Chitarwata Formation is of deltaic facies but its age is disputed (see Section 3.3.1). Alternatively, the routing of the river may have differed from present day. Earlier routing may have been through the proto-Chaman fracture zone into the Katawaz basin, Kirthar region and Makran — all these regions have suitable Palaeogene deposits, and the provenance, where documented, is recycled orogenic (see Section 5.3.1), Later uplift of the Baluchistan and Kirthar ranges in the Miocene could have pushed the river east towards its current location (Qayyum et al., 1997; Smewing et al., 2002).This reconstruction is however disputed by Sinclair and Jaffey (2001) who point out that provenance considerations render it unlikely that the Katawaz basin rocks are the downstream equivalent of the Indus Molasse since the former are of recycled orogenic provenance and the latter of magmatic arc. They propose that the Palaeogene Indus Molasse of the suture zone is the product of internally drained basins, rather than the deposits of a through-flowing palaeo-Indus, as argued by e.g., Searle et al. (1990), Clift et al. (2000) and Clift (2002) (see Section 5.2). Sinclair and Jaffey (2001) consider that the palaeo-Indus drainage along the suture zone was initiated during the Neogene, after deposition of the Indus Molasse, due to differential uplift resulting from backthrusting of the Indian margin against the Ladakh batholith, along the suture zone. In such a drainage configuration, Palaeogene deposits of the Katawaz basin, Sulaiman and Kirthar ranges, Makran and Arabian sea were sourced by a “palaeo-Indus” with a drainage route in its upper reaches which was radically different to that of Neogene times.

Routing of the palaeo-Indus in its upper reaches is controversial. The upper reaches of the palaeo-Indus may have drained west along the suture zone and not cut transversely through the orogen into the foreland basin until river capture in the Neogene (Brookfield, 1998). Shroder and Bishop (2000) suggest that the cross-cutting of the river transverse through the orogen may have been achieved by river-piracy channeled along tectonic lineaments such as the MMT and later the MBT. Movement on the MMT lineament ceased by 20–18 Ma (Treloar et al., 1989, 1991; Chamberlain et al., 1991; Chamberlain and Zeitler, 1996), co-incident with sedimentological and provenance changes which suggest drainage of the palaeo-Indus in the foreland basin at this time (Section 5.1.1). This interpretation does not support the model of Zeitler et al. (2001a,b) and Shroder and Bishop (2000). They suggest that river capture of the Indus from its upper reaches along the suture zone, transverse through the arc adjacent to the Nanga Parbat region, took place at ca. 11 Ma, coincident and linked to metamorphism and rapid exhumation of Nanga Parbat at that time (ca. 10 Ma — Section 2). However, Najman et al. (2003a) provide evidence that a Kamlial Formation’s source region was exhuming extremely rapidly by 20 Ma (Section 5.1.1). This source could be the Nanga Parbat–Haramosh massif and, coupled with other lines of evidence for early exhumation of the massif (Schneider et al., 1999; Treloar et al., 2000; Pecher et al., 2002) suggests that it is possible that the suggested feedback between Indus River incision and Nanga Parbat tectonics could have occurred, but earlier than previously believed.

6.2.3. Constraints from the sediment record on crustal deformation and accommodation of convergence

A variety of contrasting mechanisms of crustal deformation and accommodation of convergence exist, for example continental underthrusting, homogenous thickening, and lateral extrusion (see Section

2). These models differ in the relative importance and timing of events such as uplift of the Tibetan plateau and crustal thickening, on which the sediment record can provide constraint.

DeCelles et al. (1998a) explain the Late Eocene–Oligocene foreland basin unconformity (Section 3.2) as the result of migration of a peripheral forebulge over backbulge depozone deposits. Using the duration of the unconformity and the modelled width of the forebulge, the migration rate of the orogen and basin can be calculated for that time interval. Combining this with Neogene and modern day rates allowed DeCelles et al. (1998a) to determine that only a third to a half of India–Asia convergence could be accounted for by shortening of the Himalayan fold belt. Mechanisms other than thrust-belt development must have accommodated the remainder.

