TECTONICS, STRUCTURE, AND METAMORPHIC EVOLUTION OF THE
HIMALAYAN FOLD-THRUST BELT, WESTERN BHUTAN

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ABSTRACT

Field mapping in western Bhutan in combination with U-Pb ages, geochemical data, stratigraphic columns, mineral assemblages and reaction textures, micro- and macro-scale structural observations, and balanced cross sections have allowed us to: 1) evaluate the use of detrital zircon and geochemical signatures for tectonic interpretation, 2) define tectonostratigraphy of litho-units in western Bhutan, particularly the Paro Formation, 3) produce pressure-temperature paths of deformed rocks, and 4) evaluate the magnitudes and rates of shortening through this portion of the Himalayan orogen.

We divide the Lesser Himalayan (LH) section into four map units that range from Paleoproterozoic to Ordovician in age. The Paro Formation is interpreted as the distal equivalent of the Jaishidanda Formation based on a similar structural position immediately below the Main Central thrust (MCT) as well as similarity in detrital zircon signatures. Th-Pb ages of metamorphic monazite from Greater Himalayan (GH) rocks and a single age from the upper LH rocks bracket the minimum age of the MCT displacement between 20.4±1.0 and 15.1±0.4 Ma. Young monazite ages indicate that GH rocks continued to cool even until ~10 Ma. A total displacement of ~230 km achieved over 5 Myr yields a long-term horizontal shortening rate of 4.3±1.2 cm/yr.

In western Bhutan, patterns of metamorphic isograds show an inversion of metamorphic field gradient extending from the upper LH section to the higher structural levels of GH right below the lower-South Tibetan Detachment. In the GH section, deformation postdates peak metamorphic conditions that prevailed at ~20 Ma. In the Paro Formation, the presence of deformed kyanite at the base of the section and
presence of undeformed sillimanite at the upper part of the section suggests burial to the kyanite stability field and syn- to post-deformational growth of sillimanite.

A balanced cross-section across western Bhutan illustrates three endmember solutions for the geometry of deformation and highlights implications for shortening magnitude, kinematics and rates. Retro-deformation of cross sections indicates a minimum of ~466-566 km crustal shortening (72-77%). A comparison of structural geometry, thrust kinematics, shortening, and the rate of shortening highlights significant variation along-strike between eastern Bhutan and Sikkim.
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My long journey to Princeton highlights the plight of a young rural Bhutanese, who despite tremendous hardship defied the odds to attend one of the prestigious schools in the world. Born to illiterate parents who constantly struggled to make ends meet, the hope of obtaining even the basic education was a far-fetch reality and to say the least, a dream of one day obtaining a world-class education was virtually non-existent. I have always tried to follow the footsteps of those who embarked on the trail-less trails and left the trails behind to learn more about the world around me and practice good human values. My hard work and self-determination brought me closer to many inspirational people who have had a great impact on my career advancement. Today, I look back in time to sum up the distance I have traveled so far and realized that I have traveled far beyond my capacity, thanks to many inspirational individuals which I will briefly touch upon the ways they have influenced my life in a chronological order.

In 2001, Kate Miller (then Chair, Department of Geological Sciences, University of Texas at El Paso and now Dean at Texas A&M) and Lincoln Hollister (Professor, Department of Geosciences, Princeton) arrived in Bhutan to help the Geological Survey of Bhutan (GSB) install a temporary seismic network. I was then a fresh graduate from Australia just waiting for my first assignment as a geologist. After having been an office geologist for a few months, it was time for my first assignment i.e. to work as a counterpart with Kate Miller and Lincoln Hollister. It was a great privilege to work and get to know about their background. They talked science (about wiggles and reading rocks) and I felt I was learning a lot from them already. My interaction with them set the stage for my graduate studies at the University of Texas
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Linc had already established a web of connections in Bhutan through friends of Bhutan particularly the Hochs, well before I met him in 2001. He had developed special attachment with Bhutan while leading several geologic expeditions to crack the mountains of Bhutan. He also took serious efforts to transfer geologic knowledge to Bhutanese native and that is how two of us who benefitted from his initiatives are now the leading geologists in Bhutan. So on behalf of the royal government of Bhutan, I would like to take this opportunity to thank him and his group for all their contributions to Bhutan Geology and for their initiative to train Bhutanese geologists. I came to Princeton in fall 2006 with a plan to take up a research project in western Bhutan which required sound knowledge in structural geology and metamorphic petrology. Linc immediately figured out my weakness in metamorphic petrology and was ready to help me the best possible ways both in the lab and field. We spent many hours together going over metamorphic petrology and thin sections to make sure I get a good grip to continue my research. He knew I also needed his help in the field and that is why he made himself available to help me when he accompanied the Princeton alumni group in 2007. I am very thankful to Linc as a mentor and a friend. Linc’s association with Bhutan stems from his interaction with the Hochs who have been special friends of Bhutan. Although I haven’t had the privilege to meet the Hochs (late Frank and Lisina) personally, I have been benefitted indirectly by their support to Linc’s group. Without their help, our geoscientific collaboration would be very unfulfilling. So thank you for being a bridge between scientists and non-scientists.
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Chapter 1: Introduction and Overview

1.1. Acknowledgements

The Kingdom of Bhutan, which is bordered by India to the south and China to the north, has remained in isolation for many years. This isolation has adversely affected geoscientific studies in this part of the Himalaya, and has allowed a tremendous gap to develop in our knowledge of the evolution of the eastern part of the Himalayan fold-thrust belt. As the country slowly opened its borders, access for geologic research was at first limited to geologists from the Geological Survey of India between 1960 and 2002. These early geologists were later followed by western scientists including Swiss geologist Augusto Gansser in the 1960’s, Princeton University professor Lincoln Hollister in the 1980’s, and Dalhousie University professor Djordje Grujic in the 1990’s. The Geological Survey of Bhutan (GSB) came into existence in 1982. However, since only a handful of Bhutanese geologists currently working in the field, many modern concepts in geology still remain foreign. Institutional development has been thwarted by a lack of qualified personnel and laboratory facilities, which restricts the function of GSB to geologic mapping and mineral exploration. Because of these factors, original, geoscientific research by Bhutanese natives has remained out of reach. Geoscientific collaboration with western institutions is thus critical in helping the sustainability of the GSB and the training of its geoscientists. My association with both the GSB and the Department of Geosciences at Princeton has proven to be a positive step for research collaboration, and has ensured unprecedented support from the GSB and facilitated access to every nook and corner of the Bhutan Himalaya. Above all, the effort,
cooperation, and dedication of members of the ‘Team Princeton’ (Nadine McQuarrie, Sean Long, Catherine Rose, and Tobgay) field team were key to fulfilling geoscientific goals of our research.

The research submitted in my dissertation makes a significant contribution to: (1) improving our understanding of the evolution of the Himalaya orogen by filling in missing information in the eastern portion of the mountain belt; (2) transferring the technical knowledge and expertise to Bhutanese geologists and ensuring the sustainability of the GSB; and (3) fully understanding the structural setting and its linkage to earthquake hazards. The current effort of devising safe building codes in Bhutan is being hindered by the lack of comprehensive knowledge on active fault systems. This research will make a tremendous contribution to the above on-going effort in Bhutan. The implementation of safe building codes is heightened by two recent events (6.3 and 6.9 magnitude) that caused significant damages to lives and property. These events are a wake-up call for Bhutan to prepare for big earthquakes that could strike Bhutan anytime in the near future.

1.2. Introduction

The active Himalayan fold-thrust belt is the result of collision and continued convergence between India and Eurasia that began ca. 50 Ma (Heim and Gansser, 1939; Gansser, 1964; LeFort, 1975; Hodges, 2000). Representing the type example of a collision between continents, the Himalaya is the birthplace of tectonic models that are often exported and used to explain modern and ancient mountain building processes throughout the world. However, because of the challenge in accessing many portions of this mountain range, the Himalayas still remain one of the least understood mountain belts in the world.
Rocks in the Himalaya originated as sediments that were once deposited at or near the northern margin of India between Paleoproterozoic and Paleocene time. During the Himalayan event, they were metamorphosed and progressively deformed in a series of south-vergent thrust sheets, accommodating a large amount of crustal shortening (Heim and Gansser, 1939; Gansser, 1964; Powell and Conaghan, 1973; LeFort, 1975; Mattauer, 1986; Hodges, 2000; DeCelles et al., 2002; Murphy and Yin, 2003; Yin, 2006). These rocks are repositories of the pressure (P) and temperature (T) conditions under which they were formed, as well as the P-T paths they followed as they were transported to the surface via faults and ductile shear zones. Until now, many of the first-order observations on stratigraphy, structure, timing, deformation, and shortening magnitude have been focused in northwest India and central Nepal, which are relatively easy to access (e.g. Hodges et al., 1996; Searle et al., 1997; DeCelles et al., 1998; 2000; 2001; 2002; Vannay and Hodges, 2003; Richards et al., 2005; Robinson et al., 2006). Only recently have studies expanded to the eastern Himalaya in Sikkim (Bhattacharyya and Mitra, 2009, 2011), Bhutan (Grujic et al., 2002; McQuarrie et al., 2008; Long and McQuarrie, 2010; Long et al., 2011a), and Arunachal Pradesh (e.g. Yin et al., 2010). Therefore, additional study on the stratigraphy, structure, metamorphism, and deformation in Sikkim, Bhutan, and Arunachal Pradesh is necessary to better understand the evolution of the Himalaya. The research presented in this dissertation fills a significant data gap in the eastern Himalaya, by presenting new data from geologic mapping, U-Pb detrital zircon geochronology, geochemistry, lithostratigraphy, petrologic study, Th-Pb geochronology of metamorphic monazite, and structural datasets (at micro-, macro, and regional scales) from western Bhutan. The major contributions of this research include: 1) re-evaluating the stratigraphy of the fold-thrust belt in western Bhutan.
This is key for answering the fundamental geoscientific questions on kinematics and tectonic evolution; 2) constraining the age and rate of displacement on the Main Central thrust (MCT); 3) establishing pressure-temperature-time paths for deformation of LH and GH rocks; 4) constraining deformation temperatures experienced by thrust sheets; 5) estimating the amount and rate of crustal shortening in western Bhutan. This research will make a significant contribution in improving our understanding of the Himalayan geology.

1.3. Chapter organization and relation of chapters

This dissertation contains three main chapters (2, 3, and 4), each dealing with specific aspects of geology of western Bhutan, but each acting as complimentary to each other.

Chapter 2 (Using isotopic and chronologic data to fingerprint strata: The challenges and benefits of variable sources to tectonic interpretations, the Paro Formation, Bhutan Himalaya) focuses on defining the stratigraphy of the Lesser Himalayan section in western Bhutan, in particular clarify the long-standing question on the tectonostratigraphy of Paro Formation. Through detailed geologic mapping, and collection of U-Pb detrital zircon ages and geochemical signatures, we were able to highlight for the first time that: a) there are pitfalls of the now common practice of using geochronologic and isotopic techniques for differentiating Himalayan lithologic units; b) the Paro Formation is a ~5.5 km thick straight stratigraphic section equivalent to the Jaishidanda Formation, ranging between Neoproterozoic and Ordovician in age; c) the Paro and Jashidanda Formations contain a detrital record of a Cambrian-Ordovician orogenic event that affected northern India, which had not
been previously recognized in LH rocks; and d) Himalayan tectonostratigraphy has substantial along-strike variability.

Chapter 3 (The age and rate of displacement on the Main Central thrust in the western Bhutan Himalaya) focuses on constraining the timing, amount, and rate of displacement on what is arguably the most important structure in the Bhutan Himalaya. Previous studies in other parts of the Himalaya show a wide range of motion ages, which vary between ~25 Ma to the present, that are interpreted as the age of activity for the MCT. In addition, studies that employ the same methods to the same structural levels in different areas imply as much as 5 Myr differences in the timing of metamorphism and motion on the MCT. The work presented here allows us to evaluate if the MCT was active at different times along the orogen and uniquely determine the rate of displacement along it. By presenting geologic mapping, metamorphic data, chemically-characterized Th-Pb ages (prograde ages from high-Th cores and retrograde ages from high-Y rims) of metamorphic monazite from GH rocks, along with balanced cross-sections, we conclude that: a) the age of displacement on the MCT in western Bhutan is between 20 and 15 Ma which is same as the age of shearing across an outer segment of the South Tibetan detachment; b) the rate of displacement on the MCT in western Bhutan is between 3.0 and 5.5 cm/yr, which approaches plate tectonic rates over this window of time; c) the age of displacement on the MCT varies across the Bhutan Himalaya.

Chapter 4 (Metamorphism, thrust kinematics, and shortening in the Himalayan fold-thrust belt, western Bhutan) focuses on; a) characterizing metamorphic mineral assemblages across western Bhutan to estimate pressure and temperature conditions; b) studying metamorphic reaction textures and microfabrics to relate metamorphism and deformation; c) constraining the deformation temperature
ranges of LH and GH thrust sheets in western Bhutan; d) evaluating the implications
of using different geometries to fill the space between surface geology and basement
on shortening magnitude and kinematics; e) illustrating the structural geometry and
estimating the rate of crustal shortening in western Bhutan. By combining geologic
mapping, metamorphic mineral assemblages, mineral reaction textures, quartz
deformation microfabrics, micro- and macro-scale structures, and balanced and
restored geologic cross-sections, we conclude that: a) there is an inverted
metamorphic field gradient with metamorphic grade increasing from the lower
structural levels in the LH section to higher metamorphic grade in the structurally-
higher GH section; b) the initial peak metamorphic reaction in the higher structural
levels of the GH section has been overprinted by reactions (garnet + muscovite =
sillimanite + biotite) that indicate decompression and cooling; c) the Jaishidanda
Formation, Paro Formation, and GH section have been deformed at high temperatures
(500-700°C); d) retro-deformation of a balanced cross-section indicates a minimum
of 466-566 km crustal shortening (72-77%) in western Bhutan, and that variations in
rates in crustal shortening exist between eastern and western Bhutan.

1.4. Publication of chapters and coauthor contributions

Each of the three main chapters in this dissertation represents a stand-alone
article that is either published, in revision in a peer-reviewed journal, or to be
submitted at a later date after my defense. I am the lead author of all three chapters, I
collected and analyzed the data, and I wrote the text and drafted the figures. My
advisor Nadine McQuarrie is listed as second co-author for all the chapters because
she thoroughly edited the text in all manuscripts and provided valuable feedback.
Field data presented in geologic maps in all three chapters was collected by me, Nadine McQuarrie, and former fellow Princeton graduate student Sean Long. The individual contributions of additional coauthors are listed below for each chapter.

Chapter 2 was published in *Tectonics* in late 2010, and is co-authored by Nadine McQuarrie, former fellow Princeton graduate student Sean Long, and collaborators Mihai Ducea and George Gehrels (University of Arizona). Sean Long assisted in the collection of U-Pb data. Mihai Ducea conducted the geochemical analyses necessary for the whole rock Nd and Sm measurements, while George Gehrels facilitated our collection of U-Pb data and helped with the interpretation.

Citation:

Chapter 3 is currently in press in *Earth and Planetary Science Letters* and is scheduled for publication in the beginning of 2012. It is co-authored by Nadine McQuarrie, former fellow Princeton graduate student Sean Long, and Matthew Kohn and Stacey Corrie (Boise State University). Sean Long assisted in the collection of Th-Pb data. Matthew Kohn and Stacey Corrie helped with the chemical systematics of metamorphic monazite and interpretation of ages.
Citation:


Chapter 4 is currently in preparation for submission to Tectonics and is co-authored by Nadine McQuarrie, and former fellow Princeton graduate student Sean Long. Although not a co-author, Lincoln Hollister is gratefully acknowledged for his input on thin-section analysis and his guidance on relating metamorphism to deformation and other processes.

Citation:


1.5. Related publications

Other publications that I am listed as co-author on that closely relate to the research presented in this dissertation are listed below. These include five journal articles with primary authors Sean Long and Nadine McQuarrie, and three abstracts accompanying conference presentations that I have made.
1.5.1. Journal article citations


1.5.2. Abstract citations


1.6. References Cited


Grujic, D., Casey, M., Davidson, C., Hollister, L.S., Kundig, R., Pavlis, T., and Schmid, S., 1996, Ductile extrusion of the Higher Himalayan crystallines in


Chapter 2: Using isotopic and chronologic data to fingerprint strata:
The challenges and benefits of variable sources to tectonic interpretations, the Paro Formation, Bhutan Himalaya

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2.1. Abstract

We combine detrital zircons (DZ) and εNd signatures with field mapping in the Paro Formation in western Bhutan. DZ age spectra are strongly variable and display signatures that have been used to uniquely identify both Greater Himalayan (GH) and Lesser Himalayan (LH) strata. DZ age peaks from six quartzite samples require sources for ca. 0.5, 0.8, 1.2, 1.4, 1.7, 1.8 and 2.5 Ga zircons in the Paro Formation.
The youngest (~0.5 Ga) zircons argue for a Cambrian maximum deposition age. Two samples have a youngest 1.8 Ga peak typically attributed to Paleoproterozoic LH rocks. A ~450 Ma crystallization age from two granite samples constrains the minimum deposition age as Ordovician. New $\varepsilon_{\text{Nd}}$ signatures from 6 detrital samples from the Paro Formation show significant variation with lithology. Schists have $\varepsilon_{\text{Nd}}(0)$ values between -12.0 and -16.9 while quartzite values vary between -18.8 and -24.5. These data imply that the Paro Formation was derived from both young and old sources, with DZ and $\varepsilon_{\text{Nd}}$ values obtained from the same quartzite samples requiring old detritus while the $\varepsilon_{\text{Nd}}$ values obtained from interbedded schist require younger detritus. Using published isotopic and chronologic definitions of Himalayan strata, schist-rich layers would be considered GH while the interbedded quartzite would be LH. Thus, the Paro Formation refutes the generally accepted notion that different Himalayan tectonostratigraphic zones have unique DZ and $\varepsilon_{\text{Nd}}$ signatures. Our data recommend caution in the use of DZ and $\varepsilon_{\text{Nd}}$ signatures for tectonic interpretation especially when making correlations with studies that extend 1000’s of km along-strike.

2.2. Introduction

The U-Pb ages of detrital zircons (DZ) from a sedimentary rock combined with whole-rock epsilon Neodymium ($\varepsilon_{\text{Nd}}$) values provide a proxy for the age of formation of the original source materials that make up the sedimentary rock [McCulloch and Wasserburg, 1978; McLennan et al., 1989; Gehrels et al., 1999; 2000; 2006; DeGraaff-Surpless et al., 2003; Mapes, 2009]. Studies that use geochemical tools for reconstructing the provenance of sedimentary rocks are based
on the premise that the signature of measured DZ ages and whole-rock $\varepsilon_{Nd}$ values provide records of identifiable provenance [Gleason et al., 1994; Sircombe, 1999; Dickinson and Gehrels, 2003; Sircombe and Hazelton, 2004; Barbeau et al., 2005]. In tectonics research, particularly in the Himalaya, geochemical signatures from DZ and/or $\varepsilon_{Nd}$ of three lithotectonic units (Lesser, Greater, and Tethyan Himalaya) have been used to define stratigraphic horizons, help identify important geologic structures and determine unroofing histories [e.g. Parrish and Hodges, 1996; Robinson et al., 2001; DeCelles et al., 2004; Martin et al., 2005; Richards et al., 2005; 2006; Imayama and Arita, 2008; Myrow et al., 2009; 2010]. Because these lithotectonic units record the early geologic history of the Himalayas and allow us to reconstruct their original depositional relationships, understanding the provenance and original geometry of the Lesser, Greater, and Tethyan Himalayan basins is critical for identifying and estimating the displacement magnitude along the major thrust sheets in the Himalayan orogen.

Even though early DZ and isotopic analyses from the central Himalaya have been successfully applied to distinguish different Himalayan lithotectonic units, new data from the western Bhutan Himalaya indicate that the use of DZ and $\varepsilon_{Nd}$ for fingerprinting formations is not straightforward. There are several potential hazards to this approach. First, depositional systems are sensitive to climatic and geologic factors, and sediment composition is subject to alteration by surface processes that occur during erosion, transportation, and after deposition [Savage and Potter, 1991; Nesbit and Young, 1996; Potter et al., 2001]. Second, DZs from a single sedimentary succession can have multiple sources with temporally varying influences, producing heterogeneity of age distributions, which can obscure provenance characterization [e.g. Gehrels 2000; DeGraaff-Surpless et al., 2003; Mapes, 2009]. DZ signatures may
reflect combined sources, complicating provenance determinations and the whole rock $\varepsilon_{\text{Nd}}$ values would, at best, reflect an average of contributing source components [Gleason et al., 1994], but may be dominated by a source that is preferentially confined to shale horizons [McLennan et al., 1989]. Third, sedimentary sorting, weathering of labile source rocks and diagenesis can affect Nd isotope ratios of coarse and fine grained components differently [Frost and Winston, 1987; McLennan et al., 1989, 1990; Awwiller and Mack, 1989, Ohr et al., 1991]. Thus, the combination of these factors suggests the need to understand the limitations of each technique prior to interpretation.

Using U-Pb and isotopic techniques, Parrish and Hodges [1996] recognized important distinctions between Lesser Himalayan (LH) and Greater Himalayan (GH) rocks in the Annapurna region of the Nepal Himalaya. Based on the presence of 0.8-1.0 Ga (Neoproterozoic) zircons and less negative $\varepsilon_{\text{Nd}}(21)$ values in GH rocks, and the presence of >1.8 Ga (Paleoproterozoic) zircons and more negative $\varepsilon_{\text{Nd}}(21)$ values in LH rocks, the authors were the first to suggest that GH and LH rocks were derived from different source areas. Since then, many studies have examined the DZ ages of GH and LH rocks and used differences in age spectra to define unique characteristics of different Himalayan strata [Parrish and Hodges, 1996; DeCelles et al., 2000; 2004; Gehrels et al., 2003; Martin et al., 2005; Yin et al., 2006; McQuarrie et al., 2008; Myrow et al., 2009]. To a first order, $\varepsilon_{\text{Nd}}(0)$ values of LH and GH rocks support this distinction [e.g. Parrish and Hodges, 1996; Whittington et al., 1999; Ahmad et al., 2000; DeCelles et al., 2004; Martin et al., 2005; Richards et al., 2005; 2006].

Conclusions that have been drawn from U-Pb age spectra and $\varepsilon_{\text{Nd}}$ values of rocks in northwest India and Nepal include: 1) GH are distinct from LH rocks based on DZ spectra and $\varepsilon_{\text{Nd}}$ values [Parrish and Hodges, 1996; Ahmad et al., 2000;
DeCelles et al., 2000; 2004; Robinson et al., 2001; Martin et al., 2005]. 2) GH rocks did not receive detritus from India and thus were separated by some unknown distance from the Indian margin [DeCelles et al., 2000; Gehrels et al., 2003; Martin et al., 2005]. 3) DZ spectra are relatively homogeneous within a given stratigraphic unit and the youngest zircons are generally (but not always) representative of the age of the strata. [Myrow et al., 2003; 2009; DeCelles, 2004; Martin et al., 2005; Yin, 2006; Yin et al., 2010].

In this paper, we present new U-Pb geochronologic and Nd isotopic data from the western Bhutan Himalaya, particularly from the Paro Formation. These data show strong variability in DZ spectra as well as $\varepsilon_{\text{Nd}}(0)$ values in schist and phyllite that are different (more positive) by 6-12 $\varepsilon_{\text{Nd}}$ units than those of adjacent quartzite. The observed variations complicate the premise that specific detrital zircon signatures or $\varepsilon_{\text{Nd}}$ values can uniquely define tectonostratigraphic packages in the Himalaya. This complication allows us to evaluate the benefits and limitations of such techniques for distinguishing strata, distinguishing variations in basin geometry from variations in provenance, and defining large-scale orogenic structures that have telescoped a margin with a long and varied deformation history.

2.3. Himalayan Geologic Background

The Himalayan orogenic belt has been divided into four tectonostratigraphic zones, which are each bound by major structures that are continuous along-strike across the majority of the orogen [Heim and Gansser, 1939; Gansser, 1964, 1983; Le Fort, 1975]. The tectonostratigraphic units and bounding structures, from south to north, are the Main Frontal Thrust (MFT), Subhimalayan thrust system, Main
Boundary Thrust (MBT), Lesser Himalayan (LH) zone, Main Central Thrust (MCT), Greater Himalayan (GH) zone, South Tibetan Detachment (STD), and Tethyan Himalayan (TH) zone (Figure 2.1). The Himalayan tectonostratigraphic zones were originally defined with respect to significant changes in metamorphic grade [Heim and Gansser, 1939; Gansser, 1964]. Abrupt juxtaposition of higher grade rocks over lower grade rocks was used to define orogen-scale structures such as the MCT and MBT [Heim and Gansser, 1939; Gansser, 1964; Le Fort, 1975]. Thus to a first-order, Himalayan tectonostratigraphy can be uniquely identified based solely on lithologic and/or petrologic characteristics. However, since LH, GH, and TH strata all contain similar protoliths, variations that include highly metamorphosed LH strata or less metamorphosed GH strata can lead to difficulties in uniquely identifying and correlating rocks [Whittington et al., 1999; Hodges, 2000; Argles et al., 2003; Martin et al., 2005].

Protoliths of metasedimentary rocks of the GH are largely Neoproterozoic in age [Parrish and Hodges, 1996; DeCelles et al., 2000; Gehrels et al., 2003; Martin et al., 2005; Yin, 2006], and are separated from underlying LH rocks by the MCT. LH rocks represent a thick succession of clastic and carbonate rocks that blanketed the northern portion of the Indian craton during Proterozoic and Paleozoic time [Upreti, 1999; Myrow et al., 2003; Yin, 2006; Robinson et al., 2006; McQuarrie et al., 2008]. In the Nepal Himalaya, LH rocks are proposed to be entirely of Paleoproterozoic and Mesoproterozoic age [DeCelles et al., 2000; Martin et al., 2005], although in northwestern India and Bhutan, LH rocks extend into the Cambrian [Myrow et al., 2003; Richards et al., 2005; McQuarrie et al., 2008]. The TH zone, which lies structurally above the STD, represents a composite section including two superimposed rift to passive margin sequences. TH strata extend from Neoproterozoic
to Cretaceous in age [Gaetani and Garzanti, 1991; Brookfield, 1993; Garzanti, 1999; Yin et al., 2006]. Recent work has shown temporal overlap between the TH and LH zones in the Cambrian, Ordovician, and in the Permian through Paleocene [Myrow et al., 2003; 2009; Yin, 2006; McQuarrie et al., 2008; Long et al., 2011].

DZ geochronology and isotopic studies have been widely used in many parts of the Himalaya to characterize and distinguish litho-packages of LH, GH, and TH from one another [Parrish and Hodges, 1996; Gehrels et al., 1999; Whittington et al., 1999; Ahmad et al., 2000; DeCelles et al., 2000]. LH rocks had a sedimentary provenance that included a source of much older (>1.8 Ga) DZs [DeCelles et al., 2000; Parrish and Hodges, 2000; Robinson et al., 2001; Martin et al., 2005; Richards et al., 2005]. In contrast, GH rocks had a source containing generally younger DZs that range from 1.6-0.6 Ga [Martin et al., 2005]. Similarly, the Tethyan sedimentary rocks have detrital zircons showing broad peaks at 1050 Ma and ca. 530 Ma [Myrow et al., 2003; 2009; DeCelles et al., 2004; Martin et al., 2005] and may contain detrital zircons as young as 460 Ma [Gehrels et al., 2003; Long and McQuarrie, 2010]. Nd isotopic signatures of Himalayan rocks follow a similar pattern where GH and TH rocks have an average $\varepsilon_{\text{Nd}}(0)$ value of -15 (less negative) while their Lesser Himalayan counterparts have an average value of -23 (more negative) [DeCelles et al., 2000; Robinson et al., 2001; Martin et al., 2005; Richards et al., 2005; Imayama and Arita, 2008].
Figure 2.1. Simplified geologic map of Bhutan and surrounding region modified from Long and McQuarrie [2010]. Inset maps show the location and the international boundary of the Kingdom of Bhutan. Our study area (Figure 2.2) is shown by a rectangular box. Abbreviations; LH: Lesser Himalaya, GH: Greater Himalaya, TH: Tethyan Himalaya, STD: South Tibetan Detachment, KT: Kakhtang Thrust, MCT: Main Central Thrust, MBT: Main Boundary Thrust, MFT: Main Frontal Thrust, LS: Lingshi Klippe, PW: Paro Window, TCK: Tang Chu Klippe, UK: Ura Klippe, SK: Sakteng Klippe, LLW: Lum La Window from Yin et al. [2010]. Map of Sikkim region is modified from McQuarrie et al. [2008]. Map projection: geographic lat/long (WGS84).
2.4. Bhutan Tectonostratigraphy

In Bhutan, the four Himalayan tectonostratigraphic zones are present (Figure 2.1). We describe these zones from south to north, along with their lithologic characteristics, DZ ages, and $\varepsilon_{Nd}$ values in the following sections.

2.4.1. Subhimalayan Zone (Siwalik Group)

The Siwalik Group represents synorogenic sediments deposited in the foreland basin of the Himalayan fold and thrust belt in mid-Miocene to Pliocene time [Quade et al., 1995; DeCelles et al., 1998, 2001]. In Bhutan, the Siwalik Group is exposed in discontinuous patches [Gansser, 1983; Lakshminarayana and Singh, 1995], with the missing sections either covered by Quaternary sediment, overridden by the MBT, or never deposited. In eastern Bhutan the Siwalik Group is up to ~5.5 km thick [McQuarrie et al., 2008; Long et al., 2011] (Figure 2.1). However, in southwestern Bhutan, near the town of Phuentsholing, the Siwaliks section is either absent or covered by Quaternary sediment (Figure 2.1).

2.4.2. Lesser Himalayan Zone

LH strata above the MBT and below the MCT are composed of low-grade metasedimentary rocks, including quartzite, phyllite, and limestone [Gansser, 1983; Bhargava, 1995; McQuarrie et al., 2008]. In western Bhutan, the LH section can be divided into three units. From old to young these are the Daling-Shumar Group, Baxa Group and the Jaishidanda Formation.
2.4.2.1. Daling-Shumar Group

In Bhutan, the Daling-Shumar Group makes up the Paleoproterozoic LH section. In eastern Bhutan, the Daling-Shumar Group is separated into two distinct formations, the Daling Formation and the underlying Shumar Formation [McQuarrie et al., 2008; Long et al., 2011]. This 2-part stratigraphy is also observed in western Bhutan.

The Daling Formation consists of schist and phyllite with quartzite interbeds, and its lower contact is defined by the upsection transition from quartzite to phyllite and schist [McQuarrie et al., 2008; Long et al., 2011]. Also characteristic of the Daling Formation is the presence of bodies of mylonitized orthogneiss containing distinctive feldspar augen at different stratigraphic levels (Figure 2.2). The Shumar Formation consists of fine-grained quartzite, with medium to thick planar bedding [Dasgupta, 1995; McQuarrie et al., 2008; Long et al., 2011]. Quartzite layers are often separated by cm- to m-scale interbeds of phyllite and schist. In western Bhutan (Figure 2.2), the Shumar Formation is ca. 2.6 km thick, and the combined thickness of Shumar and Daling Formations is ca. 4.0 km.

