1 **How Does the Mid-crust Accommodate Deformation in Large, Hot Collisional**
2 **Orogens? A review of recent research in the Himalayan Orogen**

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Abstract

The presence of hot, weak crust is a central component of recent hypotheses that seek to explain the evolution of continent-continent collisions, and in particular may play an important role in accommodating the >3000 km of convergence within the Himalaya-Tibetan collision over the last ~55 Myr. Models that implicate flow of semi-viscous midcrustal rocks south toward the front of the Himalayan orogen, ‘channel flow’, are able to account for many geologic observations in the Himalaya, while alternative models of collision, particularly ‘thrust-wedge taper’, demonstrate that much of the observed geology could have formed in the absence of a low-viscosity mid-crustal layer. Several recent studies, synthesized here, have prompted a shift from initial assumptions that channel flow and thrust-wedge taper processes are by definition mutually exclusive. These new studies reveal the presence of several tectonometamorphic discontinuities in the midcrust that appear to reflect a continuum of deformation in which both channel- and wedge-type processes operate in spatially and temporally distinct domains within the orogen, and further, that the system may migrate back and forth between these types of behavior. This continuum of deformation styles within the collisional system is of crucial importance for explaining the evolution of the Himalayan orogen and, hence, for understanding the evolution of Earth’s many continent-continent collision zones.

Keywords: Continent-continent collision, Himalaya, channel flow, critical taper,

1. Introduction

Continent-continent collision zones are amongst the most dynamic and complex tectonic systems on Earth. In addition to remolding the earth's crust and altering mantle dynamics, collisional orogens also have profound effects on its biosphere, hydrosphere, and atmosphere. These margins are dynamic systems that react to both large-scale changes in plate motions and local processes such as focused erosion. The dynamic nature of orogens results in the preservation of a complex set of spatially and temporally distinct deformational and metamorphic episodes. Knowledge of the inter-relationships among these sometimes seemingly disparate geologic episodes is critical to elucidating the processes that govern their evolution.

Scientific understanding of how mountain belts evolve is ever changing, as for example, ideas of geosynclines and vertical tectonics gave way to the plate tectonic paradigm. More recently, our understanding of plate tectonics has also progressed from a model, quite literally, of rigid plates floating atop a convecting mantle to the recognition of significant non-rigid intra-crustal deformation. In this context, large hot orogens can be defined as convergent lithospheric regions with both high temperature (T) and high magnitude (M) in T-M space (i.e. Fig. 1 of Beaumont et al. (2006)). The role of weak layers in the lithosphere during the evolution of large, hot orogens (e.g., Bird, 1991; Grujic et al. 1996; Clark and Royden, 2000; Beaumont et al. 2001) has become a focus for modern tectonics research (*see e.g.* Law et al. 2006). The potential for orogen-scale ductile behavior of the crust has also proven a catalyst for major questions about upper crustal and geomorphological responses to large-scale (on the order of 100s of kilometers) horizontal flow at depth (e.g. Clark et al. 2005; Royden et al. 2008). These

relatively recent ideas about the rheologic behavior of the lithosphere during collision highlight the need for tectonics research directed at understanding the behavior of the crustal lithosphere as an orogenic system evolves. The present nexus of abundant, high quality geologic data, particularly pressure-temperature-time-deformation (PTtd) data, and sophisticated orogenic modeling focused on the Himalayan orogen provide the basis for this review of current and developing ideas.

While several ancient large-scale collisional convergent margins are preserved worldwide (the Proterozoic Trans-Hudson and Grenville orogens in North America, or the Caledonides in Europe for example), the upper crust is generally poorly preserved in those locations. In contrast, the Cenozoic Himalayan front, a component of the Himalaya-Tibet orogenic system, presents more than 2000 km of along-strike exposure of upper, middle, and locally lower-crustal rocks (Figure 1) facilitated by dramatic uplift and erosion, producing spectacularly incised topography. As such, the Himalayan front has become the type-model for continent-continent convergent systems. Moreover, because it is an active system, it is possible to link seismological and other geophysical data from the Himalaya directly to geologic observations, allowing the study of deformation, metamorphism, exhumation and geomorphological development of a collisional orogen in four dimensions for multiple levels of the lithosphere, as well as the asthenosphere. The ideas and models developed in this natural laboratory can be, and have been, exported and applied to ancient systems around the globe. In this context, a review and synthesis of recent key studies and associated data that explore the behavior of the midcrust during convergence in the Himalaya provides a timely opportunity to assess the current understanding of how the lithosphere behaves during collision.

It is the goal of this review to synthesize several recent studies that show the role that recently identified tectonometamorphic discontinuities play in the complex record of deformation accommodation processes in the core of the Himalaya. These studies are discussed in the context of the main tectonometamorphic features of the well-documented Cenozoic history of the Himalayan collision, summarized below in sections 2 and 3 (readers are referred to Hodges (2000) and Yin (2006) for broader reviews of the orogen). Furthermore, we consider how this complex record may be reconciled in a simplified conceptual tectonic model for plate convergence accommodation within the mid- and upper-crust through time.

