Chapter 1: Introduction

1.1 Outline of the problem

India-Asia collision resulted in enormous crustal shortening and thickening (Johnson, 2002) and formation of the Himalayan mountain chain in the early part of Cenozoic (~50 Ma, Green et al., 2008). Evolutionary models derived from analogue and numerical experiments, in agreement with the petrologic and geochronologic data, propose an early soft (~55-42 Ma, ocean-ocean to ocean-continent) collisional phase followed by a hard (~42 Ma onwards, continent-continent) collisional phase (Chemenda et al., 2000; Kohn and Parkinson, 2002; Leech et al., 2005). While the tectonothermal and geodynamic evolution of the Himalayan mountain chain is reasonably well documented for two specific phases (viz. 55-42 Ma and 25 Ma to present), the events within the time segment between 42 Ma and 25 Ma is very poorly constrained (Guillot et al., 2003; Najman et al., 2004). Sedimentary records obtained from the peripheral foreland basin, bordering the Himalayan mountain chain in the south, suggest a delayed (<30 Ma) proto-Himalayan thrust propagation and detritus shedding in the foreland basin compared to the early initiation of the collision at ~55 Ma (Najman et al., 2004). Only a generalized interpretation of a slow exhumation of the hinterland and development of an unconformity in the foreland spanning over ~10 Myr have been suggested (Guillot et al., 2003; Najman et al., 2004; DeCelles et al., 2004) for the aforementioned period.

As fold-thrust belt and bordering peripheral foreland basin start forming at a convergent plate margin, mechanical coupling between the two responds, in concert, to the sequential changes in the collisional tectonics (Ziegler et al., 1995, 1998). Variation in the flexural wavelength (governed by several geodynamic parameters e.g., flexural rigidity of the lithosphere, heat production, slab break-off, lithospheric underplating) during continued collision results in varying temporal responses of the foreland basin (DeCelles and Giles, 1996). It is generally agreed that the downward flexural warping of the subducting slab, upon which the foreland basin forms, is guided by both supra-lithospheric (topographic or static load) and sub-lithospheric loading (subduction or dynamic load, Fig. 1.1a). Topographic loading i.e., the thrust loading generates high-amplitude- short-wavelength (~100-300 km wide basin)- flexure closer to the orogenic front, whereas the subduction load i.e., pull of the subducting slab due to eclogitization of the lower crust generates long-wavelength (~1000 km wide basin) subsidence (Fig. 1.1b, Mitrovica et al., 1989; Gurnis, 1992; Holt and Stern, 1994; DeCelles and Giles, 1996; Catuneanu et al., 1997; Jácome et al., 2003; Catuneanu, 2004).

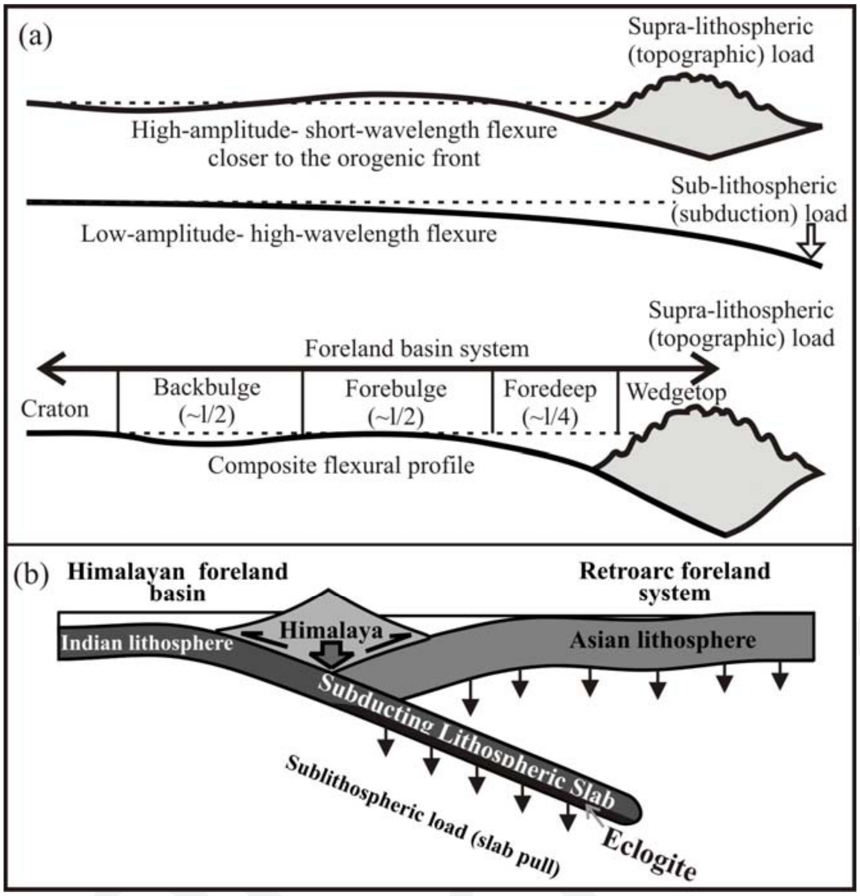

Figure 1.1: (a) Figure showing the characteristic flexural profile developed by the supra- and sub-lithospheric loading and composite flexural profile with four characteristic depozones of a foreland basin system (DeCelles and Giles, 1996). (b) The Himalayan foreland basin and mountain chain between Indian and Asian convergent plate margin.

Thus, sediment accommodation space in the foreland basin (particularly the foredeep) is essentially the manifestation of the thrust loading that results in great thickness (2-8 km) of foreland fill (Jordan, 1995; DeCelles and Giles, 1996). Foreland basin fills are classically divided into sub-aqueous flysch (under-filled stage) and sub-aerial molasse (over-filled stage) deposits (DeCelles and Giles, 1996; Sinclair, 1997). The underfilled depositional realm (underfilled trinity, c.f., Sinclair, 1997) is again sub-divided into (1) a lower unit of carbonate/coal sediments deposited on the passive cratonic margin at the beginning of the marine transgression, (2) a middle unit of offshore hemipelagic sediments, and (3) an upper unit of deep-water turbidite sediments nearer to the flexed orogenic margin, characterized by progressive thrust loading (Sinclair, 1997). Interestingly, in case of Himalayan foreland basin the middle and upper deep-marine sedimentary units (*sensu* Sinclair, 1997) are reportedly absent and shallow marine (inner-shelf) sediments are directly overlain by overfilled- molasse fluvial sediments (Guillot et al., 2003; Najman et al., 2004). Except generalized mentions of apparent absence of subsidence or lack of deposition or, widespread pedogenesis (evidence of sub-aerial exposure of basin) etc. no convincing reason or mechanism have been proposed for this anomaly which differentiates the Himalayan foreland so markedly from that of the Alps (Sinclair, 1997; Guillot et al., 2003; Najman et al., 2004). Few speculative mechanisms put forward are rigid nature of the thick- old- cold- Indian crust or subduction docking, consecutive uplift and high erosion/bypass of the foredeep sediments or low erosion and uplift in arid climatic condition (Najman et al., 2004). It is indeed interesting to know why foreland basin subsidence is low till Miocene (up to ~20 Ma) along the entire Himalayan foreland from north Pakistan in the west to Nepal in the east and deep water facies is apparently absent (Garzanti et al., 1987; Guillot et al., 2003; Najman et al., 2004). Important in this context is the nature of the flysch to molasse transition in a foreland basin which marks an important evolutionary stage of the basin from geodynamic perspective. While the existing models for this changeover varies from (1) progressive thrust propagation towards the foreland and continentalization of the basin (DeCelles and Giles, 1996) to (2) dominance of static loading over dynamic loading in the late phase of foreland cycle (Catuneanu, 2004) or (3) slab break-off followed by isostatic rebound of the subducting slab (Sinclair, 1997), little is understood in the Himalayan context. A model of this kind, however, needs to be supported by the chronology of collision and exhumation of the Himalayan metamorphic core, for which considerable debate exists (Rowley, 1996, 1998; Searle et al., 1997). Opinions vary between diachronous (Ypresian in the west to Lutetian in the east; Uddin and Lundberg, 1998) and synchronous collision and deposition of exhumed material (~31 Ma or Rupelian) through the entire orogenic front (Najman et al., 2005; Acharyya, 2007). Further, any of these mechanisms will have visible imprint on the pattern of continental sedimentation especially the fluvial architecture. The observed change in fluvial architecture from meandering to braided type deposits in the overfilled depositional realms in the Himalayas and change in provenance from a low-grade to progressively higher- grade metamorphic detritus, marking the exhumation of deeper metamorphic core of the orogen, has only been qualitatively linked to progressive thrust propagation (Najman, 2006). Answers to many of these geodynamic questions specially exhumation/erosion history of the orogen lie in the sedimentary record of the Himalayan foreland deposited during the 42-25 Ma period.