U-Pb ages of DZs from both the Daling and Shumar Formations show strong peaks at ~1.9 Ga, however, the weighted mean age of the youngest 3 concordant zircons yield maximum deposition ages of $1865\pm47$ Ma, and $1816\pm49$ Ma [Long et al., 2011]. Igneous rocks within the Daling Formation have crystallization ages of 1.78-1.9 Ga, 1.79-1.89 Ga and 1.76-1.84 Ga. [Daniel et al., 2003; Richards et al., 2006; Long et al., 2011]. These data together suggest a deposition age between 1.8-1.9 Ga for the Daling-Shumar Group. Five samples from Daling-Shumar Group analyzed by Richards et al. [2006], have $\varepsilon_{\text{Nd}}(0)$ values of -27.0, -26.6, -27.3, -25.9,
and -32.3. The Daling and Shumar Formations have very negative $\varepsilon_{Nd}$ values typical of old source terrane, which is consistent with the older DZ populations.

2.4.2.2. Baxa Group

The Baxa Group is interpreted to stratigraphically overlie the Daling Formation in Bhutan [Tangri, 1995], and this contact is observed along-strike in Sikkim [Bhattacharyya and Mitra, 2009]. In eastern Bhutan, the Baxa Group is primarily fine to medium grained, locally pebbly to conglomeratic quartzite, with jasper and rose quartz clasts [McQuarrie et al., 2008; Long et al., 2011]. However, the Baxa Group displays significant lateral variations in lithology. In western Bhutan, near Phuentsholing, the Baxa group is divided into lower (Phuentsholing) and upper (Pangsari) formations [e.g., Tangri, 1995]. The Phuentsholing Formation consists of dark gray to black slate and phyllite with interbeds of limestone, creamy dolomite, and thin beds of fine-medium grained quartzite, while the Pangsari Formation consists of gray to green, locally talcose phyllite interbedded with red to pink marble and thin beds of fine to medium grained greenish quartzite.

In eastern Bhutan, the Baxa Group yields DZ age spectra showing Cambrian (ca. 520, 525 Ma) youngest peaks, with older peaks between 1.0 Ga and 1.7 Ga [McQuarrie et al., 2008; Long et al., 2011]. Phyllite samples collected within the Baxa Group in eastern Bhutan yielded $\varepsilon_{Nd}(0)$ value of -21 [McQuarrie et al., 2008].
Figure 2.2. Geologic map of part of the western Bhutan Himalaya showing detailed lithologic units in Paro Window. Strike and dip symbols indicate our mapping. Locations of DZ and $e_{Nd}$ samples are shown. Samples of Richards et al. [2006] are shown by * (Bh10, Bh12, B85, and B88). Dashed line in Baxa Group represents boundary between Phuenstholing Formation and upper Pangsari Formation. Map projection: geographic lat/long (WGS84).
2.4.2.3. Jaishidanda Formation

Beneath the MCT, a ~1.0 km-thick interval of biotite-rich, locally garnet-bearing schist with interbeds of biotite-rich quartzite is present. These strata have been interpreted as part of the structurally overlying GH zone [Janpangi, 1974; Segupta and Raina, 1978; Trichal and Jarayam, 1989], as the uppermost part of the Daling-Shumar Group [Guha Sarkar, 1979; Gansser, 1983; Ray, 1989], and as a unique lithotectonic unit called the Jaishidanda Formation (JF) mapped in thrust contact over the Daling-Shumar Group [Dasgupta, 1995]. However, based on distinct lithologies and a lack of fault rocks at the base of the unit, McQuarrie et al. [2008] and Long et al., [2011] mapped the JF in depositional contact with the underlying Daling Formation. In western Bhutan, the Jaishidanda Formation contains biotite-rich, garnet-bearing schist with quartz vein boudins interbedded with light-gray quartzite, with abundant crenulated, biotite-rich laminations and thin interbeds of biotite schist. The contact between the Jaishidanda Formation and the Daling Formation is distinguished by a downsection transition from gray, lithic clast-rich, and biotite-rich quartzite of the Jaishidanda Formation to clean, tan, thinly-bedded quartzite containing rare biotite of the Daling Formation.

DZs from the JF display peaks between ca. 475 Ma and 530 Ma, 0.9 Ga and 1.7 Ga, and a prominent peak at 2.5 Ga [McQuarrie et al., 2008 and Long et al., 2011]. This DZ age range, minus the young Ordovician to Cambrian peaks, is similar to the DZ data from strata immediately below the MCT in Arunachel Pradesh [Yin et al., 2006; 2010]. The presence of DZ peaks as young as ~475 Ma indicates an Ordovician maximum deposition age.
2.4.3. Paro Formation

The rocks of the Paro, Haa, Bunakha, and Thimphu regions (Figure 2.2) are referred to as the Paro Formation [Gokul et al., 1976; Jangpangi, 1978, 1980; Gansser, 1983; Dasgupta, 1995; Ikemoto, 2000; Koike, 2002]. The Paro Formation consists of high-grade metasedimentary and calcareous rocks, including calc-silicate rocks, marble, quartzite, quartz-garnet-staurolite-kyanite schist with subordinate feldspathic schist and bodies of two mica granite-composition orthogneiss [Gansser, 1983; Dasgupta, 1995; Ikemoto, 2000]. Although the Paro Formation has been mapped in thrust contact with the overlying GH paragneiss [Gansser, 1983; Dasgupta, 1995], the exact tectonic affinity of the unit is still ambiguous. Jangpangi [1978, 1980] correlated the upper quartzite-rich section and the lower carbonaceous section with the Shumar and Baxa Formations of the LH sequence, respectively, based purely on lithology and lower metamorphic grade relative to the overlying GH rocks. Based on the presence of garnet- and staurolite-bearing muscovite-biotite schist, calc-silicate bands, and actinolite amphibolite lenses, Gansser [1983] designated upper greenschist to amphibolite facies for Paro Formation, and mapped the unit as a metasedimentary unit within the GH sequence. Later, Dasgupta [1995] redesignated part of the Paro Formation as Jaishidanda Formation. Richards et al. [2006] analyzed three schist samples from the Paro Formation and reported less negative $\varepsilon_{\text{Nd}}(0)$ values (of -14.2, -16.4, and -16.9). Based on these values, Richards et al. [2006] mapped Paro rocks as part of the GH sequence. Our observations describing the complete stratigraphic section of the Paro Formation, along with detailed lithologic descriptions from our field mapping, are discussed in section 5.
2.4.4. Greater Himalaya

In Bhutan, the GH section consists of orthogneiss and metasedimentary rocks that are separated from underlying LH rocks by the MCT [Gansser, 1983; Golani, 1995; Grujic et al., 2002; Long and McQuarrie, 2010] (Figure 2.1). The GH rocks are intruded by both Miocene leucogranites related to the Himalayan orogenic event and Cambrian-Ordovician granites related to an older magmatic event [Daniel et al., 2003; Gehrels et al., 2003; Richards et al., 2006; Cawood et al., 2007]. The GH section is divided into a lower structural level above the MCT and below the Kakhtang thrust (KT), and a higher structural level above the KT [Grujic et al., 2002] (Figure 2.1). The lower structural level GH rocks overlie the Paro Formation. Here, GH rocks are paragneiss (muscovite-biotite-garnet-kyanite-fibrolite) interlayered with quartzite, both of which contain partial melt textures (cm-scale, granite-composition leucosomes, which are often deformed and sheared approximately concordant to primary foliation, and are generally distributed throughout the rock) (Figures 2.1 and 2.2). In the field, GH rocks can be differentiated from the Paro Formation based on the presence of partial melt textures and paragneiss indicating a distinct difference in metamorphic grade. This contrast in lithology, combined with pervasively sheared rocks present at the contact suggests that the GH is in thrust-shear zone contact above the Paro Formation.

DZ age spectra from three quartzite samples in the metasedimentary unit of the structurally lower GH section yielded a Neoproterozoic (ca. 900 Ma) youngest peak for one sample and Cambrian and Ordovician (500 and 460 Ma) youngest peaks for two samples [Long and McQuarrie, 2010]. An orthogneiss at the base of the GH section in eastern Bhutan yielded a 487±7 U-Pb (zircon) crystallization age [Long and
These data suggest an age range of Neoproterozoic to Ordovician for the GH metasedimentary unit in Bhutan [Richards et al., 2006; Long and McQuarrie, 2010]. The lower and upper structural levels of the GH section have less negative average $\varepsilon_{\text{Nd}}(0)$ values of -14.5 and -15.4, respectively [Richards et al., 2006].

2.4.5. Tethyan Himalaya

Throughout Bhutan, rocks of the TH zone are preserved above the GH section in synforms [Gansser, 1983; Grujic et al., 2002; Kellett et al., 2009]. The basal TH rocks, referred to as Chekha Formation, consist of quartzite, shale, siltstone, sandstone, and conglomeratic quartzite [Tangri and Pande, 1995]. These are overlain by Maneting Formation phyllite and fossiliferous limestone of Late Cambrian age locally known as Wachi La Formation [Tangri and Pande, 1995; Hughes et al., 2010].

Detrital zircon spectra of TH rocks from Bhutan all display peaks between 800 and 900 Ma [McQuarrie et al., 2008; Long and McQuarrie, 2010; Hughes et al., 2010]. In addition, samples from the Wachi La region (Deshichilling Formation and Quartzite formation) in western Bhutan have a significant late Cambrian peak and unique late Cambrian fossils [Hughes et al., 2010; Myrow et al., 2010]. However, detrital zircon ages from the Chekha Formation in Central Bhutan (Zhemgang region) yield peaks as young as ~460 Ma, which indicates an Ordovician or younger depositional age for TH rocks there [Long and McQuarrie, 2010]. No Nd isotope data exist for the TH rocks of Bhutan. However, based on the stratigraphic and
structural position of TH rocks in relation to GH rocks along-strike, similar, less negative $\varepsilon_{Nd}$ values are expected.

2.5. Methods

2.5.1. Field Mapping

Map data included in this paper come from our geologic mapping at 1:50,000-scale along roads and trails in western Bhutan (Figure 2.2). Mapping was focused on the Paro Formation although lithologic and structural data were also collected from LH and GH units between the MFT in the south to the MCT in the north. Our map data lie primarily in six transects (Figure 2.3a): 1) four from Chuzom to Bunakha, Paro, Tsaluna, and Genekha, 2) one along the road between Betekha and Chele La, 3) one along the Phuentsholing-Thimphu highway between Chukha and Taktikoti. Our mapping was integrated with published geologic maps of Bhutan (Gansser, 1983; Bhargava, 1995; Grujic et al., 2002), to help trace contacts along strike. The level of rock exposure in western Bhutan was good because of fresh roadcuts.
Figure 2.3a. Stratigraphic sections of the Paro Formation measured along different traverses (See Figure 2.2 for locations). Note thicknesses of units are in meters and the thickness of marble band I is ~10 m. The section has sample locations (labeled A-F) collected at different structural levels with their $\varepsilon_{Nd}(0)$ values next to it (A-F). The $\varepsilon_{Nd}(0)$ values marked with one asterisk are from Richards et al. [2006]. Two asterisks denote crystallization ages of orthogneiss samples (G = BU07-83 & H = BU08-128).
2.5.2. U-Pb geochronology

DZ age populations reflect the age of the igneous or metamorphic protolith in which the zircon crystallized [McCulloch and Wasserburg, 1978; McLennan et al., 1989]. The advent of high precision analytical systems capable of producing high quality and quantity datasets necessary for provenance studies has revolutionized DZ geochronology and established it as a powerful tool for reconstructing the provenance of sedimentary rocks, particularly when primary depositional information has been erased by subsequent metamorphism [Gehrels et al., 1999; 2000; 2006; DeGraaff-Surpless et al., 2003; Mapes, 2009].

U-Pb geochronologic analyses were conducted on individual zircon grains using laser-ablation multicollector inductively coupled plasma-mass spectrometry (LA-MC-ICP-MS) at the University of Arizona LaserChron Center. See Appendix A for a detailed discussion of the methodology of this laboratory. Approximately 100 grains were dated per sample. We analyzed 6 detrital samples and 2 granite samples collected at different stratigraphic levels from the Paro Formation in order to provide age control on unit deposition, and to compare the isotopic variations. Sample locations and lithologies are listed in Table 2.1, and map locations are shown on Figure 2.2. The 505 U-Pb zircon analyses (after the removal of analyses with less than 70% concordance) are shown for each sample in Figure 2.3b in relative age probability plots (1σ errors) and data and measurement (analytical) errors (1σ) for individual analyses are listed in Table A-1, and shown in Pb/U concordia plots for each sample in Figure A-1.
Table 2.1. Sample Locations for Western Bhutan.

<table>
<thead>
<tr>
<th>Sample</th>
<th>°N (dd.ddddd)</th>
<th>°E (dd.ddddd)</th>
<th>Analysis</th>
<th>Formation</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>NBH-22</td>
<td>27.43297</td>
<td>89.64098</td>
<td>$\varepsilon_{Nd}$</td>
<td>Paro</td>
<td>Schist</td>
</tr>
<tr>
<td>BU07-84</td>
<td>27.42561</td>
<td>89.41311</td>
<td>DZ</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-77</td>
<td>27.45644</td>
<td>89.52264</td>
<td>DZ &amp; $\varepsilon_{Nd}$</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-73</td>
<td>27.02925</td>
<td>89.58050</td>
<td>DZ &amp; $\varepsilon_{Nd}$</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-75</td>
<td>27.15200</td>
<td>89.54175</td>
<td>DZ &amp; $\varepsilon_{Nd}$</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-76</td>
<td>27.22606</td>
<td>89.51678</td>
<td>DZ &amp; $\varepsilon_{Nd}$</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-83</td>
<td>27.26186</td>
<td>89.35925</td>
<td>Zircon &amp; $\varepsilon_{Nd}$</td>
<td>Paro</td>
<td>Orthogneiss</td>
</tr>
<tr>
<td>BU07-79</td>
<td>27.33758</td>
<td>89.57497</td>
<td>DZ</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU08-128</td>
<td>27.33325</td>
<td>89.57353</td>
<td>Zircon</td>
<td>Paro</td>
<td>Orthogneiss</td>
</tr>
</tbody>
</table>
Figure 2.3b. Probability plots of samples A-F (see Figure 2.3a) showing U-Pb ages of detrital zircons.
In general, $^{206}\text{Pb}^* / ^{238}\text{U}$ (asterisk denotes correction for common Pb; see A.1.2 for details on this correction; all ages described in the text have had this correction) ratios were used for ages younger than ~1.0 Ga, and $^{207}\text{Pb}^* / ^{206}\text{Pb}^*$ ratios were used for ages older than ~1.0 Ga, because this is the approximate cross-over in precision for these two ages. Age cutoffs used for each sample are listed in Table A-2. Sources for systematic error, which include contributions from the fractionation correction, composition of common Pb, age of the calibration standard, and U decay constants (see footnotes of Table A-1 for these values), are not added into the errors shown in Table A-1 (which includes only measurement errors at $1\sigma$), and could collectively shift age-probability peaks by up to ~3% ($2\sigma$). Total systematic errors for each sample are listed at $2\sigma$ in Table A-2 along with records of standards run for each individual sample. Note that the uncertainty of the weighted mean is based on the scatter and precision of the set of concordant ages, weighted according to their measurement errors. The systematic errors are then added to this measurement error quadratically to calculate the total error. Systematic errors include contributions from the fractionation correction, composition of common Pb, age of the calibration standard, and U decay constants.

2.5.3. Epsilon Neodymium isotope geochemistry

$\epsilon_{\text{Nd}}$ values give information on the age the source material since its extraction from the mantle [McCulloch and Wasserburg, 1978; McLennan et al., 1989] and thus to a first order provide a proxy for the age of formation of the original source materials that make up the sedimentary rock. The utility of $\epsilon_{\text{Nd}}$ for provenance studies is founded on the coherent behavior of rare earth elements during sedimentary
processes, and that sedimentary processes average the various sources from which the sediment was derived [Boghossian et al., 1996].

Nd isotopic ratios and the elemental concentrations of Sm and Nd were measured by thermal ionization mass spectrometry at the University of Arizona, following procedures reported in Otamendi et al. [2009]. Details on analytical errors and blanks specific to this study are reported in Appendix A; section A.2. Analyses were done on six samples (five quartzite and one igneous) that were also analyzed for U-Pb ages. One additional sample came from a schist interbed close to a quartzite sample. Each sample is from a specific stratigraphic horizon, and reflects the $\epsilon_{\text{Nd}}$ value at that point. The 7 samples analyzed span 3560 m of exposed Paro Formation section (from bottom to the top) (Figure 2.3). Three samples represent lower, middle, and upper sections of fine-grained quartzite unit that measures 1960 meters thick. An igneous sample is from a 225 meters thick orthogneiss present in the mid-section of the fine-grained quartzite unit. One sample is from the upper stratigraphic level of a 1600 m thick coarse-grained quartzite. For each sample, the average of 100 isotopic ratios was taken to calculate the $\epsilon_{\text{Nd}}$ value. Analyzed samples have estimated analytical $\pm 2\sigma$ uncertainties of $^{147}\text{Sm}/^{144}\text{Nd} = 0.4\%$, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.0012\%$, which corresponds to an $\epsilon_{\text{Nd}}$ error of $\pm 0.5$. Table 2.2 contains measured Nd isotopic data (present day values). Although the maximum and minimum depositional age, as well as the ages of potential source regions for LH and GH rocks can be inferred with varying degrees of confidence, they are essentially unknown for the Paro Formation. Thus rather than bias $\epsilon_{\text{Nd}}$ values with an assumed age for the source rock, we report Nd isotopic composition using $\epsilon_{\text{Nd}}(0)$ values. This is a common practice for Himalayan rocks and facilitates comparison with previous work in Bhutan and along the Himalayan arc.
Table 2.2. Epsilon Neodymium Isotopic Analyses.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Sm (ppm)</th>
<th>Nd (ppm)</th>
<th>$^{143}$Nd/$^{144}$Nd (0)</th>
<th>std err %</th>
<th>$\varepsilon_{\text{Nd}}$(0)</th>
<th>Formation</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>NBH-22</td>
<td>6.62</td>
<td>24.86</td>
<td>0.511995</td>
<td>0.0016</td>
<td>-12.5</td>
<td>Paro</td>
<td>Schist</td>
</tr>
<tr>
<td>BU07-73</td>
<td>2.282</td>
<td>8.165</td>
<td>0.511545</td>
<td>0.0010</td>
<td>-21.3</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-75</td>
<td>7.552</td>
<td>28.970</td>
<td>0.511677</td>
<td>0.0009</td>
<td>-18.8</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-76</td>
<td>1.014</td>
<td>5.283</td>
<td>0.511402</td>
<td>0.0009</td>
<td>-24.1</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-77</td>
<td>0.848</td>
<td>3.235</td>
<td>0.511382</td>
<td>0.0008</td>
<td>-24.5</td>
<td>Paro</td>
<td>Quartzite</td>
</tr>
<tr>
<td>BU07-83</td>
<td>4.492</td>
<td>17.027</td>
<td>0.511978</td>
<td>0.0008</td>
<td>-12.9</td>
<td>Paro</td>
<td>Orthogneiss</td>
</tr>
</tbody>
</table>

2.6. Results

2.6.1. Field Mapping

2.6.1.1. Map view description

The Paro Formation crops out in a ~700 km$^2$ window in western Bhutan and contains metasedimentary rocks intruded by thin, 100-200 m sills of orthogneiss (Figure 2.2). Our mapping confirms that the Paro Formation is exposed as a tectonic window and that it is separated from the overlying GH section by a top-to-the-south thrust contact. Bedding in quartzite and marble and foliation in schist horizons dip north, south, east, and west at the northern, southern, eastern, and western edges, respectively, which defines a structural dome with the structurally deepest part exposed at Sisina (Figure 2.2).

To the south of the main window, the Paro Formation is exposed again in a thrust sheet that is ~2.0 km wide (N-S) and ~50 km long (E-W) on the map with a
structural thickness of ~1.5 km, bounded on both sides by north-dipping thrusts. Here the thrust and GH section overlying the Paro Formation has been folded into a synform between the northern and southern exposures of the Paro Formation. The southern boundary of the southern Paro Formation exposure is a north-dipping thrust fault that places the Paro Formation over GH rocks. On both sides of the fault, bedding planes and foliations are steeply north-dipping (50°-56°) and are separated by a 2-3 meter thick zone of gouge and altered rock. There is a sharp contrast in lithology with melt-bearing paragneiss in the footwall and quartzite with schist interbeds in the hanging wall. Asymmetric feldspar σ-clasts and macroscopic fold vergence show a top-to-the-south shear sense. We interpret this contact as an out-of-sequence thrust (Figure 2.2) that placed the Paro Formation over GH rocks.

### 2.6.1.2. Stratigraphic description

Rocks of the Paro Formation can be divided into several thin (2 km to 100’s of meters) but mappable units. We describe these units from the base of the Paro Formation to the top and argue that the relationships between the mapped units are stratigraphic. The base of the Paro Formation is a 600 m-thick schist unit, which contains muscovite-biotite-garnet-staurolite-kyanite mineral assemblages. Kyanite is present within quartz veins, and rarely with biotite, but textures do not indicate the presence of any in-situ partial melt. Also, within this schist unit is a two mica, granitic gneiss (ca. 110 m thick) exposed on the either side of river at Sisina (Figures 2.2 and 2.3a).

The schist unit is overlain by a gray-colored, thin-bedded, fine-grained micaceous quartzite unit that is up to 1960 m thick (Figure 2.3a). The contact between
the two units is gradational with thin quartzite layers within the schist becoming more prevalent and thicker up section (Figure 2.4). Within the quartzite unit there are two marble bands. Marble band I is ~10 m thick and is stratigraphically lower, and marble band II is ~195 m thick and is stratigraphically higher. These two marble bands are recognized in seven different transects that extend across the Paro Formation and can be mapped as continuous bodies in the north (Haa, Chele La, Paro) and south (Chuzom, Betekha). The uniformity between the sections (Figure 2.3a) allows us to correlate the marble horizons around the full extent of the main window and emphasize a continuous stratigraphy through the Paro Window (Figures 2.2 and 2.3a). Along the south-western edge of the window, near Betekha and Shari are additional orthogneiss bodies. Near these intrusions are kyanite and fibrolite schist interbeds within the fine-grained micaceous quartzite unit (Figures 2.2 and 2.3a). Further upsection (Figure 2.2), we map small calc-silicate bodies associated with marble band III at Chele La, Tsaluna, and marble band II south of Thimphu (Figure 2.2). Overlying the fine-grained quartzite unit is a tan-gray colored, coarse-grained, thin-medium bedded, biotite-rich quartzite, which is ~1600 m thick, and contains biotite-muscovite-garnet schist partings. The coarse-grained quartzite unit is beneath GH paragneiss at the eastern, western, and northern edges of the Paro window (Figure 2.2). Within the coarse-grained quartzite of the Paro Formation is marble band III (ca. 255 m thick) and two granite intrusions (ca. 255 m thick), one just north of Paro and the other at Tsaluna (Figures 2.2 and 2.3a). Marble III is not exposed across the entire window and instead pinches out to the east, south of Genekha, and to the southwest of Betekha (Figure 2.2).
Figure 2.4. (a) Simplified geologic map of the Paro Window with hypothetical faults (1, 2, and 3) drawn based on DZ and $\varepsilon_{\text{Nd}}$ data. Solid and dashed rectangles are areas with and without evidence of fault, respectively. (b) Stratigraphic section of Paro Formation with hypothetical faults at various levels. Numbers 1, 2, and 3 correspond to permissible order of post MCT faulting if one assumes GH strata are north of LH strata [e.g., DeCelles et al., 2000]. Note the boundary between fine- and coarse-grained quartzite can be either stratigraphic or the axis of a recumbent syncline that repeats lithological packages on both sides of axis (marble bands and orthogneiss), but the DZ and $\varepsilon_{\text{Nd}}$ values do not match across limbs. Grey boxes to the right show temperature range of Paro Formation obtained from quartz deformation fabrics i.e. subgrain rotation (SGR) - 400-500°C; grain boundary migration (GBM) - 500-700°C; chessboard extinction - 650-700°C [Stipp et al., 2002]. (c) Photo of gradational contact between schist (sch) and quartzite (qtz) near Sisina. (d) Photo of undeformed coarse-grained quartzite. Lithologic colors same as Figure 2.2.
Quartz micro-fabrics (amoeboid grains with large amplitude sutures and very large recrystallized grain sizes) associated with grain boundary migration recrystallization suggest the deformation temperatures of coarse-grained quartzite are ~500-700°C, and the coarse grained texture is a metamorphic fabric [Stipp et al., 2002] (Figure 2.4). These high temperatures in the Paro Formation correlate with higher temperatures (>700°C) in the immediate hanging wall of the overlying thrust. The fine-grained quartzite in the south indicates temperatures of 400-500°C (associated with subgrain rotation recrystallization) and is overlain by cooler GH rocks whose peak temperature reached 600-700°C. In general, foliation is parallel to small (schist partings and interbeds) and large (first order changes from schist to quartzite to marble) changes in lithology. We observe that foliation is parallel to the original sedimentary bedding throughout the Paro Formation, and that the planar (flattening) fabric is due to tectonic loading. Meso-scale folding is present in more incompetent lithologies, such as marble and interbeds of quartzite and schist. These are most prevalent at the base of Paro Formation near Chuzom, around the orthogneiss near Betekha and locally within marble band II (Figure 2.2). Folding is asymmetric and indicates top-to-the-south shear. The lack of foliation parallel to axial planes as well as lack of axial planar cleavage suggest that the development of foliation fabric parallel to bedding predates folding, that bedding orientations are right side up, and that the original, first-order stratigraphic succession is not repeated by isoclinal folding.

In the southern exposure, the Paro Formation consists of metaquartzite and muscovite-biotite-garnet schist (Figure 2.2). This section is predominantly medium-coarse grained, micaceous quartzite (1500 m) overlain by interlayered fine grained quartzite and schist above a thin band of marble (250 m) (Figure 2.2). The northern contact of the southern exposure is fine-grained quartzite overlain by GH paragneiss
(muscovite-biotite-garnet-kyanite-fibrolite) containing partial melt textures. Based on the contact between the fine grained quartzite and the overlying GH as well as the thickness of both the marble band and the underlying quartzite in the Paro Formation exposed in the south, we correlate the marble in the southern window with marble II in the northern window (Figure 2.3a).

The stratigraphy we define above, schist overlain by fine-grained quartzite with marble bands I and II, overlain by coarse-grained quartzite with marble III, is recognized in each mapped transect across the Paro window. Because of this, stratigraphic sections drawn between Chuzom and Bunakha, Chuzom to Paro, Chuzom-Bemay Rong Chu, and Chuzom to Genekha are very similar except small, local variations in thickness (Figure 2.3a). In total, the average thickness of the Paro Formation in the Paro Window is ca. 5.5 km, which is three times the exposed thickness of the fine-grained quartzite and marble band in the linear window further to the south (Figure 2.3a).

2.6.2. U-Pb geochronology

Sample BU07-79 is from quartzite interbedded within the staurolite-kyanite-schist unit at the base of the Paro Formation. This sample has a prominent youngest peak at 0.8 Ga with older, smaller peaks at 1.7 Ga and 2.5 Ga. BU08-128 is a granitic gneiss that is exposed at Sisina within the basal Paro Formation schist unit. Its likely age of crystallization is between 400 and 500 Ma (refer to Data Repository Discussion 2 for further discussion).

Four samples (BU07-73, BU07-75, BU07-76, and BU07-83) collected from the fine-grained quartzite unit were analyzed. Sample BU07-73 is from quartzite that
we interpret as correlative to the fine-grained quartzite unit of the Paro Formation, from the southern exposure south of Chukha (Figure 2.2). It yielded a prominent peak at 1.7 Ga along with a series of smaller peaks at 0.50 Ga, 0.70 Ga, 0.90, 1.25 Ga, and 2.5 Ga (Figure 2.5). Sample BU07-76 which is stratigraphically above sample BU07-73 (based on correlating the marble in the southern exposure with marble II in the north), but in the main window, yielded a prominent peak centered at 1.8 Ga with two smaller and older peaks at 2.2 Ga and 2.5 Ga. Sample BU07-75 which is from the uppermost section of the fine-grained quartzite unit, has a strong peak at 1.7 Ga and a minor peak at 0.8 Ga. Sample BU07-83 is from an orthogneiss collected along the road to Haa near Shari (Figure 2.2). Similar to sample BU08-128, the likely crystallization age of this granite is between 400 to 500 Ma (refer Discussion 2 in the Data Repository section).

Two samples (BU07-77, and BU07-84) collected from the coarse-grained quartzite unit were analyzed. Samples BU07-77 and BU07-84 are from the upper and lower parts of the section respectively. BU07-77 has a strong peak at 1.8 Ga. Sample BU07-84 has a multiple-grain peak at ca. 1.7 Ga and smaller peaks at 1.0 Ga and 1.4 Ga.

In addition, a sample from the Jaishidanda Formation (BU08-135) in western Bhutan was analyzed. DZ ages from this sample display a series of peaks between 475 Ma and 530 Ma, 1.0 Ga and 1.7 Ga, and a prominent peak at 2.5 Ga (Figure 2.6).
Figure 2.5. U-Pb concordia plot for 84 single zircon analyses of sample BU07-73 collected from the southern Paro Formation exposure south of Chukha, correlative to the base of the fine-grained quartzite unit of the Paro Window. This sample has a prominent peak at 1.7 Ga with a series of smaller peaks at 0.50 Ga, 0.70 Ga, 0.90, 1.25 Ga, and 2.5 Ga. The inset shows that two 500 Ma grains (highlighted in red) are not pulled off the concordia line and that the young peak is significant and robust. Error ellipses are shown at the 1σ level (68.3% confidence).
2.6.3. Epsilon Neodymium isotope geochemistry

McQuarrie et al. [2008] reported $\varepsilon_{Nd}$ values from four LH samples collected from eastern Bhutan, two from the Jaishidanda Formation, in the immediate footwall of the MCT, one from Baxa Formation, and one from the Diuri Formation. These samples have $\varepsilon_{Nd}(0)$ values of -19.4, -21.4, -21.4, and -19.8, respectively.

Richards et al., [2006] reported $\varepsilon_{Nd}(500)$ values of 21 samples collected from the Bhutan Himalaya, out of which 6 were from LH rocks, 12 were from GH rocks, and 3 were from the Paro Formation. Here, we recalculate these as $\varepsilon_{Nd}(0)$ values. The six LH samples (one garnet-schist, two quartzite and three phyllite samples) were from the Jaishidanda Formation and the Shumar-Daling group south of the MCT. The 5 Shumar-Daling samples have an average $\varepsilon_{Nd}(0)$ value of -25.9 typical of the lower LH. Sample Bh12, located at the base of the Jaishidanda Formation in western Bhutan, has an $\varepsilon_{Nd}(0)$ value of -16.6. The less negative value led Richards et al., [2006] to interpret this sample as part of the GH section. Three garnetiferous schist samples (B85, B88, and Bh10) sampled from the base of the Paro Formation (Figure 3), have $\varepsilon_{Nd}(0)$ values of -14.2, 16.9, and 16.3.