2. Overview Of The Himalayan Orogen¹

The Himalaya-Tibet orogen is the result of the continental collision of India with Asia and continued convergence following closure of the intervening Tethyan Ocean at ca. 55-50 Ma (Rowley, 1996; DeCelles et al. 2004; review in Yin 2006; Green et al. 2008; Najman et al., 2010 and references therein). The southern flank of the orogen, hereafter referred to as the Himalaya, or the Himalayan front, is a large-scale system of predominantly north-dipping structures that affect upper and midcrustal rocks (Figure 1). These rocks are dominated by those originally deposited as part of a southward-tapering supracrustal sedimentary wedge on the northern margin of the Indian continent (Searle et al. 1987), but also include remnants of lower crustal rocks (Groppo et al. 2007; Cottle et al. 2009a; Corrie et al. 2010; Warren et al., 2011; Kellett et al., 2014a).

The Himalaya has been broadly divided into a series of fault-separated lithotectonic assemblages (Figure 1; Le Fort, 1975; Nelson et al. 1996). The northernmost

¹ many of the references cited in this section are the primary/original ones only.

assemblage, the unmetamorphosed to very low-grade Tethyan sedimentary sequence (also referred to as the Tibetan Series), is bound to the north by the Indus-Tsangpo suture (also variably referred to as the Yarlung-Tsangpo suture), the boundary between the Indian and Asian continents (Searle et al., 1997), and to the south by a network of top-to-the-north sense shear zones and brittle faults, the South Tibetan detachment system (Burg 1983, Burchfiel et al. 1992). The footwall of the South Tibetan detachment system consists of high-grade metamorphic rocks of the Greater Himalayan sequence (also referred to as High Himalayan crystallines or Tibetan slab). The unmetamorphosed-to-middle metamorphic grade Lesser Himalayan sequence lies structurally below the Greater Himalayan sequence, and is separated from it by the Main Central thrust (Heim and Gansser, 1939; Gansser, 1964). The lower boundary of the Lesser Himalayan sequence, the Main Boundary thrust, separates it from the subjacent Subhimalaya, the deformed foreland basin of the orogen. The youngest and most foreland-ward thrust of the Himalayan orogen is the presently active Main Frontal thrust (Figure 1).

Metamorphic and associated anatectic rocks exposed along the Himalayan orogen have been variably ascribed to either the Greater Himalayan sequence or Lesser Himalayan sequence, depending on the inferred location of the Main Central thrust. The nature and location of the Main Central thrust, which was originally defined as separating Greater Himalayan sequence rocks in its hanging wall from Lesser Himalayan sequence rocks in its footwall (Heim and Gansser, 1939; Gansser, 1964), has been controversial. Different researchers have used different criteria to interpret and map this thrust (e.g., Searle et al., 2008). Consequently, major differences exist in the published literature where similar, or even identical, rocks have been ascribed to different units based on their

interpreted structural position (e.g. Goscombe et al., 2006; Searle et al., 2008; Larson and Godin, 2009; Larson et al., 2010; From and Larson, 2014). The lack of a single accepted interpretation has made reconciling data and models for the evolution of the mid-crust difficult (e.g., Kohn et al., 2002; Searle et al., 2008). For simplicity, and to focus on the goal of general applicability of the Himalayan collision to other orogens, we do not emphasize the classification of rocks as either Greater Himalayan sequence or Lesser Himalayan sequence, but herein use the term Himalayan metamorphic core (HMC) to refer to all rocks that record evidence of Cenozoic mid-crustal deformation, metamorphism, melting and cooling.

3. P-T-t-D evolution of the HMC

Assessing the convergence accommodation processes responsible for the evolution of the Himalayan mid-crust over its ~55 Myr history requires a detailed reconstruction of metamorphism and deformation through time. Here we summarize existing constraints on the pressure, temperature, deformation and time paths that rocks deformed and metamorphosed in the mid-crust followed since the onset of collision. For a comprehensive review of metamorphism within the Himalayan system, readers are referred to Kohn (2014).

3.1 Early stage of collision: crustal thickening, burial and prograde metamorphism of the mid-crust

The mid-crust of the Himalayan orogen is typical of large, hot orogens in that the early prograde metamorphic and deformation histories are pervasively overprinted by the

development of later, high-temperature metamorphic assemblages and fabrics, and the formation and emplacement of anatectic melts (e.g. Williams and Jiang, 2005; Jessup et al., 2008). Moreover, since the rocks available for study are now at Earth's surface, there they may also have been modified by exhumation processes and/or metamorphic retrogression during ascent. Fortunately, several data sets exist for constraining the general prograde history of the Himalayan mid-crust, where these issues are either absent or have been resolved.

The lid to the Himalayan metamorphic core, the Tethyan sedimentary sequence (Figure 1), is a package of predominantly Paleozoic to Mesozoic sedimentary units originally deposited in the former Tethys ocean. This sedimentary package was deformed during crustal shortening into a fold and thrust belt that extends from the South Tibetan detachment system in the south to the Indus-Tsangpo suture in the north. Similarities in Nd isotopes and detrital zircon populations (e.g. Parrish and Hodges, 1996; Whittington et al., 1999; Ahmad et al., 2000; DeCelles et al., 2000; Miller et al., 2001; Robinson et al., 2001; Argles et al., 2003; Richards et al., 2005; Richards et al., 2006; Kellett, 2010; Long and McQuarrie, 2010; Myrow et al., 2010) indicate a common sediment source from a portion of the HMC. This affinity has been considered evidence for a “continuous margin” origin for the now adjacent rock packages, rather than HMC rocks representing an accreted terrane or crystalline axis origin (see DeCelles et al., 2000; Myrow et al., 2003). Shortening of the Tethyan sedimentary sequence following collision is thought to have doubled its thickness from ~10 km to ~20 km resulting in regional folding, fabric development and very low-grade (~300 – 400 °C) metamorphism (Schneider & Masch, 1993; Garzanti et al., 1994; Ratschbacher et al., 1994; Guillot et al., 1995; Willems et al.,