Additionally, the knowledge of the exhumation/erosion history of the Himalayan mountain chain, based on the foreland sedimentary archive for this period, is critical to major questions on Cenozoic climate change viz. (i) if the mountain uplift and enhanced silicate weathering are causally linked to observed atmospheric CO_2 drawdown, consequent global transition from greenhouse to icehouse condition and rapid increase in marine $^{87}Sr/^{86}Sr$ ratio across the Eocene-Oligocene boundary (EOB; ~33 Ma; Zachos et al., 1999) and (ii) if the continent-continent convergence and consequent uplift of Tibetan plateau are linked to initiation/intensification of south Asian monsoon. Below some detail of these important questions linked to Himalayan orogeny has been discussed, which were the motivation behind present study, an attempt to seek answer to yet an unresolved question of "Tectonic forcing on late Cenozoic climate" (Raymo and Ruddiman, 1992).

1.1.1 *The greenhouse to icehouse transition and marine $^{87}Sr/^{86}Sr$ ratio*

The transition from early Eocene ice-free greenhouse to early Oligocene icehouse globe with polar ice caps is a dramatic climatic change in the Cenozoic. The rise in atmospheric CO_2 concentration in the early Palaeogene (Palaeocene-Eocene boundary; ~55 Ma), triggered by the destabilization of marine gas hydrate reservoir (Dickens, 2003), and its subsequent removal by silicate weathering during Eocene through EOB (Raymo and Ruddiman, 1992) are current popular hypotheses to explain this dramatic climatic shift. While the incorporation of methane derived CO_2 in the global carbon cycle is well documented by large carbon isotopic depletion in both marine and terrestrial proxy records (Sluijs et al., 2007; see Fig. 1.2 red arrow), the mechanisms behind the gradual drawdown of atmospheric CO_2, linked deep sea cooling, formation of Antarctic ice cap and increasing seawater strontium isotopic composition are still contentious issues mainly due to the inconsistencies among the proxy records (Pearson et al., 2009). The rise in sea water $^{87}Sr/^{86}Sr$ ratio from ~40 Ma onwards has long been interpreted as the indication of enhanced silicate weathering (responsible for the drawdown of atmospheric CO_2) due to exhumation/erosion of the Himalayan mountain chain (Fig. 1.2; Zachos et al., 1999).

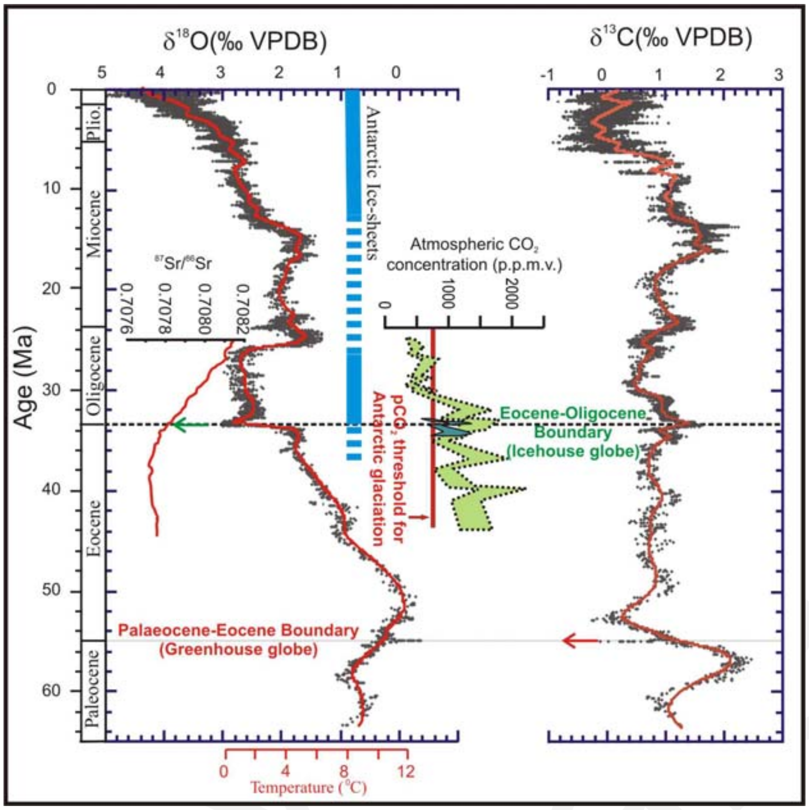

Figure 1.2: Cenozoic deep sea oxygen and carbon isotope record constructed from long-lived benthic taxa, *Cibicidoides* and *Nuttallides* (Zachos et al., 2001) showing depletion in carbon isotopic record (red arrow) across Palaeocene-Eocene Boundary and fall in deep sea temperature (green arrow) across Eocene-Oligocene Boundary (EOB). Marine $^{87}Sr/^{86}Sr$ ratio (Zachos et al., 1999) shows the characteristic rise from ~40 Ma onwards. The long term atmospheric CO_2 concentration (green envelop, Pagani et al., 2005) and across the EOB (blue envelop, Pearson et al., 2009) show a general declining trend with few aberrations. The blue bar indicates the temporal change in the extent of Antarctic ice sheet (Zachos et al., 2001).

The argument for this hypothesis was the fact that, the two major rivers viz. Ganges and Brahmaputra, draining the Himalaya today, are responsible for supplying over 25% of the total sediments and ~9 % of dissolved solid entering into the oceans with anomalously high $^{87}Sr/^{86}Sr$ ratio (Edmond, 1992; Palmer and Edmond, 1992). Such explanation has, however, been questioned by various workers. First, at ~40 Ma, when marine $^{87}Sr/^{86}Sr$ ratio began to rise, the Himalaya was a small scale topographic feature (Rea, 1992; Quade et al., 1997; Rea et al., 1998) and rapid exhumation of Himalayan Mountain chain began only in early- late Miocene (Harrison et. al., 1992; Clift et al., 2008). Hence, contribution of sediments as well as dissolved radiogenic Sr load to the ocean was presumably less than what has been estimated for the last 10 Ma (Galy et al., 1999). Second, studies from the headwater of Yamuna (Dalai et al., 2002), Ganges (Bickle et al., 2003) in India and Arun and Seti river of Nepal (Quade et al., 2003) show that the elevated $^{87}Sr/^{86}Sr$ ratios in these river waters are mainly due to the weathering of Himalayan metacarbonates and not so much from silicate weathering. Third, although the evidence of small transient alpine type ice sheet in Antarctica around 38-39 Ma has been used in favor of silicate weathering and linked rise of marine $^{87}Sr/^{86}Sr$ hypothesis (Zachos et al., 1999) strong reservation exist on the possibility of stable ice sheet formation before the EOB (Pekar and Christie-Blick, 2008). This is also supported by recent isotope capable global climate/ice sheet model which suggest a lower (~750 p.p.m.v.) atmospheric CO_2 concentration as threshold value for the development of stable Antarctic ice sheet (DeConto and Pollard, 2003; DeConto et al., 2008). Although not persistent, many CO_2 proxy data on the other hand, show a high atmospheric CO_2 level during the late Eocene (see Pagani et al., 2005; Peason et al., 2009).