The predominantly old DZ ages that we obtained for most of our Paro Formation samples contrast strongly with the less negative $\varepsilon_{Nd}$ values obtained by Richards et al., [2006]. To evaluate whether the Paro Formation has a mixture of old and young sources and to examine possible sediment sorting effects, we measured Sm and Nd isotopic ratios on the same samples we analyzed for U-Pb geochronology (i.e. BU07-73, 75, 76, 77, 83, and 84). In addition, we also collected one sample (NBH-22) (schist) at the same stratigraphic level as BU07-77 and BU07-84 (quartzite) to look at the isotopic signatures of sandstone-shale pairs.
Figure 2.6. U-Pb detrital zircon age spectra of sample BU08-135 from western Bhutan and a composite U-Pb detrital zircon age spectra of Jaishidanda Formation (compiled from data presented in McQuarrie et al., [2008] and Long et al., [2011]) from eastern Bhutan displaying a series of peaks between 475 Ma and 530 Ma, 1.0 Ga and 1.7 Ga, and a prominent peak at 2.5 Ga. Refer Figure 2.2 for sample locations. Sample Bh12 of Richards et al. [2006] collected from the same stratigraphic level as BU08-135 yield an $\varepsilon_{Nd}(0)$ value of -16.6.
Sample BU07-76, collected from the lower section of the fine-grained quartzite has an \( \varepsilon_{\text{Nd}}(0) \) value of -24.1 (Figure 2.3a), while the garnetiferous schist at the same stratigraphic level has \( \varepsilon_{\text{Nd}}(0) \) value of -16.3 [Richards et al., 2006]. Sample BU07-73 was also collected from the lower section of the fine-grained quartzite, but from the southern exposure, has an \( \varepsilon_{\text{Nd}}(0) \) of -21.3 (Figure 2.3a). Sample BU07-75 collected from the upper section of the fine-grained quartzite has \( \varepsilon_{\text{Nd}}(0) \) value of -18.8. The orthogneiss (BU07-83) within this quartzite unit has \( \varepsilon_{\text{Nd}}(0) \) value of -12.9.

A schist sample (NBH-22) adjacent to BU07-84 and BU07-77 was collected from the lower section of the coarse-grained unit and has an \( \varepsilon_{\text{Nd}}(0) \) value of -12.5, while BU07-84 has an \( \varepsilon_{\text{Nd}}(0) \) value of -24.0 (Figure 2.3a). Sample BU07-77, collected from the upper part of this coarse-grained quartzite unit has an \( \varepsilon_{\text{Nd}}(0) \) value of -24.5.

2.7. Discussion

With many previous studies claiming that GH and LH rocks in the Himalayas are distinguishable by distinct DZ signatures and isotopic composition, the U-Pb geochronologic and isotopic data from the Paro Formation presented in this study provides us with a unique opportunity to test these hypotheses. This study shows that the Paro Formation can be interpreted as either 1) a single sedimentary succession in the western Bhutan Himalaya that requires both old (typically associated with LH strata) and young (typically associated with GH or TH strata) detritus (Figure 2.7) or 2) a complex package of isoclinally folded and/or thrusted strata where the fingerprints of Himalayan strata (LH, GH, TH) are unique and have been tectonically interleaved (Figure 2.4).
2.7.1. Provenance and age of the Paro Formation

Schist within the Paro Formation shows less negative $\varepsilon_{\text{Nd}}(0)$ values typical of sediments derived from relatively young sources that are consistent with the young (500 Ma – refer Figure DR3 in Data Repository section) DZ peak obtained from the Paro Formation in the southern exposure and the Cambrian-Ordovician granitic gneiss within the Paro Formation (note the similarity between the $\varepsilon_{\text{Nd}}(0)$ value of the granite (-12.9) and the $\varepsilon_{\text{Nd}}(0)$ of the schists (-12.0 to -16.9)). However, several DZ samples spread throughout the Paro Formation display a significant peak at 1.8-1.9 Ga with minor young peaks (Figure 2.8) requiring that Paro Formation largely received sediment from very old sources. Assuming that the tectonostratigraphic zones throughout the Himalaya are distinguishable by the distinct DZ signatures and/or isotopic compositions described above (section 1), then using only the $\varepsilon_{\text{Nd}}$ values obtained from schist, the Paro Formation would be correlated to the GH zone [Richards et al., 2006]. However, if our view of the Paro Formation was only from DZ samples BU07-76, 77, and 84 we would interpret the Paro Formation as Paleoproterozoic to Mesoproterozoic LH rocks. The presence of sediment ages that have previously been attributed to GH, LH and TH rocks within the Paro Formation lead us to propose that the Paro Formation received sediment from multiple source areas, both young and old. The combination of old DZs and less negative $\varepsilon$Nd values in the Paro Formation suggest that both LH and GH rocks are likely sources. Young Cambrian DZs from fine-grained quartzite sample BU07-73 which correlates to rocks in the immediate footwall of the MCT in the southern exposure, requires a Cambrian source.
Figure 2.7. Composite stratigraphic section of the Paro Formation (~5.5 km thick) showing samples at various stratigraphic levels. Next to the section are detrital zircon peaks, $\varepsilon_{\text{Nd}}$ values, and metamorphic mineral assemblages (bt, biotite; grt, garnet; plg, plagioclase; ky, kyanite; sil, sillimanite; st, staurolite). Both detrital zircon peaks and $\varepsilon_{\text{Nd}}(0)$ show no coherent pattern stratigraphically from low to high. Numbers with asterisk are $\varepsilon_{\text{Nd}}(0)$ values from Richards et al. [2006]. All samples have muscovite and quartz. Lithologic colors same as Figure 2.2.
DZ spectra from the Bhutan Himalaya indicate that GH, TH and upper LH strata all contain Cambrian-Ordovician DZs with the exception of lower LH strata that only contain $\geq 1.8$ Ga DZs [McQuarrie et al., 2008; Long and McQuarrie, 2010; Long et al., 2011]. In addition, the presence of 1.8 Ga DZs in the Paro Formation requires a 1.8 Ga source for grains with just 1.8 Ga zircons. Thus the Paro Formation could be completely sourced from upper LH rocks or the source rocks for LH strata, limiting a unique provenance interpretation.

Arranging the Paro Formation as a single stratigraphic succession, and placing DZ and $\varepsilon_{Nd}$ analyses in stratigraphic order (Figure 2.7) suggests that there is no coherent pattern in the age distribution of DZ and $\varepsilon_{Nd}$ signatures. The DZ age signatures vary significantly within the stratigraphic section, suggesting that the Paro Formation was accumulated in a depositional setting characterized by abrupt changes in detrital input. This can result from a combination of active tectonics or a heterogeneous source region, as well as limited mixing of sediment within the basin [DeGraff-Surpless et al., 2003]. One possibility is the Paro Formation represents a basin caught between LH rocks on the northern margin of India and GH rocks undergoing deformation [e.g. Valdiya, 1995; Gehrels et al., 2003; 2006; Cawood et al., 2007] to the north. Limited mixing is still required to explain the $\varepsilon_{Nd}$ values. DZ ages and $\varepsilon_{Nd}$ values analyzed on the same quartzite samples are consistent with each other while the $\varepsilon_{Nd}$ values on schist are always significantly less negative, suggesting that the whole rock $\varepsilon_{Nd}$ might have been affected by sedimentary sorting. Thus to determine the age spectrum that represents the entire Paro Formation requires combining the spectra from individual samples. A combination of several DZ spectra from a sedimentary succession creates a pattern that can be more confidently linked to
source regions than any single spectrum [Gehrels, 2000; DeGraff-Surpless et al., 2003].

The youngest DZ peak in the Paro Formation exposed in the main window is 800 Ma, requiring the Paro Formation to be Neoproterozoic (~800 Ma) or younger in age. However, the presence of Cambrian (~500 Ma) DZs in the Paro Formation in the southern exposure as well as less negative $\varepsilon_{\text{Nd}}(0)$ values of schist in spite of at least one source of old sediment, suggests that the Paro Formation is possibly Cambrian or younger. We obtained an $\varepsilon_{\text{Nd}}(0)$ value of -12.9 on Silurian-Ordovician granite that intrudes the Paro Formation. Ashes (a highly labile component) from this (or slightly older Cambro-Ordovician magmatic events) could provide the young source material that is being preferentially concentrated in the schist of the Paro Formation [McLennan et al., 1989]. The granite that intrudes the Paro Formation (BU08-128) limits the youngest deposition age at between 400 and 500 Ma.

2.7.2. Stratigraphic section or structural repetition

Because the Paro Formation is always in the footwall of a folded thrust that places GH rocks over either the Paro Formation or the Jaishidanda Formation, we suggest that this thrust is the MCT. Except for an abrupt transition in metamorphic grade, the rocks above and below the MCT in western Bhutan are broadly similar in composition. Schist of the GH and Paro Formation have similar $\varepsilon_{\text{Nd}}$ values and they share all of the same DZ peaks, except for the presence of ~1.8-1.9 Ga peaks in the Paro Formation. Although the metamorphic grade of the Paro Formation is distinctly lower than the over riding GH rocks, most notably the presence of paragneiss with partial melt textures in the GH rocks [Swapp and Hollister, 1991; Davidson et al.,...
1997], the garnet + biotite ± staurolite, kyanite and/or sillimanite metamorphic grade of the Paro Formation is similar to GH sections farther east in Bhutan [Long and McQuarrie, 2010]. The strongest argument for the thrust above the Paro Window being the MCT is mapping that links the GH rocks above the Paro Formation to GH rocks in the south that are clearly in the hanging wall of the MCT [Gansser, 1983; Bhargava, 1995; our mapping).

The structural setting of the Paro Formation, beneath the MCT may be used as an argument for structural repetition within the Paro Formation to explain varying DZ and \( \varepsilon_{Nd} \) signatures. Requiring a given DZ or \( \varepsilon_{Nd} \) signature to identify specific Himalayan units would require a complicated pattern of thrust faults and shear zones within the Paro Window to explain geochronologic and geochemistry data. Figure 2.4 highlights the positions of hypothetical structures that could potentially interleave strata. Note that neither folds nor faults separate “matching” packages of rocks. In addition, our mapping argues against structural repetition, which is best shown by the contact between the lower schist unit and the fine-grained quartzite unit. At this boundary the change from schist to quartzite is gradational, with quartzite layers becoming more persistent up-section (Figure 2.4). There are no fault rocks or small-scale ductile structures (crenulations and shear bands) at this boundary. The same observations (no concentration of small-scale structures, faults, etc.) hold true for all of our hypothetical tectonic boundaries (Figure 2.4A). In addition, no change in metamorphic grade (with exception to the Betekha area which shows signs of contact metamorphism near an intrusive orthogneiss body) or strain gradient are observed. In contrast, the contact between the Paro Formation and the overlying GH rocks we map has highly altered, and pervasively sheared rocks and boudinaged and folded quartz veins in upper most Paro Formation schist interbeds. These asymmetric boudins, C-S
fabrics, and shear bands consistently show top-to-the-south shear sense. In addition, there is an abrupt change in metamorphic grade traversing from the upper part of Paro Formation to the lower section of GH paragneiss.

2.7.3 Provenance and age of the Jashidanda Formation

The combined DZ age spectra for the Jaishidanda Formation and Paro Formation, represented by 698 and 505 zircon grains, respectively, are almost identical (Figure 2.8). Both have prominent peaks at 1.25 Ga, 1.7 Ga, and 2.5 Ga, along with several zircons between 900 Ma and 480 Ma (Figure 2.8). While Mesoproterozoic-Neoproterozoic DZs could have the same sources as that of Paleozoic LH rocks, the Jaishidanda and Paro Formations are unique in that they are the only strata that are directly below the MCT whose youngest DZ population is Cambro-Ordovician. Although there is nothing limiting the Jaishidanda Formation from being younger than Cambro-Ordovician, the presence of Silurian-Ordovician granite in the Paro Formation requires it to be older than ~400-500 Ma. We suggest that the source region for the young detritus in the Jaishidanda Formation and Paro Formation is the same as that proposed for similar age detritus in GH and TH rocks, which is from uplift and erosion of the GH section during the Cambro-Ordovician deformation and magmatic event that is recorded across the Himalaya [Valdiya, 1995; Gehrels et al., 2003; 2006; Cawood et al., 2007]. In addition to the similar DZ signature, both the Paro Formation and Jaishidanda Formation occupy the same structural level below the MCT. An important difference is thickness; the Paro Formation measures 5.5 km and the Jaishidanda Formation measures ~1.0 km.
Figure 2.8. Composite U-Pb detrital zircon age spectra of all lithologic packages (Shumar-Daling Group, Paleozoic LH units, Jaishidanda Formation, Paro Formation, GH metasediments, and Tethyan sediments) in the Bhutan Himalaya. Other than data for the Paro Formation, all other data are compiled from McQuarrie et al. [2008], Long and McQuarrie [2010], and Long et al. [2011]. The number (n) besides each plot gives the total number of detrital zircon grains analyzed. Note the close resemblance of DZ spectra of the Paro Formation and the Jaishidanda Formation.
8. Conclusions

The use of DZ and $\varepsilon_{\text{Nd}}$ signatures has been critical in defining lithotectonic units, aiding in deciphering the tectonic history in the central and western parts of the Himalaya. However, new data from the western Bhutan Himalaya are in marked contrast to the increasingly accepted notion that different Himalayan tectonostratigraphic zones always have unique DZ and Nd isotopic signatures. Because these tectonostratigraphic zones record the pre-collisional geologic history of the Himalayas, the accurate reconstruction of their original depositional relationships is fundamentally important. Inaccuracies in correlations due to limited sampling of DZ spectra may have serious implications for the models of India-Asia collision.

Our data show that DZ spectra may not be homogeneous within a given stratigraphic unit and that the youngest zircons do not necessarily represent the deposition age of the strata. Hence the necessity for multiple samples from the same unit to define variability and substantiate provenance conclusions as limited sampling potentially biases both provenance and tectonic interpretations.

The Paro Formation in western Bhutan is a good example of a heterogeneous basin that contains both young (possibly sourced from the orogenesis preserved in GH rocks in the north) and old (from the Paleoproterozoic-Neoproterozoic sources in the south) material. In light of this new data we recommend caution in the use of DZ and $\varepsilon_{\text{Nd}}$ signatures for tectonic interpretation especially when making correlations with studies that extend 1000’s of km east or west of the original study sites.

A strong detrital zircon signal at ~800 Ma implies that the Paro Formation must be Neoproterozoic in age, or younger. However, the presence of a ~500 Ma DZ peak in the southern exposure together with less negative $\varepsilon_{\text{Nd}}(0)$ values suggests that
the Paro Formation is Cambrian or younger. Granite intrusion crystallization ages bracket the youngest permissible deposition age of the Paro Formation as Cambro-Ordovician.

Because of the similarity in structural position and DZ age spectra, we correlate the Paro Formation with the Jaishidanda Formation, allowing us to align the formations (north to south respectively) in reconstructing the pre-collisional Indian margin. Both these formations are recognized to occupy the same stratigraphic level (below the MCT). We suggest that the Paro and Jashidanda Formations contain a detrital record of the Cambrian-Ordovician orogeny, which has not been previously recognized in LH rocks.

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Chapter 3: The age and rate of displacement along the Main Central Thrust in the western Bhutan Himalaya

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3.1. ABSTRACT

In western Bhutan, the Main Central thrust (MCT) is broadly folded, creating multiple exposures of the fault surface over a ~70 km across-strike distance. This unusual map pattern presents a unique opportunity to map the MCT and document both the magnitude and age of displacement. In situ Th-Pb (SIMS and LA-ICP-MS) geochronology of metamorphic monazite from the immediate hanging wall of the MCT indicates that prograde monazite growth in Greater Himalayan (GH) rocks
continued until 20.8±1.1 Ma, whereas crystallization of \textit{in situ} melts, characterized by high Y monazite overgrowths, occurred during cooling from ca. 15-10 Ma. Prograde monazite growth at 15 Ma in Lesser Himalaya (LH) rocks in the immediate footwall requires that LH footwall strata began to be buried at this time, and the MCT had reached its southernmost, exposed extent. By combining prograde monazite ages in the immediate hanging wall and footwall, the duration of MCT displacement is bracketed between 20.8±1.1 and 15.0±2.4 Ma. Immediately north of our study area, a published estimate of shearing along the outer-South Tibetan detachment (STD) argues for displacement between 20-15 Ma, coeval with the age range for MCT displacement that we document in this study. However retrograde monazite grains as young as 10 Ma suggest that GH rocks were cooling until ~10 Ma, 5 Myr later than motion on the outer-STD immediately to the north. This cooling was either the result of continued displacement on the MCT, or growth of a duplex that passively folded the MCT. Using a sequential reconstruction, we estimate a total displacement of ~230 km, which is the sum of displacements on the MCT and the structurally-lower Paro thrust, over a duration of 5.8±2.6 Myr. This indicates a horizontal shortening rate of 4.0±3.2/-1.3 cm/yr, which exceeds present rates estimated from geodetic measurements across the Himalaya, and MCT displacement rates (c. 2 cm/yr) inferred from petrologic and thermal models in central Nepal but is indistinguishable from plate convergence rates calculated for eastern Bhutan between 23-20 Ma (3.3±0.7 cm/yr). Our study highlights that displacement on the MCT alone achieved plate velocity rates in western Bhutan, and that the age and rate of MCT displacement varied significantly across the Himalaya.
3.2. INTRODUCTION

The Himalayan fold-thrust belt formed in response to ongoing continental convergence between India and Asia that began at ca. 50 Ma (e.g. Hodges, 2000). Today, shortening in the Himalaya is estimated to take up nearly one-third of the 5.8±0.4 cm/yr India-Asia convergence rate (e.g. Bilham et al., 1997; Larson et al., 1999). Spanning the entire length of the orogen, the Main Central thrust [MCT; or thrust contact between Greater and Lesser Himalayan (GH and LH) rocks] has accommodated a large percentage of the total shortening in the Himalaya (e.g. Hodges, 2000; Yin and Harrison, 2000) (Fig. 3.1).

Numerous studies have focused on the timing of MCT displacement, which strongly influences views of orogenic evolution. For example, were displacement rates and timing constant (within error) across strike, or did they vary? If they varied, did they do so systematically or chaotically, and how was overall convergence otherwise partitioned? Taken together, the broad sweep of monazite ages in the Himalaya indicates earliest MCT movement in the early Miocene (e.g. Hodges et al., 1996; Harrison et al., 1998; Catlos et al., 2001; 2004; Kohn et al., 2004; Chambers et al., 2011), but inconsistent levels of detail hide potential variations in displacement rates along the MCT. To differentiate rates and their tectonic drivers more targeted comparisons are required. For example, data from central Nepal indicate ca. 5 Ma differences in timing for MCT movement (Kohn, 2008; Corrie and Kohn, 2011) and possible large variations in displacement rates (Kohn et al., 2004). Comparable studies elsewhere in the Himalaya are as yet lacking, prohibiting orogenic generalization.
Figure 3.1. Simplified geologic map of Bhutan and surrounding region modified from Long et al. (2011d). Upper-left inset shows generalized geologic map of the Himalayan orogen (modified from Gansser, 1983). Upper-right inset shows the international boundary of Bhutan. Our study area (Fig. 3.3) is shown by a rectangular box. Abbreviations: TH - Tethyan Himalaya; GH - Greater Himalaya; LH - Lesser Himalaya; STDo – outer-South Tibetan detachment; STDi – inner-South Tibetan detachment; KT - Kakhtang thrust; MCT - Main Central thrust; MBT - Main Boundary thrust; MFT - Main Frontal thrust; ST - Shumar Thrust; RT – Ramgarh thrust; PW - Paro window; LS – Lingshi syncline; TCK - Tang Chu klippe; UK - Ura klippe; SK - Sakteng klippe; LLW - Lum La window (location from Yin et al., 2010). Map of Sikkim region modified from McQuarrie et al. (2008). Baxa Formation in the footwall of RT in Sikkim from Bhattacharyya and Mitra (2009). Map projection - geographic lat/long (WGS84).
Few data are available in Bhutan to constrain the timing of MCT movement. In eastern Bhutan MCT movement is thought to have occurred between 22 and 16 Ma, based on bulk monazite ages from GH gneisses and LH schist collected immediately adjacent to their mutual contact (Daniel et al., 2003). In western Bhutan, no information exists on the actual age of MCT displacement. However, multiple exposures of the MCT and varying structural levels of the GH section in its hanging wall allowed us to sample over a ~70 km across-strike distance. Few other locations in the Himalaya expose such large across-strike distances for any major shear zone, and none has yet been explored chronologically for the unique tectonic information available from such exposures. In principle, across-strike exposures allow resolution of timing and rates of shear zone movement independent of thermal-mechanical models that other studies have relied upon (e.g., Kohn et al., 2004; Kohn, 2008; Corrie and Kohn, 2011).

Here, we present chemically-defined Th-Pb in situ monazite ages from GH rocks immediately above the MCT in western Bhutan in combination with a balanced geologic cross-section to define a minimum amount of displacement. These data allow us to specifically evaluate both the timing and rates of displacement on the MCT. By combining our data with published monazite data further north in Bhutan (Kellett et al., 2010), we compare the age of MCT displacement to proposed coeval deformation of the GH section and displacement on the South Tibetan detachment (STD). We also assess along-strike variations in initiation of MCT motion in comparison with data ~150 km to the east (Daniel et al., 2003).
3.3. GEOLOGIC BACKGROUND AND SAMPLES

The western Bhutan Himalaya was chosen for this study because; i) GH rocks in western Bhutan have experienced high deformation temperatures. GH rocks exhibiting partial melt textures (leucocratic segregation surrounded by melanosome; Fig. 3.2) extend southward to within 15 km of the deformation front. The presence of deformed kyanite together with partial melt distributed through most of the section (Fig. 3.2) constrains a minimum temperature of >700°C (e.g. Spear et al., 1999). Chemical reactions associated with partial melting are important for understanding chemical systematics of monazite (Spear et al., 1999; Spear and Pyle, 2002; Kohn et al., 2005). ii) Folding of the MCT and overlying GH thrust sheet has resulted in multiple exposures over an across-strike distance of ~70 km, which provides a unique opportunity to map and sample the MCT and GH section in multiple locations.

The first-order tectonostratigraphic subdivisions and major bounding faults from south to north in western Bhutan are the Main Frontal thrust (MFT), the Subhimalayan zone (a.k.a. Siwalik Group), Main Boundary thrust (MBT), Lesser Himalayan (LH) zone, Main Central thrust (MCT), Greater Himalayan (GH) zone, South Tibetan detachment (STD), and the Tethyan Himalayan (TH) zone (e.g. Gansser, 1964, 1983; Le Fort, 1975; Hodges, 2000; Yin, 2006) (Fig. 3.3). The Siwalik Group consists of Miocene to Pliocene synorogenic deposits that are bound at their base by the MFT (e.g. Gansser, 1964, 1983; Tukuoka et al., 1986; DeCelles et al., 2004).

The LH zone consists of clastic and carbonate metasedimentary rocks and is divided into two stratigraphic successions: the Paleoproterozoic lower Lesser Himalayan section, and the Neoproterozoic-Paleozoic upper Lesser Himalayan
Figure 3.2. A-D) Photomicrographs of GH rocks taken in plane polarized light showing breakdown of garnet (rounded or replaced grains), which is the likely source of Yttrium (Y) in monazite, to form sillimanite and biotite. In these rocks, the likely retrograde reaction is grt + kfs + melt = bt + sil (melting reaction 12 of Spear et al. (1999) in retrograde sense). Refer to Fig. 3.3 for sample locations and Supplementary Table B1 for mineral assemblages. A) Sample BU08-4, with assemblage qtz + bt + grt + pl + sil + kfs. B) Sample BU08-130, with assemblage qtz + bt + grt + pl + ky + sil + kfs + st. C) Sample BU08-131, with deformed kyanite in presence of qtz + bt + grt + pl + sil + kfs. D) Photograph showing leucocratic segregation (leucosome + biotite = gneiss) (Longitude: 89.25575°E, Latitude: 27.24893°N). Mineral Abbreviations: qtz – quartz, ms – muscovite, bt – biotite, grt – garnet, st – staurolite, ky – kyanite, sil – sillimanite, kfs – K-feldspar, pl – plagioclase.
section (McQuarrie et al., 2008; Long et al., 2011a). The lower Lesser Himalayan section consists of the Shumar Formation (containing fine-grained and medium- to thick-bedded, cliff-forming quartzite with schist and phyllite interbeds) and Daling Formation (schist and green phyllite with quartzite interbeds), which are collectively referred to as the Daling-Shumar Group. The Neoproterozoic-Paleozoic upper Lesser Himalayan section consists of three units, the Baxa Group, Jaishidanda Formation, and Paro Formation. The Baxa Group consists of dark gray to black slate and phyllite, creamy dolomite, white to pink marble and fine to medium grained quartzite. The Baxa Group is separated from the underlying Siwaik Group by the MBT (Fig. 3.3). The Jaishidanda Formation is a thin (~0.5-1 km) interval of light-gray, biotite-rich quartzite, interbedded with biotite-garnet schist that is exposed immediately below the MCT (Fig. 3.3). The Paro Formation consists of quartzite, quartzite-garnet-schist, marble, and minor calc-silicate rocks. It is intruded by two mica-garnet orthogneiss and is interpreted as the northern (distal) equivalent of the Jaishidanda Formation (Tobgay et al., 2010).

The GH zone in western Bhutan consists of lower metasedimentary and orthogneiss units, and an upper metasedimentary unit exposed only in the northern and eastern portions of the map area (e.g. Long et al., 2011d) (Fig. 3.3). The lower metasedimentary unit is Neoproterozoic-Cambrian in age and consists of paragneiss containing staurolite, kyanite, and sillimanite, muscovite-biotite-garnet schist, and quartzite. In western Bhutan it is ~5.0-6.0 km thick and dominates GH rock exposure (Long et al., 2011d). This lower metasedimentary unit is separated from the underlying Jaishidanda Formation in the south and Paro Formation in the north by the MCT (Fig. 3.3). The orthogneiss unit is granitic in composition and represents a
Figure 3.3. Geologic map of western Bhutan (modified from Long et al., 2011b) (see Fig. 3.1 for location and structure abbreviations). Solid stars are our samples with Th-Pb ages of monazite enclosed in white boxes. Open stars are samples from Kellett et al. (2010) in the north. Note that the GH section is divided into three mappable units (GHmu - upper metasedimentary unit, GHo - orthogneiss unit, and GHml - lower metasedimentary unit) that are not separated by faults. A-A’: cross section line of Fig. 3.6E. See Fig. 3.2 for mineral abbreviations. Map projection - geographic lat/long (WGS84).
deformed Cambro-Ordovician granite pluton that intruded GH sedimentary protoliths (Long and McQuarrie, 2010).

The TH zone is separated from the underlying GH zone by the outer-STD. The outer-STD is a ductile shear zone with top-to-the-north shear sense that is located closer to the orogenic front (farther south) than the inner-STD along the high Himalayan peaks (e.g. Grujic et al., 2002; Fig. 3.1). The TH zone consists of Neoproterozoic to Mesozoic sedimentary rocks that were deposited on the distal part of the northern Indian margin (e.g. Garzanti, 1999).

The tectonostratigraphy described above is overprinted by a progressive increase in metamorphic grade, from garnet-biotite-muscovite in the LH Jaishidanda Formation beneath the MCT in the south, to staurolite-muscovite, kyanite-muscovite, sillimanite-muscovite, and finally sillimanite-K-feldspar (muscovite out) in GH rocks above the MCT in the north. This apparent inversion of metamorphic grade occurs within an across-strike distance of <15 km from the southern trace of the MCT. The MCT is commonly mapped as the boundary between GH and LH rocks based on both a lithologic and metamorphic contrast, although penetrative top-to-the-south shear may extend structurally beneath the MCT into the LH section. In the field, we mapped the southernmost trace of the MCT as a boundary that separates ortho- and paragneiss exhibiting partial melt textures from the underlying LH Jaishidanda Formation, which contains biotite-rich, garnet-bearing schist interbedded with quartzite. Further to the north, the MCT is mapped as a boundary that separates GH rocks from the underlying Paro Formation (Fig. 3.3), and again this contact coincides with a sharp lithologic and metamorphic change. Immediately above the MCT, GH rocks consist of partially melted kyanite-bearing paragneiss. Here, the additional presence of sillimanite in association with K-feldspar and leucosome indicates the muscovite dehydration-
melting reaction (ms + pl + qtz = sil + kfs + melt) and suggests that these rocks have attained a minimum temperature of 700°C at a pressure of 8 kbar (e.g. Spear et al., 1999; Daniel et al., 2003). Rocks of the Paro Formation immediately below the MCT are predominantly quartzite with schist interbeds (Tobgay et al., 2010). In addition to accessory minerals, rocks contain quartz, muscovite, biotite, garnet, and rare staurolite with kyanite only in the lowest part of the section. Together with quartz recrystallization microstructures that indicate deformation temperatures of ~500-630°C, these assemblages suggest distinctly lower metamorphic grades in the footwall of the MCT (Tobgay et al., 2010).

Samples were collected from the GH section, with one additional sample from the Jaishidanda Formation just beneath the MCT, along two N-S transects in western Bhutan. Sampling transects extended between the immediate footwall of the southernmost trace of the MCT to north of the GHS-Paro Formation contact (Fig. 3.3). GH samples for Th-Pb geochronology include paragneiss and schist collected within ~650 m structural distance above the MCT, and a LH schist sample collected from ~50 m below the MCT (Fig. 3.3; Suppl. Table B-1).

3.4. MONAZITE CHEMISTRY

We focused on monazite, a LREE-phosphate, because it incorporates significant amounts of U (up to 1 wt. %) and Th (up to 10 wt. %) (Spear and Pyle, 2002), contains little or no common Pb (<1 ppm; Parrish, 1990), and is resistant to radiogenic Pb-loss via diffusion during metamorphism (Parrish, 1990; Smith and Giletti, 1997; Catlos et al., 2002; Harrison et al., 2002; Cherniak et al., 2004). As has
been discussed extensively, monazite growth and chemistry in metapelitic rocks are directly linked to reactions involving silicate minerals, particularly garnet (Spear and Pyle, 2002; Pyle and Spear, 2003; Pyle et al., 2001; Wing et al., 2003; Kohn and Malloy, 2004; Corrie and Kohn, 2008). Key to interpreting monazite ages is the recognition of Yttrium (Y) and Thorium (Th) as chemical tracers that are strongly and systematically zoned in response to metamorphic reactions (Pyle and Spear, 1999, 2003; Spear and Pyle, 2002; Kohn and Malloy, 2004; Kohn et al., 2004, 2005).