1996; Godin, 2003; Godin et al., 1999a, b; 2001; Kellett and Godin, 2009; Cottle et al.,
2011). The timing of shortening and thickening within the Tethyan sedimentary sequence
is poorly constrained but predates Miocene structures (Kellett and Godin, 2009), and is at
least locally >44 Ma (Aikman et al., 2008, Aikman et al., 2012). It seems reasonable to
conclude, therefore, that during the early stage of collision, the core of the orogen was
buried by at least 10 km simply by ‘thin-skinned’ thickening and shortening of the upper
crust, in addition to any internal thickening occurring within the mid-crust itself.

Perhaps the best constraints on the early record of convergence in the core of the
Himalayan system come from high-pressure and ultra-high pressure (UHP) metamorphic
assemblages preserved in exhumed continental margin rocks exposed in Kaghan,
northern Pakistan and surrounding Tso Moriri in northwest India (Figure 1).
Thermobarometric data outline a metamorphic history that involves initial UHP
conditions of 2.7–3.6 GPa and 640–760 °C, followed by decompression to 0.9–1.7 GPa
and 500–640 °C and a final phase of reheating at 0.8–1.2 GPa and 650–720 °C (as
summarized in Donaldson et al., 2013). Careful geochronology has outlined similar
protracted metamorphic histories in the two exposures that include peak UHP
metamorphism between ca. 47 and 43 Ma (O’Brien et al., 2001; Sachan et al., 2004; de
Sigoyer et al., 1997, 2000, 2004; Kaneko et al., 2003; Leech et al. 2005; O’Brien, 2006;
Parrish et al., 2006; Guillot et al., 2007; Wilke et al., 2010; St. Onge et al. 2013;
Donaldson et al., 2013), consistent with deformation of the upper crust in southern Tibet
that predates leucogranite intrusion at 44 - 42 Ma in the Tethyan sedimentary series (e.g.
Aikman et al., 2012). These records of crustal shortening and high-pressure

metamorphism indicate that Indian continental subduction initiated sometime earlier, between 51-47 Ma (Donaldson et al., 2013).

Rocks preserving relic evidence for high-pressure metamorphism have also been identified across a substantial portion of the eastern Himalaya, from Ama Drime to Bhutan (Figure 1). Now hosted in granulite-facies rocks within the HMC, these relics indicate equilibration under eclogite-facies conditions with pressures of at least 1.5 GPa while garnet Lu-Hf ages show that eclogite-stable conditions occurred in the lower crust of the Himalaya by mid-Eocene (ca. 38 Ma), indicating that at least that portion of the orogen had reached a crustal thickness of at least 60 km by that time (Lombardo and Rolfo, 2000; Groppo et al., 2007; Rolfo et al., 2005; Cottle et al., 2009a; Corrie et al., 2010; Grujic et al., 2011; Warren et al., 2011; Kellett et al., 2014a). These rocks demonstrate a burial and exhumation history distinct from the subduction-related UHP eclogites in the northwestern Himalaya. They are instead, interpreted to have formed during crustal thickening of the Indian plate accompanying the early stages of collision (e.g., Lombardo and Rolfo, 2000; Kellett et al., 2014a).

In the HMC itself, pervasive ductile deformation at peak and post-peak metamorphic conditions has virtually obliterated evidence of prograde fabrics, although inclusion trails and inclusion-rich cores are observed locally in garnet grains. A recent investigation of the timing of prograde garnet growth in structural windows into the HMC in Tibet using Lu-Hf geochronology, yielded ages ranging between 54-49 Ma (Smit et al., 2014). P-T-t studies that rely on monazite geochronology (often preserved in garnet cores), however, reveal kyanite-stable and locally sillimanite-stable metamorphic conditions of 550–680 °C and 9–11 kbar during the Eocene and Oligocene (ca. 40 – 30

Ma) (e.g. Walker et al. 1999; Simpson et al., 2000; Godin et al., 2003; Searle et al., 2003; Kohn et al., 2005; Cottle et al. 2009b; Kellett et al., 2010; Rubatto et al., 2013; Montomoli et al., 2013). Trace element distribution (i.e. HREE, Gd/Yb, Y) in monazite generally support crystallization coeval with prograde garnet growth during the Eocene and early Oligocene. Rare ca. 35–26 Ma kyanite-bearing leucosomes and granites (e.g. Coleman 1998; Thimm et al. 1999; Godin et al. 2001; Prince et al. 2001; Viskupic and Hodges, 2001; Zhang et al. 2004; and Lee and Whitehouse 2007; Cottle et al., 2009a; Groppo et al., 2010; Imayama et al., 2012; Rubatto et al., 2013) provide evidence of partial melting within the HMC as early as the Oligocene. Taken together, the mid-crust of the Himalayan orogen experienced protracted prograde metamorphism between 54 – 28 Ma with sporadic partial melting during initial crustal shortening. The accumulated crustal thickness that resulted from this early shortening set the conditions that controlled the processes responsible for much of the subsequent Miocene evolution of the orogen.