A serious problem in this climate reconstruction is the paucity of high resolution atmospheric CO_2 data those can be compared with the ice volume record (Pearson et al., 2009). Recent studies show that during this climate shift an initial global cooling triggered the formation of small scale ice cap that was followed by a rapid ice sheet growth amplified by both orbital forcing and positive ice albedo feedback (DeConto and Pollard, 2003; Coxall et al., 2005; Pälike et al., 2006; Lear et al., 2008; Katz at al., 2008; DeConto et al., 2008). The atmospheric CO_2 concentration however show substantial fluctuation during this period so much so that survival of ice sheet in a post transition high (~1500 p.p.m.v.) CO_2 atmosphere has also been proposed (Pearson et al., 2009). The two most important atmospheric CO_2 records i.e. marine Alkenone (Pagani et al., 2005) and foraminiferal boron isotope (Pearson et al., 2009) based pCO_2 estimates show several short lived episodes (aberrations) of low pCO_2 (Fig. 1.2) those prevailed, both in the late Eocene and early Oligocene, over a general declining trend of atmospheric CO_2 concentration when it changed from ~>1000 p.p.m.v. level to the characteristic low-Miocene ~<500 p.p.m.v value. These evidences suggest that in reality perhaps climate system behaves in a more complicated way than was thought before. Hence, making Himalayan exhumation/erosion/silicate weathering (tectonic forcing) as the sole mechanism for driving atmospheric CO_2 drawdown, increase in marine $^{87}Sr/^{86}Sr$ ratio and eventual formation of Antarctic ice cap formation might be too simplistic. It is equally possible that a brief tectonic forcing could have triggered the late Cenozoic climate change and was eventually boosted by other multiple positive feedback parameters. One way to address this problem is to retrieve the exhumation/erosion history of Himalayas during this time period (42-25 Ma; see discussion above) for which not much data base, either in terms of chronology or foreland sediment record, exist. Also an exciting task would be to find out a continent based CO_2 proxy right in the foreland sediments and estimate the pCO_2 level. Till then the causal link between tectonics forcing and climate change remains poorly understood.

1.1.2 *Plateau uplift and initiation/intensification of South Asian Monsoon System*

Constraining the India-Asia Collision at ~50 Ma, subsequent uplift of the Tibetan Plateau (Garzione et al., 2000a, b; Spicer et al., 2003; Cyr et al., 2005; Currie et al., 2005; Rowley and Currie, 2006; DeCelles et al., 2007; Saylor et al., 2009) and its impact on the setup of a prototype monsoon system (Kutzbach et al., 1989; Molnar et al., 1993; Wei et al., 2006; Harris, 2006; Clift et al., 2008) is an area of intense research for past few decades. While a critical height of the Tibetan Plateau (TP) was long held responsible for the initiation of South Asian Monsoon System (SAMS, Prell and Kutzbach, 1992), recent work on upper atmospheric temperature estimates suggests that the low pressure zone, which drags the monsoonal wind, is largely created over the continental north India and not over the TP as thought earlier (Boos and Kuang, 2010). This, along with General Circulation Model (GCM) simulation with removal of the entire TP shows that SAMS can be still activated by the narrow orography of the Himalayas alone thus questioning the connection between plateau uplift and monsoon (Boos and Kuang, 2010).

Till this hypothesis gets a firm ground, number of studies attempted independent estimation of past plateau height to assess the applicability of models involving crust-mantle dynamics (England and Housemann, 1986; Molnar et al., 1993; Tapponnier et al., 2001; Beaumont et al., 2004) and climate change (Harrison et al., 1995; Zheng et al., 2000; Dettman et al., 2001, 2003). Several attempts have been made to constrain the palaeo-altitude of TP which used variety of proxy records viz. pollen (Dupont-Nivet et al., 2009), leaf physiognomy (Spicer et al., 2003), C_4 vegetation abundance (Wang et al., 2006), fossil faunal and floral data (Mercier et al., 1987) as well as empirically derived (Garzione et al., 2000a, b; DeCelles et al., 2007; Saylor et al., 2009) and atmospheric thermodynamics based altitude-dependent stable (oxygen) isotope fractionation (Rowley et al., 2001; Cyr et al., 2005; Currie et al., 2005; Rowley and Currie, 2006) models. Marked discrepancy, however, exists between, for example, the faunal/floral based and stable isotope based palaeo-altitude (see Fig. 1.3 top inset).