Continental subduction is used by Guillot et al. (2003) to explain the discrepancy between the amount of Himalayan shortening calculated by mass balanced cross-sections and that deduced from plate reconstruction. In this scenario, the foreland basin regional unconformity may have developed as a response to the change from the continental subduction regime to the continental collisional regime, when full thickness crust reached the subduction zone at 47 Ma (de Sigoyer et al., 2000; Najman and Garzanti, 2000). Such an event may have been accompanied by slab break-off, resulting in isostatic uplift and possible rejuvenation of the load, by which processes unconformities could have developed in the foreland basin (Guillot et al., 2003; Najman et al., 2004). The contrast in sedimentation rates, facies and petrography below and above the unconformity may be explained by the change from the continental subduction-related regime with limited development of topographic relief, to the continental collisional regime where relief became well developed.

Johnson (1994) compared the total volume of sediments eroded from the Himalaya with probable volumes for restored sections of the Himalaya, taking into account uncertainties in crustal thickness and values of shortening (see Section 4.1). Volume balancing between the repositories and the Himalaya only worked if the lowest values for shortening and the lowest value for volume of restored section was used, of which the latter is likely to be an underestimate. From this he concluded that mechanisms other than erosion (see Section 4.1) need to account for between one third and a half of shortening since the onset of collision, although errors are large in both estimations of sediment and Himalayan volume, as discussed in Section 4.1.

The contrasting Palaeogene Indus Fan datasets of Clift et al. (2001b) and Metivier et al. (1999), and thus their conflicting conclusions on the dominance of lateral extrusion or crustal thickening during this period, have already been discussed in Section 5.3.1. Data from the Bengal Fan, (Copeland and Harrison, 1990; Harrison et al., 1992 and Galy et al., 1996) (see Section 5.3.2) led these authors to conclude that their evidence of a high and rapidly eroding Himalaya by the Early Miocene, requires an early age for the development of crustal thickening in the region. In addition, Copeland and Harrison (1990) consider their evidence inconsistent with models in which the Himalaya and Tibet are uplifted either uniformally over the last 40 My or predominantly in the last 2–5 My. By contrast, Rea (1992), examining the rate of sediment accumulation in the Indian ocean (see Section 5.3.1), suggests it provides strong support for theories of Late Cenozoic rapid uplift of Himalaya and Tibet, and does not support theories of widespread uplift and erosion in the Early Miocene.

Caution should be used however when interpreting uplift from the sediment record, since evidence of rapid exhumation need not imply increase in surface uplift (England and Molnar, 1990; Molnar et al., 1993), although Searle (1996) points out that it would be difficult to obtain high rates of erosion without development of topography.

6.3. Uplift of the Tibetan Plateau and climate change

In addition to providing an archive of orogenesis, the sediment record also provides history of climate change. Although this is beyond the scope of this paper, it is proposed that a major reorganisation of atmospheric circulation patterns and intensification of the monsoon resulted from uplift of the Tibetan Plateau (e.g., Ruddiman and Kutzbach, 1991; Prell and Kutzbach, 1992; Kutzbach et al., 1993). Hence the sediment record of climate change has been used to interpret the timing of plateau uplift.

Sedimentological, isotopic and mineralogical changes in the sediment record between ca. 9–7 Ma have been used as evidence of intensification of the monsoon, thereby implying uplift of the Tibetan plateau above a certain threshold by this time. Changes in clay mineralogy in the Bengal Fan (Bouquillon et al., 1990, (see Section 5.3.2) are interpreted as the result of longer storage and more intense erosion after this time (France-Lanord and Derry, 1994). Changes in carbon isotopes in the sediment succession of the foreland basin and Bengal Fan are broadly co-eval with the time of change in clay mineralogy (Quade et al., 1989,1995; Quade and Cerling, 1995; Harrison et al., 1993). The stable isotopic composition of organic matter and carbonate in paleosols can be used to reconstruct palaeovegetation. C_3 plants, which favour a cool growing season, fractionate carbon isotopes differently from C_4 plants which favour a warm growing season. Hence variation in δ^{13}C values in paleosols reflects change in the relative contributions of these vegetation types, which relates to climate. Measurements of oxygen isotopes in soil carbonates in the foreland basin also show broadly co-eval changes. However, further use of the oxygen isotope approach provides evidence of climatic variation prior to this time, both in the foreland basin sediment record (e.g., Dettman et al., 2001) and fluvial and lacustrine sediments to the north of the thrust belt (e.g., Garzione et al., 2000; Dettman et al., 2003). δ^{18}O values of lacustrine carbonates can be taken as a proxy for values of palaeometeoric water. Dettman et al. (2003) used values from lacustrine carbonates to suggest that a change in rainfall values occurred at 12 Ma, which he interpreted as the result of climate change resulting from uplift of the Tibetan plateau. Garzione et al. (2000) used the relationship between δ^{18}O values of meteoric water and elevation to show that by 11 Ma, the southern Tibet plateau was at high elevation. This, plus evidence from other studies including leaf physiognomy (Spicer et al., 2003), and loess deposits (Guo et al., 2002) adds to the growing evidence that the Tibetan Plateau was at high elevation since at least 15 Ma, 22 Ma or perhaps by the Eocene (Tapponnier et al., 2001) or even earlier (Murphy et al., 1997). Thus, other explanations may be sought to explain the climatic changes shown by the sediment record during more recent times.