3.2 Transition: decoupling of upper crust from mid-crust, peak metamorphism and increased melt production in the mid-crust

In the late Oligocene to early Miocene, several changes occurred within the Himalayan orogen. The Tethyan sedimentary sequence fold and thrust belt was overprinted and transposed by the Miocene South Tibetan detachment system (Godin, 2003; Kellett et al., 2009). The Tethyan belt was also affected by Miocene extension and doming during formation of the North Himalayan gneiss domes (e.g., Lee et al., 2000; 2006; Larson et al., 2010b). Subsequent deformation affecting the Tethyan sedimentary sequence includes crenulation, low-wavelength regional buckling, and E-W extensional

grabens and related structures (e.g., Coleman and Hodges, 1995; Godin 2003; Gleeson and Godin, 2006; Kellett et al., 2012; Mitsuishi et al., 2012). This evidence of continued deformation in the upper crust, while prevalent, is relatively insignificant in terms of accommodating the relative tectonic plate motion between India and Asia of ~5 cm/yr during the Miocene (e.g. Molnar and Stock, 2009). This transition and stalling of convergence-related deformation in the Tethyan sedimentary sequence during late Oligocene/early Miocene (ca. 23 Ma) may in part be attributed to short-term decoupling of the upper crust from the mid-crust across the South Tibetan detachment system (Burg and Chen, 1984; Burchfiel et al., 1992; Hodges et al., 1992; Godin et al., 2006 and references therein, Cottle et al., 2007; Kellett and Godin, 2009; Kellett et al., 2010; 2013; Chambers et al., 2011; Law et al., 2011). Following this decoupling, the upper crust became, in effect, a passive lid to the orogen (e.g. Grasemann et al., 1999; Searle et al., 2003; Law et al., 2004; Webb et al., 2011; Kellett and Grujic, 2012) though it continued to respond locally to midcrustal processes throughout the Miocene (e.g., Mitsuishi et al., 2012).

In contrast to the upper crust, convergence accommodation continued to drive metamorphism and deformation in the mid-crust during the late Oligocene and early Miocene. Most of the HMC reached peak metamorphic conditions (peak T), ranging from amphibolite to granulite facies, during this time period as part of a clockwise P-T loop (Figure 2; Hodges et al., 1988; Goscombe and Hand, 2000; Kohn et al., 2001; Searle et al., 2003; Kohn, 2008; Cottle et al., 2009; Groppo et al., 2009; Kellett et al., 2010, 2013; Streule et al., 2010; Warren et al., 2011; Corrie et al., 2012; Larson et al., 2013; see review in Kohn, 2014).

Coeval with high temperatures and upper crustal decoupling, the HMC experienced significant anatexis, producing a large volume of two-mica $\pm$ tourmaline $\pm$ peritectic garnet and cordierite-bearing leucogranite (e.g. Godin et al., 2006 and references therein; Searle et al., 2010; Groppo et al., 2012; Lederer et al., 2013; Rubatto et al., 2013). Petrologic modeling indicates that Himalayan leucogranites likely formed via a combination of vapor-absent incongruent melting of muscovite at P-T conditions of 650–750 °C and 4–9 kbar (Scaillet et al. 1990, 1996; Harris & Massey 1994; Patino-Douce & Harris 1998; Prince et al. 2001) and biotite at higher temperatures (Groppo et al., 2012). Together, these processes had the potential to produce a significant (15–20 volume %) amount of melt (Groppo et al. 2012). In addition, because vapor-absent dehydration melting reactions typically have positive P–T slopes, the melts had the potential to rise a considerable distance in the crust before reaching their solidi. Such a process allowed melt to pool in large quantities towards the top of the mid-crust, consistent with the occurrence of several large, laccolithic Miocene intrusions along the Himalaya including Shivling, Bhagirathi, Thalay Sagar, Manaslu, Makalu, Shisha Pangma, Kangchenjunga, the base of the Everest–Nuptse–Lhotse massif, Khula Kangri, Chomolhari and Masang Kang.

Deformation of the mid-crust at this time was pervasive (Grujic et al., 1996; Martin et al., 2005; Jessup et al., 2006; Larson and Godin, 2009; Larson et al., 2010a; Larson and Cottle, 2014) with southern translation (Law et al., 2004, 2011, 2013), and partial exhumation of the HMC (Kellett et al., 2009; Cottle, 2009). Vorticity estimates from the HMC, potential methodological issues notwithstanding (Li and Jiang, 2011; Stahr and Law, 2011), are consistent with a general shearing across most of the HMC

306 transitioning to more simple shearing toward the upper and lower boundaries (Law et al.
307 2004, 2013; Carosi et al., 2006; Jessup et al., 2006; Larson and Godin, 2009; Langille et
308 al. 2010; Larson et al., 2010a) and potentially along a decelerating strain-path through
309 time (Grasemann et al., 1999). When combined with available strain ratio data, those
310 vorticity estimates correspond to 10 – 44% shortening perpendicular to the flow plane
311 and, assuming approximately plane strain deformation, stretches of 10 – 78% parallel to
312 the flow direction (Law et al. 2004, 2013; Carosi et al., 2006; Jessup et al., 2006; Larson
313 and Godin, 2009; Langille et al. 2010).