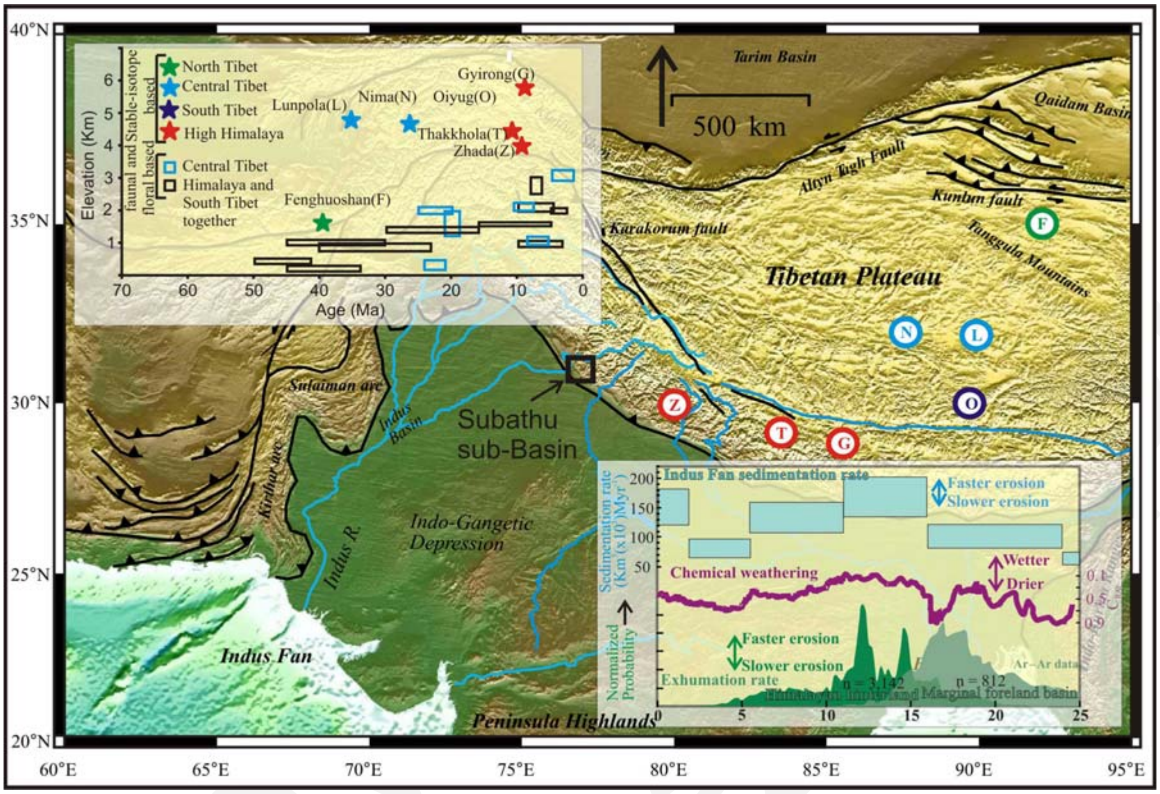

Figure 1.3: Topographic map of south Asia showing India and Tibetan plateau with locations of the Cenozoic sedimentary basins i.e., Zhada (Z), Thakkhola (T), Gyirong (G), Oiyug (O), Nima (N), Lunpola (L), and Fenghuoshan (F) from where palaeo-altitude data are available. Top inset showing the elevation of the respective basins versus age (Stars: stable isotope based estimates, Boxes: faunal and floral based estimates, Rowley and Garzione, 2007; Mercier et al., 1987). Note large discrepancy between these two independent proxy records. The bottom inset shows the exhumation-erosion proxy record of the Himalayan mountain chain with purple line indicating the oldest monsoonal proxy record going maximum up to ~23 Ma (see Clift et al., 2008).

Interestingly, several recent works propose that the erosion-driven exhumation of the Himalayan orogen is genetically linked to the Neogene monsoonal activity thus making the causative mechanism reversed (climate forcing on tectonics and not the tectonic forcing on climate as discussed earlier, see Figs. 1.3, 1.4; also see Wang et al., 2005; Wei et al., 2006; Clift et al., 2008 for detail review). Compared to the TP, however, palaeo-altitude of the Himalayan range is poorly documented.

1.1.3 *Tectonic forcing or climate forcing – a chicken-egg debate*

The link between tectonic and climate has often been considered as one dimensional vector (Fig. 1.4), with the climate change preceded by the significant uplift. But, recent studies from large scale topographic analysis and field evidences have shown that climatic variation can be a first order control on the morphology and uplift rate of the mountain belt (Montgomery et al., 2001; Thiede et al., 2009). Although spatial correlation between precipitation and rapid rock uplift has been recognized for the recent past, more compelling evidences for climate forcing on tectonics on a million year time scale is yet to be convincingly documented (see Whipple, 2009 for details). The question of climate driven tectonics becomes more important in case of large orogenic belts like Himalaya, lying in zone of a major climate-switch like monsoon, where intense precipitation indeed causes rapid erosion. A unique proposition is that the orogenic margin (locale of the precipitation) eventually drags large volume of mid-crustal rocks to establish the so-called crustal channel flow and uplift (Beaumont et al., 2004; Jamison et al., 2004). It is known that the channel flow model is capable of explaining many features of the Himalayan mountain chain e.g. layered crustal structure, mid-crustal low seismic velocity zone, inverted metamorphic field gradient etc. (Klemperer, 2006). Nevertheless, in the context of climate-tectonics connection, the validity of the model critically depends upon the exogenous factor like focused orographic precipitation on a previously established orographic gradient, rather than the endogenous factors e.g. age of the initiation of collision, convergence rate, relative strength of the crustal layers etc (Beaumont et al., 2004; Jamison et al., 2004). High resolution palaeo-monsoon record from marine (south China and Arabian sea) and Chinese loess sediments from Indo-Asian region, indicate a strong South Hemispheric (Antarctic ice sheet) forcing on Indian monsoon between 5 to 2.75 Ma, a period when North Hemispheric (Arctic) ice cap was minimal (Clemens et al., 2008). If true a positive feedback of southern hemisphere glaciation on the onset of Indian monsoon, consequent erosion and channel flow induced uplift during the Oligocene can be a distinct possibility. Hence understanding the relative timing between the onset of precipitation and channel flow (or uplift) is the key issue to solve the chicken-egg debate for Himalayan exhumation and initiation/intensification of south Asian monsoon.

If the climate forcing on tectonics be true then an early (Palaeogene) upliftment history (DeCelles et al., 2007; Rowley and Currie, 2006, Cyr et al., 2005) of the TP (or Himalayas in general) demands a concomitant onset/intensification of monsoonal activity way back into the Palaeogene for which no data exist till date. The oldest ocean based monsoon record is available up to 23 Ma only (Fig. 1.3). As discussed above, some stable isotope data about past plateau altitude already exists. However, based on the foreland and TP soil carbonate stable isotope data serious doubt has recently been raised about the pristine nature of these orogenic belt carbonates and their use in palaeo-altitude estimations (Leier et al., 2009; Snell et al., 2010). Below the information available on the limited stable isotope data on foreland sediments have been reviewed.

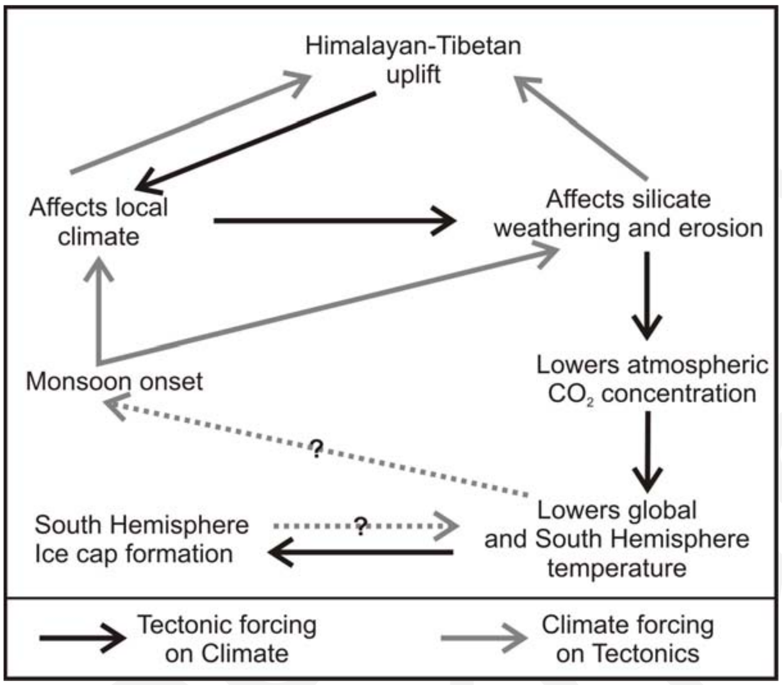

Figure 1.4: Schematic diagram showing the two possible ways by which tectonics (Himalayan orogeny) and Cenozoic climate change can be related.