7. Future directions

In the current economic climate, there can be intense pressure on academics to strive exclusively towards funding-friendly research; "high profile" new models and concepts, preferably produced rapidly within the life-span of a grant-proposal. Where this is achievable, this is excellent, but we should not lose sight of the need to continue exploration in areas where basic data do not yet exist. This fundamental research provides the building blocks upon which our concepts are based and tested, and without which our models may stand on shaky foundations. It should be our goal to work towards greater integration of all data types, using model–data comparisons to constrain our theories of orogenesis.

As has been outlined in this paper, the extent of our knowledge of the material eroded from the mountain belt is spatially and temporally variable. In some areas we have yet to decide even if the sediments are Himalayan derived. For example, only relatively recently was it proposed that the Katawaz basin rocks in Pakistan are of Himalayan origin, the provenance of the Palaoegene Bengal Basin rocks are still disputed, and little is known of the provenance of the rocks deposited further east — for example in Nagaland, India. A more accurate knowledge of the volume of sediment eroded from the orogen is required before we can consider with greater clarity the relative importance of erosion in accommodating India–Asia convergence. Provenance studies of surface exposures, and greater integrated use of well and seismic data would allow us to identify basins filled with Himalayan detritus.

Where such basins have been identified, the age and stratigraphy of the rocks is often still poorly resolved, for example in the Palaeogene records of the suture zone and foreland basin. Considerable effort has been put in to documenting the provenance of these deposits, using both traditional and more recently evolved techniques. Yet the full benefit of this information to unravelling collisional and tectonic events cannot be realised until the rocks can be dated and placed in stratigraphic context. This is a significant challenge, given the often unfossiliferous, deformed and fragmentary nature of the deposits.

The Neogene record, both onland in the foreland basin, and offshore in the deep sea fans, is generally well dated. To these well dated sections, our host of newly developed as well as traditional analytical techniques can be applied with most benefit. We need to continue to improve our analytical techniques in order to be able to identify source areas with more precision. Single grain techniques provide the best way forward since they are not influenced by the effect of mixing of sources. Methodologies which allow more than one type of analysis to be carried out on the same grain are of particular value, allowing most information to be extracted from one grain.

Whilst significant gaps in our basic knowledge still remain, it is prudent to be cautious with our interpretations. Once a comprehensive overview of the sedimentary record has been achieved, more ambitious questions can be tackled more rigorously. For example:

- How much sediment has been eroded from the orogen in total, and remembering that there will be subducted and underthrust sediment unaccounted for:
  - what proportion of accommodation of convergence can be attributed to erosional processes?
- What is the variation in the proportion of sediment eroded from the mountain belt over time?
  - If the lack of substantial Palaeogene orogenic erosion is substantiated, why is there a significant delay between the proposed time of collision, and the onset of major erosional exhumation, in spite of the fact that crustal thickening was clearly occurring by the Oligocene?
  - Which mechanisms of accommodation of convergence, which differ in both their predicted erosional output and relative importance through time, best fit with the sediment record?
  - Erosion of the Himalaya, or specific units within it, is proposed as a cause of sharp rises in the marine $^{87}Sr/^{86}Sr$ value since ca 40 Ma (e.g., Raymo and Ruddiman, 1988; Krishnaswami et al., 1992; Richter et al., 1992; Derry and France-Lanord, 1996; Bickle et al., 2001; Quade et al., 2003). How well does the start of major Himalayan erosion, and the start of erosion of specific units within the Himalaya, correlate with timings of increase in the marine $^{87}Sr/^{86}Sr$ record?
  - How well does the start of major erosion fit with the time of Cenozoic global cooling, proposed to be the result of increased erosion due to the attendant increased drawdown in atmospheric CO_2?

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