314 Available quartz petrofabric orientation (CPO) data from the HMC are typically
315 extracted from the peripheries of the mid-crustal core with only sparse examples from the
316 interior regions (e.g. Grujic et al., 1996; Law et al., 2004; 2011; 2013; Larson and Godin,
317 2009; Langille et al., 2010; Larson et al., 2010a, b; Kellett et al., 2014b). Available data
318 consistently outline top-to-the-north sense shear at and near the top of the HMC, which
319 transitions down structural section to top-to-the-south sense shear throughout the majority
320 of the rock package. Deformation temperatures estimated from CPO fabric opening
321 angles increase up structural section from the base of the HMC (Larson and Godin, 2009;
322 Larson et al., 2010a and reference therein; Law et al., 2013) and down structural section
323 from the top of the HMC (Law et al., 2004; 2011). While detailed characteristics from
324 within the interior of the HMC are rare, the limited data indicate the presence of
325 otherwise cryptic structural discontinuities. The presence of such structures, discussed
326 further below, indicate that the HMC, which for some time had been treated as a
327 homogeneous body (e.g. Brunel, 1986; Grujic et al., 1996; Robinson et al. 2006; Searle et
328 al., 2006), is significantly more structurally complex than previously thought.

329

330 **4. Convergence Accommodation in the Himalayan Mid-crust**

331 The midcrustal response to convergence between India and Asia has been the focus of
332 many scientific studies examining the Himalayan system. This matter became very
333 topical-after the synthesis of geophysical (e.g. Nelson et al., 1996; Hauck et al., 1996)
334 and geological data (e.g. Grujic et al, 1996) into the channel flow hypothesis of
335 Beaumont et al. (2001), with further elaboration in Beaumont et al. (2004) and Jamieson
336 et al. (2004). The thermo-mechanical models outlined in those studies provided a viable
337 explanation for many of the known geological and geophysical constraints extracted from
338 the mountain belt. Importantly, they provided predictions against which new data could
339 be quantitatively tested (see summary in Gervais and Brown, 2011). These studies
340 invigorated Himalayan research and inspired scientists to return to the field, resulting in
341 an influx of empirical data that researchers have used to assess the validity of the thermo-
342 mechanical model predictions. This generation of Himalayan research typically
343 considered two possible end-members for accommodation of crustal shortening in the
344 Himalayan orogen (Figure 3a, b). In simplified terms:

- 345 1. *development of a **fold-and-thrust belt** within a **frictional critical taper wedge***
346 *(e.g., Robinson et al. 2006; Kohn, 2008; Sachan et al. 2010)*
347 2. *development of **gravity-driven lateral mid-crustal flow** (e.g., Grujic et al. 1996;*
348 *Beaumont et al. 2001; Searle et al., 2003; Jamieson et al. 2006).*

349

350 *4.1 End-member models*

Critical taper theory was developed to describe deformation in thin-skinned foreland fold-and-thrust belts and accretionary wedges (e.g., Elliott, 1976; Chapple, 1978; Boyer and Elliott, 1982; Davis et al, 1983). In this model, deformation occurs within a wedge defined by a hinterland-dipping basal detachment and foreland-inclined topography, and is commonly, though not exclusively, dominated by foreland-propagating stacked thrusts and fault-bend folds (Figure 3a) (Elliott, 1976; Chapple, 1978; Davis et al, 1983; Dahlen 1990). Friction along detachments, erosion, addition of material, rock strength, and rate of shortening all interact to influence the nature of wedge evolution and deformation (Dahlen 1990). The orogen is predicted to respond to perturbations in the taper angle, which can occur as a result of changes of any of the above parameters, by propagating deformation to a new fault in the foreland, in the case of an increase in the taper angle, or forming an out-of-sequence thrust in the interior of the orogen (in the case of a decrease in the taper angle) to reestablish the critical taper angle. With respect to convergence accommodation, this model predicts that given a persistently layered lithostratigraphy, crustal shortening is accommodated dominantly along relatively weak thrusts, with comparatively insignificant internal deformation within thrust sheets.

Critical taper principles have been applied in several kinematic models of the Himalaya (e.g., Avouac 2003; Robinson et al. 2003; Bollinger et al. 2006; Robinson, 2008, Long et al., 2011), particularly in an effort to describe the inverted metamorphism and deformation of the exhumed metamorphic core and movement across its bounding structures (Figure 3a, Harrison et al. 1998; Bollinger et al. 2006; Robinson et al. 2006; Sachan et al. 2010).

The channel flow hypothesis evolved out of extruding wedge models that sought to explain coeval activity on the structures that bound the mid-crustal core of the Himalaya, the South Tibetan detachment system and the Main Central thrust (e.g. Burchfiel et al. 1992; Grujic et al. 1996, Figure 3b). Subsequent re-interpretation of geophysical data proposed that the upper boundary of the mid-crustal core may flatten at depth to the north such that domes of exhumed mid-crust may be continuous with mid-crustal rocks exposed at the orogenic front (Figure 3b, Nelson et al. 1996). This structural framework, along with available geologic data, including coeval motion across the South Tibetan detachment system and Main Central thrust, were compared by Beaumont et al. (2001) with an analogous 2D model orogen created within a finite-element thermo-mechanical framework. The observed evolution of the HT1 model led to their proposal that in a large, hot orogen such as the Himalaya, melt-weakened material bounded by stronger upper and lower crustal layers will flow if subjected to a lateral lithostatic pressure gradient. In the case of the Himalaya, the Tibetan plateau and the Indian foreland provide the lateral lithostatic gradient. Focused surface denudation can provide a positive feedback mechanism to maintain the pressure gradient and sustain flow, but is not a requisite for flow (e.g. Royden et al., 1997; Clark and Royden, 2000; Beaumont et al., 2004). In this model, significant convergence between India and Asia is accommodated by large-scale ductile flow and extrusion of the mid-crust, rather than being confined to displacement of multiple thrust sheets across a series of discrete thrust faults or shear zones.