1.1.4 *An appraisal of the stable isotope systematics of Himalayan foreland sediments*

Compared to large body of geochemical and chronological data (see Najman, 2006 for review), stable isotope data on foreland sediments are scanty mainly coming from the Subathu sub-Basin, India and Nepal (Fig. 1.5; Singh and Lee, 2007; Singh et al., 2007, 2009; Leier et al., 2009). These studies included the marine carbonate shells of Palaeocene-Eocene Subathu Formation (Leier et al., 2009) and soil carbonates of Oligocene Dagshai and Dumri Formations (Singh and Lee, 2007; Singh et al., 2007, 2009; Leier et al., 2009; see later discussion on detail geology of Himalayan foreland including various stratigraphic Formations). These studies were mainly focused on the diagenetic alteration of the foreland carbonates and possible implications in retrieving the atmospheric CO_2 concentration during the Oligocene Dagshai/Dumri rocks (Singh and Lee, 2007; Leier et al., 2009). When critically looked into, these data present contrasting views. Figure 1.5 shows plot of available oxygen and carbon isotope data of foreland sediments (soil carbonates) which shows large ranges in both isotopes. While Singh and Lee (2007) explained their crude positive correlation as a signature of biogenic fractionation of stable isotope suggesting preservation of original isotopic signature, extremely depleted oxygen isotope values even in micritic carbonates were explained by complete diagenetic overprinting of all foreland sediments during high temperature burial diagenesis (Najman et al., 2004; Leier et al., 2009; Fig.7) thus making them unsuitable for any palaeo-climatic interpretation. Because sediments in most orogenic belts undergo deep burial and diagenesis, it is possible that the original signatures may be considerably lost. If true then all stable isotope based palaeo-altitude estimation from the Himalaya must be reassessed (Snell et al., 2010). This suggests that further fabric selective stable isotope data are needed to see the efficacy of stable isotope based altimetry in an orogenic belt.

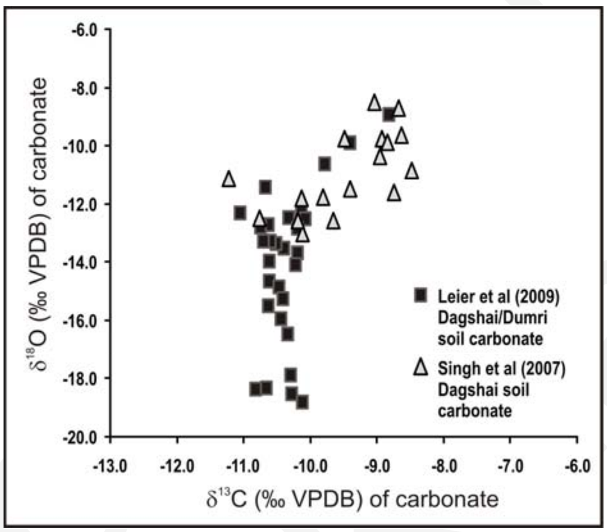

Figure 1.5: Stable oxygen and carbon isotopic compositions of foreland soil carbonates (data source: Singh et al., 2007; Leier et al., 2009).

Another major problem in the stable isotope based palaeo-altimetry is the correct assessment of the meteoric water isotopic composition of coexisting low altitude site that fell on the past moisture transport direction (Rowley et al., 2001; Rowley and Currie, 2006; DeCelles et al., 2007) against which palaeo-data are compared. Oxygen isotope based palaeo-altimetry equation of Rowley and Currie (2006) has the following form:

$$H = -6.14\times10^{-3}\Delta(\delta^{18}O_p)^4 - 0.6765\Delta(\delta^{18}O_p)^3 - 28.623\Delta(\delta^{18}O_p)^2 - 650.66\Delta(\delta^{18}O_p)$$

Where, the H (altitude) is a function of $\Delta(\delta^{18}O_p)$ (difference in oxygen isotopic composition of precipitation between that at sea level and at elevation). In absence of proper meteoric water stable oxygen isotopic composition of low altitude site (e.g. data from the Himalayan foreland) all the earlier studies on the Palaeogene palaeo-altitude of Tibet (Fig. 1.3 top inset) i.e. from Fenghuoshan Basin (39-36 Ma, Cyr et al., 2005), Lunpola Basin (35±5 Ma; Rowley and Currie, 2006), and Nima Basin (~26 Ma, DeCelles et al., 2007) used oxygen isotope data of either fluvial Miocene Siwalik carbonates (that presumably deposited at an ancient low altitude level) or modern day New Delhi precipitation for which long term average $\delta^{18}O$ value is known. Because the nature of monsoon trajectory and sources of vapour which control the eventual isotopic composition of precipitation (see Sengupta and Sarkar, 2006 for stable isotope signature of modern monsoon system) could have been drastically different during the Cenozoic, an element of speculation always remains in altitude calculation. Unless resolved by more robust estimation the question of antiquity of Asian monsoon system (late Oligocene or early Miocene; Clift et al., 2008) remains unresolved. Again the sedimentary and isotope records in foreland sediments are promising archives to unravel some of these mysteries.

It is now obvious that the sedimentary archive of Himalayan foreland might be a treasure trove for finding answers to many of the questions raised above. At this stage it, therefore, prudent to discuss the sedimentology of the Himalayan foreland sediment particularly the Palaeogene part.

1.2 Review of the sedimentology of the Palaeogene Himalayan foreland sediments

1.2.1 *Litho-tectonic setup of the Himalayan mountain chain and occurrence of Palaeogene foreland sediments*

The Himalaya consists of six litho-tectonic zones (Gansser, 1964). From north to south, these belts are as follows. (i) The Trans-Himalayan zone: situated north of the Indus suture zone, mainly consists of late Cretaceous to Eocene calc-alkaline plutons (Le Fort, 1996). (ii) The Indus suture zone: the collisional zone between India and Asia mainly consists of deep water sedimentary accretionary wedge complex sediments, ophiolites, ophiolitic mélange, and island arc volcanic rocks (Searle, 1983; Robertson and Degnan, 1993). (iii) The Tethyan Himalayan zone: lies between Indus suture zone and South Tibetan Detachment (STD) is composed of Cambrian to Palaeocene sediments, deposited on the Indian passive continental margin (Gaetani and Garzanti, 1991). (iv) The Greater

Himalaya zone: lies between STD and Main Central Thrust (MCT) mainly consists of metamorphosed Indian continental crust and overlying sediments (late Proterozoic–Cambrian age) (Parrish and Hodges, 1996) and intruded by leucogranite of Miocene age (Treloar and Searle, 1993, and papers therein). (v) The Lesser Himalayan zone: bounded by MCT to the north and Main Boundary Thrust (MBT) to the south, composed of non-metamorphosed or weakly metamorphosed Indian continental crust and sedimentary rocks (middle Proterozoic to Palaeozoic) (Valdiya, 1980; Parrish and Hodges, 1996), and sparse streaks of Palaeogene foreland basin sedimentary rocks (Srikantia and Bhargava, 1967; DeCelles et al., 1998a) and (iv) The Sub-Himalayan Zone: bounded by MBT to the north and Main Frontal Thrust (MFT) to the south, composed of Palaeogene-Neogene foreland basin sediments eroded from the rising mountain chain (Najman et al., 1993; Critelli and Garzanti, 1994; DeCelles et al., 1998a; Fig. 1.6).

As stated earlier, the foreland basin fills are classically subdivided into underfilled (marine) and overfilled (fluvial) phases depending on the sediment deposition history (Sinclair, 1997; DeCelles et al., 1998b). The Palaeogene foreland basin of the Himalaya, studied during the present work, consists of underfilled marine and overlying continental fluvial overfilled sediments. These sediments are now exposed along the Sub-Himalayan belt of entire northwest Himalaya (Fig. 1.6). Among the pioneer workers, Medlicott (1864) mapped the area between rivers Ravi and Ganga and proposed a three-fold subdivision of these lower Tertiary sequence i.e. marine Subathu Formation (≡ Patala in Pakistan, ≡ Subathu in India, ≡ Bhainskati in Nepal), continental molasse Dagshai (≡ Balakot in Pakistan, ≡ Lower Murree in Jammu, ≡ Lower Dharamsala in Kangra, ≡ Dumri in Nepal) and Kasauli (≡ Upper Murree in Jammu, ≡ Upper Dharamsala in Kangra) Formation (see Fig. 1.6). This stratigraphic classification has generally been accepted by all subsequent workers viz. Pilgrim and West (1928), Auden (1934), Wadia (1957) and Krishnan (1958) as well as most recent workers. The Neogene part of the foreland deposits consists entirely of continental Siwalik Group sediments (Sanyal et al., 2005) and is not discussed in the present work.