In principle, Th and Y systematics in monazite can depend on numerous minerals, including garnet, xenotime, and allanite. Allanite has been observed in some LH rocks and may be an important precursor to monazite (e.g. Catlos et al., 2002). Theoretical analysis (Spear, 2010) demonstrates that during prograde metamorphism, allanite breaks down abruptly to form monazite, and this may explain how prograde monazite formed in the LH Jaishidanda Formation. In contrast, we observed no allanite or xenotime in GH rocks, either as matrix grains or inclusions, nor did we find chemical evidence for their former stability (e.g. high-Y garnet cores: Pyle and Spear, 1999, 2003; Spear and Pyle, 2002). Allanite is rarely identified in GH metapelites (allanite as inclusions within garnet was identified by Chambers et al., 2011). Xenotime is more common and was reported within the matrix of GH samples below the STD in western Bhutan (Kellett et al., 2010), as well as in GH samples from eastern Bhutan (Daniel et al., 2003). For rock compositions without allanite or xenotime, such as those in this study, monazite Y content links directly to prograde and retrograde reactions involving garnet (Pyle and Spear, 1999, 2003; Pyle et al., 2001; Kohn et al., 2004, 2005). Monazite that grows before prograde garnet or during garnet breakdown is characterized by high Y content, whereas monazite that grows in the presence of stable garnet tends to have low Y content. This occurs because, much
like Mn, Y is sequestered in growing garnet, reducing Y contents of later-grown minerals (Pyle and Spear, 1999, 2003; Pyle et al., 2001; Foster et al., 2002; Spear and Pyle, 2002; Kohn et al., 2005). When partial melting occurs, monazite dissolves into the melt, while garnet continues to grow. Upon retrograde cooling and melt recrystallization, monazite grows either as new grains or as overgrowths on old (low-Y, low-Th) monazite cores while garnet dissolves. Y content in the monazite overgrowth should be high because the dissolution of garnet releases Y into the melt that is subsequently sequestered in monazite overgrowths (Pyle and Spear, 2003; Kohn et al., 2005).

In light of this well-understood chemical behavior, we selected GH samples that exhibited leucocratic segregations of quartz and feldspar in the expectation that these represented partial melts (Fig. 3.2; Daniel et al., 2003), and that monazite grains would have core-rim chemical systematics that could be related to partial melting reactions. Typical mineral assemblages are garnet + plagioclase + sillimanite + biotite + quartz + either muscovite or K-feldspar; several rocks contain kyanite or staurolite. In many rocks, garnets are rounded and replaced by sillimanite and biotite (Fig. 3.2), which we interpret as the retrograde melt crystallization reaction: garnet + K-feldspar + melt = biotite + sillimanite (reaction 12 of Spear et al., 1999, in a retrograde sense). In such rocks we would anticipate finding monazite with prograde low-Y cores mantled by retrograde high-Y rims (Pyle and Spear, 2003; Kohn et al., 2004, 2005), allowing a direct link between monazite chemistry and the heating vs. cooling of rocks.
3.5. METHODS

3.5.1. Monazite mapping

All monazite grains were identified in polished thin-sections by reconnaissance mapping on the electron microprobe housed at the RUMrunner facility, Rutgers University, New Jersey. Garnet was present in all samples, but inclusions of monazite in garnet were uncommon and too small for analysis. For each monazite grain located, a backscattered (BSE) image was collected at low magnification to determine its textural relationship. Monazite grains (~ 20-100 \( \mu \text{m} \)) were either at the grain boundaries or inside of mica and feldspar crystals (Fig. 3.4A-F, Suppl. Figs. B-1 and B-2). Two to five grains from each sample were then mapped for elemental distribution (Y, Th, U, P, Ce, and Si) by electron microprobe using an accelerating voltage of 15 kV, a cup current of 200 nA, and a time per pixel of 30 ms. Considering the smaller size of grains and edge effects, beam mapping was preferred over stage mapping (which was performed only on two grains that were between ~100 and 200 \( \mu \text{m} \)). We used Y and Th mapping to guide the selection of spots for Th-Pb analyses because their zoning pattern correlates with different generations of metamorphic growth in a single monazite grain (e.g. Foster et al., 2002; Spear and Pyle, 2002; Gibson et al., 2004; Kohn and Malloy, 2004; Kohn et al., 2005) (see section 3 above).

3.5.2. Monazite Th-Pb geochronology

Monazite grains mapped for Y, Th, U, Ce, and P were relocated in thin section then drilled out using a micro-diamond drill corer (1/4” or 1/8”). Grains were mounted with UCLA 554 monazite standard in epoxy rounds, and analyzed for U-Th-Pb isotopes using secondary ion mass spectrometry (SIMS) at the Department of
Earth and Space Sciences, University of California, Los Angeles. Analysis of monazite followed analytical protocols described in Harrison et al. (1995). One to five analyses per grain were possible depending on crystal size. We used a primary beam intensity of 2 nA ($^{16}$O$^-$), a lateral spot size of ~10 µm, and a total analysis time of 10 minutes, which equates to a crater depth of ~0.5 µm. $^{204}$Pb intensities were corrected for $^{144}$NdThO$_2$$^{++}$ interference by using measured $^{143}$NdThO$_2$$^{++}$ intensities (average 0.161 counts per second), and a $^{144}$Nd/$^{143}$Nd isotopic abundance ratio of 1.95. Common Pb corrections were based on Pb compositions typical for southern California ($^{208}$Pb/$^{204}$Pb = 38.34; Sanudo-Wilhemy and Flegal, 1994). Th-Pb relative sensitivity calibration is based on a 554 monazite standard with $^{208}$Pb/$^{232}$Th age of 45 Ma (Harrison et al., 1999). Three to ten bracketing standard analyses were collected on the same mounts as the unknowns using a linear regression with a fixed slope (0.122) in measured ThO$_2$/Th vs. Pb-Th relative sensitivity for 554 standard. Th/U ratios are based on uncorrected secondary ion intensities. Th-Pb ages are reported with 2σ uncertainties (Table 3.1).

Because of questions relating to core-rim ages, we further analyzed two grains for Y and U-Th-Pb isotopes using a New-Wave UP-213 laser interfaced with a Thermo XSeries2 Quadrupole ICP-MS at Boise State University. Laser conditions included a spot size of 8 µm, repetition rate of 5 Hz, and fluence of c. 12 J/cm$^2$. Isotopes analyzed and their count times in ms were $^{44}$Ca (10), $^{89}$Y (10), $^{202}$Hg (80), $^{204}$Hg+$^{204}$Pb (80), $^{206}$Pb (120), $^{207}$Pb (100), $^{208}$Pb (150), $^{232}$Th (10) and $^{238}$U (40). $^{204}$Pb was too low and uncertain for robust common Pb corrections, whereas $^{207}$Pb/$^{235}$U ages were highly uncertain (≥25%), and $^{206}$Pb/$^{238}$U ages appeared biased from excess $^{206}$Pb. Consequently we focus on $^{208}$Pb/$^{232}$Th ages. All ages were standardized against 44069 monazite (Aleinikoff et al., 2006).
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BU10_81 (grain 2)  3  15.0  2.0  0.00074  0.00010  2.59  86.5  11.6  LR
BU10_81 (grain 1)  2  14.5  2.0  0.00072  0.00010  2.45  80.8  12.6  LR

Early prograde age in the Jaishidanda Fm.;
BU10_79 (grain 1)  1  15.0  2.4  0.00074  0.00012  2.43  63.5  13.6  EP

Remarks:
1. $^{204}\text{Pb}$ intensities corrected for $^{144}\text{NdThO}_2^{++}$ interference by Using measured $^{144}\text{NdThO}_2^{++}$ intensities (average 0.161 counts per second), and $^{144}\text{Nd}/^{142}\text{Nd}$ isotopic abundance ratio = 1.95.
4. Pb-Th relative sensitivity calibration for three to ten bracketing standard analyses on the same mounts as the unknowns using a linear regression with a fixed slope (0.122) in measured ThO$_2$/Th vs. Pb-Th relative sensitivity for 554 standard.
5. Th/U ratios are based on uncorrected secondary ion intensities "primary beam intensity 2 nA (10$^6$O), lateral spot dimensions ~10 μm, depth of analysis crater ~0.5 μm. For additional instrumental parameters, see Harrison et al. (1995), Earth Planet. Sci. Lett., vol. 133 (3-4), 271-282.

For comparison to the LH monazite and other regional ages, we identified the youngest prograde and oldest retrograde mean ages for GH rocks by iterating on population averages until all ages within a mean were consistent to within 95% confidence. This resulted in averaging 5 ages for the youngest prograde mean age (used for calculating thrust rates below), and 18 ages for the oldest retrograde mean age.

3.6. RESULTS

3.6.1. Map Pattern of the MCT

In west-central Bhutan, a tectonic window through the overlying GH section exposes the Paro Formation below a top-to-the-south thrust contact (Figs. 3.1, 3.3). South of this tectonic window, the Paro Formation is exposed again, and is bounded on both sides by north-dipping thrusts (Fig. 3.3). The southern of these two thrusts is an out-of-sequence structure that places the Paro Formation over GH paragneiss (Tobgay et al., 2010). Between the two exposures of the Paro Formation the GH section has been folded into a synform (Fig. 3.3). The southernmost MCT trace is located just ~15 km north of the Main Frontal thrust (MFT), the southernmost structure of the Himalayan system. At its southernmost trace, the E-W striking and N-dipping MCT has GH rocks in its hanging wall and the LH Jaishidanda Formation in its immediate footwall.

3.6.2. Monazite Chemical Zonation Mapping

Most zoning observed in GH monazite grains conforms to expectations regarding prograde melting and retrograde melt crystallization reactions, including
low Y cores, high Y rims, and inverse correlation between Th and Y. Monazite grains from the same sample do not necessarily possess similar zoning patterns. Variations include grains exhibiting patchy Y or Th zoning, or strong zoning in one element and not another. Different chemistries and zoning patterns have been observed in other Himalayan anatetic rocks (e.g. Kohn et al., 2005; Corrie and Kohn, 2011), but typically the chemistries and ages are reconcilable. For example, homogeneously low-Y grains may coexist with homogeneously high-Y grains and grains with low-Y cores and high-Y rims. This simply reflects heterogeneities of mineral growth: some prograde monazite grains serve as nuclei for retrograde rims (zoned grains), others do not (homogeneously low Y), whereas still others nucleate and grow afresh during cooling (homogeneously high Y). This fact is further demonstrated by age distributions.

3.6.3. Monazite Th-Pb geochronology

Monazite ages obtained from multiple spots within individual grains show a wide range of Th-Pb ages with low-Y cores (high Th) typically about 10 Ma older than high-Y rims (e.g. Figs. B-1a and B-2c). Single samples can contain monazite cores as old as 24 Ma, as well as rims and entire grains that are as young as 10 Ma (e.g. BU08-1, Fig. B-1a; BU10-61, Fig. B-2d). Low Y (high Th) cores typically yielded the oldest ages while high Y (low Th) rims had the youngest ages. Many samples illustrate this, such as BU10-55, which exhibits a high Y, 14-16 Ma rim mantling a low Y, 20-22 Ma core. The seemingly anomalous 16.0 Ma age in a low-Y “core” region reflects the different regions sampled by X-rays (surface ~1 µm) vs. the laser (5-10 µm). Our LA-ICP-MS analysis at this spot almost immediately ablated through a thin low-Y, high $^{208}{\text{Pb}}/^{232}\text{Th}$ shell into a uniformly high-Y domain, from
which the age was derived. That is, the X-rays represent the edge of the core, whereas the age represents the underlying rim. The northernmost two samples (BU08-1, and BU08-124; Fig. 3.3) also show high Y, 10-13 Ma rims around low-Y 21-24 Ma cores (Figs. B-1a and B-1c). Unlike other samples, BU08-4 exhibits a low Y core with a high Y rim, but the core and rim ages of 13-15 Ma are the same as retrograde ages from both laterally-equivalent samples and samples further to the south (Figs. 3.3 and 3.4D). For example, sample BU08-130 has grains with ages of $14.0 \pm 0.6$, $15.4 \pm 0.6$ and $13.4 \pm 0.6$ Ma. We interpret the entire BU08-4 grain as post-anatetic. As a whole, monazite grains from the central part of western Bhutan (Fig. 3.3; Figs. B-1 and B-2) have high Y content and correspond to growth ages that range from $15.4 \pm 0.6$ to $9.5 \pm 0.8$ Ma. Monazite grains sampled from rocks close to the southernmost trace of the MCT (Figs. 3.3, B-2d, e, f, g, and h) record prograde ages of $20.3 \pm 1.4$ Ma, $18.7 \pm 3.4$ Ma, and possibly as young as $16.9 \pm 1.4$ Ma and retrograde rims between $15.0 \pm 2.0$ Ma and $10.7 \pm 1.0$ Ma. Monazite in the Jaishidanda Formation sample (BU10-79; Fig. 3.4F), which is located immediately below the MCT, yielded an age of $15.0 \pm 2.4$ Ma, which is similar to the oldest retrograde age of $15.0 \pm 2.0$ Ma immediately above the MCT.

3.7. INTERPRETATIONS: AGE AND RATE OF THRUSTING

An age probability diagram of monazite ages from western Bhutan (Fig. 3.5A) emphasizes the age of the last growth of prograde sub-solidus monazite (youngest high Th and low Y core) and the age of final melt crystallization during cooling (high Y rims). A population of youngest prograde ages (5 analyses) obtained from low-Y cores (high-Th) together give a mean and 2σ uncertainty of $20.8 \pm 1.1$ Ma (Fig. 3.5B).
Figure 3.4A-F. Elemental maps of Y and Th in monazite, illustrating distinct chemical domains. Representative spot analyses are inset in back-scattered electron images that show textural context of *in-situ* monazite (abbreviated mnz). White circles are chronologic analysis spots with ages. Some of these spots are LA-ICP-MS analyses, not SIMS analyses (listed with 2σ uncertainties). High vs. low concentrations of Y and Th in monazite are denoted by “hot” vs. “cold” colors. Refer Fig. 3.2 for mineral abbreviations, Fig. 3.3 for sample locations, and Figs. B-1 and B-2 for supplementary data.
A mean of oldest retrograde ages (18 analyses) suggests that in situ melt crystallization (high-Y monazite rims) commenced at 15.1±0.4 Ma (Fig. 3.5C).

The young population of GH monazite ages, (15.0±2.4 to 9.5±0.8 Ma) are from monazite with high Y rims or high homogeneous Y. Our sampling criteria (garnet in the presence of partial melt) together with garnet-reaction textures (Fig. 3.2) observed in thin sections, confirm that the dissolution of garnet in GH rocks is the source of Y that is sequestered in monazite during melt crystallization (Spear et al., 1999; Pyle and Spear, 1999). Partial melt is ubiquitous in GH rocks in western Bhutan, and because the MCT emplaces hot GH rocks (at least 700-750°C; Davidson et al., 1997; Daniel et al., 2003; Corrie and Kohn, 2011) on cold (ca. 400-450°C) LH rocks, the initiation of thrusting along the MCT and emplacement of GH rocks would be predicted to trigger a thermal perturbation in rocks in the immediate footwall (Jaishidanda Formation), driving prograde monazite growth in the footwall while simultaneously terminating prograde monazite dissolution and driving retrograde growth in the immediate hanging wall. Under this definition, the prograde monazite grains in GH rocks are pre-kinematic (i.e. associated with structural burial), while retrograde monazite grains in GH rocks are syn-kinematic to post-kinematic with respect to south-directed thrusting. Therefore, the youngest prograde age of GH monazite grains limits the oldest possible initiation of south-directed thrust motion on the MCT. The age of prograde monazite growth in the Jaishidanda Formation (~15 Ma) constrains the time at which the MCT is emplaced over LH rocks at the southernmost extent of its trace. Thus the minimum duration of thrusting on the MCT in western Bhutan is from 20.8±1.1 to 15.0±2.4 Ma. Thrusting along the MCT is commonly interpreted to be coeval to north-vergent normal shearing along the South Tibetan detachment (STD) (e.g. Burchfiel et al., 1992; Hodges et al., 1992; Godin et
Figure 3.5. A) Probability distribution plot of Greater Himalayan (GH) and Jaishidanda Formation (JF) monazite ages, with prograde and retrograde (cooling) divisions based on monazite chemistry. GH rocks contain compositionally-distinguishable prograde and retrograde monazite domains. B) A population of youngest prograde ages (5 analyses) obtained from low-Y cores (high-Th) giving a mean age of 20.8±1.1 Ma (2σ uncertainty). C) A population of oldest retrograde or early crystallization ages obtained from high-Y rims (18 analyses) from GH rocks, giving a mean age of 15.1±0.4 Ma (2σ uncertainty).
In the Lingshi syncline (LS; Figs. 3.1, 3.3) in northwestern Bhutan, U-Pb ages of high Y monazite overgrowths in GH rocks in the immediate footwall of the outer-STD constrained the timing of displacement (shearing) to ca. 20-15 Ma (Kellett et al., 2010), which is equivalent to the minimum duration for MCT motion. The cessation of north-directed shearing on the outer-STD argues for termination of retrograde monazite growth at 15 Ma. However, young retrograde monazite grains between 15 and 9.5 Ma at the base of the GH section suggests that GH rocks were crystallizing and cooling until ~10 Ma, 5 Myr longer than motion on the outer-STD in the Lingshi syncline. Because cooling of the GH section is commonly thought to be driven by transport along the MCT, or underlying thrusts, it is possible that displacement on the MCT may have continued until ~10 Ma. This continued displacement must postdate motion on the outer-STD and may be linked to the development of a hinterland duplex that passively folded the northern MCT and the outer-STD, as discussed below (Fig. 3.6). This geometrical argument allows for limited transport along the MCT until ~10 Ma. Alternatively, growth of a duplex, directly under the Paro window would have focused erosion in this region, facilitating erosional cooling of the GH rocks while inhibiting continued thrusting along the MCT.

Sequential restoration of a balanced cross-section across western Bhutan allows estimation of displacement, which we relate to the timing of monazite growth. The two most important constraints in determining displacement magnitudes are the amount of overlap of GH rocks over LH rocks and the amount of overlap of the Paro Formation over more frontal LH rocks. The present-day north to south map extent of GH rocks is 140 km. Taking into account map-scale folding, the total overlap of GH rocks over LH rocks is 174 km (Figs. 3.3 and 3.6). The Paro Formation is underneath
Figure 3.6. Sequentially-restored cross-section illustrating the emplacement of GH rocks along the Main Central thrust (MCT) in western Bhutan. A) Pre-20 Ma distribution of LH rocks and GH protoliths. B) Displacement along the MCT between 20-15 Ma that places GH rocks over a ~174 km-long Paro Formation section. C) Continuing motion on the MCT and displacement along the structurally-deeper Paro thrust (PT) together bury a ~58 km-long Jaishidanda Formation section, resulting in prograde monazite growth in the Jaishidanda Formation at 15 Ma. D) Initiation of displacement along the Shumar thrust (ST) and development of the Paro duplex in the hinterland between 15 and 10 Ma. E) Geologic cross-section (A-A’ in Fig. 3.3) across western Bhutan. Prograde (p) and retrograde (r) monazite ages are plotted across-strike. p* and r* denote prograde and retrograde ages, respectively, from Kellett et al. (2010).
and parallel to GH rocks for a north-south distance of 50 km. Again taking into account map-scale folds, the minimum displacement on the Paro thrust is 58 km. Thus, the collective amount of displacement on the MCT and Paro thrust needed to bury the LH Jashidanda Formation and grow monazite is 232 km (Figure 3.6 A-C).

Determining the amount of shortening within LH rocks beneath the Paro thrust and the geometry of the duplex which folds the Paro thrust and MCT are subject to greater uncertainties. Shortening estimates from balanced cross sections are fundamentally controlled by the area between the mapped surface geology and the basal decollement, and the stratigraphic thickness of the rock units that are structurally repeated to fill this area. As a consequence, depending on the structural level of mapped exposure, shortening estimates for the same map data may vary. There are three possible scenarios for accommodating the area under the Paro Window, which include filling this space by repeating horses of the LH: 1) Baxa Group, 2) Daling-Shumar Group, or 3) Paro Formation. Because the Daling-Shumar Group and the Paro Formation are thicker than the Baxa Group, filling space beneath the Paro window with these formations decreases the total amount of shortening within this portion of the Himalayas. However, since the MCT and Paro thrust must be emplaced over the total restored lengths of both the Paro Formation and the Daling-Shumar Group to bury the LH Jaishidanda sample, most of the shortening within the thrust belt in these 2 scenarios must predate 15 Ma. The first option, filling the space by repeating the Baxa Group, limits the amount of displacement on the MCT to the map constraints described previously. However, filling the space below the Paro window with thrusts that repeat the Paro Formation and the Daling-Shumar Group increases the distance the MCT must travel to reach the southernmost, lower LH rocks by 15 Ma. For the scenario that repeats the Daling-Shumar Group within the duplex, that distance is
~340 km, while the scenario that repeats the Paro Formation requires a distance of 325 km. Filling the space by repeating Baxa Group horses is preferred because: 1) it matches structural repetition of the Baxa Group observed in a similar tectonic window to the west in Sikkim (Fig. 3.1); and 2) it minimizes the rate of displacement on the MCT, keeping it at or below plate tectonic rates.

Combining a sequential reconstruction through western Bhutan (Fig. 3.6) with ages of monazite growth (Fig. 3.3) provides age estimates for thrust initiation, cessation and thrusting rates. From 26-21 Ma the GH section was buried, most likely by shortening and thickening in the overriding TH section (Patel et al., 1993; Vannay and Hodges, 1996; Robinson et al., 2001, 2006; Murphy and Yin, 2003; Ding et al., 2005; Aikman et al., 2008), promoting growth of prograde monazite. Displacement was initiated on the MCT sometime after 20.8±1.1 Ma (Fig. 3.6B), resulting in the cessation of prograde monazite growth through the GH section and eventual burial of LH rocks at 15 Ma. Continued displacement along the frontal part of the MCT was accompanied by motion on the Paro thrust (PT), which places the Paro Formation and the over-riding GH section over at least 232 km of LH rocks, in order to place the GH section over the Jaishidanda Formation sample (BU10-79) at 15 Ma (Fig. 3.6C). Based on this total displacement of ~230 km in 5.8±2.6 Ma, we calculate an average rate of 40 mm/yr with lower and upper limits between 27 and 72 mm/yr. We suggest that between 15 and 10 Ma, the Shumar thrust (ST) placed the Daling-Shumar Group (lower LH) over the Baxa Group (upper LH) in the foreland, and that continued motion occurred along the MCT as a duplex formed in the Paro Formation in the hinterland (Fig. 3.6D). This hinterland duplex passively folded overlying GH and TH rocks while sending an additional ~28 km of displacement towards the foreland on the MCT (Fig. 3.6D and E). This combined shortening of ~124 km along the Shumar
thrust (96 km) and Paro duplex (28 km) from 15-10 Ma provides a shortening rate of ~26 mm/yr. We propose that post-10 Ma shortening was accommodated by the formation of a duplex in the Baxa Group underneath the Paro Window and in front of the ST. The remaining shortening (165 km) from 10-0 Ma suggests continued slowing to rates of 16.5 mm/yr. The long-term average shortening rate from 20 Ma to present is ~28 mm/yr.

The rate of thrusting on the MCT in western Bhutan is higher than both the modern rates of convergence in the Himalaya, which is estimated at ~20 mm/yr from geodetic measurements across the Nepal Himalaya (Bilham et al., 1997; Larson et al., 1999; Cattin and Avaouc, 2000; Bettinelli et al., 2006), and an MCT displacement rate of 22±7 mm/yr, which is inferred from petrologic and thermal models in central Nepal (Kohn et al., 2004). With displacement rates possibly higher than 50 mm/yr, the MCT and Paro thrust rates would be indistinguishable from the rate of relative convergence between the eastern edge of the Indian plate and the Asian plate at this time (van Hinsbergen et al., 2011). The rate we present for MCT displacement is strongly dependent on what loaded and buried the LH Jaishidanda Formation. Although it is possible that the thickened TH section that buried GH rocks to their peak P and T (Patel et al., 1993; Vannay and Hodges, 1996; Robinson et al., 2001), may have also contributed to the burial of the Paro and Jaishidanda formations to the south, we prefer to relate the prograde growth of monazite in LH (and GH) rocks to the timing of peak metamorphism. Mapped inverted temperature gradients in LH rocks in eastern Bhutan (Daniel et al., 2003; Long et al., 2011d) and the Paro Formation (Tobgay et al., 2010) strongly support the hypothesis that peak temperatures were reached via burial by a hot GH section carried by the MCT. In addition, if the Jaishidanda Formation section was initially buried by southward-
displaced, cooler TH rocks, then we would expect prograde metamorphism to continue as the GH rocks were emplaced over the Jaishidanda Formation. The lone prograde monazite age at 15 Ma suggests that the prograde burial path did not continue significantly past this time. Increasing the duration of heating of the Jaishidanda Formation would lower displacement rates for the MCT while maintaining the long-term shortening rate.

3.8. DISCUSSION

A suite of data from this study and from previous studies in the Bhutan Himalaya allow us to evaluate variations in the duration and rate of thrusting along the MCT. In eastern Bhutan, prograde GH monazite grew from 26-~23 Ma (Chambers et al. 2011). Monazite from the immediate hanging wall of the MCT has a U-Pb age of 22±1 Ma (not linked to monazite chemistry) that is interpreted to represent the initiation of displacement on the MCT (Daniel et al., 2003). In the MCT footwall, U-Pb ages of monazite between 20 and 18 Ma suggest a prograde age as a result of burial due to emplacement of the MCT over LH rocks (Daniel et al., 2003). This combination of monazite ages in the immediate hanging wall and footwall imply that MCT displacement in eastern Bhutan occurred between 23-20 Ma, with continued motion (until ~18 Ma) possibly linked to duplex formation in LH rocks (Long et al., 2011b). U-Pb ages of monazite from deformed leucogranite and migmatite higher in the GH section in eastern Bhutan suggest continued internal deformation and shearing within the GH section between 18 and 16 Ma as other parts of the system moved (Daniel et al., 2003). In the Sakteng klippe (SK; Fig. 3.1) in eastern Bhutan, initiation of north-directed normal shearing across the outer-STD would have cooled GH rocks in the immediate footwall and terminated prograde
monazite growth. As a result, prograde GH monazites (26-~23 Ma) are pre-kinematic with respect to the north-directed shear across the STD and prograde monazite from the TH Chekha Formation (~23-21.5 Ma) are syn-kinematic with respect to shear across the STD (Chambers et al., 2011). This timing is consistent with the 3 Myr duration of MCT slip in eastern Bhutan from 23-20 Ma. Using 80-120 km of MCT displacement measured from balanced cross sections across eastern Bhutan (Long et al., 2011b), and a 3 Myr duration of displacement, we calculate an average motion rate of 3.3±0.7 cm/yr, with lower and upper limits of 2.6 and 4 cm/yr.

The preceding compilation of monazite ages from eastern Bhutan highlights an east-west younging trend in initiation and duration of MCT displacement from 23-20 Ma in eastern Bhutan to ~20-15 Ma in western Bhutan. Although the MCT in western Bhutan post-dated displacement in eastern Bhutan by at least 3 Myr, the rate of displacement in western Bhutan (3-7 cm/yr) has substantial overlap with rates from eastern Bhutan (2.6-4 cm/yr). Our displacement rates of 3-7 cm/yr on the MCT are notably higher than the long-term average shortening rate (2.2±0.5 cm/yr) within Bhutan. In addition, MCT displacement rates in Bhutan are higher than the ca. 2 cm/yr rate calculated for central Nepal based on geochronologic and thermobarometric data tied to thermal-mechanical models (e.g., Kohn et al., 2004; Kohn, 2008; Corrie and Kohn, 2011). A critical caveat is that model results are strongly dependent on the boundary conditions, thermal properties and convergence rates used (e.g. Corrie and Kohn, 2011). Generally, rates of displacement are held constant in models based on the similarities between geodetic convergence rates (19±2.5 mm/yr; Bettinelli et al., 2006), Holocene shortening rates (21.5±2 mm/yr; Lavé and Avouac, 2000) and rates of total shortening across the Himalaya from 25 Ma to present (18-22 mm/yr; e.g. DeCelles et al., 2001; Long et al. 2011b), thus
precluding documenting potentially significant changes in rates though time. By applying a similar approach to the Kathmandu/Langtang region as we apply here, map patterns of the MCT and Ramgarh thrust (RT) require minimum displacement amounts of 174 km and 168 km respectively (Pearson and DeCelles, 2005). Linking this displacement to Th-Pb in-situ monazite ages from the same region (Kohn et al., 2004) suggests displacement rates of 2.5-4.3 cm/yr for the MCT (between 16±1 and 10.5±0.5 Ma) and rates that range from 3.4 to as high as 8.4 cm/yr (from 10.5±0.5 to 8.9 Ma (muscovite cooling age) or 5 Ma (prograde LH duplex monazite)) for the RT. However, it is unlikely that shortening rates ever exceeded India-Asia convergence rates of 5±0.5 cm/yr (Copley et al., 2010; van Hinsbergen et al., 2011). Combining these data from Bhutan and Nepal confirms that the MCT, which extends throughout the length of Himalayan orogen, may not have one unique age and rate of displacement. In addition, although long-term shortening rates are remarkably similar to those measured through GPS, we suggest these long-term shortening rates may be averaging periods of fast and slow shortening and potentially obscuring tectonically significant changes in rate.

3.9. CONCLUSIONS

The unique across-strike exposures in western Bhutan combined with chemically-defined Th-Pb in-situ monazite ages from GH rocks immediately above the MCT in western Bhutan, allow us to constrain the age of displacement on the MCT, and with the aid of sequential cross-section reconstruction, estimate the rate of this displacement. The following conclusions can be drawn from our study:

1. The age of displacement on the MCT in western Bhutan is between 20 and 15 Ma, as defined by the youngest prograde GH monazite in the hanging wall and
a prograde age of LH monazite in the footwall. This is also the same age
duration as displacement and shearing along the outer-STD.
2. The calculated rate of displacement on the MCT in western Bhutan is between
3 and 7 cm/yr, which is similar to estimated plate tectonic rates over this
window of time.
3. The age of displacement on the MCT varies across the Bhutan Himalaya.
There is a ~3 Myr delay between the initiation of MCT displacement in
eastern Bhutan and the initiation of MCT displacement in western Bhutan.
4. Variations in age, rate, and duration of MCT displacement may be the rule
rather than the exception. Documenting how these rates vary in space and
time will provide critical insight into the processes that govern the
accommodation of plate convergence.

3.10. ACKNOWLEDGEMENTS

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3.11. REFERENCES CITED


Chapter 4. Metamorphism, thrust kinematics, and shortening in the
Himalayan fold-thrust belt, western Bhutan

4.1. ABSTRACT

Mineral assemblages of metapelites in western Bhutan show an inverted metamorphic sequence, starting with biotite zone in the upper Shumar-Daling Group, which transitions into garnet zone in the Jaishidanda Formation and the lower Greater Himalayan (GH) section, and then into the sillimanite and sillimanite-K-feldspar zones in higher structural levels of GH section. While metamorphic conditions of ~400-500°C and ~5-7 kbar are prevalent in the biotite zone, the equilibrium mineral assemblage in the garnet zone indicates pressure (P) and temperature (T) conditions of ~5-12 kbar and ~550-650°C, respectively. In the lower part of the GH section right above the garnet zone, the staurolite and kyanite zones show higher temperatures (~570-675°C) with pressures (~6-9 kbar) identical to that of the garnet zone. The sillimanite zone indicates attainment of P-T conditions of ~5-9 kbar and ~575-700°C. In the sillimanite-K-feldspar zone, the presence of deformed kyanite together with melt constrains peak P and T of ≥8 kbar and ≥700°C. This peak metamorphic mineral assemblage is overprinted by sillimanite + K-feldspar that constrain P to 3-4 kbar. Quartz microstructures also record an increase in deformation temperatures from the upper Shumar-Daling Group (400-500°C), Jaishidanda Formation (~500°C), and up through the GH section (~500-700°C).