More recently, it has become apparent that these two ‘end-member’ accommodation processes should not be considered as mutually exclusive processes

where evidence for one negates the possibility of the other (Beaumont and Jamieson, 2010; Larson et al., 2010, 2013; Corrie et al., 2012; Jamieson and Beaumont 2013). Though not commonly acknowledged following its publication, the thermo-mechanical model HT1 of Beaumont et al. (2001) originally predicted formation of a critical taper wedge in the model orogen foreland while the deep hinterland simultaneously deformed by lateral mid-crustal flow. Moreover, orogenic-scale wedge taper models have also predicted lateral flow of the mid-crust in the hinterland (e.g., Platt, 1986). Considering the evolving nature of an orogen, it can be anticipated that any crustal accommodation processes are dynamic, shifting in both space and time. Thus, the finite deformation recorded in any particular rock may record a complex history which could, theoretically, include initial deformation in a critical-taper wedge during initial underthrusting, subsequent incorporation into a mid-crustal channel flow, and then finally deformation in a critical-taper wedge during exhumation (Figure 3c).

4.2 Spatial and temporal transition zones

Conceptually, the relationship between channel flow and critical taper processes can be presented as *extending flow* vs. *compressing flow* (Figure 4), as developed for geologic analysis by Kanizay (1962) and Price (1972). It is worth noting from the outset that use of the term ‘flow’ here refers to particle movement and is not necessarily synonymous with the ‘flow’ in channel flow, because we use the original terms as defined by Kanizay (1962) for historical continuity. Compressing flow is defined as shortening in the direction of flow with diverging streamlines in the flow direction and thrust-sense slip lines, that is typical of shallow foreland regions. In contrast, extending

flow is extension in the direction of flow with converging streamlines in the flow direction and normal-sense slip lines which is, typical of deep hinterland regions (Figure 4, Price, 1972). The transition between the two manifests as streamlines that converge in the direction of flow in the extending flow regime, enter the compressing flow regime, and diverge (Figure 4). The orientations of sliplines in the same conceptual system are flat at the transition between regimes and at opposite orientations within either regime, where extension across a slipline in extending flow becomes shortening in compressing flow (Figure 4), maintaining internal kinematic compatibility. Geologic evidence for both compressing flow (e.g. vertical thickening via basal accretion) and extending flow (e.g. vertical thinning – recumbent folds, over-steepened apparent pressure gradients) is recorded within the exhumed HMC (e.g. Larson et al., 2010; 2011; 2013; Yakymchuk and Godin, 2012). This record of deformation is most easily interpreted as follows: mid-crustal material was extruded laterally from the deep hinterland of the orogen into the foreland under extending flow conditions. As it intruded the foreland, the extruded material cooled, became more viscous, and began to deform as part of a tectonic wedge (e.g. Price, 1986, Webb et al, 2007; 2011). The deformation history experienced by the extruded mid-crustal material progressed from extending flow - vertical thinning/horizontal stretching, characteristic of hinterland-style deformation, to compressing flow - vertical thickening/horizontal compression, characteristic of fold-thrust-type deformation, as it transferred into the foreland (Figure 4, e.g. Larson et al., 2010; Larson and Cottle, 2014). As the orogen evolved, therefore, the material that was once situated at mid-crustal depths in the deep hinterland of the orogen translated through time and space to become incorporated into the foreland-propagating wedge. This process

resulted in preservation of a complex spatial and temporal transition between different deformation types preserved as the record of finite strain in the rock package. Because of variable exposure levels within the orogen, the overprinting relationships associated with this transition are locally dominated by different deformation regimes (extending vs. compressing). As a consequence, depending on the study location and investigation, strong evidence can be found for one, or both of the end-members (Larson et al., 2010; 2013; Chambers et al., 2011; Yakymchuk and Godin, 2012). However, a geologic basis is lacking for mutual exclusivity of the critical taper and channel flow end-members, and many of these findings have unnecessarily polarized the Himalayan research community. Complex spatial and temporal transitions that record this composite deformation were largely ignored and have only recently been documented at multiple points across the orogen (e.g., Montomoli et al., 2014).