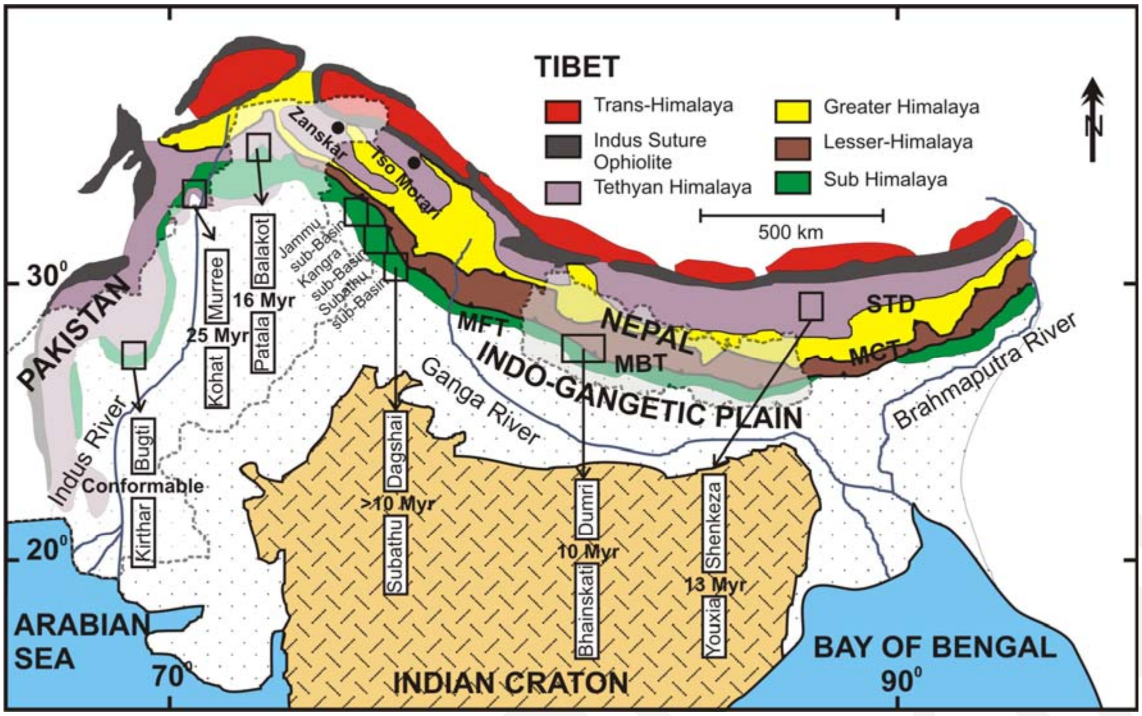

Figure 1.6: Litho-tectonic subdivision of Himalaya showing the disposition of major thrusts and six litho-tectonic belts. The boxes indicate the important outcrop belts of Palaeogene sediments in Pakistan, India, and Nepal along with approximate duration of unconformity between the marine and continental deposits at each location (see chapter 3 for detailed explanations).

1.2.2 *Distribution, sedimentology, and depositional age of the Palaeogene foreland sediments in Indian Himalaya*

The Palaeogene successions in Indian part of the Himalayan foreland are well exposed along the foothill belts of Jammu, Kangra, and Simla Himalaya of NW India and extend through Nepal to NE India where it is drained by river Brahmaputra. In India, the continental Palaeogene sediments generally overlie the Precambrians (Sirban Limestone, Bandla Limestone, Simla Slate) with pronounced unconformity marked by oxysol and/or bauxite and underlain by Neogene Siwalik sediments (Bhatia, 1982, Najman et al., 1993; Singh, 2003; DeCelles et al., 2004). The cumulative thickness of the Palaeogene succession i.e., Subathu, Dagshai, and Kasauli Formation rocks exceed over ~3600 m (Karunakaran and Ranga Rao, 1979; Fig. 1.7). Stratigraphically older Subathu Formation rocks start with a chert breccia, bauxite, successively overlain by carbonaceous coal/shale (Singh, 2003), cyclic succession of green/yellow shale, and carbonate (Singh and Andotra, 2000). While the lower chart breccia, bauxite, and black/gray coal/shale have been interpreted as early transgressive deposits over Precambrian basement, the overlying cyclic occurrence of green/yellow shale and oyster bearing carbonates have been interpreted as product of a barrier bar-lagoonal setup (Singh and Andotra, 2000; Singh, 2003). Red shale with fossiliferous carbonate lenses occupying the top most part of Subathu succession has variously been described as red Subathu and/or passage beds (*sensu* Bhatia and Mathur, 1965). Based on the presence of volcanic lithic fragments, serpentine schist, chart, and Cr spinel within the coarser clastic interbeds in red Subathu rocks, Najman and Garzanti (2000) interpreted a mixed source for terminal marine Subathu sediments. Further, geochemically, high Ni and Cr content, whole rock ε_{Nd} values of ~–9 and $^{87}Sr/^{86}Sr$ ratios of ~0.710-0.715 suggest significant input from ophiolites of Indus suture zone (Najman et al., 2000; Najman, 2006).

Formation	Member or equivalent	Age: Najman, 2006	Age: Jain et al., 2009	Age: Bhatia and Bhargava, 2006	Thickness	Facies
Kasauli Formation	Kasauli	<22Ma	~21Ma ~25Ma	*Early-Miocene*	2100m	Alluvial
Dagshai Formation	Kumarhatti-Solan	~27 Ma		*Late Bartonian to Rupelian (~39 to 31 Ma)*	900m	Alluvial
	Main Dagshai	<28 Ma	~29Ma			
	Lower Dagshai	~28 Ma	~31Ma			
Unconformity	White Sandstone/ Basal Dagshai	~31Ma				
Subathu Formation	Red Subathu/ Passage beds	*~44Ma*		*Late Lutetian to Middle Bartonian (~45 to ~39 Ma)*	600m	Shallow marine
	Black/Gray Subathu		~49Ma	*Late Thanetian to Middle Lutetian (~56 to ~45 Ma)*		
Simla slate or Precambrian basement						

Figure 1.7: General stratigraphy of Palaeogene foreland sediments and depositional age (biostratigraphic ages shown in italics). Note the unconformity placed at the base of the White Sandstone; however, if the White Sandstone is to be marine the unconformity should be placed atop it (see text for details). Data compiled from Mathur (1978, 1979), Karunakaran and Ranga Rao (1979), Najman et al. (1997, 2004), Singh and Andotra (2000), Singh (2003), Najman (2006), Bhatia and Bhargava (2006), and Jain et al. (2009).

The terminal red Subathu shale in the entire north-western Himalaya has been found underlain by a characteristic greenish gray, quartz rich sandstone unit described as White Sandstone unit. The White Sandstone unit, representing the topmost part of Subathu succession, marks the termination of marine sedimentation in the Himalayan foreland, is laterally persistent on basin scale and has widely been used for regional correlation of the lithounits in an otherwise high structurally disturbed terrain (Najman and Garzanti, 2000; Najman et al., 2004; Sangode et al., 2005). Based on the presence of clean white nature, well sorted- medium sized- quartz grains, characteristic sedimentary structure, and facies association this unit has been considered as coastal barrier or beach deposits by Srivastava and Casshyap (1983). Najman et al. (2004 and reference therein), however, included this unit into the overlying fluvial Dagshai Formation rocks based on its erosive base, presence of red shale clasts in the basal part, distinctive isotope, trace element, and detrital composition. Thus the interpretation of depositional environment for this important marker horizon posed a major stratigraphic problem to researchers working on the Palaeogene Himalayan foreland sediments (see later discussion on chronology and unconformity).