In the Paro Formation, the presence of staurolite-kyanite at the base of the section and sillimanite in the upper part of the section indicates an upright
metamorphic sequence. Peak metamorphic minerals suggest constant temperature (~575-700°C) throughout the Paro Formation, with pressure decreasing upsection from ~5-12 to ~4-7 kbar. Quartz deformation microstructures define a temperature profile of ~500-700°C in the basal section, ~500°C in the middle section, and ~500-700°C in the upper section.

Prograde metamorphism of metapelites of the GH section continued until ~20 Ma, when prograde metamorphism was initiated in the Paro Formation as a result of structural burial related to thrust-sense emplacement (shear) of GH rocks over the Paro Formation along the Main Central thrust (MCT). This emplacement of GH rocks recorded by synmetamorphic microfabrics such as shear bands, and helicitic inclusion trails of garnet porphyroblasts, within the main foliation with top-to-the-south shear sense overprint the earlier tectonic foliation.

Continuing motion on the MCT together with displacement on the Paro thrust, a structure interpreted at the base of the Paro Formation, triggered prograde metamorphism in the Jaishidanda Formation at ~15 Ma, via burial under the frontal portion of the GH section. Therefore, a large percentage of overall shortening (~230 km) in the Bhutan thrust belt was accommodated by displacement on the MCT and PT between ~20 Ma and ~15 Ma, which yields long-term shortening rates of ~30-55 mm/yr. Displacement along the Paro thrust is described by microstructures (C-S fabric, shear bands, σ-type garnet porphyroblasts) that overprint the main matrix foliation. The burial of the Jaishidanda Formation by continued south-directed motion of the MCT resulted in the development of shear bands and crenulations that overprint the schistosed matrix fabric. All kinematic indicators show top-to-the-south shear sense. Mineral stretching lineations in the GH section and Paro Formation trend NNE-
NNW and SSE-SSW directions that correspond to a maximum stretching parallel to the principal direction of the Himalayan shortening.

From ~15 Ma to present, shortening was accommodated by displacements on the Shumar thrust, Main Boundary thrust, Main Frontal thrust, the upper and lower Lesser Himalayan duplex which structurally repeats horses of Baxa Formation, Shumar-Daling Group, and Paro Formation. From the minimum shortening of 466-566 km accommodated between the MFT and South Tibetan detachment from ~20 Ma to present, a long-term shortening rate of 25-28 mm/yr is calculated for western Bhutan. A comparison of structural geometry, thrust kinematics, shortening, and the rate of shortening highlights a significant variation across the Himalayan fold-thrust belt between eastern Bhutan and Sikkim which may imply variability of convergence and strain rates on timescales of Myr even within an along-strike distance of ~250 km.

4.2. INTRODUCTION

The Himalayan orogen is a striking example of a young active continent-continent collision zone, where the original positions and contacts of sedimentary rocks that were deposited at or near the northern margin of India between early Proterozoic and Paleocene time have been profoundly metamorphosed and progressively deformed and translated in a series of large, south-vergent thrust sheets. These rocks, which are now exposed on the surface, record the pressures (P) and temperatures (T) under which they were formed and the P-T paths they followed on their way to the surface. Because P-T paths are closely linked to the faults and ductile shear zones that transport these metamorphic rocks, a study that integrates
metamorphic data with the sequence of deformation defined from the observed geometries of major structures is necessary to document the key processes that drive deformation and metamorphism during mountain building. The Himalayas provide a unique opportunity to study the temporal evolution of deep rocks in an orogen that is still actively building mountains.

Although major structures in the Himalaya extend for 1000’s of kilometers, there are uncertainties as to how far the motion age of these structures, the P-T conditions of rocks they carry, and their displacement magnitudes can be extrapolated along-strike. For example, a compilation of previous studies revealed significant along-strike differences in absolute ages of the Main Central thrust (MCT), one of the most important structures in the Himalaya (Tobgay et al., in revision). Despite these differences, similarities of peak P-T conditions and displacement rates along-strike are used to assert that strain measurements in one part of the orogen can be realistically extrapolated over hundred kilometers (e.g. Corrie and Kohn, 2011). Until now, many of the first-order observations for the Himalaya are focused in areas that are relatively easy to access, such as central Nepal, and only recently have studies expanded to areas of the eastern Himalaya, including the kingdom of Bhutan (Grujic et al., 2002; Hollister and Grujic, 2006; McQuarrie et al., 2008; Bhattacharyya and Mitra, 2009, 2011; Long and McQuarrie, 2010; Yin et al., 2010; Long et al., 2011a; Tobgay et al., in revision). This growing body of work across the Himalaya allows us to compare similarities and differences along-strike over 100’s of kilometers, and evaluate what stays the same, what changes, and why. Being able to examine major structures and P-T paths at the scale of the Himalayan orogen elucidates key orogenic processes, and the rates at which these processes occur.
Bhutan is located in the eastern Himalaya between the Indian states of Sikkim (to the west) and Arunachal Pradesh (to the east). Much of the early work in Bhutan focused on the tectonometamorphic evolution of the thick metamorphic core of the orogen known as the Greater Himalayan sequence (GHS) (Swapp and Hollister, 1991; Davidson et al., 1997; Daniel et al., 2003). More recent studies have elucidated the stratigraphy, structure, and kinematics of the Lesser Himalayan sequence (LHS) in the frontal part of the thrust belt (McQuarrie et al., 2008; Tobgay et al., 2010; Long et al., 2011a, 2011b, 2011c). Even within Bhutan, studies on the evolution of the MCT highlight significant variations in the age, duration, and kinematics of slip over an along-strike distance of <250 km (Daniel et al., 2003; Kellett et al., 2009, 2010; Chambers et al., 2011; Tobgay et al., in revision). In addition, GH rocks in the west are preserved within 15 km across-strike distance of the Main Frontal thrust (MFT), whereas in the east they are 50-90 km further to the hinterland.

The western Bhutan Himalaya was chosen for this study because: (1) the cumulative effects of deformation and erosion has resulted in multiple exposures of the LH-GH contact in several places across-strike. These exposures provide a unique opportunity to sample at similar structural levels in multiple places across-strike, which enables evaluation of how P-T conditions evolved from hinterland to foreland; (2) the western Bhutan Himalaya records significantly hotter temperature conditions both above and below the MCT than equivalent structural levels in eastern Bhutan; and (3) the multitude of structural levels exposed in western Bhutan allows for establishment of the relationship between deformation and metamorphism in the LH, GH and Tethyan Himalaya (TH).

In this chapter, I present thin-section observations (mineral assemblages, reaction textures, micro-structures, and quartz deformation microtextures) of LH
rocks (Baxa, Shumar-Daling, Jaishidanda and Paro Formations) and GH rocks (starting from the exposure of the southernmost trace of the MCT to immediately below the outer-South Tibetan detachment), in combination with balanced cross-sections to elucidate the relationships between deformation geometry and metamorphism in the western Bhutan Himalaya. Interpretations of balanced cross-sections provide shortening estimates across the western Bhutan Himalaya, and linking these shortening estimates to geochronologic and thermochronologic data allow us to determine the rates of shortening.

4.3. HIMALAYAN GEOLOGIC BACKGROUND

The Himalayan orogenic belt stretches between the Namche Barwa syntaxis in Tibet and the Nanga Parbat syntaxis in Pakistan, and is the result of active convergence between the Indian and Eurasian plates. This active convergence is accommodated by crustal shortening, which has been accommodated by a south-vergent fold-thrust system that involves sedimentary rocks that were deposited on the northern margin of India between Proterozoic and Paleocene time (Gansser, 1964; Powell and Conaghan, 1973; Mattauer, 1986; Dewey et al., 1988; Hodges, 2000; DeCelles et al., 2002; Murphy and Yin, 2003). This fold-thrust system has been divided into several tectonostratigraphic zones bounded by major brittle faults and ductile shear zones that can be recognized along the entire length of the orogen (Gansser, 1964, 1983; LeFort, 1975; Hodges, 2000; Yin, 2006). From south to north, the tectonostratigraphic subdivisions and major bounding faults include the Indo-Gangetic foreland basin, the MFT, Subhimalayan rocks, the Main Boundary thrust (MBT), LH rocks, the MCT, GH rocks, the southern Tibetan Detachment system (STDs), and the overlying TH strata (Fig. 4.1).
The Subhimalayan zone consists of synorogenic sedimentary deposits exposed along the entire length of the orogen (Gansser, 1964). These synorogenic deposits are collectively named the Siwalik Group, and are Miocene to Pliocene in age (Gansser, 1964; Tukuoka et al., 1986; Harrison et al., 1993; Quade et al., 1995; Burbank et al., 1996; DeCelles et al., 1998, 2001, 2004; Ojha et al., 2000; Huyghe et al., 2005; Robinson et al., 2006). The Siwalik group is bound at its base the Main Frontal thrust (MFT), which coincides with the present Himalayan topographic front.

The LH zone, which sits structurally above the Subhimalayan zone across the Main Boundary thrust (MBT), contains clastic and carbonate sediments metamorphosed to green-schist facies, which were deposited on the northern portion of the Indian craton during Proterozoic and early Paleocene time (Gansser 1964; Schelling and Arita, 1991; Parrish and Hodges, 1996; Kumar, 1997; Upreti, 1996, 1999; DeCelles et al., 2000; Richards et al., 2005; McQuarrie et al., 2008; Kohn et al., 2010). In Nepal, Lesser Himalayan strata are Paleoproterozoic and Mesoproterozoic in age (DeCelles et al., 2000; Martin et al., 2005), although they are as young as Neoproterozoic to Cambrian in northwest India (Myrow et al., 2003; Azmi and Paul, 2004; Richards et al., 2005), Nepal (Brunnel et al., 1985; Valdiya, 1995), Bhutan (McQuarrie et al., 2008; Long et al., 2011a), and Arunachal Pradesh (Tewari, 2001; Yin, 2010). Permian glacial deposits (diamictite), the Permian to Eocene Gondwana sequence, and Eocene foreland basin deposits constitute the youngest Lesser Himalayan rocks (Najman et al., 2006; Robinson et al., 2006; Yin, 2006).

The GH zone consists of metasedimentary and metaigneous rocks, along with Miocene leucogranite, and is separated from the underlying LH zone by the MCT (Heim and Gansser, 1939; Gansser, 1964; LeFort, 1975). GH rocks have attained the highest metamorphic grades of any rocks in the orogen (upper amphibolite to
granulite facies) and represent the metamorphic core of the Himalaya (e.g. Grujic et al., 2002).

The Tethyan Himalayan zone which both structurally (separated by the STDs) and stratigraphically overlies the Greater Himalayan zone, represents Neoproterozoic to Mesozoic sediments that were deposited on the distal northern Indian margin of the Tethyan ocean basin (Gaetani and Garzanti, 1991; Brookfield, 1993; Garzanti, 1999).
Figure 4.1. Simplified geologic map of Bhutan and surrounding region modified from Long et al. (2011d). Upper-left inset shows generalized geologic map of the Himalayan orogen (modified from Gansser, 1983). Upper-right inset shows the international boundary of Bhutan. Our study area (Fig. 4.2) is shown by a rectangular box. TH - Tethyan Himalaya, GH - Greater Himalaya, LH - Lesser Himalaya, STDo – outer-South Tibetan detachment, STDi – inner-South Tibetan detachment, KT - Kakhtang thrust, MCT - Main Central thrust, MBT - Main Boundary thrust, MFT - Main Frontal thrust, ST - Shumar thrust, RT – Ramgarh thrust, PW - Paro window, LS – Lingshi syncline, TCK - Tang Chu klippe, UK - Ura klippe, SK - Sakteng klippe, LLW - Lum La window (location from Yin et al., 2010). Map of Sikkim region modified from McQuarrie et al. (2008). Baxa Formation in the footwall of RT in Sikkim (from Bhattacharyya and Mitra, 2009). Map projection - geographic lat/long (WGS84).
The four Himalayan tectonostratigraphic units are described here from south to north, based on 1:50,000-scale geologic mapping along roads and trails in western Bhutan performed by our research group. Mapping involved collection of lithologic and structural data from all structural levels between the MFT and the outer-STD (Fig. 4.2). Our map data lie along several roads and trails in western Bhutan (Fig. 4.2). Our mapping was integrated with previously mapped data from Gansser (1983), Bhargava (1995), and Grujic et al. (2002) to help trace contacts both along and across-strike. The level of rock exposure was good where there were fresh road cuts but along the semi-used trails rock remained covered by thick vegetation.

4.4.1. The Subhimalayan Zone

The map pattern of the Siwalik Group varies significantly from east to west across the map area, from a limited 3.0 km wide N-S exposure of the middle Siwalik section in the southeast corner (SE of Phuentsholing) to no exposure in the southwest corner (Fig. 4.2). The discontinuous outcrop pattern of the Siwalik Group throughout Bhutan has been explained with missing sections being: 1) covered by Quaternary sediment; 2) overridden by the MBT; or 3) never deposited (Gansser, 1983; Bhargava, 1995). Where the Siwalik Group is best exposed in the east it is subdivided into lower, middle, and upper members (Gansser, 1983; Acharyya, 1994; Lakshminarayana and Singh, 1995).
Figure 4.2. Geologic map of part of the western Bhutan Himalaya showing detailed lithologic units (modified from Long et al., 2011d). Strike and dip symbols indicate our mapping. Map projection: geographic lat/long (WGS84).
East of the border town of Phuentsholing (Fig. 4.2), only the middle section of the Siwalik Group is exposed over a ~2.5 km N-S extent. The middle member of the Siwalik Group is composed of tan to gray, medium to coarse-grained sandstone and pebble to cobble conglomeratic sandstone (Lakshminarayana and Singh, 1995). Aerial photography data suggest that in this location the Siwalik Group strikes E-W with an average dip between 30° and 50°N (Gansser, 1983). Across the majority of the Himalaya, the MFT is identified by the southernmost exposure of the Siwalik Group, and since these rocks are not exposed in western Bhutan, location of this structure is difficult. We suggest that the best estimation for the MFT location is immediately south of uplifted river terraces mapped by Gansser (1983) ~20 km south of Phuentsholing. This implies the presence of a thicker, buried section of Siwaliks, possibly similar to the thickness of Siwaliks exposed in eastern Bhutan (Long et al., 2011a).

4.4.2. The Lesser Himalayan Zone

The LH zone can be divided into two stratigraphic successions: (1) the Paleoproterozoic lower Lesser Himalayan section, and (2) the Neoproterozoic-Paleozoic upper Lesser Himalayan section.

Daling-Shumar Group: The Paleoproterozoic lower Lesser Himalayan section is separated into two distinct formations, the Shumar Formation and overlying Daling Formation. The Shumar Formation consists of fine-grained and medium to thick bedded, cliff forming quartzite with schist and phyllite interbeds. The Shumar Formation is overlain by green phyllite and quartzite interbeds of the Daling Formation along a gradational contact. The presence of a mylonitized orthogneiss body containing distinctive feldspar augen northeast of Phuentsholing is characteristic
of the Daling Formation. These units have a combined thickness of ~3.5 km. The map pattern for the Shumar-Daling Group is consistent along-strike for a distance of ~60 km. To the east and west the Shumar Formation pinches out.

The upper Lesser Himalayan section consists of three map units, the Baxa Group, Jaishidanda Formation, and Paro Formation (Figs. 4.2 and 4.3). Two Permian LH units, the Diuri Formation and Gondwana sequence, are limited to eastern Bhutan, and are not exposed in our map area (Fig. 4.2).

**Baxa Group:** The Baxa Group is ~1.3 km thick, and is differentiated into lower (Phuentsholing) and upper (Pangsari) formations near the town of Phuentsholing (e.g. Tangri, 1995). To the west, between Phuentsholing and Samtse, the middle rock unit of the Baxa Group, the Manas Formation is exposed. The Phuentsholing Formation is characterized by dark gray to black slate and phyllite with interbeds of limestone, creamy dolomite, and thin beds of fine- to medium-grained quartzite. The Pangsari Formation consists of gray to green, locally talcosed phyllite interbedded with characteristic white to pink marble and thin beds of fine- to medium grained greenish quartzite. The Manas formation is primarily fine to medium grained, locally pebbly to conglomeratic quartzite interlayered with gray phyllite and gray dolostone, which occurs in lenses up to 2 km thick. All three Baxa Group formations are locally repeated by thrust faults. These thrusts were identified based on stratigraphic (lithologic repetition) and structural observations (localized zones of deformation, including highly-sheared and folded rocks, brecciation, and the presence of hot springs and tufa precipitation). The predominant lithology of the Baxa Group varies significantly along strike. In western Bhutan and Sikkim the Baxa group is dominated by phyllite and dolomite (Bhattacharyya and Mitra, 2009), while in central and eastern Bhutan the lithology is predominantly fine to medium grained quartzite,
Figure 4.3. A) Column showing tectonostratigraphy of western Bhutan with generalized lithology. B) The same section with Jaishidanda Formation correlated to the Paro Formation, its distal equivalent. Note: above column B), rest of the TH section is not shown. Unit ages of Jaishidanda and Paro Formations are from Tobgay et al., (2010). Refer Fig. 4.1 for structure abbreviations.
interbedded with dolomite and phyllite (Long et al., 2011a). Just within western Bhutan, the Baxa Group shows significant along-strike variation in across-strike width of exposure and lithology. In the southeast corner of the map (Fig. 4.2), only the Manas Formation is exposed, across a 15 km wide N-S distance. Near the town of Phuentsholing (Fig. 4.2), only the Phuentsholing and Pangsari formations are exposed, over N-S width of ~5.0 km. To the west of Phuentsholing, all three Baxa Group formations are exposed over a N-S width of ~15.0 km. These three Baxa Group formations are lumped as the tan Baxa Group shown on cross sections discussed under section 4.5.3.3. In the southwestern corner of the map area, a thin band of the Phuentsholing Formation is exposed along the northern edge of uplifted Quaternary river terraces and is present as strath terraces underneath Quaternary deposits.

**Jashidanda Formation:** The Jashidanda Formation is a ~0.5 km-thick interval of biotite-rich, garnet-bearing schist with common quartz vein boudins, interbedded with light-gray, biotite-rich quartzite (Figs. 4.2 and 4.3). Schist of the Jashidanda Formation displays abundant biotite-rich laminations and crenulations. The map pattern of the Jashidanda Formation is approximately 4-5 km thick across-strike, and is separated from GH rocks to the north by the MCT. To the south the Jashidanda overlies Paleoproterozoic lesser Himalayan rocks across a depositional (disconformable) contact. This map pattern is continuous along-strike for 150 km to the east without structural repetition.

**Paro Formation:** The Paro Formation is a ~5.5 km thick straight stratigraphic section that not repeated by structures (Fig. 4.3) (Tobgay et al., 2010). It consists of high-grade metasedimentary and calcareous rocks including calc-silicate rocks, marble, quartzite, quartz-garnet-schist, and two mica-garnet orthogneiss (Tobgay et
Staurolite and kyanite are only present at the base of the section. The 5.5 km thickness is comparable to the 5.0 km thick Paro Formation mapped by Mitra et al. (2010) to the west in Sikkim. The Paro Formation exhibits a consistent internal stratigraphy, and was divided into several thin but mappable units (described in Tobgay et al. (2010)). Because of their similarity in structural position below the MCT, the Paro and Jaishidanda formations were interpreted by Tobgay et al. (2010) as correlative, and representing sedimentary rocks deposited on distal and proximal parts of the precollisional Indian passive margin, respectively (Tobgay et al., 2010) (Fig. 4.3). The Paro Formation is exposed in two windows through the overlying GHS. The map pattern of the main Paro window in the north is a broad ~40 km diameter dome complicated by 2nd order open anticlines and synclines (Fig. 4.2). To the south of the main window, the Paro Formation is exposed for ~2.0 km in the N-S direction and at least 25 km in the E-W direction. The northern boundary of this exposure is a north dipping, low angle shear zone that places GH rocks over the Paro Formation, while the southern boundary is an abrupt, brittle fault dipping ~45° to the north that places Paro Formation over GH rocks.

4.4.3. The Greater Himalayan Zone

The GH zone is ~7-10 km thick and is divided into three mappable units; lower (GHml) and upper (GHmu) metasedimentary units and an intervening orthogneiss (GHo) unit (Figs. 4.2 and 4.3). These three units together constitute the structurally-lower GH section and their relationships are argued to be original (stratigraphic and intrusive) and not repeated by structures (Long et al., 2011d). The lower metasedimentary unit (Neoproterozoic-Cambrian) consists of quartzite, biotite-
muscovite-garnet schist, and paragneiss with common staurolite, kyanite, or sillimanite and ubiquitous partial melt textures (leucosomes) that are deformed.

The orthogneiss unit (GHo) is characteristically cliff-forming, massive weathering, and granitic in composition. It exhibits leucosomes and abundant cm-scale, feldspar porphyroclasts. This orthogneiss unit is interpreted as deformed Cambro-Ordovician granite plutons that intruded GH sedimentary protoliths (Long and McQuarrie, 2010).

The majority of western Bhutan is dominated by exposures of the lower metasedimentary unit (GHml), which crop out over a north-south distance of 75 km between the southernmost trace of the MCT and an area ~5 km south of the outer-STD (Fig. 4.2). All GH rocks immediately above the MCT have a pervasive foliation that generally strikes E-W and dips 25°-35° north. In the central part of this exposure, foliation in GH rocks defines a broad dome, in which the main window of the Paro Formation is exposed.

4.4.4. The Tethyan Himalayan Zone

In northwestern Bhutan, the basal TH map unit, the Chekha Formation, is predominantly tan, cliff-forming marble with lesser gray phyllite and dark phyllitic quartzite. This 2.2-4.0 km thick Chekha Formation is overlain by ~4.0 km of Paleozoic TH strata, which are dominated by medium gray, cliff forming, thin bedded, silt lamination-rich, fossiliferous limestone and brown pebble-clast diamictite. Above this is a ~2.0 km thick Triassic-Jurassic section that contains dark gray, tan weathering shale and fine-grained sandstone that form diagnostic tan slopes. This is overlain by a Cretaceous map unit (~600 m thick), which dominated by dark gray to black, brown weathering, carbonaceous shale and brown sandstone. The
Tethyan rocks are preserved in a broad (~40 km diameter) structural basin (Fig. 4.2), called the Lingshi syncline (Fig. 4.1). The boundary that separates TH rocks from GH rocks immediately below is a top-to-the-north, normal-sense shear zone, called the outer southern Tibetan detachment (STDo) (e.g. Grujic et al., 2002; Kellett et al., 2009).

4.5. METAMORPHISM

Metamorphic studies in the Bhutan Himalaya have focused primarily on GH rocks in eastern and central parts of the country as well as GH rocks at the highest structural levels in northwestern Bhutan (Swapp and Hollister, 1991; Davidson et al., 1997; Daniel et al., 2003). These studies have divided the GH section into several metamorphic zones including a garnet-staurolite zone immediately above the MCT, followed by a kyanite zone and then by a sillimanite zone a few hundred meters upsection. The deepest and hottest rocks the Bhutan thrust belt, which are in the sillimanite-k-feldspar zone, are exposed only between a large out of sequence structure in northern Bhutan, the Kaktang thrust (KT), and the trace of the STD at the Bhutan/Tibet border. The arrangement of metamorphic zones, with increasing grade upsection within the GH section, defines an apparent inverted metamorphic field gradient. Outside of the original pioneering work of Gansser (1983), no metamorphic study has examined the LH and the structurally lower GH unit in western Bhutan. As a result, the work presented below from western Bhutan will: 1) add to the growing body of information on metamorphism in the Bhutan Himalaya; 2) provide the data to evaluate the similarities and/or dissimilarities in the metamorphic histories of rocks in
eastern and western Bhutan; and 3) allow us to evaluate the resulting tectonic implications.

4.5.1. Metamorphic minerals and textures

This section contains descriptions of thin section observations (mineral assemblage and reaction textures) according to their structural level, including the LH section (Shumar-Daling Group, Jaishidanda Formation, and Paro Formation), and the GH section, in order to understand the relationship between metamorphic mineral growth and fabric development and to constrain P-T paths in western Bhutan (Fig. 4.4).

Within western Bhutan a complete sequence of all the Barrovian metamorphic zones are present within a map distance of ~5-20 km, with metamorphism progressing in a direction from biotite-zone in the Shumar-Daling Group to sillimanite-k-feldspar-zone within the GH section. These metamorphic zones are generally parallel to the MCT and to the strike of the lithologic layering. However, zones that include kyanite, sillimanite, and sillimanite-K-feldspar in the higher structural levels of GH section are progressively condensed from east to west (Fig. 4.5), indicating that metamorphic isograds are not always concordant to lithologic layering or structures. Major silicate phases present in each zone are presented in Tables 4.1-4.3 and metamorphic reactions based on equilibrated mineral assemblages are described in section 4.4.1.1 below. The stability fields of mineral assemblage characteristic of each metamorphic zone are described using a petrogenetic grid from Davidson et al. (1997) and Daniel et al. (2003), and are discussed in section 4.4.2.
Figure 4.4. Map of western Bhutan showing sample locations. Dark stars are samples from this study and open stars are samples from Kellett et al. (2010). Map legend same as Fig. 4.2.
Metamorphic grade starts with lower greenschist facies in the upper Shumar-Daling Formation that transitions into upper greenschist facies in Jaishidanda Formation and lower GH section, and then to amphibolite-granulite facies higher within the GH section (Fig. 4.5). A gentle increase in metamorphic grade from the upper Shumar-Daling section through the Jaishidanda Formation becomes a steep gradient (garnet to sillimanite-k-feldspar zone) within the GH section over a structural distance of ~5 km. This transition zone coincides with a major tectonic structure, the MCT.

Samples from the Shumar-Daling, Jaishidanda, and Paro Formations and the GH section all display dynamically-recrystallized quartz deformation microstructures (e.g. Grujic et al., 1996; Long and McQuarrie, 2010; Long et al., 2011c) with quartz crystallographic preferred orientation (CPO) and/or grain shape preferred orientation (GSPO) showing N-S transport direction consistent with mineral stretching lineation (Fig. 4.6).

4.5.1.1. The Biotite Zone (the Shumar-Daling Group)

Baxa quartzite below the Shumar thrust (ST) and Shumar-Daling quartzite collected within the first ~200 m above the Shumar thrust (ST) primarily contain quartz, muscovite, and chlorite (Fig. 4.7). Above this, biotite is present (marked as biotite-in isograd on Fig. 4.5), with its abundance increasing in the upper part of the Daling Formation section. Therefore, the area between the biotite-in isograd and the top of the Shumar-Daling Group constitutes “biotite zone”.

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Table 4.1. Mineral assemblages of LH samples. All samples have quartz. Mineral abbreviations; ms - muscovite, bt - biotite, grt - garnet, st - staurolite, ky - kyanite, pl - plagioclase, sil - sillimanite, cl - chlorite. Cross indicates mineral presence in the sample.

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<th>Sample</th>
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<td>x</td>
</tr>
<tr>
<td>BU10-77 (-20 m*)</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td></td>
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</tr>
<tr>
<td>BU07-66 (-50 m*)</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td></td>
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<tr>
<td>BU10-78 (-50 m*)</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td></td>
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</tr>
<tr>
<td>BU10-79 (-50 m*)</td>
<td>x</td>
<td>x</td>
<td>x</td>
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<tr>
<td>BU08-135 (-450m*)</td>
<td>x</td>
<td>x</td>
<td>x</td>
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<td>Shumar-Daling Group;</td>
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<tr>
<td>BU07-63 (+500 m**)</td>
<td>x</td>
<td>x</td>
<td></td>
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<tr>
<td>BU07-59 (+500 m**)</td>
<td>x</td>
<td>x</td>
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</tr>
<tr>
<td>BU07-57 (+250 m**)</td>
<td>x</td>
<td>x</td>
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<td>Baxa Group;</td>
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<td>BU08-137 (-100 m**)</td>
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</tr>
<tr>
<td>BU10-70 (-1300 m **)</td>
<td>x</td>
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<td></td>
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<td>x</td>
</tr>
<tr>
<td>BU07-60 (-2.5 km**)</td>
<td>x</td>
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<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
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</table>

* Approximate distance in meters structurally below the MCT
** Approximate distance in kilometers or meters structurally above and below the ST
Table 4.2. Mineral assemblages of Paro Formation. All samples have quartz. Mineral abbreviations; ms - muscovite, bt - biotite, grt - garnet, st - staurolite, ky - kyanite, sil - sillimanite, and pl - plagioclase. Cross indicates mineral presence in the sample.

<table>
<thead>
<tr>
<th>Sample</th>
<th>ms</th>
<th>bt</th>
<th>grt</th>
<th>st</th>
<th>ky</th>
<th>sil</th>
<th>pl</th>
</tr>
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<tbody>
<tr>
<td>BU08-77 (~5.0 km)</td>
<td>x</td>
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</tr>
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<td>BU07-84 (~4.0 km)</td>
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</tr>
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<td>BU08-118 (~3.7 km)</td>
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<td>TTS07-14 (~3.3 km)</td>
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<td>BU07-75 (~3.2 km)</td>
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<td>BU08-129 (~3.2 km)</td>
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</tr>
<tr>
<td>PF08-3 (~3.0 km)</td>
<td>x</td>
<td>x</td>
<td>x</td>
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<td></td>
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<td>PF08-4 (~3.1 km)</td>
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<tr>
<td>TTS07-4 (~2.7 km)</td>
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<td>TTS07-16 (~2.7 km)</td>
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<td>TTS07-11 (~2.5 km)</td>
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<tr>
<td>TTS07-6 (~2.4 km)</td>
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<td>TTS07-28 (~2.0 km)</td>
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<td>TTS07-19 (~2.0 km)</td>
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<td>TTS07-18 (~1.75 km)</td>
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</tr>
<tr>
<td>TTS07-17 (~1.75 km)</td>
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<tr>
<td>TTS07-20 (~1.7 km)</td>
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<td>x</td>
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<tr>
<td>TTS07-29 (~1.7 km)</td>
<td>x</td>
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<tr>
<td>BU08-133 (~1.7 km)</td>
<td>x</td>
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<td>x</td>
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</tr>
<tr>
<td>BU07-73 (~1.5 km)</td>
<td>x</td>
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</tr>
<tr>
<td>PF08-2 (~1.2 km)</td>
<td>x</td>
<td>x</td>
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</tr>
<tr>
<td>BU07-76 (~1.2 km)</td>
<td>x</td>
<td>x</td>
<td></td>
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</tr>
<tr>
<td>BU07-80 (~900 m)</td>
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<tr>
<td>PF08-1 (~800 m)</td>
<td>x</td>
<td>x</td>
<td></td>
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<td></td>
<td>x</td>
</tr>
<tr>
<td>TTS07-23 (~800 m)</td>
<td>x</td>
<td>x</td>
<td>x</td>
<td></td>
<td></td>
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<td>x</td>
</tr>
<tr>
<td>BU08-127 (~800 m)</td>
<td>x</td>
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<td>x</td>
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<tr>
<td>TTS07-10 (~700 m)</td>
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<tr>
<td>BU08-3 (~700 m)</td>
<td>x</td>
<td>x</td>
<td>x</td>
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</tr>
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<td>BU08-2 (~500 m)</td>
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</tr>
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<td>TTS07-15 (~150 m)</td>
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</tr>
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<td>BU07-79 (~150 m)</td>
<td>x</td>
<td>x</td>
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</table>

Note: Approximate distance in kilometers or meters structurally above the Paro thrust.
Table 4.3a. Mineral assemblages of GH samples. All samples have quartz (qtz).