4.3 Tectonometamorphic Discontinuities

The recognition and investigation of newly recognized, widespread tectonometamorphic discontinuities within the exhumed metamorphic core (Figure 5; see summary in Table 1 in Montomoli et al., 2014) is essential to advancing our understanding of the tectonic evolution of the Himalaya based on a composite deformation model. Such features have been identified through detailed P-T-t-D studies, and are typically cryptic, or nearly impossible to identify in the field. The term “tectonometamorphic discontinuity” reflects the observation that adjacent rock packages of similar lithology have contrasting tectonometamorphic histories, identified through comparison of their detailed P-T-t-D paths, indicating a discontinuity in rock evolution

from one package to the next. We intentionally avoid labelling these features as “faults” or “shear zones” because faults or shear zones should be identifiable in the field as mappable physical breaks, offsets, or high strain zones. Even faults and shear zones that are not exposed (i.e. inferred) typically have clear, contrasting features on either side that separate the footwall from hanging wall, whether it is rock type, metamorphic grade, basement inheritance, deformation style, etc. Since these features do not obviously demonstrate any of these classical differences, their definitive classification is still a work in progress.

The Tama Kosi region of east-central Nepal provides a well-documented example of a tectonometamorphic discontinuity that was identified through a detailed P-T-t-D study. The constructed P-T-t-D paths presented in Larson et al. (2013), and shown in Figure 2B and C, were obtained from adjacent rock packages in a continuous, apparent Barrovian sequence of para and orthogneiss within the exhumed Himalayan metamorphic core. The markedly different P-T-t-D paths demonstrate a tectonometamorphic discontinuity between rocks at lower structural levels that were deformed and metamorphosed at staurolite grade over a timespan of only ~2 myr along a burial path and rocks at higher structural levels that record a kyanite to sillimanite grade metamorphic history lasting at least 6 myr with peak P-T conditions and monazite growth along a variable decompressional heating path, and late, post-peak metamorphism deformation. The age of kyanite growth above the discontinuity is constrained by ~18.5 Ma monazite that is interpreted to have grown with kyanite and is now preserved within it (Larson et al., 2013). This timing is in contrast to the age of staurolite growth structurally below the discontinuity, which is ~10 myr younger. The difference in

ages demonstrates that the staurolite–kyanite grade rocks did not form during a typical Barrovian metamorphic cycle. The existence and location of the tectonometamorphic discontinuity has been further documented using quartz *c*-axis orientation fabric data, which record a sharp break in deformation temperatures between the staurolite and kyanite grade rocks (Larson and Cottle, 2014).

Documented tectonometamorphic discontinuities to date include features that are interpreted as early stage in-sequence thrusts, late-stage out of sequence thrusts, and other equivocal structures, and have long been inferred in the Kali Gandaki region of west central Nepal (Vannay and Hodges, 1996; Godin et al., 2001), the Langtang Himal, central Nepal, (Reddy et al., 1993; Fraser et al., 2000), and the Khumbu region of east-central Nepal (Searle et al., 1999). Researchers have more recently described similar structures in far west Nepal (Yakymchuk and Godin, 2012), western Nepal (Carosi et al., 2010; Montomoli et al., 2013), west-central Nepal (Kohn et al., 2005; Larson et al., 2010; 2011; Corrie and Kohn, 2011), east-central Nepal (Larson, 2012; Larson et al., 2013; Wang et al., 2013; From and Larson, 2014; Larson and Cottle, 2014), far east Nepal (Goscombe et al., 2006; Ambrose et al., In Press), Sikkim (Rubatto et al., 2013; Mottram et al., 2014), and Bhutan (Grujic et al., 2011; Warren et al., 2011) (see Figure 5). Where age data are available, these structures are generally either interpreted as early (ca. 27 – 23 Ma), in-sequence thrust faults that drove extrusion in the hanging wall and metamorphism in the footwall (e.g. Carosi et al. 2010; Montomoli et al., 2013) or younger (ca. 16 – 12 Ma), post-peak metamorphic, sometimes out-of-sequence, structures (e.g. Warren et al., 2011; Grujic et al., 2011; Warren et al., 2014). See Montomoli et al.

(2014) for a recent review of these various structures along the Himalaya and Larson and Cottle (2014) for further discussion.

Based on available structural, thermobarometric and geochronologic data, we interpret the Tama Kosi and other discontinuities to record the transition from deep hinterland-style deformation (pervasive ductile flow) to shallower foreland-style deformation (discrete deformation zones). The timing of that transition may be a function of the nature of the local characteristics and geometry of the HMC. In regions with relatively thin structural sections of metamorphic rocks, such as west-central Nepal, the discontinuities are manifested as thrust faults that bury and drive metamorphism in the footwall (e.g. Carosi et al., 2010; Montomoli et al., 2013; Montomoli et al., 2014), implying the timing of movement along the structure is coeval with footwall metamorphism (e.g. Long et al., 2011). In these regions lateral extrusion of the hinterland is thought to be minor or even non-existent (Kohn and Corrie, 2011; Carosi et al., 2013), such that the development of these thrust structures within the HMC may indeed be the main mechanism by which convergence is accommodated. In regions with comparatively thick structural sections of metamorphic rocks such as east central Nepal, Sikkim, and Bhutan, the development of the discontinuities in the HMC appears to have dominantly post-dated peak metamorphism (Warren et al., 2011; Grujic et al., 2011; Rubatto et al., 2013; Larson and Cottle, 2014; Warren et al., 2014). The apparent difference in the relative timing of metamorphism and development of discontinuities at higher structural levels may be rheologically driven. Thinner sections can be expected to have carried less thermal energy/heat and cooled faster and sooner, thereby transitioning to foreland-style

compressing flow and discrete deformation earlier. Whereas thicker sections were hotter, cooled more slowly, and would have transitioned to a discrete deformation regime relatively later in the orogen's evolution.