The Dagshai Formation rocks, overlying the White Sandstone unit, are composed of alternating red/brown sandstone (thickness not exceeding 10 m), mudstone, and mature calcrete. The fining upward cycles constituted of sandstone-mudstone couplets are interpreted as of fluvial channel-levee-flood plain origin (meandering river deposits; Najman et al., 2004). Kasauli Formation rocks, overlying the Dagshai Formation rocks, consist dominantly of gray-colored multistoried sandstone beds (thickness always >10 m) separated by minor gray/green and occasionally red mudstone- siltstone units. The mudstone-siltstone units are thin- to thick- bedded, and locally contain well-preserved fossil leaves. Kasauli Formation rocks have been interpreted as braided river product (Najman et al., 2004) due to its high sand:mud ratio than Dagshai Formation with near absence of floodplain deposits. Presence of low grade metamorphic rock fragments in the basal part of Dagshai succession (including White Sandstone unit) and progressive increase in the metamorphic grade of the rock fragments in stratigraphically higher sections (up to Kasauli Formation rocks) have been interpreted as the signal for shedding from more and more deeper parts of exhumed Himalayan metamorphic core (Najman et al., 1997). This was also supported by whole rock ε_{Nd} values (ranging between ~−12 to −18 and $^{87}Sr/^{86}Sr$ ratios of ~0.755-0.775; Najman et al., 2000) of these continental rocks. Thus available sedimentological, petrographic, and geochemical or isotope data indicate a major change in both style of sedimentation (e.g. marine to continental) and provenance from Subathu to Dagshai/Kasauli Formation.

Such changes from marine to continental set up can be much better understood by applying modern sequence stratigraphic analysis that deals with the spatial and temporal change in accommodation space, in response to orogeny driven subsidence/upliftment of the foreland basin and/or eustacy (Catuneanu, 2006). Additionally, the change in fluvial depositional type from meandering to braided style across Dagshai-Kasauli transition is not yet studied from sequence stratigraphic appraisal. For example, sequence stratigraphic analysis, carried out in other foreland basin of the world, e.g., Karoo (South Africa) and Western Canada foreland Basin (North America), have been used for understanding the flexure control on the foreland basin sedimentation and depositional cycle (Catuneanu, 2004). Model involving the tectonic loading, creation of accommodation space, and depositional pattern of foreland basin depicts an increasing rate of accommodation space creation during the thrust loading phase and deposition of orogeny derived material to foreland adjacent to the orogenic front. On the other hand, erosion induced rebound cause destruction of accommodation space in the proximal part of the orogenic belt and major sediment bypassing across the foreland. Sediment deposition on the distal part of the foreland basin increases during this period and marked by unconformity development in the proximal part. These tectonic responses together result in unconformity bounded fining up fluvial succession from high energy, coarse-grained braided to low energy, fine grained meandering style deposits in the proximal part of the foreland basin and thus help to understand the chronology and amplitude of thrust propagation (Catuneanu and Sweet, 1999; Catuneanu and Elango, 2001; Catuneanu and Bowker, 2001; Catuneanu, 2004; Catuneanu, 2006). No such attempt, however, has been made to understand the dynamic evolution of Himalayan foreland basin and most sedimentological studies were confined either to mere facies analysis or general provenance study. Because no sequence stratigraphic data are available, the major problems of Himalayan foreland sedimentation e.g. where the marine sedimentation starts (the 'White Sandstone' debate) or what is the nature and duration of the unconformity (between marine and continental deposits), what drove the transition from meandering to braided fluvial style still remain unresolved. Below some of these pertinent problems of the Himalayan foreland stratigraphy has been discussed.

1.3 Review of the chronology of Himalayan foreland sediments and span of unconformity between marine and continental deposits

The Palaeogene sequence comprising the Subathu, Dagshai, and Kasauli Formations was initially considered as a continuous and conformable succession (Medlicott, 1864). Based on the occurrence of Shallow Benthic Zone (SBZ of Serra-Kiel et al., 1998) indicators *Daviesina garumnensis* (SBZ-4) in Jammu area and other diagnostic foraminifera of SBZ-5 to SBZ-12, the lower and middle part of the Subathu Formation were dated as Palaeocene to early Eocene age (Bhatia and Syed, 1988; Mathur and Juyal, 2000). From the presence of larger benthic foraminifer *Assillina spira abrardii* and *A. exponens–A. papillata–Nummulites discorbinus* assemblage (SBZ-13 and SBZ-14) bearing beds the terminal Subathu Formation rocks of the Simla Hills and Jammu region were assigned an early to late Lutetian age, the upper limit being ~ 44 Ma (Fig. 1.7; Mathur, 1978; Bhatia and Bhargava, 2006). From the occurrence of several sporadic records of palynomorphs, leaf impressions, molluscs, ostrocodes, and vertebrates Dagshai Formation rocks have been assigned a late Eocene to Oligocene age. These fossils include *Linderina* (from basal Dagshai; late Bartonian to Priabonian), and charophyte assemblage of *Harrischara* cf. *vasiformis, Nitellopsis (Tectochara), latispira-Rhabdochara* sp., and *Chara* sp. (from the Dharamsala area: middle Bartonian to middle Priabonian), and few palynomorphs (Mathur et al., 1996; Bhatia and Bhargava, 2006). However, because both *Lindarina* and the reported charophytes and palynomorphs are not very specific age characteristics and most of these faunal data are not formally published, the above deposition age of Dagshai Formation was strongly challenged by Najman (2007). Based on the palynological and the charophyte data (e.g. *Meyerepollis*: late Eocene–Oligocene, *Chara microcera*, *Stephanochara ungeri-Chara notata:* late Rupelian–middle Chattian and *S. ungeri*: early Miocene; Mathur, 1984; Arya, 1997) the Kasauli Formation rocks have been assigned an Oligo-Miocene age. With the Lutetian or early Eocene age of the terminal part of marine Subathu and Oligo-Miocene age of the overlying continental molasse sediments, these dates envisaged a major unconformity in between. The scarcity and doubtful nature of Dagshai body fossils, as discussed above, made the chronology of this important transition more contentious.