Mineral abbreviations; ms - muscovite, grt - garnet, st - staurolite, ky - kyanite, pl - plagioclase, sil - sillimanite, kfs - k-feldspar, and chl - chlorite. Cross indicates mineral presence in the sample. Above the southernmost MCT trace;

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<th>st</th>
<th>ky</th>
<th>sil</th>
<th>pl</th>
<th>kfs</th>
</tr>
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<td>Western side;</td>
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<td></td>
<td></td>
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</tr>
<tr>
<td>BU10-57 (~1.7 km)</td>
<td>x</td>
<td></td>
<td>x</td>
<td>x</td>
<td>x</td>
<td></td>
<td></td>
<td>x</td>
</tr>
<tr>
<td>BU10-56 (~1.7 km)</td>
<td>x</td>
<td>x</td>
<td></td>
<td>x</td>
<td>x</td>
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<tr>
<td>BU10-55 (~1.5 km)</td>
<td>x</td>
<td></td>
<td></td>
<td>x</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BU10-58 (~1.5 km)</td>
<td></td>
<td>x</td>
<td>x</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>BU10-59 (~650 m)</td>
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<td>x</td>
<td>x</td>
<td>x</td>
<td></td>
<td></td>
<td>x</td>
</tr>
<tr>
<td>BU10-60 (~400 m)</td>
<td>x</td>
<td>x</td>
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<tr>
<td>BU10-61 (~220 m)</td>
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<tr>
<td>BU10-62 (&lt;100 m)</td>
<td>x</td>
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<td>Eastern side;</td>
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<tr>
<td>BU10-87 (~4.0 km)</td>
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<td>x</td>
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<td>x</td>
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</tr>
<tr>
<td>BU10-88 (~3.5 km)</td>
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<td>x</td>
<td>x</td>
<td>x</td>
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</tr>
<tr>
<td>BU10-86 (~3.0 km)</td>
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<td>x</td>
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<td>TTS07-1 (~1.7 km)</td>
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<td>TTS07-2 (~1.7 km)</td>
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<tr>
<td>TTS07-26 (~1.7 km)</td>
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<td>x</td>
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<td>BU10-85 (~1.5 km)</td>
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<tr>
<td>BU10-84 ~1.5 km)</td>
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<tr>
<td>BU10-75 (~700 m)</td>
<td>x</td>
<td>x</td>
<td>x</td>
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<td>BU10-76 (~600 m)</td>
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<td>BU10-83 (~550 m)</td>
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<td>BU10-81 (~400 m)</td>
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<tr>
<td>BU07-79 (~400 m)</td>
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<td>x</td>
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<tr>
<td>BU08-134 (~400 m)</td>
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<td>BU10-74 (~400 m)</td>
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<td>BU10-82 (~100 m)</td>
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<td>BU10-80 (~100 m)</td>
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<td>North of the sil-kfs-in isograd;</td>
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<td>BU07-74 (~800 m)</td>
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<td>BU10-52 (~800 m)</td>
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<td>BU10-51 (~600 m)</td>
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<td>BU10-53 (~600 m)</td>
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<td>BU10-54 (~500 m)</td>
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<td>CG09-2 (~500 m)</td>
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</tr>
<tr>
<td>BU08-132 (~50 m)</td>
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<td>x</td>
<td>x</td>
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<td>BU08-131 (~50 m)</td>
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<td>x</td>
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<td>x</td>
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</tr>
</tbody>
</table>

(Approximate distance in kilometers or meters structurally above the MCT)
Table 4.3b. Mineral assemblages of GH samples. All samples have quartz (qtz).

Mineral abbreviations; ms - muscovite, grt - garnet, st - staurolite, ky - kyanite, pl - plagioclase, sil - sillimanite, and kfs - k-feldspar. Cross indicates mineral presence in the sample. Around the edges of Paro window:

<table>
<thead>
<tr>
<th>Sample</th>
<th>ms</th>
<th>bt</th>
<th>grt</th>
<th>st</th>
<th>ky</th>
<th>sil</th>
<th>pl</th>
<th>kfs</th>
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<tbody>
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<td>BU08-116 (~6.0 km)</td>
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<td>BU08-115 (~5.25 km)</td>
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<td>x</td>
</tr>
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<td>BU08-114 (~5.0 km)</td>
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<td>x</td>
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</tr>
<tr>
<td>BU08-113 (~4.3 km)</td>
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<td></td>
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(Approximate distance in kilometers or meters structurally above the MCT)
Figure 4.5. Map of western Bhutan showing metamorphic isograds (bt-in, grt-in, st-in, ky-in, ky-out, sil-in, and sil-kfs-in) and the distribution of kyanite and sillimanite. Sillimanite is also present either as kyanite psuedomorph or in association with k-feldspar and muscovite. Refer Table 4.1-4.3 for mineral abbreviations. Sillimanite location with asterisk from Gannser (1983).
4.5.1.2. The Garnet Zone (The Jaishidanda Formation and base of GH)

Garnet first begins in the Jaishidanda Formation and is present throughout the GH section. Therefore, garnet zone extends from ~500 m below the MCT to ~100-650 m above the MCT (the staurolite isograd) (Fig. 4.5). Garnet co-existing with quartz, muscovite, biotite, plagioclase, and chlorite in absence of chloritoid suggests garnet was a likely product of a continuous reaction; chlorite + muscovite + quartz = garnet + biotite + water (Fig. 4.8).

Garnet crystals were not analyzed for their mineral chemistry (Mn, Fe, Mg, and Ca), and thus nothing can be stated on core-rim zoning patterns of these endmember compositions. However, most garnets contain inclusions of opaque minerals and to a lesser extent quartz, often mantled by thin inclusion-free rims (Fig. 4.7c and d). These inclusions are either in the form of straight or curved lines within the garnet poikiloblasts and are largely discordant to the main matrix foliation.

4.5.1.3 The Staurolite Zone (GH section)

The staurolite zone is present between 400m and 600m above the MCT in the eastern portion of our study area (north of Phuentsholing) but is restricted to only 100m and 200m above the MCT along our westernmost transect (Fig. 4.5). Staurolite grains are pale-yellow and range in size from one to several millimeters. Grains are subhedral or anhedral (six-sided crystals are embayed) and are sub-parallel to the surrounding foliation. The mineral assemblage for this zone includes quartz, muscovite, biotite, garnet, staurolite, plagioclase, and chlorite (Fig. 4.9). Although there is no obvious reaction texture, staurolite poikiloblasts that contain quartz inclusions suggest that staurolite is formed by a common discontinuous reaction: garnet + chlorite + muscovite = staurolite + biotite + water (Fig. 4.10).
Figure 4.6. Map of western Bhutan showing the trend and plunge of mineral stretching lineations and sense of shear. Mineral stretching lineations generally trend NNE-NNW and SSE-SSW directions with predominantly top-to-the-south shear sense. From ~100-200m below the South Tibetan detachment (STDo), kinematic indicators show top-to-the-north shear sense.
Figure 4.7. Photomicrographs (A and B) taken in cross-polarized light. A) Sample BU07-59: Quartzite from Shumar-Daling Group containing muscovite, biotite, and chlorite. B) Sample BU07-57: Quartzite from Shumar-Daling Group containing only muscovite. Other minerals in A) and B) include quartz, muscovite, and biotite with chlorite only in B). Photomicrographs (C and D) taken in plane-polarized light. C) Sample BU10-63: Subhedral garnet crystals with inclusion-rich cores and inclusion-free rims from the upper part of the Jaishidanda Formation. D) Subhedral garnet porphyroblast with inclusion trails from the lower part of the Jaishidanda Formation.
Figure 4.8. Petrogenetic grid showing the stability fields of the progressive metamorphic mineral assemblages observed within the Lesser Himalayan strata beginning from the middle Shumar Formation to the Jaishidanda Formation immediately below the MCT. Circled letters A and B refer to mineral assemblages in Thompson (1957) AFM diagrams depicted in the lower part. Discontinuous metamorphic reaction curves after Daniel et al. (2003). An arrow in the upper right diagram shows the metamorphic field gradient.
The stability of staurolite requires pressures of ~6-9 kbar and temperatures of ~570-620°C (e.g. Davidson et al., 1997; Holdaway, 2000; Daniel et al., 2003; Dasgupta et al., 2004). This indicates that staurolite is only stable over a very restricted domain of temperature (~50°C). Garnet grains are also subhedral or anhedral and contain spiral-shaped inclusion trails and straight inclusions which are oblique to the main foliation.

4.5.1.4. The Kyanite Zone (GH section)

Kyanite first appears ~200m and ~700m above the MCT in our western and eastern transects, respectively. The kyanite zone is defined as the region above the first appearance of kyanite and below the first appearance of sillimanite. It is characterized by the following mineral assemblage: kyanite, staurolite, garnet, biotite, muscovite, plagioclase, and quartz (Fig. 4.9d). Kyanite exists as elongated tabular crystals with corroded edges, which exhibit both bent and kinked crystal forms. The kyanite crystals are generally aligned sub-parallel to the matrix foliation, and define the mineral lineation in areas where there is no obvious stretching lineation. With no obvious reaction textures, there are three possible kyanite producing reactions: (1) staurolite + chlorite = kyanite + biotite, (2) staurolite + muscovite + quartz = kyanite + biotite + garnet + water, and (3) staurolite = kyanite + garnet + biotite + water (staurolite-out isograd) (Fig. 4.10). Among these possibilities, reaction (2) is favored because kyanite is observed to co-exist with staurolite and garnet with chlorite.
Figure 4.9. Photomicrographs of lower GH samples taken in plane-polarized light. A-C). Garnet zone ~100-400 m above the MCT. D). Staurolite and deformed kyanite from the kyanite zone above the staurolite-in isograd. E-F). Sillimanite in association with kyanite, biotite, garnet, muscovite, and quartz within the sillimanite zone.
4.5.1.5. The Sillimanite Zone (GH section)

Sillimanite (fibrolite) is the high temperature \( \text{Al}_2\text{SiO}_5 \) polymorph that marks the highest grade zone in the Barrovian sequence. Fine needles and bundles of fibrolite first appear \(~1.7\) km and \(~400\)m above the MCT along the eastern and western traverses in western Bhutan, respectively (Figs. 4.5 and 4.9e-f). Fibrolite is nucleated within mica (both muscovite and biotite). The mineral assemblage in this zone is muscovite, biotite, garnet, kyanite, fibrolite, and plagioclase with fibrolite partially replacing bent and broken kyanite (Table 4.3a). These textures suggest that fibrolite is the result of the discontinuous reaction staurolite + muscovite + quartz = sillimanite + biotite + garnet + water, and a polymorphic conversion of kyanite to sillimanite (Fig. 4.10). The absence of staurolite in the mineral assemblage indicates that staurolite has been completely consumed by the first discontinuous reaction to produce sillimanite.

4.5.1.6. The Sillimanite-K-feldspar Zone (GH section)

The first appearance of sillimanite in association with K-feldspar marks the second sillimanite isograd. This “sillimanite-kfs zone” extends from \(~5-15\) km north of the southernmost trace of the MCT to \(~100\)m below the outer-STD, and includes GH rocks immediately above the MCT in the region surrounding the Paro window (Fig. 4.2). In this zone, the mineral assemblage includes quartz, muscovite, biotite, garnet, sillimanite, K-feldspar, plagioclase, and kyanite (Figs. 4.5 and 4.11a-f; Table 4.3b). However, a drastic reduction in muscovite abundance in this zone is
Figure 4.10. Petrogenetic grid for metapelitic rocks showing the stability fields of the progressive metamorphic mineral assemblages observed within the GH section (above the MCT and below the outer-STD). A) Discontinuous metamorphic reaction curves after Davidson et al. (1997) with Almandine = 0.74 isopleth and $P_{H2O} = P_{total}$. 1. Decompression reaction $\text{Grt} + \text{Ms} = \text{Sil} + \text{Bt}$; 2. Retrograde reaction $\text{Grt} + \text{Kfs} + \text{Vapor} = \text{Sil} + \text{Bt}$. B) Decompression and retrograde reaction curves (dashed grey lines) with varying $\text{Fe}/(\text{Fe} + \text{Mg})$ in garnet after Spear et al. (1999). Dark dashed line is Almandine = 0.74 isopleth. 1. Decompression reaction $\text{Grt} + \text{Ms} = \text{Sil} + \text{Bt}$; 2. Retrograde/decompression reaction $\text{Grt} + \text{Kfs} + \text{Melt} = \text{Sil} + \text{Bt}$. Arrows show the metamorphic field gradient. Note reaction curves in presence of vapor and melt with Almandine = 0.74 isopleth. Circled letters C-G refer to mineral assemblages in Thompson (1957) AFM diagrams depicted in the lower part.
characterized by proportionate increase in Kfs abundance and coarse-grained, prismatic sillimanite. Muscovite that is present in trace amounts in this zone is aligned parallel to the foliation and increases in abundance ~500m below the outer-STD. Throughout the sillimanite-kfs zone, Kyanite is preserved as both euhedral grains, as well as grains that have been bent and corroded. Garnets are anhedral and are partially replaced by sillimanite and biotite. The intergrowth of K-feldspar in plagioclase is observed in some samples collected in the western section (Fig. 4.11b). Cordierite, which generally forms at low pressures (~4 kbar) and high temperatures (>600°C) (e.g. Davidson et al., 1997) was not observed in any of the GH samples. The above mineral assemblage and reaction textures suggest at least two reactions. The reaction of muscovite with quartz (known as the muscovite dehydration reaction) to produce sillimanite, K-feldspar, and water confirms the attainment of peak metamorphic conditions at the second sillimanite isograd (Fig. 4.10). This reaction coincides with the demise of the coarse-grained muscovite, and the release of water can trigger partial melting. However, the presence of muscovite in trace amounts concordant to the foliation suggests that muscovite was not fully consumed (original muscovite is still in equilibrium). In addition, anhedral garnet crystals partially replaced by sillimanite and biotite suggest garnet reacting with muscovite to form sillimanite and biotite. There are two possible interpretations for such a texture, depending on the scale of observation; (1) a decompression reaction where T remains nearly constant as pressure drops (Davidson et al., 1997; Daniel et al., 2003; Hollister and Grujic, 2006) or, (2) a retrograde reaction: garnet + k-feldspar + melt = sillimanite + biotite, characterized by a drop in T (cooling) (Spear et al., 1999). At thin section scale, we do not observe the presence of melt, which suggests the first reaction (Fig. 4.11e).
Figure 4.11. Photomicrographs taken under cross polarized light (A and B) and under plane polarized light (C, D, and E) showing the presence of K-feldspar in association with sillimanite and muscovite. Mineral assemblage in A), B, C, D, and E include bt, grt, pl, sil, kfs, and ms with ky present only in C). Exsolved lamellae (i.e. K-feldspar growing inside plagioclase) present in B). Sillimanite wrapping around garnet (E) suggests sillimanite as a product of garnet dissolution. F) Outcrop photo showing leucosome surrounded by biotite-rich melanosome inside garnet-rich paragneiss.
However, at an outcrop scale, we do observe sillimanite and biotite wrapping around garnet, and flanked by leucosomes and K-feldspar porphyroblasts (Fig. 4.11f). Thus, the mechanism for producing the observed texture (decompression or cooling) is dependent on the mobility of the reaction (and the required proximity to melt). In either case, the first peak metamorphic reaction characterized by the presence of K-feldspar together with sillimanite, deformed kyanite, and melt is overprinted by a decompression reaction (Fig. 4.11e) and/or a retrograde reaction where garnet reacts with muscovite to form sillimanite and biotite.

A mineral K-feldspar is hard to distinguish from quartz and plagioclase. In some samples, K-feldspar was identified by its characteristic cross-hatched twinning and the presence of exsolved lamellae. In samples without obvious cross-hatching twinning and exsolution lamellae, Becke Line method (relative relief) was used to distinguish K-feldspar from quartz and plagioclase. Because no probe work was done to identify K-feldspar, there are uncertainties in identifying K-feldspar in several of my samples.

4.5.1.7. The staurolite-kyanite-sillimanite zone (Paro Formation)

The 5.5 km-thick Paro Formation is divided into several mappable units. From lower to higher structural levels, these units include a basal schist unit, a fine-grained quartzite unit, which exhibits distinctive marble interbeds (10’s of meters thick; marble I and marble II), and a coarse-grained quartzite unit with a thick marble interbed (marble III) (Tobgay et al., 2010). In the basal schist unit and in the schist interbeds within the fine-grained quartzite unit, mineral assemblages include quartz, muscovite, biotite, garnet, staurolite, kyanite, and plagioclase (Table 4.2). Garnet porphyroblasts are generally inclusion-rich, with inclusion patterns forming either
straight or curved lines (Fig. 4.12a and e). Staurolite is rare and when present it is 
dismembered. The kyanite crystals are comparatively more abundant than staurolite 
and are generally dismembered, bent, or broken. The peak metamorphic mineral 
assembleage (quartz, muscovite, biotite, garnet, plagioclase, staurolite, and kyanite) 
(Table 4.2) suggests that kyanite was a product of the reaction: staurolite + muscovite 
+ quartz = kyanite + biotite + garnet + water (Fig. 4.13). At higher structural levels 
within the fine grained quartzite, kyanite disappears and fibrolite emerges together 
with garnet, biotite, plagioclase, muscovite, and quartz. Fibrolite needles radiate in all 
directions and in several cases fibrolite wraps around garnets (Fig. 4.12d). Samples 
collected from the coarse-grained quartzite unit at the top of the Paro Formation 
contained no fibrolite, although the presence of graphitic sillimanite within the coarse-
grained quartzite near Haa was reported by Gansser (1983).

4.5.2. P-T Conditions and Grade of Metamorphism

Metamorphic mineral reactions plotted on petrogenetic grids provide the range 
of pressure and temperature conditions at which the observed mineral assembleges are 
stable, but do not precisely constrain peak P-T conditions (Figs. 4.8, 4.10, and 4.13).

In the biotite zone (i.e. Daling-Shumar Group), co-existing quartz, muscovite, 
biotite, and chlorite, indicates lower greenschist facies metamorphism corresponding 
to ~400-500°C, as a result of both burial to depths of 15-20 kms and heating from 
higher-grade rocks in the hanging wall of the overriding MCT (Daniel, et al., 2003; 
Celerier et al., 2009; Long et al., 2011c) (Fig. 4.8).
Figure 4.12. Photomicrographs taken in plane polarized light showing the mineral assemblages of Paro Formation. A). Subhedral garnet from schist interbed within the fine grained quartzite unit. B-C). Dismembered and deformed kyanite crystals at the base of the fine-grained quartzite. D). Sillimanite (and most likely cordierite) wrapping around garnet. E-F). Dismembered kyanite from the basal schist unit.
In the garnet zone immediately below and above the MCT, the equilibrium mineral assemblage suggests P and T conditions of ~4-12 kbar and ~550-650°C, respectively (Figs. 4.8 and 4.10). These estimates overlap with the peak metamorphic P-T conditions (~9-12 kbar and ~650-675°C) calculated based on the equilibrated mineral assemblage of garnet-biotite-muscovite-plagioclase-quartz for the same zone in eastern Bhutan (Daniel et al., 2003). Although peak metamorphic pressures across the MCT remain invariant, metamorphic temperatures increase progressively from the footwall into the hanging wall (Daniel et al., 2003). While this inverted temperature gradient is still recorded by metamorphic mineral zones in western Bhutan, the data are not available to constrain peak metamorphic pressures. In the staurolite and kyanite zones, the equilibrium mineral assemblage suggests slightly higher temperature conditions of 570-675°C with pressure conditions (~6-9 kbar) remaining similar to that of the garnet zone (Figs. 4.9 and 4.10). Further upsection into the sillimanite zone, the mineral assemblage gives pressure conditions of ~5-9 kbar and a temperature range of ~575-700°C. In the sillimanite-Kfs zone, the presence of deformed kyanite together with partial melt constrains a minimum pressure of 8 kbar at temperature ~700°C. The overlapping pressure ranges combined with a notable increase in minimum required temperatures strongly suggests that the GHS is a straight, non-faulted section and that the inverted metamorphic gradient documented across western Bhutan is a function of increasing temperatures from 550-650 °C at the base to ~700°C within ~7 km. Petrologic evidence (dissolving garnet mantled by sillimanite and biotite) shows that this peak metamorphic mineral assemblage later experienced either low pressure (<5 kbar) metamorphism while maintaining high temperatures (>700°C ), or that it underwent cooling (T = <725°C, P = 4-6 kbar).
Figure 4.13. Petrogenetic grid for metapelitic rocks showing the stability fields of the progressive metamorphic mineral assemblages observed within the Paro Formation. Circled letters A-D refer to mineral assemblages in Thompson (1957) AFM diagrams depicted in the lower part. Discontinuous metamorphic reaction curves after Daniel et al. (2003). Arrows in the upper right diagram shows the metamorphic field gradient
The mineral assemblage in the basal schist unit and the lower section of fine-grained quartzite unit of the Paro Formation shows burial of rocks to the kyanite stability field \((\text{staurolite} + \text{muscovite} + \text{quartz} = \text{kyanite} + \text{biotite} + \text{garnet} + \text{water})\), corresponding to P-T conditions of \(\sim5\)-12 kbar and 575-700°C (Fig. 4.13). In the upper section of the Paro Formation, the existence of fibrolite with garnet, biotite, plagioclase, muscovite, and quartz shows shifting of into the sillimanite stability field \((\text{staurolite} + \text{muscovite} + \text{quartz} = \text{sillimanite} + \text{biotite} + \text{garnet} + \text{water})\). The sillimanite stability field corresponds to P-T estimates of \(~4\)-7 kbar and 575-700°C (Fig. 4.13). The range of pressures and temperatures estimated in the kyanite and sillimanite stability fields show obvious reduction in pressure in the upper section while temperature conditions remained almost constant. This top-bottom pressure gradient can be attributed to the thickness (5.5 km) of the Paro section. The juxtaposition of rock containing undeformed fibrolite near the deformed Cambrian-Ordovician granite intrusion near Shari (Fig. 4.2) rules out intrusion as the likely source of heat for sillimanite-grade metamorphism as proposed in higher structural levels of GHS in central Nepal (e.g. Godin et al., 2001). The presence of graphitic sillimanite in the uppermost section of the Paro Formation is reported by Gansser (1983). Here, the alignment of sillimanite needles suggests pre- to syn-kinematic growth of sillimanite. All these observations in combination with the mineral assemblage (muscovite, biotite, garnet, staurolite, kyanite, sillimanite, and plagioclase) in the lower and upper parts of the section lead to the conclusion that the Paro Formation has attained upper amphibolite facies metamorphism.
4.6. STRUCTURAL GEOLOGY

4.6.1. Map view structures

While the MFT is easily recognized in eastern Bhutan as the contact between the Miocene-Pliocene Siwalik Group and Quaternary foreland basin sediment, the location of the MFT in western Bhutan is less clear. With much of the frontal part covered by Quaternary sediment, the position of the MFT is inferred from the boundary of uplifted fluvial terraces mapped by Gansser (1983). This places the MFT about 10 km south of the southernmost exposure of LH rocks (Gansser, 1983; Long et al., 2011d). The orientation of the MFT is represented by the orientation of Siwalik Group rocks that strike E-W and dip north with dip amounts ranging between 30° and 50°.

The exact location of the MBT in western Bhutan is also uncertain. To the east, the MBT is clearly mapped as the contact between the upper Siwalik Group and the upper LHS (i.e. Gondwana succession). East of Phuentsholing, the MBT is mapped as a north-dipping fault that places the Baxa Group (Neoproterozoic-Cambrian) over the Siwaliks. To the west, the Siwaliks and part of the Baxa Group are not exposed. The position of the MBT is extrapolated from the Siwaliks-Baxa contact east of Phuentsholing to the southern limit of exposed Baxa group strata between Phuentsholing and Samtse. Bedding and foliation in the Baxa Group generally strike E-W and dip north with dip amounts ranging between 25° and 50°. The map view extent of the Baxa Group is much greater than its estimated thickness of 1.3 km. This map view extent together with lithologic (dominantly phyllite and dolomite) repetition, and the presence of faults identified in the field argue that the Baxa Group is repeated by a series of intraformational thrust faults (Fig. 4.2).
The contact between the Baxa Group and structurally higher (but older) Daling-Shumar Group is the Shumar thrust (ST). Bedding and foliation of rocks in its hanging wall are parallel those in its footwall. In addition, foliation is approximately parallel to bedding as indicated by the preservation of original sedimentary layering (Fig. 4.2).

The MCT is exposed in multiple locations from north to south across western Bhutan. In the south, the MCT is mapped as the contact between the Jaishidanda Formation and GH rocks. Field measurements of tectonic foliation show that GH rocks are parallel to bedding and foliation in the underlying Jaishidanda Formation. This contact approximates the minimum southern extent of the MCT, which is exposed only 15 km north of the MFT. To the north, the MCT is located between the Paro Formation and GH rocks. The southern extent of the Paro Formation is constrained by its exposure near Chukha (Fig. 4.2). Across their structural contact throughout the map area, foliation within the GHS and bedding and foliation within the Paro Formation are parallel. In the Chukha region, GH foliation has been warped into a regional anticline-syncline pair. The fault that places the Paro Formation over GH rocks along the southern window is correlated to the northern anticline of this regional anticlinal-syncline pair (Long et al., 2011D). This map pattern constrains the southern-most extent of the Paro Formation south of Chukha but north of the MCT.

The syncline located between the southern extent of the MCT and the anticline that exposes the Paro Formation at Chukha constrains the shallowest depth to the regional decollement in western Bhutan. At the axis of the syncline a total thickness of GH (lower metasedimentary unit), the Jaishidanda Formation, the Shumar-Daling Group, and the Baxa group must fit above the regional decollement between the Baxa Group and the underlying Daling-Shumar Group (Figs. 4.1 and 4.2). This is because
the amount of shortening in the thrusts carrying the Baxa Formation is greater than the
distance from the MBT to the axis of the syncline south of Chukha.

Because of the first-order geometry of the Himalayan fold-thrust belt
described above, the minimum depth to the basal decollement for western Bhutan is
approximately at the stratigraphic top of the Daling Formation, which is ~6 km deep
in the foreland, and dips 4° to the north (Alsdorf et al., 1998; Hauck et al., 1998; Mitra
et al., 2005; Schulte-Pelkum et al., 2005; Singh et al., 2010). The folded MCT and the
underlying Paro Formation requires shortened LH formations to fill the space between
the base of the Paro Formation and the basal decollement (McQuarrie et al., 2008;
Bhattacharyya and Mitra, 2009; Mitra et al., 2010; Long et al., 2011b).

4.6.2. Micro- and meso-scale structures

4.6.2.1. The Baxa Group

The Baxa Group quartzite displays flattened (median ellipsoid ~2:2:1) relict
detrital grains with neoblastic muscovite in pressure shadows between grains (Long et
al., 2011C). The neoblastic muscovite, original detrital mica, and the flattened detrital
quartz grains define a weak foliation that is parallel to primary sedimentary layering.
Baxa Group quartzite displays evidence for subgrain grain rotation localized at grain
boundaries which is indicative of bulging recrystallization and corresponds to a
deformation temperature range of ~280°C and 400°C (Stipp et al., 2002) (Fig. 4.14A-
C).
4.6.2.2. The Shumar-Daling Group

Quartzite of the Daling-Shumar Group has a dominant, bedding-parallel fabric (S1) that is represented by alignment of neoblastic muscovite and biotite (Fig. 4.14D-F). In thin section, quartz grains are observed to be equidimensional and contain polygonal subgrains, typical of subgrain rotation (SGR) recrystallization (Grujic et al., 1996; Stipp et al., 2002) (Fig. 4.14D-F) which corresponds to deformation temperatures between 400- and 500°C (Stipp et al., 2002).

4.6.2.3. The Jaishidanda Formation

There are three main fabrics (S1, S2, and S3) preserved in microstructures of the Jaishidanda Formation (Fig. 4.14G-I). S1 is defined by inclusions in the garnet porphyroblasts while S2 is defined by schistosed matrix fabric (alignment of mica grains). S3 is the overprinting fabric on S2 such as shear bands and crenulation cleavage (Fig. 4.14G-I). Crenulation is more prevalent in the upper section near the Jaishidanda-GH boundary.

Quartz grains exhibit variable grain sizes and formation of amoeboid ‘island’ grains up to ~0.5 mm in length, enclosed by smaller subgrains (Fig. 4.14G-I). Such deformation fabrics are interpreted to represent a transitional texture between SGR and grain boundary migration (GBM) recrystallization (~500°C) (Stipp et al., 2002). Grains have shape preferred orientation parallel to the alignment of micas.
Figure 4.14. Photomicrographs illustrating quartz deformation microstructure observed in the Baxa Group, Shumar-Daling Group, and the Jaishidanda Formation (under cross-polarized light). All thin sections cut perpendicular to quartzite bedding and parallel to mineral stretching lineation. A-C) Quartzite from the Baxa Group exhibiting flattened relict detrital grains with neoblastic muscovite in pressure shadows between grains. Subgrain rotation localized at grain boundaries is indicative of bulging recrystallization. D-F) Quartzite from the Shumar-Daling Group showing equidimensional and polygonal subgrains characteristics of SGR recrystallization. S1 represents a fabric defined by alignment of neoblastic muscovite and biotite. G-I) Quartzite from the Jaishidanda Formation exhibiting variable grain size and formation of amoeboid shaped grains surrounded by smaller subgrains, characteristics of a transition between GBM and SGR recrystallization. Also present are isolated, plastically elongated, non-recrystallized quartz grains surrounded by biotite-muscovite matrix. Here S2 is a pervasive fabric defined by alignment of micas while S3 is the overprinting fabric on S2 defined by shear bands and crenulations.
4.6.2.4. The Paro Formation

The Paro Formation is characterized by both small (schist partings and interbeds) and major changes in lithology (marble, quartzite, schist) that are interpreted as original sedimentary bedding ($S_0$). The basal section of the Paro Formation records three main fabrics; $S_1$, $S_2$, and $S_3$ (Fig. 4.15). $S_1$ is defined by straight but rotated inclusions in garnet porphyroblasts while $S_2$ is represented by the main matrix foliation (schistosity) as well as spiraling inclusions in garnet porphyroblasts (Fig. 4.15a and b). $S_3$ is the overprinting fabric shown by shear bands and crenulation cleavage on $S_2$ (Fig. 4.15b, c, and f). These shear bands and this micro-folding are more common at the base of Paro Formation. The distribution of bedding-parallel foliation mostly in the upper-middle section and concentration of sheared fabrics at the lower section infer the presence of a bounding fault at the base which is referred to as Paro thrust (PT). Where $S_1$ is preserved as straight inclusions in some garnet porphyroblasts, it is generally discordant to $S_2$. More importantly, $S_2$ which is the more pervasive planar fabric, is oriented parallel to the original sedimentary bedding ($S_0$). In addition, the regional foliation and isograds are folded into a regional-scale, doubly-plunging anticline defining what is known as the Paro dome. The axial plane of this dome constitutes $S_4$. These observations suggest that a regional deformation folded the regional foliation ($S_2$) and metamorphic isograds.
Figure 4.15. Photomicrographs from Paro Formation taken in plane-polarized light illustrating top-to-the-south shear sense. All thin sections cut perpendicular to quartzite bedding or tectonic foliation and parallel to mineral stretching lineation. A) Syn-tectonic garnet porphyroblast with its helicity showing top-to-the-south shear sense. B) Helicity of syn-tectonic garnet from schist interbed within the fine-grained quartzite unit. C) Microfold vergence at the southwestern edge of Paro window. D) Shear sense criteria deduced from quartz veins developed inside garnet, parallel to the greatest stress direction. E) C-S fabric and helicity of garnet inclusion trails from schist interbed within the fine-grained quartzite unit. F) Shear band at the base of the Paro Formation. Three main fabrics at the base of the Paro Formation include; S₁ defined by straight (rotated) inclusions in garnet porphyroblast, S₂ represented by schistosity (matrix foliation) and spiraling garnet porphyroblasts, and S₃ defined by shear bands and crenulation cleavage overprinted on S₂. G) Meso-scale folded lineations at the base of the Paro Formation. The hinge of the fold plunges 18° towards 350°. H) Folded upright foliations (near H) with axial planes sub-parallel to the Swiss army knife (used for scale). I) A meso-scale sheath fold at the southwestern edge of the Paro window near Betekha (latitude 27.409167°, longitude 89.413250°; hammer for scale and refer Fig. 4.2 for location).
In thin section, quartz grains in the coarse-grained quartzite unit in the upper section of Paro Formation are amoeba-shaped (multiple mm in size), with large-amplitude sutures (Fig. 4.16a-d). Grains display undulose extinction where subgrains are coalesced to form large recrystallized grains. These micro-textures are characteristics of deformation by grain boundary migration recrystallization, which occurs between deformation temperatures of ~500-700°C (Stipp et al., 2002). The crystallographic preferred orientations examined by inserting a gypsum plate, together with the alignment of micas (predominantly biotite) are parallel to S2. A sample (Fig. 4.16e and f) located ~500 m above the base of the coarse-grained unit has plastically elongated, non-recrystallized quartz clasts in the muscovite-biotite matrix. Here quartz clasts (S-fabric) are subtly inclined to the flat-lying micas (C-fabric), which together constitute a C-S fabric that shows a top-to-the-south shear sense (Fig. 4.15e).

Quartz grains in the fine-grained quartzite unit exhibit deformation microtextures that are a hybrid between SGR and GBM. In the lower part of the section, quartz grains are largely equidimensional, polygonal, and have shape preferred orientation characteristics of SGR recrystallization which occurs between ~400°C and 500°C (Fig. 4.16g). In the upper part of this unit, amoeboid grains with highly irregular and serrate grain boundaries become dominant (Fig. 16d)). These SGR and GBM microtextures together put the fine-grained quartzite unit as a transition zone that records deformation temperatures just under 500°C (Stipp et al., 2002).
Figure 4.16. Photomicrographs taken under the crossed polarized light showing quartz deformation microfabrics of the Paro Formation. A-D). Grain boundary migration (GBM: ~500°-700° C) - amoeboid grains with large amplitude sutures and very large recrystallized grain sizes. Samples A, B, C, and D are from the coarse grained quartzite unit of the Paro Formation. E) Transition between subgrain rotation (SGR) and GBM - majority of grains are equal in size and have regular grain boundaries. F) SGR - grains almost of same size. G) SGR: 400°-500° C. Grains are polygonal, have sharp boundaries and equant grain sizes. Few amoeboid shaped grains are also present. H) GBM - grain boundaries are irregular and interfingering. I) GBM - Grains are amoeboid shaped and have irregular boundaries (500°-700°C). Chessboard extinction with subgrain boundaries parallel to prism and basal planes present that constrain the minimum deformation temperature of ca. 630-650°C. These samples are arranged from the top to the bottom of the section.
In the basal section of the Paro Formation, quartz grains are (~0.5-3 mm) amoeba- with sutured grain boundaries (Fig. 4.16h) characteristic of GBM recrystallization deformation which occurs at temperatures between ~500 and 700°C (Stipp et al., 2002). In addition, several quartz grains display chessboard extinction, a type of GBM recrystallization (Kruhl, 1996; Stipp et al., 2002) characterized by square subgrains with boundaries roughly parallel to grain boundaries (Fig. 4.16i). The chessboard extinction constrains minimum deformation temperature at ~630-650°C (Stipp et al., 2002). Therefore, quartz deformation microstructures define a deformation temperature profile within the Paro Formation that with GBM in the basal section, a SGR-GBM transition zone in the middle section, and GBM in the uppermost section (Figs. 4.17 and 4.18).

Micro-scale kinematic indicators throughout the Paro Formation include c-s fabric, shear bands, asymmetric folding, mica fish, asymmetric σ-clasts, asymmetric boudins, σ-type garnet porphyroclasts, and helicity of garnet. All of these indicators consistently show top-to-the-south shear sense (Figs. 4.6 and 4.15).

Meso-scale structures and kinematic indicators in the Paro Formation include the following. In the lower section of the Paro Formation, within interbedded quartzite and schist near Sisina (Fig. 4.2), the outcrop-scale fold that plunges south has folded lineations (Fig. 4.15g). In the southwestern part of the main Paro window (latitude 27.409167°, longitude 89.413250°), a folded schist layer has axial planes sub-horizontal (Fig. 4.15h) and in the same layer but down-section, there is an outcrop scale sheath fold with sub-horizontal axial plane (Fig. 4.15i). In this schist layer, alignment of kyanite defines lineations that run parallel to hinge lines and plunge 30° towards 130°-140°.
Figure 4.17. Deformation temperature profile through the Paro Formation along different traverses in western Bhutan, as constrained by quartz deformation microstructures. Note traverses radiate from Chuzom (refer Fig. 4.2 for location). Gray boxes show a range of deformation temperatures for individual unit in each section. In the basal section, some quartz crystals exhibit chessboard extinction (CBE), a grain boundary migration deformation that occurs at ca. 630-650°C temperature. Quartz recrystallization abbreviations: SGR - subgrain rotation, GBM - grain boundary migration.
4.6.2.5. The GH Section

Three main fabrics preserved in the microstructures of GH samples include S₁, a foliation defined by inclusions in garnet porphyroblasts, S₂, the main matrix foliation (gneissosity or schistosity) defined by phyllosilicates, and S₃ a shearing and crenulation of S₂ (Fig. 4.19). S₁ is preserved primarily in the garnet zone and represents a relict early foliation that is discordant to the main matrix foliation (S₂) (Fig. 4.19a). S₂ is the most pervasive planar fabric in all zones, being schistosity in the garnet to sillimanite-muscovite zones, and gneissosity in the sillimanite-Kfs zone (Fig. 4.19b and c). Mineral growth in all zones is dominantly syn-S₂. Metamorphic isograds are generally sub-parallel to S₂. S₃ is defined by the crenulation of S₂ (Fig. 4.19c). Like the Paro Formation S₄ is later-stage folding of original layering, foliation and metamorphic isograds into both broad domes as well as shorter wavelength anticlines and synclines (Fig. 4.2).

In all examined GH samples, quartz grains display microstructures related to deformation by grain boundary migration. Quartz grains are amoeba-shaped with highly-irregular, interfingering boundaries, and multiple-mm in size (Fig. 4.19d-f) (Stipp et al., 2002). Such microstructures constrain the deformation temperatures of GH rocks between ~500°C and 700°C (Stipp et al., 2002) which overlap with the temperature estimates from mineral assemblages.

Micro-scale kinematic indicators from GH samples that include shear bands, mica fish, asymmetric folds, and asymmetric σ-clasts consistently show top-to-the-south shear sense (Fig. 4.19a, b, c, h, and i). In addition, shear sense criteria deduced
Figure 4.18. Composite stratigraphic section of the Paro Formation (~5.5 km thick). Next to the section are zonal distributions of metamorphic minerals (solid dark line represents mineral presence) and deformation temperatures obtained from quartz microfabrics.
from the helicitic inclusion trails of garnet porphyroblasts record top-to-the-south rotation of >180° (Fig. 4.19a). Thus, all kinematic indicators record one phase of south-directed ductile deformation within the GH section.

Isoclinal folds and in some case refolded folds within the GH section were observed at several localities. The lack of three dimensional exposures of these folds made it difficult to determine relative orientations of axes. Several folds that have an “eye-ball” shape in cross section and sheath-like shapes in three dimensions were found with isoclinal folds southwest of Haa (latitude 27.25158°, longitude 89.25644°). Isoclinal fold vergence, shear bands, and asymmetric leucocratic σ- and δ-clasts show top-to-the-south shear sense (Fig. 4.19h and i).

4.6.3. Balanced Cross-Sections

4.6.3.1. Rationale

Geologic mapping establishes the first-order structural framework of the Himalayan fold-thrust belt in western Bhutan. Combining field mapping with regional balanced cross-sections elucidates viable sub-surface structural geometries, quantifies possible shortening amounts by restoring deformed rocks to their original state, and provides testable kinematic scenarios for how the fold-thrust belt evolved with time.

At the Himalayan front, bedding within LH rocks, GH foliation and the major thrust faults (MBT, ST and MCT) all strike E-W. Mineral stretching lineation that generally trends N-S not only argue for a limited flow of material in and out of the plane of cross-sections, but also correspond to a maximum stretching in the N-S transport direction, which is the principal direction of the Himalayan shortening. Therefore, a cross section line (A-A’ in Fig. 4.2) that extends from south of the MFT
Figure 4.19. Photomicrographs illustrating quartz deformation microstructure (cross-polarized light) and kinematic indicators (plane-polarized light) observed in the GH section. All thin sections cut perpendicular to quartzite bedding or tectonic foliation and parallel to mineral stretching lineation. A) Top-to-the-south shear sense deduced from the helicity of garnet inclusion trails. B-C) Shear bands indicating top-to-the-south shear sense. Fabrics include; S₁ defined by inclusions in garnet porphyroblast, S₂, the main matrix foliation (gneissosity or schistosity) defined by phyllosilicates, and S₃ represented by shearing and crenulation of S₂. D-E) Amoeba-shaped grains with large amplitude and irregular grain boundaries characteristic of GBM recrystallization. F) GBM chessboard extinction in GH schist constrains minimum deformation temperatures of ca. 630-650°C. G) Sigma-clast in GH schist near the northern GH-Paro contact (latitude 27.46518°, longitude 89.51974°) shows top-to-the-south shear sense. H) GH paragneiss containing deformed leucosome and biotite-rich melanosome. Delta-clast shows top-to-the-south shear sense. I) Isoclinally folded leucosomes (hammer for scale).
to north of the outer-STD is justified because it is perpendicular to the structural trend and is parallel to N-S mineral stretching lineation.

The warping of the MCT and overlying GH foliation together with earthquake seismology, which constrains the depth and dip of the basal decollement, provides the critical first-order constraints on the geometry of the balanced cross-sections. The section has a 4° dip for the flats along the basal decollement, imparting an average ~5° dip to the Main Himalayan thrust (MHT). This geometry is constrained by the active seismic reflection from the International Deep Profiling of Tibet and the Himalaya (INDEPTH) project (Alsdorf et al., 1998; Hauck et al., 1998) and passive seismology (Singh et al., 2010; Mitra et al., 2005; Schulte-Pelkum et al., 2005).

Balanced cross-sections were constructed using the kink band method (e.g. Suppe, 1983), and conservation of line lengths of thrust sheets on the deformed and restored section (e.g. Dahlstrom, 1969). The cross section line was divided into dip domains based on the surface data and gaps in surface data were filled by calculating apparent dips from the nearest surface data and then extrapolating along-strike to their equivalent position on the line of section. The orientations of fold axial surfaces were determined by bisecting the angle between the two fold limbs with the assumption that fold limbs contain uniform dips, and are separated by distinct kinks (e.g. Suppe, 1983). A pin line was fixed in the foreland just south of the MFT as a reference point for restoration. Local folding including sheath folding visible in at outcrop-scale at the base of GH section near Haa (Fig. 4.15i) and within the Paro Formation near Betekha (Fig. 4.2) and crenulations observed in the thin section in LH, GH, and Paro rocks are impossible to accurately represent on the scale of the cross section. Therefore, no attempt was made to incorporate small-scale deformation or ductile deformation into the restoration of LH and GH rocks.
Because the positions of the hanging wall cutoffs of Subhimalayan, LH, and GH thrust sheets that have passed through the erosion surface are unknown, they pose an uncertainty in the balancing process and affect the estimation of shortening magnitudes. However, the placement of hanging-wall cutoffs just above the erosion surface is adopted as a conservative way of estimating shortening magnitudes. The northern most footwall ramp through the Paro Formation on the restored cross-sections marks the northern extent of LH units and the southern most point that GH rocks may be restored to (Fig. 4.20)

Shortening magnitudes within balanced cross-sections are dependent on: (1) the area between mapped surface geology and the depth to the decollement that needs to be filled; and (2) the thickness of unit(s) that are structurally repeated to fill that space. Thus in the case of western Bhutan where the warping of the GH and Paro Formation suggest duplex formation, similar to what is exposed to the east (Long et al., 2011c) and west (Mitra et al., 2010; Bhattacharyya and Mitra, 2011), but the geometry of that duplex is not exposed, shortening estimates depend on which units are repeated within the duplex. Due to this uncertainty we discuss three possible scenarios.

4.6.3.2. The Main Frontal thrust, Main Boundary thrust, and Lesser Himalayan thrust sheets

As stated in section 4.3, the Siwaliks are not fully exposed in western Bhutan, thus, making the location of the MFT uncertain. However, the presence of uplifted terraces ~9.0 km south of Phuentsholing provides the best estimation of the location of MFT. This location together with dips of ~40-45° for the Siwaliks in central and eastern Bhutan suggest that thickness of the Main Frontal thrust sheet is ~6.0 km,
similar to the thickness in eastern Bhutan (Long et al., 2011c). Because there is no Siwalik exposure along the cross section line, the hanging-wall cutoff is assumed to have passed through the erosion surface based on geometries further to the east. The minimum amount of slip in all three scenarios required to pass the hanging-wall cutoff through the erosion surface is measured to be ~10.0 km (Figs. 4.20 and 4.21).

The Baxa Group is in thrust contact over the Siwaliks across the MBT. The ~5-12 km N-S extent of Baxa Group exposure, combined with field observations that define fault locations that divide the mapped extent of the Baxa group into 1.3 km-thick thrust sheets that are structurally repeated ~4 times (Fig. 4.2 and 4.20). This original thickness of the Baxa group is similar to that documented in eastern Bhutan (1.5- to 2.6 km) and Sikkim (1.2 km thick) (Long et al., 2011a, Bhattacharyya and Mitra, 2009). Between the MBT and ST, five structural repetitions (horses) of the 1.3 km thick Baxa Group are interpreted (Fig. 4.20). Out of these five horses, four (#1-#4) have their hanging-wall cutoffs above the erosion surface while one (#5) is inferred in the subsurface. The folding observed in the overlying Daling-Shumar Group is explained by the presence of Baxa horse #5 below the surface. The geometry of the Baxa duplex (upper LH) is hinterland dipping, with four surface breaking horses dipping between 30° and 40° to the north.

The lower Lesser Himalayan thrust sheet exposed over a N-S distance of ~5.0 km is bound below by the ST and above by the MCT, and is comprised of the Daling-Shumar Grou and Jaishidanda Formation which have a combined thickness of ~4.0 km (Figs. 4.20 and 4.21). This thrust sheet represents a simple stratigraphic section that is not repeated by faults. The length of the Shumar-Daling Group shown above the erosion surface is roughly equivalent to length of its footwall cutoff located in the sub-surface. This thrust sheet that was initially sub-parallel to the Baxa duplex, only
to be passively folded by motion on Baxa horse #5 which postdated motion on the ST (Fig. 4.20).

The exposure of the Paro Formation in the southern window near Chukha requires and out-of-sequence fault that places Paro over GH rocks. The magnitude of shortening along this thrust is the minimum required to expose the Paro Formation at the surface. Because it is out-of-sequence, the structure must project from the regional, active decollement to the exposed fault that bounds the southern window.

4.6.3.3. The Paro duplex

The amount of area between the mapped surface geology and the basal decollement and how that space is filled have implications on shortening estimates in any thin-skinned fold-thrust belt. No drill-hole or seismic reflection data are available across the western Bhutan Himalaya to guide how to best fill the space above the basal decollement. However, the geometry and strata involved in exposed duplexes to the east and west provide insights into realistic solutions. We highlight three end-member scenarios and assess the pros and cons associated with filling the space with (1) the Baxa Group, (2) the Shumar-Daling Group and (3) the Paro Formation. In all three scenarios, the minimum MCT overlap measures 174 km and the geometry of the front of the fold-thrust belt remains the same.

Scenario I: Filling the space by duplexing of the Baxa Group: In this scenario, the Baxa Formation (1.3 km thick) is repeated to fill the space beneath the Paro anticlinal window (Fig. 4.20a). Since the ST overlies the Baxa Group in the foreland, it must also overlie the duplex of Baxa group under the Paro Window. Thus, the structurally-repeated Baxa Group forms a duplex system with the ST as the roof thrust. This geometry implies that the ST equivalent to Ramgarh thrust in Nepal and
Sikkim, and requires that duplexing of the Baxa Group passively folds and post-dates the MCT and PT but concurrent to the ST. In this scenario the eleven Baxa horses in the hinterland are imbricated first and in-sequence, and then the imbrication of the remaining five foreland horses with the last horse being the MBT and placing the Baxa group over the Siwalik Group. In the hinterland, the MCT and TH section are passively folded by late duplexing of the Paro Formation. These later thrusts feed slip into the folded (and possibly eroded) MCT. We propose that this extra slip is transferred to the surface as out-of-sequence motion on the MCT, possibly in the area immediately north of the Paro window. Because the Baxa Group is comparatively thinner, its restored length extends well into the distal part of the passive margin. As required by balanced sections, the ramp that places the Daling-Shumar Group over the Baxa Group must be north of the total restored length of Baxa group horses. This implies that the underlying Daling-Shumar Group should restore to a distal part of the passive margin.

Due to the thickness of Baxa horses (1.3 km) this scenario produces the maximum shortening amongst the three possible scenarios (Table 4.4). This estimate is dependent on how well the Baxa thickness is constrained. The measured thickness and the interpretation of its structural repetition is consistent with observations in eastern and central Bhutan (Long et al., 2011a) as well as in Sikkim (Bhattacharyya and Mitra, 2009; Mitra, 2010). The total shortening for this scenario is 566 km.

Scenario II: Filling up space by duplexing the Daling-Shumar Group: As described in section 4.3, the Shumar-Daling Group together with the Jaishidanda Formation measures ~4.0 km in thickness. Filling the space by duplexing of the Shumar-Daling Group restricts the Baxa Group to only those thrust sheets identified in the foreland, thereby dramatically reducing the amount of shortening
accommodated by Baxa Group duplex (Table 4.4). Shortening within the Daling-Shumar Group is divided among 4 thrust faults. There are three Shumar-Daling horses interpreted beneath the Paro window, and then the ST sheet, which places the Shumar-Daling Group over the Baxa group in the foreland (Fig. 4.20b). Emplacement of the Shumar-Daling horses beneath the Paro Window passively folds and post-dates both the MCT and the PT, but predates motion on the ST proper. As in the case of scenario I, in the hinterland, the MCT sheet and TH rocks are passively folded by late duplexing of the Paro Formation, with out-of-sequence motion on the MCT (Fig. 4.20b). Although filling the space by the Shumar-Daling Group does not violate any surface observations, there are implications for the kinematics of the shortening and shortening magnitudes. The largest implication is for the timing of duplex formation and continual motion on the MCT. The roof thrust for the Shumar duplex is the PT, which tips out and joins the MCT below the syncline south of Chukha (Figs. 4.2 and 4.20b). Space constraints beneath the PT allow only the Shumar duplex without requiring a Shumar roof thrust (the geometry of the ST described for the baxa duplex). Thus with the PT and MCT as the roof thrust for the duplex, the southern most outcrop of the MCT would have remained an active thrust while the duplex was forming. The ST and the 5 Baxa horses are then the youngest structures in the system. The total shortening for the cross section is ~507 km.

**Scenario III: Filling up space by duplexing of the Paro Formation:** Filling the space by duplexing the Paro Formation limits the restored lengths of Baxa and Shumar-Daling Groups and minimizes the amount of shortening (Fig. 4.20c). This is because the Paro Formation is thicker and thus requires fewer horses to fill the space. Although filling the space by duplexing the Paro Formation does not violate surface observations, there are implications; the first is that the magnitude of shortening
within the Baxa and Shumar-Dahling Groups is dramatically reduced while shortening within the Paro Formation is increased (Fig. 4.20b). This reduction in shortening translates into a significant reduction in the across-strike width of the restored Baxa and Shumar-Daling basins in western Bhutan. The second implication is that formation of the structural high that becomes the Paro Window predates motion on the Shumar thrust, as in scenario b. Thus the frontal expression of the MCT is active while the duplex forms. The total shortening is estimated at 466 km, which is obtained from scenario III, and represents the least amount out of three scenarios (Table 4.4).
B) Scenario II: Filling space by the Shumar-Daling Group

Notes for Shumar-Daling Duplex
i. Length of eroded Shumar-Daling that has passed through erosion surface is equal to its footwall cutoff.
ii. Length of eroded GH section that has passed through erosion surface is equal to footwall cutoff of the Paro Formation.
iii. Shumar-Daling horses in the lower LH duplex feed slip onto horse #4 which feeds slip onto the Paro thrust. All of the shortening in the Shumar duplex (except for the Shumr thrust) must predate burial of the Jaishidanda Formation at 15 Ma.
iv. Paro horses in the hinterland feed slip onto the MCT.
C) Scenario III: Filling space by the Paro Formation

Notes for Paro Formation Duplex
i. Length of eroded Shumar-Daling that has passed through erosion surface is equal to footwall its cutoff.
ii. Length of eroded GH section that has passed through erosion surface is equal to footwall cutoff of Paro Formation.
iii. The Paro horses in the hinterland feed slip onto the MCT. All of the shortening in the Paro duplex must predate burial of Jaishidanda Formation at 15 Ma.

Notes common to all three cross sections (i.e. A, B, and C)
i. Location of the MFT taken from Gansser (1983) based on uplifted river terraces.
ii. 4°N dip along the basal decollement is based on reflection seismology from the International Deep Profiling of Tibet and the Himalaya (INDEPTH) project (Alsdorf et al., 1998; Hauck et al., 1998) and passive seismology (Singh et al., 2010; Mitra et al., 2005; Schulte-Pelkum et al., 2005).
iii. Lighter shading on deformed cross section represents material that has passed through the erosion surface.
iv. The Jaishidanda Formation is shown pinching out to south because it is not observed in the upper LH section in the foreland. It is also shown to progressively thicken in the hinterland and transitions into the Paro Formation.
v. Length of the eroded Shumar-Daling Group above the erosion surface is at least equal to its footwall cutoff.
vi. The MFT displacement (~10 km) is the amount necessary to move Siwalik Group hanging wall cutoff through
Figure 4.20. Balanced cross section of western Bhutan fold-thrust belt drawn along the section line A-A’ in Fig. 4.2, showing the space beneath Paro window being filled by: A) the Baxa Group, B) the Shumar-Daling Group, and C) the Paro Formation. No vertical exaggeration.
Table 4.4. Shortening and fault displacement estimates for western Bhutan.

<table>
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<th>Baxa duplex</th>
<th>Shumar-Daling duplex</th>
<th>Paro duplex</th>
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Subhimalaya

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Shortening in Baxa Group

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<td>Total shortening (%)</td>
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Shortening in Shumar-Daling Group

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Shortening in Paro Fm.***

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<td>Total shortening (%)</td>
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Paro Thrust

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<td>Min. Paro thrust overlap (km)</td>
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Greater Himalaya

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<td>Min. MCT overlap (km)</td>
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<td>Total shortening (%)</td>
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<td>MBT displacement (km)</td>
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<td>ST displacement (km)</td>
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<td>PT displacement (km)</td>
<td>50.0</td>
<td>87.0</td>
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* LH length and shortening estimates include Subhimalaya
** Includes restored length from LH-SH and MCT overlap
*** Does not include Paro thrust
4.7. DISCUSSION

4.7.1. Timing constraints

Th-Pb geochronology of metamorphic monazite is used to place constraints on the timing of peak metamorphism and age of displacement on the MCT and STDo. Ages obtained from high Th (low Y) cores indicate that prograde metamorphism in the immediate hanging-wall of the MCT was underway by ca. 23.1±0.9 Ma and that peak metamorphism was reached at ca. 20.4±1.0 Ma (2σ uncertainty) (Tobgay et al., in revision). In addition, ages obtained from high Y overgrowths suggest crystallization of in-situ melts at 15.1±0.4 Ma, which continued until 10 Ma. A sample from the Jaishidanda Formation, in the immediate footwall of the MCT, argues for prograde metamorphism of LH rocks at 15 Ma. A combination of prograde monazite ages in the hanging-wall and footwall of the MCT brackets the age of the south-directed MCT displacement between 20.4±1.0 and 15.1±0.4 Ma. In addition, a combination of prograde (high Th cores) and retrograde (high Y rims) ages obtained from metamorphic monazite in the immediate footwall of the STDo immediately north of the study area, constrain the timing of north-directed motion on the STDo at ca. 20-15 Ma (Kellett et al., 2010). Despite the coeval nature of the MCT and the STDo, the presence of retrograde monazite as young as 10 Ma in the GH section suggests that GH rocks continued cooling for 5 Myr even after the motion on the STDo ceased. This cooling of GH rocks is either a result of continued displacement on the MCT, which is not compatible with the Daling-Shumar Group or Paro Formation duplex models, or cooling due to duplex growth and increased exhumation without requiring motion on the MCT (Tobgay et al., in revision).
4.7.2. Thrust kinematics in western Bhutan

Sequential reconstruction, which illustrates the development of the south-vergent fold-thrust system in western Bhutan, is important in order to link deformation and metamorphism. GH rocks, which restore much further into the hinterland than LH rocks, have undergone metamorphism due to structural burial from crustal shortening and thickening in TH rocks (e.g. Hauck et al., 1998; Hodges, 2000; DeCelles et al., 2002; Murphy and Yin, 2003). This metamorphism is recorded by index minerals (garnet, staurolite, kyanite, sillimanite, and sillimanite-k-feldspar) and by chemical zoning of monazite grains (characteristic high Th content in the core). The development of the most pervasive fabric (S\textsubscript{2}), which is represented by schistosity in the garnet to sillimanite-muscovite zones and gneissosity in the sillimanite-kfs zone, is coeval with the progressive metamorphism. Therefore, the growth of the index minerals and prograde monazite within the GH section is syn-S\textsubscript{2}. Prograde monazite ages indicate that prograde metamorphism occurred prior to peak metamorphism (partial melting) at least by 20 Ma. The peak metamorphism was followed by a phase of deformation where hot (at least 700-750°C) GH rocks were emplaced over lower temperature Paro Formation by thrusting as well as shearing along the MCT at ~20 Ma. This high temperature deformation within the GH section is supported by the presence of quartz deformation microfabrics characteristic of GBM deformation at temperatures between 500°C and 700°C (Stipp et al., 2002). In addition, this deformation resulted in decompression and cooling of GH rocks, thus, driving monazite growth with high-Y overgrowths. Therefore, prograde monazite grains in GH rocks are pre-kinematic, while retrograde monazite grains are syn-kinematic to possibly post-kinematic with respect to deformation. This retrograde
monazite growth is accompanied by the development of another fabric (S3) in the form of shearing and crenulation of S2.

Metamorphism (staurolite-kyanite-sillimanite grade) and the development of a tectonic fabric (S2) within the Paro Formation are due to structural burial associated with the emplacement of hot GH rocks (Figs. 4.21-4.23). Simultaneously, this structural burial by hot GH rocks would have triggered prograde monazite growth in the Paro Formation (although no Th-Pb ages of monazite were obtained). Quartz deformation microstructures indicate deformation temperatures of ~500-700°C in the lower and upper sections of Paro Formation while in the middle section, quartz deformation microstructures indicate ~400°-500°C deformation temperatures. This deformation temperature profile of high temperature in the basal section, comparatively lower temperature in the middle, and then back to high temperature in the uppermost part of the section, suggests heating of the upper Paro Formation due to emplacement of hot GH rocks via the MCT, while heating of the lower Paro was due to the accompanying burial of the section. The concentration of S3 (shear bands and crenulations) overprinted on S2 at the base of the Paro Formation section associates the development of S3 with deformation along the PT. Displacement along the PT placed the Paro Formation and the over-riding GH section over the Jaishidanda Formation (Figs. 4.21-4.23). Thus, the deformation and metamorphism observed in the Jashidanda Formation is related to the combined emplacement of the MCT and the PT. This drove prograde metamorphism in the Jaishidanda Formation and syn-kinematic growth of garnet and monazite. Spiraling inclusion trails inside garnet porphyroblasts and Th-rich monazite (15 Ma) in the Jaishidanda Formation suggest contemporaneous growth of these minerals with the emplacement of the MCT and PT. This deformation led to the development of S3 (shear bands and crenulations), the
overprinting fabric on S$_2$. In addition, quartz deformation microstructures show transitional texture between SGR and GBM recrystallization, which occurs at an approximate temperature of ~500°C.