Advent of modern sophisticated dating techniques did not improve the situation either. Zircon fission track date of the lower part of the Subathu Formation yielded an age of 49.4 ± 2 Ma (Jain et al., 2009), while that for the White Sandstone unit provided a maximum depositional age of ~31 ± 2 Ma (Najman et al., 2004). For the lower Dagshai rocks laser ablation $^{40}Ar/^{39}Ar$ and fission track dates of single detrital muscovite and zircon grains suggested ages ranging from <28 Ma (Najman et al., 1997) to 31.6 ± 3.9 Ma (Jain et al., 2009) respectively (Fig. 1.7). The implication of these dates is that if an unconformity is considered at the base of the White Sandstone a large (>10 Ma) hiatus is a must. Alternatively if the unconformity is placed atop the White Sandstone, the hiatus must be of a small duration. The debate, therefore, essentially centers on the White Sandstone precisely, because earlier workers made completely divergent interpretations about its depositional environment and hence stratigraphic status (see Fig. 1.7). As mentioned earlier the beach or tidal flat (Singh and Khanna, 1980; Srivastava and Casshyap, 1983) depositional setting for this unit meant its inclusion in marine Subathu, while its geochemical (Nd Sr isotope ratios) similarities with the Dagshai rocks led Najman et al. (2000, 2004) to include it in fluvial Dagshai Formation. Thus based on the biochronological continuity, Bhatia and Bhargava (2006) made a strong campaign for a conformable contact between the marine and continental sediments but was strongly challenged by Najman (2007), who based on the absolute dates and geochemistry, suggested a large unconformity of >10 Ma (covering late Eocene to early Oligocene). Further, this unconformity was considered as a basin wide synchronous phenomena from Pakistan (between the Patala and Balakot Formations) through India (Subathu and Dagshai Formations) to Nepal (Bhainskati and Dumri Formations)—all three pairs being supposedly correlative (DeCelles et al., 1998a; Najman et al., 2005; Najman, 2006; Figs. 1.6, 1.7). Neither the palaeontological nor the absolute chronology based claim, however, is supported by any unequivocal field or sedimentological evidence or sequence stratigraphic analysis, so essential to establish the exact nature and stratigraphic status of the marine to continental transition in Himalayan foreland. Only passing references to the contact between the two formations as being "sharp and well defined, with no observable angular unconformity" has been made by earlier workers (see Najman et al., 2004). To address the issues, raised above, the present work focused on detail sedimentological, sequence stratigraphic and stable isotope investigation in parts of the Subathu sub-Basin of NW Himalayan foreland.

1.4 Brief geology of the Subathu sub-Basin

The Subathu sub-Basin is located in the west-central part of the Himalayan foreland basin (HFB) and is marked by the Yamuna Transverse Fault in the east and Fugtal–Manali–Ropar transverse fault in the west (not shown in Fig. 1.8). The basin is bounded by Main Boundary Thrust (MBT; Krol Thrust) to the north and the Main Frontal Thrust (MFT) to the south (Fig. 1.8). Both Palaeogene (Subathu, Dagshai, and Kasauli Formation) and Neogene (Siwalik Group) rocks and post-Siwalik Late Quaternary Pinjaur Dun sediments are well exposed along the Sub-Himalayan belt with a nearly NNW-SSE strike (Kumar et al., 2007). The foreland basin sediments occur in two parallel structural units in Subathu sub-Basin. The first unit (Surajpur Structural Unit) crops out between MBT and Surajpur Thrust (a major splay of MBT) and the second unit (Bilaspur Structural Unit) exposed between Surajpur Thrust and Bilaspur Thrust (another splay of MBT). The Palaeogene rocks display repetitive map pattern with dips generally exceeding 80 ° towards ENE. The repetition of Subathu-Dagshai-Kasauli Formation rocks in the Sub-Himalayan region of Simla Himalaya has variously been interpreted as either due to folding (Khanna, 1978) and/or thrusting (Raiverman, 1979). Later, Mukhopadhyay and Mishra (1999) also interpreted this repetitive map pattern as a result of numerous splay thrusting. A serious problem in studying such structurally disturbed terrain is, therefore, to erect appropriate lithologs and time stratigraphy before any other temporal history (be it sedimentology, sequence architecture or isotope based climate change) is reconstructed.

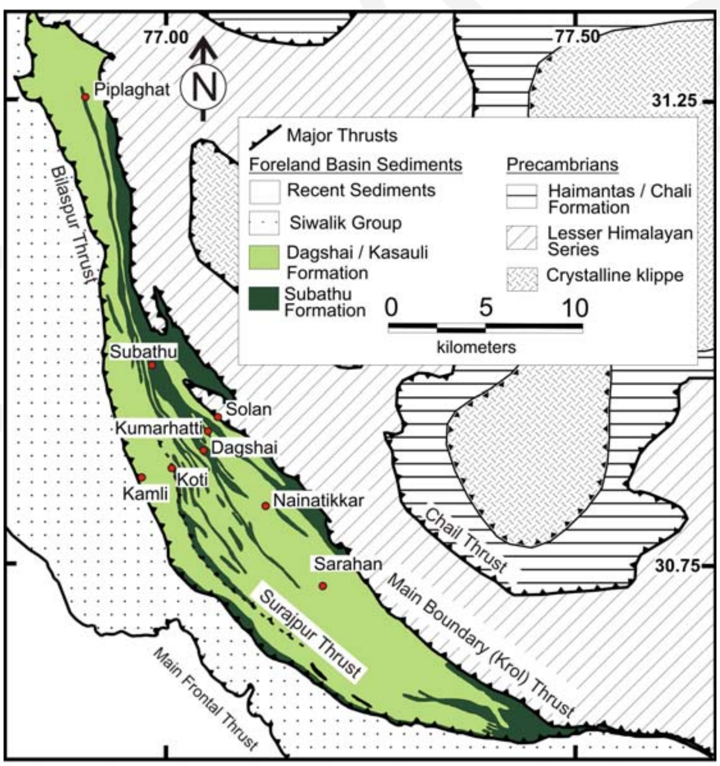

Figure 1.8: Geological map of the Subathu sub-Basin showing Palaeogene sediments and major thrusts (after Khan and Prasad, 1998). The important outcrop locations are marked by solid red circles.

1.5 Purpose of the present study

The preceding discussions raised several key issues on the interrelationship between tectonics and profound climate changes including monsoon those occurred during the Cenozoic. The major questions still need to be addressed are, what are the sequence of events either in the Himalayan orogen or its constituent foreland basin during ~40 to ~25 Ma: the question of data gap as discussed in the beginning of this chapter. Why foreland basin subsidence is low up to ~20 Ma and deep water facies is apparently absent in the entire Himalayan foreland? Which geodynamic mechanism is responsible for the flysch to molasse transition in Himalayan foreland? What is the status of Oligocene unconformity? When exactly thrust propagation and accompanied vigorous erosion began? When did monsoon system setup? How far the stable isotope data of foreland sediments are reliable for past monsoon reconstruction? Is erosion genetically linked to monsoon initiation/intensification? Do the past climate data support the channel flow model for substantiating the climate forcing hypothesis? The present work is an attempt to answer some of these questions by using variety of techniques from the Subathu sub-Basin and has been focused to study:

(1) Detailed mapping in parts of the Subathu sub-Basin to understand and delineate the three target formations i.e., Subathu, Dagshai, and Kasauli Formation rocks, regional structure (fold vs. thrust repetitions) of the NW Himalayan foreland and preparation of lithological logs of the exposed sections in different thrust slices and their correlation.

(2) Sampling, methods, and establishment of various isotope analytical protocols.

(3) Process based facies and palaeo-environmental analysis of different lithic variants in the studied Subathu-Dagshai transition to understand palaeo-environmental shifts in spatio-temporal framework in sequence stratigraphic parlance. Further identification of underfilled and overfilled segments of the foreland, status of the White Sandstone unit, and nature and placement of Oligocene unconformity between marine and continental sediments has important bearing on the basin filling history.

(4) Sequence stratigraphic analysis of the Dagshai-Kasauli fluvial molasse sequence for providing a viable sequence stratigraphy based model of basin dynamics (creation/destruction of accommodation space) in response to Himalayan orogeny and climate change.

(5) Oxygen and carbon isotope studies of soil carbonates and oxygen isotope study of tooth enamel phosphates (bio-apatites) from foreland molasse and assessment of burial diagenetic effect.

(6) Tracking down the Palaeogene monsoon record as well as palaeo-altitude of Tibetan Plateau using oxygen isotope in soil carbonates.

(7) Synthesis of sedimentological, sequence stratigraphic, stable isotope data and assessment of interrelationship between two major driving forces viz. tectonics and climate in the context of Himalayan orogeny and Cenozoic climate.