In the later stage, displacement on the ST placed the Daling-Shumar Group over the Baxa Group in the foreland post-15 Ma (Figs. 4.21-4.23). The timing for the development of the Baxa foreland duplex system depends on which of the three possible geometric models is used. For the Baxa duplex, the duplex forms post-initial emplacement of the ST, after 15 Ma (Fig. 4.22). Duplex development could be responsible for the 15-10 Ma retrograde cooling of monazite. The Shumar Duplex model and the Paro duplex model both require MCT, PT and duplex formation all to be concentrated between 20 and 15 Ma (Figs. 4.22 and 4.23). Because space constraints beneath the PT make it geometrically difficult to have a Shumar roof thrust for the Daling-Shumar Group duplex, and impossible for the Paro duplex, in these models growth of the duplex is linked to southward motion on the PT and the MCT. Thus the burial of the LH rocks by the MCT would be towards the end of duplex formation. Duplex formation in the Shumar-Daling or Paro duplex models, which would argue for continued displacement on the MCT, theoretically provides a mechanism for continued cooling to 10 Ma. However, motion on the horses is required before burial of the Jaishidanda Formation at ~15 Ma.
Figure 4.21. Sequential cross-section restoration illustrating the emplacement of GH rocks along the MCT in western Bhutan for scenario I (filling space underneath the Paro window (PW) by Baxa duplex). a) Pre-20 Ma distribution of LH and GH protoliths. b) Displacement along the MCT between 20-15 Ma that places GH rocks over the Paro Formation. c) Continuing motion on the MCT and displacement along the PT together bury Jaishidanda Formation at 15 Ma. d) Initiation of displacement along the ST and development of Paro duplex in the hinterland between 15 and 10 Ma. e) Geologic cross-section (A-A’ in Fig. 4.2) across western Bhutan.
a) Pre-20 Ma

b) 20-17 (?) Ma: MCT + PT

c) 17 (?)-15 Ma: Development of Shumar-Daling duplex

d) Present
Figure 4.22. Sequential cross-section restoration illustrating the emplacement of GH rocks along the MCT in western Bhutan for scenario II (filling space underneath the Paro window (PW) by Shumar-Daling duplex). a) Pre-20 Ma distribution of LH and GH protoliths. b) Displacement along the MCT between 20-17 (?) Ma that places GH rocks over Paro Formation followed by displacement along the Paro thrust (PT) that places Paro Formation over Jaishidanda Formation. c) Development of Shumar-Daling duplex from ~17-15 Ma allows for continuous motion on the MCT and PT to bury the Jaishidanda Formation at 15 Ma. d) Initiation of displacement along the ST, development of Baxa duplex, and development of Paro duplex in the hinterland between 15 and 10 Ma; Geologic cross-section (A-A’ in Fig. 4.2) across western Bhutan.
a) Pre-20 Ma

b) 20-17 (?) Ma: MCT + PT

Jaishidanda sample
(BU10-79)

GH sample (BU10-74, 81)
GH sample (BU10-61, 62)

Hinterland Paro duplex

17 (?) -15 Ma: Development of frontal Paro duplex and burial of Jaishidanda by GH

d) 15-0 Ma: ST, Baxa duplex, ST + hinterland Paro duplex
Figure 4.23. Sequential cross-section restoration illustrating the emplacement of GH rocks along the MCT in western Bhutan for scenario II (filling space underneath the Paro window (PW) by Paro duplex). a) Pre-20 Ma distribution of LH and GH protoliths. b) Displacement along the MCT between 20-17 (?) Ma that places GH rocks over the Paro Formation followed by displacement along the PT that places Paro Formation over the Jaishidanda Formation. c) Development of Paro duplex from ~17-15 Ma allows for continuous motion on the MCT and PT to bury the Jaishidanda Formation at 15 Ma. d) Initiation of displacement along the ST, development of Baxa duplex, and development of Paro duplex in the hinterland between 15 Ma to 10 Ma; Geologic cross-section (A-A’ in Fig. 4.2) across western Bhutan. Legend same as Figs. 4.21 and 4.22.
4.7.3. Kinematic scenarios and rates of shortening

Each of the three scenarios presented above for filling the space beneath the Paro Window have different magnitudes of shortening that must be accomplished between 20.4±1.0 and 15.1±0.4 Ma. In the first scenario (duplexing of Baxa Group horses), the magnitude of shortening achieved between 20.4±1.0 and 15.1±0.4 Ma is ~230 km and is taken as the sum of displacement lengths on the MCT and the PT. Considering ~230 km displacement in 5.4±1.1 Myr, the shortening rate is calculated between 35 and 55 mm/yr (Tobgay et al., in revision). Post 15 Ma the rate of shortening averages 22 mm/yr. In the second scenario (duplexing of Daling-Shumar Group horses), the magnitude of shortening is taken as the sum of displacements on the MCT, PT, and the shortening in the Shumar-Daling duplex, and equals ~350 km. Based on this displacement amount of 350 km in 5.4±1.1 Myr, shortening rates are estimated between ~54 and ~82 mm/yr. After 15 Ma, the rate drops to 9.7 mm/yr. In the third scenario (duplexing of Paro Formation horses), the magnitude of shortening is taken as the sum of displacements on the MCT, PT, and the shortening in the Paro duplex. In this case, the shortening magnitude of 375 km achieved in 5.4±1.1 Myr yields shortening rates between 58 and 87 mm/yr. In this scenario, the post-15 Ma shortening rate drops to 9.4 mm/yr.

The above calculations show varying shortening rates inherent to each scenario with minimum and maximum rates at 35 mm/yr and 87 mm/yr, respectively. The upper limit on the rate of motion on the MHT at this time is the plate convergence rate between India and Asia. Plate tectonic reconstructions indicate that between 20 and 15 Ma the eastern syntaxis of India was converging with Asia at a rate of 57±4 mm/yr (van Hinsbergen et al., 2011). Because of this, filling the space by duplexing Baxa Group horses underneath the Paro Formation is the preferred
geometry. This is because: (1) shortening rates of 35-55 mm/yr are the minimum out of the three scenarios; (2) displacement on the MCT, PT and duplex formation are not concentrated between 20-15 Ma, which creates a huge disparity in shortening rates before and after 15 Ma; and (3) it matches the structural repetition of the Baxa Formation observed in a similar tectonic window to the west in Sikkim. The actual geometry of the duplex may be a hybrid between the Baxa and Shumar scenarios with one or two horses of the Shumar-Daling Group in the northern portion of the duplex similar to the duplex in eastern Bhutan (e.g. Long et al., 2011a) or in Sikkim (e.g. Bhattacharyya and Mitra, 2009). A hybrid solution suggests that shortening in western Bhutan probably is between the estimate for the Baxa Group and Daling-Shumar Group duplex scenarios (i.e. most likely between 566 and 507 km), with long-term shortening rates of 25-28 mm/yr from 20 Ma to present.

4.7.4. Shortening variations

In western Bhutan, total shortening (taken as the difference between restored and deformed lengths) of the Indian margin, calculated from the three scenarios defined above (Table 4.4), ranges between ~466-566 km, with percent shortening between ~72-77%. Depending on which scenario is considered, long term shortening rates may vary between 23.3 to 28.3 mm/yr. In comparison, shortening in eastern Bhutan ranges between 344 (15.6 mm/yr) km and 405 km (18.4 mm/yr), with percent shortening that ranges from ~70-75% (Long et al., 2011a). In comparison, total shortening of 502 km is estimated in Sikkim out of which 246 km is taken up by Rangit and Daling duplexes (Bhattacharyya and Mitra, 2009, 2011; Mitra et al., 2010). 502 km of shortening in 22-20 Myr yields a long-term shortening rate of ~25-23 mm/yr. While shortening and percent shortening show no obvious east to west
trend, the minimum MCT overlap show a systematic along-strike variation even within a distance of ~250 km in Bhutan. The minimum MCT overlap in all three scenarios in the cross section of western Bhutan presented here measures 174 km, compared to ~97-156 km measured along four cross-section lines in eastern Bhutan (Long et al., 2011a). MCT overlap shows an increasing trend from the far-east to the west. The minimum MCT overlap measures 97 km in the far-east and steadily increases westwards to 107 km, 140 km, and 156 km within a distance of ~100 km (Long et al., 2011a).

4.7.5. Along-strike variations in shortening kinematics

The east-west variation of shortening estimates described in the preceding section may be linked to variations in shortening kinematics along-strike from eastern Bhutan to Sikkim. In eastern Bhutan, a minimum length of 80-120 km of the GH section is directly emplaced over the Jaishidanda Formation by the MCT (Long et al., 2011c) at 23-20 Ma with continued motion as young as ~18 (Daniel et al., 2003; Chambers et al., 2011). Continued motion along the frontal part of the MCT was accompanied by shortening within the lower LH duplex (15-70 km) until ~17 Ma. The ST placed the Daling-Shumar Group, Jaishidanda Formation, and the over-riding GH section over the Baxa Group at ~17-15 Ma. Between 14 Ma and 11 Ma, out-of-sequence motion on the Kakthang thrust (KT) in the hinterland in central and eastern Bhutan emplaced a hotter section of GH rocks on top of a comparatively colder section of GH rocks (Grujic et al., 2002; Daniel et al., 2003). This out-of-sequence motion on the KT overlapped with growth of the upper LH duplex at ~15-10 Ma. From 10-0 Ma shortening is accommodated by the thrust sheets in the upper LH (Diuri and Gondwana,) and the Siwaliks (Long et al., 2011c).
The presence of the KT in eastern and central Bhutan and presence of the PT in only the west require different kinematic models for eastern and western Bhutan. While the combined MCT and PT shortening rates for western Bhutan (35-45 mm/yr) are comparable to MCT rates for eastern Bhutan (27-40) mm/yr, the timing of motion is offset by ~3-4 million years. During rapid MCT motion in western Bhutan, eastern Bhutan was shortening at a much slower rate (~15 mm/yr) during the development of the lower LH duplex and emplacement of the Shumar Thrust. If the rapid cooling of retrograde monazite in western Bhutan from 15-9 Ma is a result of growth of the Paro duplex (Fig. 4.21), then from 15-9 Ma both eastern and western Bhutan rapidly shortened via duplex formation at rates of 34-40 mm/yr over this window of time.

In the Sikkim Himalaya, shortening is accommodated by motion on the MCT1, MCT2 (equivalent to the PT), RT (Ramgarh thrust), MBT, MFT, Daling duplex, and Rangit duplex (Bhattacharyya and Mitra, 2009, 2011; Mitra et al., 2010). Kinematic evolution began with emplacement of the GH section over the Paro-Lingtse gneiss at ~20-15 Ma along the MCT1 (Catlos et al., 2004; Bhattacharyya and Mitra, 2009, 2011; Mitra et al., 2010). Following the displacement on the MCT1, the Paro-Lingtse gneiss (equivalent to the Paro Formation) and the frontal part of the GH rocks were emplaced over the Daling Formation along the MCT2 (PT) at ~15-10 Ma (Catlos et al., 2004; Bhattacharyya and Mitra, 2009, 2011; Mitra et al., 2010). The rate of shortening over this window of time is ~24-25 mm/yr. Post 10 Ma, during the formation of the Daling and Rangit duplex, shortening rates may have increased slightly to 26 mm/yr. More detailed thermochronometry is needed to confirm whether the Sikkim Himalayas deformed at near constant rates of ~23-25 mm/yr or if there were pulses of rapid shortening as documented in Bhutan.
4.8. CONCLUSIONS

From geologic mapping and balanced cross-sections through the western Bhutan Himalaya, in combination with the new metamorphic and structural data presented here, the following conclusions about metamorphism and deformation in western Bhutan can be made:

i. A complete sequence of the Barrovian metamorphic zones are present, with lower greenschist facies conditions in the upper Daling-Shumar Group, which transitions into upper greenschist facies conditions in the Jaishidanda Formation and lower GH section, and then to amphibolite-granulite conditions in the higher structural levels of GH section. The transition between the LH and GH rocks is characterized by a progressive but a rapid increase in metamorphic grade which coincides with the MCT. Metamorphism in GH rocks, which is coeval with the development of the most pervasive fabric ($S_2$), continued until peak metamorphism and partial melting at 20 Ma.

ii. Pressure and temperature conditions in each metamorphic zone include ~5-7 kbar and ~400-500°C in the biotite zone, ~5-12 kbar and ~550-650°C in the garnet zone, ~6-9 kbar and 570-675°C in the staurolite-kyanite zone, and ~5-9 kbar and ~570-700°C in the sillimanite zones. At the higher structural levels of GH section in the sillimanite-K-feldspar zone, the presence of deformed kyanite together with melt constrains the peak pressure and temperature of >8 kbar and >700°C. This peak metamorphic mineral assemblage is later overprinted by growth of sillimanite + K-feldspar that constrains pressure to 3-4 kbar. The inversion of metamorphic field gradient in western Bhutan is mainly a function of temperature that increases from ~400-500°C at the lower section (Shumar-Daling Formation) to 700-750°C in the higher structural levels of GH section.
iii. In higher structural levels of the GH section (i.e. sillimanite-k-feldspar zone), the initial peak metamorphic reaction has been overprinted by the following reactions: (1) garnet + muscovite = sillimanite + biotite, which is characteristic of high temperature (700-750°C) decompression (to ~5.5 kbar); and (2) garnet + kfs + melt = sillimanite + biotite characteristic of a retrograde reaction (cooling).

iv. The Paro Formation has experienced lower to upper amphibolite facies metamorphism with P-T conditions of ~5-12 kbar and 575-700°C at the base and ~4-7 kbar and 575-700°C in the upper section. Metamorphism and high temperature deformation in the Paro Formation is interpreted to be related to structural burial by hot GH rocks along the MCT as well as motion on the Paro thrust that emplaced the Paro Formation and the over-riding GH rocks on top of the Jaishidanda Formation.

v. Peak metamorphism in the GH section was followed by high temperature deformation where hot GH rocks were emplaced over the lower temperature Paro Formation along the MCT at ~20 Ma. The continued motion on the MCT placed the frontal part of the MCT and the over-riding Paro Formation and GH section over the Jaishidanda Formation at ~15 Ma. Consequently, the emplacement of hot (~500-700°C) GH rocks heated the upper part of the Paro Formation and the Jaishidanda Formation. The Baxa and Shumar-Daling groups were deformed at much lower temperatures of ~280-400°C and 400-500°C, respectively.

vi. Major structural features of the frontal part of the fold-thrust belt in western Bhutan include: (a) the 6.0 km thick Main Frontal thrust sheet; (b) the Baxa duplex system which structurally repeats ~1.3 km thick horses of Baxa Group with the Shumar thrust as the roof thrust; (c) the Shumar-Daling duplex, which repeats ~4.0 km thrust sheets of Shumar-Daling Group; (d) the ~7.0 km thick structurally lower
Greater Himalayan thrust sheet above the MCT and below the outer-South Tibetan detachment.

vii. There is a variation in the amount of shortening calculated for western Bhutan, depending on how the space beneath the Paro window is filled, whether by a duplexing Baxa Group, Daling-Shumar Group, or Paro Formation horses. In all three end-member solutions, 292-392 km of minimum shortening or 62-69% is recorded in the Subhimalayan and Lesser Himalayan zones, but is partitioned differently among the different formations. The minimum MCT overlap in each scenario is ~174 km and the minimum PT overlap is between 50-93 km. Total minimum shortening between the MFT and the outer South Tibetan detachment in western Bhutan is 466-566 km or 72-77%. The Subhimalayan zone only accounts for 10 km, or ~2%.

viii. For each of the three duplexing scenarios, the amount of shortening achieved between 20.4±1.0 Ma and 15.1±0.4 Ma ranges between 250 and 365 km, which yields a range of shortening rates between 35 and 87 mm/yr. Due to the huge variation in shortening rates, it is proposed here that filling the space underneath the Paro Window by duplexing Baxa Group horses is the most likely scenario because: (1) it yields minimum shortening rates of 35-55 mm/yr, which are less than or equal to plate tectonic rates; (2) under this scenario, displacement on the MCT, PT, and shortening associated with Baxa Group duplex formation is not concentrated between 20-15 Ma, which negates the large disparity in shortening rates before and after 15 Ma introduced under the other two scenarios; and (3) this scenario best matches the structural repetition of the Baxa Group mapped in a similar tectonic window to the west in Sikkim.

ix. The actual geometry of the duplex does not have to exclusively involve Baxa Group horses. Instead, the geometry may contain one or more horses of the Shumar-
Daling Group in the northern portion of the duplex system, which is similar to
duplexes observed in eastern Bhutan and Sikkim. This suggests that shortening in
western Bhutan is probably between the estimates for duplexing of Baxa Group and
Daling-Shumar Group horses, with long-term shortening rates of 25-28 mm/yr from
20 Ma to present.

A compilation of shortening and shortening rates estimated across the Himalayan
fold-thrust belt between eastern Bhutan and Sikkim indicates variation without a
systematic trend. In addition, there are variable rates of convergence and differences
in structural geometry and shortening kinematics across the Bhutan and Sikkim
Himalaya. These may imply variability of strain rate on timescales of Myr even
within an along-strike distance of ~250 km.
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Appendix A (Accompanies Chapter 2)

This appendix is published as an online supplementary data file, which accompanies the paper presented in Chapter 2. It can be accessed in the online version of this paper at doi: 10.1029/2009TC002637. The complete citation is:


A.1. U-Pb geochronology

A.1.1. Methods of Arizona LaserChron Center U-Pb zircon dating

U-Pb geochronologic analyses were conducted on individual detrital zircon grains using laser-ablation multicollector inductively coupled plasma-mass spectrometry (LA-MC-ICP-MS) housed at the University of Arizona LaserChron Center. We analyzed 6 detrital samples and 2 igneous samples from the Paro Formation in western Bhutan. U-Pb age spectra of detrital zircons (DZ) and igneous crystallization ages are shown in age probability plots in Fig. A-1, and data for individual analyses are listed in Table A-2 (1σ). Sample locations are listed in Table A-1, map locations are shown on Figs. 2.1 and 2.2 and stratigraphic locations are shown on Fig. 2.3a.
Rock samples were crushed and pulverized to sand-size grains, separated by
density on a Wilfley table, and then separated into dense and light fractions by a 3.32
g/cc liquid separation. The dense fraction was passed through a Frantz magnetic
separator, and the zircons were mounted in epoxy plugs and then polished to half-
thickness. Photographic images were made of all samples to keep track of which
grains were dated.

Material was ablated from the sample surface using a DUV193 Excimer laser
system from New Wave Instruments. The laser operates at a wavelength of 193 nm,
and for all the samples (BU07-59, -60, -73, -76, -77, -79, -83, -84, and BU08-128) in
this study 35 micron-wide spot sizes were used, except for sample BU07-75 which
contained smaller zircons (80 grains) and was hit with a 25 micron-diameter beam
(see notes below Table A-1). For most analyses the laser was operated at minimum
output energy (~40 mJ) with a repetition rate of 8 pulses per second, which created a
~15 micron-deep pit for a typical 20 second analysis. The ablated material is carried
in helium gas into the plasma source of a multicollector inductively coupled plasma
mass spectrometer (an Isoprobe, from GV Instruments). This instrument is equipped
with nine moveable Faraday collectors and four low-side Channeltrons (ion counters).
Eight of the Faraday collectors use a $10^{11}$ ohm resistor, whereas the Faraday used for
measuring $^{207}$Pb is equipped with a $10^{12}$ ohm resistor. This configuration allows
static-mode measurement of all isotopes, using $10^{11}$ Faraday detectors for $^{238}$U, $^{232}$Th,
$^{208}$Pb, and $^{206}$Pb, a $10^{12}$ Faraday detector for $^{207}$Pb, and an ion-counting channel for
$^{204}$Pb. Each analysis consists of one 20-second integration on peaks with the laser off
(for backgrounds), 20 or 12 one-second integrations with the laser firing, and a 30
second delay to purge the previous sample and prepare for the next analysis. Each
analysis is evaluated for consistency of $^{206}$Pb/$^{238}$U and $^{206}$Pb/$^{207}$Pb ratios through the
20 seconds of data acquisition. If ratios display either a sudden change, or a gradual increase greater than ~5% for $^{206}\text{Pb}/^{238}\text{U}$, the analysis is discarded. This ensures that analyses are not compromised by crossing an age boundary.

A.1.2. U-Pb zircon data

The 893 U-Pb zircon analyses that yielded less than 30% isotopic discordance are shown for each sample in Figures 2.3, 2.4, and 2.7 in relative age probability plots (1σ errors), and data and measurement (analytical) errors (1σ) for individual analyses are listed in Table A-1, and shown in Pb/U concordia plots for each sample in Figure A-1. This large range of accepted discordance is justified because we interpret clustering as a more powerful tool than concordance for determining the reliability of ages, given that both Pb-loss and inheritance commonly move analyses along concordia. Therefore, single concordant analyses do not necessarily yield robust ages. In contrast, analyses that yield a cluster of ages are more likely robust, even if slightly to moderately discordant, because Pb loss and inheritance will always tend to increase scatter. Therefore, in this study we accept analyses with discordance up to 30%, and place the most significance on clusters (peaks) supported by at least three analyses. Peaks defined by only one or two analyses are interpreted as less significant.

Common Pb correction was accomplished by using the measured $^{204}\text{Pb}$ and assuming an initial Pb composition from Stacey and Kramers [1975]. Conservative uncertainties of 1.0 for $^{206}\text{Pb}/^{204}\text{Pb}$, 0.3 for $^{207}\text{Pb}/^{204}\text{Pb}$, and 2.0 for $^{208}\text{Pb}/^{204}\text{Pb}$ were used for the composition of the common Pb. $^{204}\text{Hg}$ present in the argon plasma gas, as well as any background $^{204}\text{Pb}$ or molecular 204, was accounted for by first measuring backgrounds in the 204 mass position, then measuring the peak 204
intensity with the laser firing, and subtracting the background intensity from the peak intensity.

Fractionation of Pb/U and Pb/Th occurs primarily in the laser pit, and is highly sensitive to the rate of carrier gas flow across the sample surface. An optimal balance between signal intensity and stability occurs at a carrier gas flow rate of 0.45 ml/minute, which generates a Pb/U sensitivity of 0.9 (e.g., a 500 Ma zircon yields a \(^{206}\text{Pb}/^{238}\text{U}\) age of 450 Ma). To correct for Pb/U and Th/U fractionation, standards were analyzed once every 5 unknowns. Fractionation standards for zircon are fragments of a large Sri Lanka zircon crystal that yields an age of 563.5±3.2 Ma (2-sigma, ID-TIMS) [Gehrels et al., 2008]. The unknowns are corrected for the closest 6 standards using a sliding window average. The error on this fractionation factor is generally ~1% (2-sigma) for \(^{206}\text{Pb}/^{238}\text{U}\) ages. Fractionation of Pb isotopes is minimal, with a maximum of ~3% fractionation of \(^{206}\text{Pb}/^{207}\text{Pb}\). This fractionation is also removed by comparison with standards, using the same procedure described above. The error on this fractionation factor is generally ~1% (2-sigma) for \(^{206}\text{Pb}/^{207}\text{Pb}\) ages. Pb/U and Pb/Th fractionation varies with depth during laser ablation, increasing by ~5% during a 20-second analysis that excavates to a depth of 15 microns. This was accounted for by monitoring the depth-related fractionation of standards, and then applying a sliding-window depth-related fractionation factor to the unknowns. Pb/U fractionation also varies by up to several percent depending on position on the mount surface, due to variations in the flow rate/pattern of the helium carrier gas across the sample surface. For this reason, all standards and unknowns are mounted close together in the central portion of the mount, and care was taken to analyze standards that are as close as possible to each unknown.
To determine accurate concentrations of U and Th, we compare intensities with the Sri Lanka standard, which has concentrations of U, Th, and Pb known to ~20%. For each zircon analysis, the errors in determination of $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ result in a measurement error of ~1-2% (at 2-sigma level) in the $^{206}\text{Pb}/^{238}\text{U}$ age. The errors in measurement of $^{206}\text{Pb}/^{207}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ also result in ~1-2% (at 2-sigma level) uncertainty in age for grains that are >1.0 Ga, but are substantially larger for younger grains due to low intensity of the $^{207}\text{Pb}$ signal. We refer to errors that arise from the measurement of $^{206}\text{Pb}/^{238}\text{U}$, $^{206}\text{Pb}/^{207}\text{Pb}$, and $^{206}\text{Pb}/^{204}\text{Pb}$ as random (or measurement) errors, because they are different for each analysis within a session. For most analyses, the cross-over in precision of these random errors for $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{207}\text{Pb}$ ages occurs at ~1.0 Ga. For this reason, $^{206}\text{Pb}/^{238}\text{U}$ ratios were considered the most representative, and were used for analyses younger than ~1.0 Ga, and $^{207}\text{Pb}*/^{206}\text{Pb}*$ ratios were considered the most representative, and were used for ages older than ~1.0 Ga. Table A-2 shows the cutoff ages used for individual samples.

Table A-1 reports analytical data at 1σ uncertainties based on the analytical (or measurement) errors. The uncertainty of the weighted mean is based on the scatter and precision of the set of concordant ages, weighted according to their measurement errors. The systematic errors are then added to this measurement error quadratically. Systematic errors include contributions from the fractionation correction, composition of common Pb, age of the calibration standard, and U decay constants. Total average systematic errors are listed for individual samples in Table A-2.

For detrital zircon samples, approximately 100 randomly-selected zircon crystals were analyzed from each sample, to identify each of the main age groups present. Data were filtered according to precision (typically 10% cutoff) and
discordance (typically 30% cutoff) and plotted on Pb/U concordia diagrams or relative age-probability plots using algorithms of Ludwig [2003] (Figure A-1). Relative age probability curves were constructed by: (1) calculating a normal distribution for each analysis based on the reported age and uncertainty, (2) summing the probability distributions of all acceptable analyses into a single curve, and (3) if normalized, dividing the area under the curve by the number of analyses.

Interpretations of U/Pb detrital zircon data are based on the view that only clusters of ages record robust sources ages. This is because a single age determination may be compromised by Pb loss or inheritance (even if concordant), whereas it is unlikely that two or more grains that have experienced Pb loss or inheritance would yield the same age. We accordingly attach age significance only to clusters defined by three or more overlapping analyses; this has particular importance for determination of the youngest age component in a detrital zircon sample, which is commonly used as a maximum depositional age.

For further discussion of the analytical methods of the University of Arizona LaserChron Center, refer to Gehrels et al. [2006; 2008].
Table A-2. Total average systematic errors (s.e.) for $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{207}\text{Pb}$ for each sample (2σ), and average $^{206}\text{Pb}/^{238}\text{U}$ ages and standard deviation calculated for all standards run for each sample (n=24 for a typical 100-analysis DZ sample). Also shown are cutoff ages for each sample; $^{206}\text{Pb}/^{207}\text{Pb}$ ages were considered the best age for grains older than the cutoff age, and $^{206}\text{Pb}/^{238}\text{U}$ ages were considered the best age for grains younger than the cutoff age (listed in Table A-1). Systematic errors not available for sample BU07-73 due to corrupt file.

<table>
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<tr>
<th>Sample</th>
<th>$^{206}\text{Pb}/^{238}\text{U}$ s.e. (%)</th>
<th>$^{206}\text{Pb}/^{207}\text{Pb}$ s.e. (%)</th>
<th>$^{206}\text{Pb}/^{238}\text{U}$ standard ave. age (Ma)</th>
<th>$^{206}\text{Pb}/^{238}\text{U}$ standard std. dev (Ma)</th>
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Figure A-1. Pb/U concordia plots of Paro Formation samples. Includes all analyses greater with concordance >70%. Data for individual zircon analyses listed in Table A-1. Error ellipses are shown at the 1σ level (68.3% confidence).
A.1.3. Determination of crystallization ages of intrusive bodies

The determination of the crystallization ages of our two granite samples (BU07-83 and BU08-128) is problematic, as concordant zircon ages for these samples are widely spread over ca. 200 Ma for BU07-83 and ca. 100 Ma for BU08-128 (Fig. A-2). As the weighted mean of all the grains may not necessarily represent the true crystallization age of these intrusive bodies, we only report their likely ages ranging between 400 and 500 Ma. From sample BU07-83, eight grains including BU0783-50, -60, -81, -70, -45, 83, -73, and -84 were considered discordant because they had % concordance less than 70. From sample BU08-128, only three grains including BU08128-2, -10, and -17 had % concordance less than 70. These discordant grains were eventually discarded.

Regrettably, no CL imaging was used to differentiate cores from rims for these two samples, and as a result, we interpret many of the older ages as the result of mixing and/or inheritance. Also, it is uncertain whether young ages may be a result of Pb loss. However, despite these issues, note that these samples intrude strata with a maximum deposition age of Cambrian, and the ages we obtain, though not ideal, do give constraint on the youngest permissible deposition age.
Figure A-2: Pb/U zircon concordia plots with insets showing range of error for individual DZ analyses. Refer Figure 2.2 for sample map locations, and Figure 2.3 for stratigraphic locations. Data for individual zircon analyses listed in Table 2.2. A) Sample BU07-83, orthogneiss from Paro Formation, near Betekha. B) Sample BU08-128, orthogneiss at the base of Paro Formation exposed at Sisina.
Figure A-3. Plots of age versus U concentration, U-Th ratio, and percentage concordance for sample BU07-73. Five 500 Ma grains define the 500 Ma peak for this sample. All have low U concentration and low U-Th ratio that suggest the young ages are unlikely from metamorphic rims or Pb loss. Two of the 5 grains are 100% concordant at 1σ.

A.2. Details of Nd isotopic analyses

The isotopic ratios of $^{143}\text{Nd}/^{144}\text{Nd}$, and the trace element concentrations of Sm, and Nd were measured by thermal ionization mass spectrometry on whole rock samples. Rock powders were put in large Savillex vials and dissolved in mixtures of hot concentrated HF-HNO$_3$ or alternatively, mixtures of cold concentrated HF-HClO$_4$. The dissolved samples were spiked with the Caltech mixed $^{147}\text{Sm-}^{150}\text{Nd}$ spike [Wasserburg et al., 1981] after dissolution. The bulk of the REEs were separated in cation columns containing AG50W-X4 resin, using 4N HCl. Separation of Sm and Nd was achieved in anion column containing LN Spec resin, using 0.1N to 2.5N HCl. Sm and Nd were loaded onto single Re filaments using platinized carbon, and resin beads, respectively. Mass spectrometric analyses were carried out at the University of Arizona on an automated VG Sector multicollector instrument fitted with adjustable Faraday collectors and a Daly photomultiplier [Otamendi et al., 2009]. Concentrations of Sm and Nd were determined by isotope dilution, with isotopic compositions determined on the same spiked runs. An off-line manipulation programs was used for isotope dilution calculations. Typical runs consisted of acquisition of 100 isotopic ratios. The mean results of five analyses of the standard nSmβ performed during the course of this study are: $^{148}\text{Sm}/^{147}\text{Sm} = 0.74880\pm0.021$, and $^{148}\text{Sm}/^{152}\text{Sm} = 0.42110\pm0.006$. 
Ten measurements of the LaJolla Nd standard were performed during the course of this study. The standard runs yielded the following isotopic ratios: $^{142}\text{Nd}^{144}\text{Nd} = 1.14184\pm 2$, $^{143}\text{Nd}^{144}\text{Nd} = 511853\pm 2$, $^{145}\text{Nd}^{144}\text{Nd} = 0.348390\pm 2$, and $^{150}\text{Nd}^{144}\text{Nd} = 0.23638\pm 2$. Nd isotopic ratios were normalized to $^{146}\text{Nd}^{144}\text{Nd} = 0.7219$. The estimated analytical $\pm 2\sigma$ uncertainties for samples analyzed in this study are: $^{147}\text{Sm}^{144}\text{Nd} = 0.4\%$, and $^{143}\text{Nd}^{144}\text{Nd} = 0.0012\%$. Procedural blanks averaged from five determinations were 2.7 pg Sm, and 5.5 pg Nd.
Table A-1. U-Pb geochronologic analyses.

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<th>Zr-Co</th>
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A.3. References Cited


Ludwig, K.J. (2003), Isoplot 3.00: Berkeley Geochronology Center Special Publication 4, 70.


Precise determination of Sm/Nd ratios, Sm and Nd isotopic abundances in
standard solutions, Geochimica et Cosmochimica Acta, 45, 2311-2323.
Appendix B (Accompanies Chapter 3)

This appendix is published as an online supplementary data file, which accompanies the paper presented in Chapter 3. It can be accessed in the online version of this paper at doi:10.1016/j.epsl.2011.12.005. The complete citation is:

Table B-1. Mineral data and sample structural positions.

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* All samples contain quartz and biotite. Note: Sample BU10-79 also has chlorite.

** Approximate distance in meters structurally above and below the MCT.
Figure B-1. Elemental maps of Y and Th in monazite, illustrating non-diagnostic zoning of Y and Th. Representative spot analyses are inset in back-scattered electron images that show textural context of in-situ monazite (abbreviated mnz). White circles with ages are SIMS analysis spots with 2σ uncertainties. High vs. low concentrations of Y and Th in monazite are denoted by “hot” vs. “cold” colors. Refer Fig. 3.2 for mineral abbreviations and Fig. 3.3 for sample locations.
Figure B-2. Elemental maps of Y and Th in monazite, illustrating nearly homogeneous or non-diagnostic distribution of Y and Th. Representative spot analyses are inset in back-scattered electron images that show textural context of in-situ monazite (abbreviated mnz). White circles with ages are SIMS analysis spots with 2σ uncertainties. High vs. low concentrations of Y and Th in monazite are denoted by “hot” vs. “cold” colors. Refer Fig. 3.2 for mineral abbreviations and Fig. 3.3 for sample locations.