4.4 Inverted metamorphism

A sequence of inverted metamorphic isograds, characterized by an upward - transition from biotite- to sillimanite-grade, occurs at the base of the HMC along the ~1700 km length of the Himalaya (e.g. Mallet, 1875; Hagen, 1954; Hashimoto, 1959, 1973; Bordet 1961). This iconic feature of Himalayan geology has received much attention since it was first documented in the 19th century, with numerous models proposed to account for its formation including: 1) the 'hot iron' model (e.g. Frank et al. 1973; LeFort, 1975), which postulates that as the HMC exhumed, heat was conducted down and away from its center toward the cooler footwall of the Main Central Thrust, producing a set of inverted isograds at the base; 2) shear heating (e.g. Arita, 1983, England & Molnar, 1993; Harrison et al. 1998; Nábělek et al., 2010; Nábělek and Nábělek, 2011; 2014) along the Main Central; 3) accretion of crustal material rich in radiogenic-heat producing elements to the base of the HMC combined with advection of heat from the system by erosion (Royden, 1993; Huerta et al. 1996); and 4) 'post-metamorphic' models in which pre-existing 'normal' isograds are subsequently folded (e.g. Searle & Rex 1989, Searle et al. 1999), transposed (e.g. Thöni, 1977; Jain and Manickavasagam, 1993; Grujic et al. 1996; Hubbard, 1996; Grasemann and Vannay, 1999), or subjected to imbricate thrusting (e.g. Brunel and Kienast, 1986; Treloar et al. 1989; Swapp & Hollister, 1991). None of the above-proposed explanations is without

problems. The former two models require inversion of the rock sequence whereas the latter does not. However, each of these models implies that metamorphism pre-dates deformation along in the vicinity of the MCT. If metamorphism and deformation are synchronous as suggested by some (e.g. Kohn, 2008) then these ‘post-metamorphic’ models do not fully explain the origin of the inverted isograds. For additional details, in depth reviews of this topic are given by Vannay and Grasemann (2001) and Yin (2006).

Recent work in Bhutan may help elucidate the development of inverted isograds. Detailed structural analysis indicates that material within the inverted metamorphic sequence was deformed and metamorphosed coeval with overthrusting by (or underplating beneath) overriding material (Fig 6; Long et al., 2011). Once the thrust propagated towards the foreland, or the underplated material was added to the hanging wall, the process would begin anew, though now at a slightly higher structural level and farther toward the foreland, assembling the inverted metamorphic sequence (Fig. 6). Such a process is compatible with geochronologic control from the inverted metamorphic sequence, which comprises monazite dates that decrease in age from the Early to Late Miocene with lower structural position and metamorphic grade (Kohn et al. 2001; Catlos et al. 2001; Larson et al. 2013; Mottram et al., 2014). Moreover, it is consistent with kinematic models that evoke underplating as a dominant process during the evolution of the Himalaya (Bollinger et al. 2006; Herman et al., 2010; Webb, 2013) and the down-structural continuation of processes that resulted in the development of thrust discontinuities in structurally higher rocks (e.g. Carosi et al., 2010; Montomoli et al., 2013; Larson et al., 2013) as discussed in the previous section.

5. Summary

The great number of studies and amount of data available concerning the evolution of the exhumed metamorphic core of the Himalaya provide the background necessary to describe its general evolution. Early crustal thickening related to shortening driven by convergence between India and Asia set up the initial conditions for the development of the HMC. Pressure-Temperature-time paths for metamorphic and igneous rocks in the upper, middle and lower crust exposed within the Himalaya outline early (Eocene-Oligocene) prograde metamorphism and variable anatexis during initial crustal thickening (Figures 7a, b). Peak (T) metamorphism occurred in the Late Oligocene to Early Miocene and was accompanied by widespread partial melting. This melting coincided with the decoupling of the upper crust from the middle crust and the southward lateral translation of the HMC out from the hinterland of the orogen and toward the foreland of the orogen (Figure 7c). This change in position within the orogen is reflected as a transition from extending flow (hinterland-style deformation) to compression flow (wedge-taper deformation), manifest in the rock record as the development of tectonometamorphic discontinuities, discrete structures that cut earlier pervasive ductile deformation and act to accommodate crustal-scale thickening and shortening within the Himalayan metamorphic core (Figure 7d). These cryptic structures are important, yet potentially often-overlooked components of the convergence accommodation process in collisional systems. Their structural position reveals important insight into fundamental changes in the style of material flow and deformation throughout an orogen whereas the timing and duration of movement places quantitative constraints on the timescales of large-scale changes in the convergence accommodation processes and perhaps orogenic

boundary conditions that influence the upper and middle crust. Detailed P-T-t-D studies of tectonometamorphic discontinuities therefore have the potential to elucidate both the spatial and temporal dynamism of strain accommodation recorded in large, hot orogens. This type of detailed, integrated work is critical to further characterizing and understanding orogen-scale kinematics that controlled the development of similar ancient mountain belts across the globe.

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Figure Captions.

Figure 1. (A) Simplified geologic map of the Himalaya modified after He et al. (in Press) and Webb et al. (2011a,b). Map location is shown in the inset regional diagram in the legend. (B) Schematic vertical geologic section drawn across the central Nepal Himalaya. Location of section is shown in (A). Relict eclogite localities are marked by red stars (K: Kaghan; Tso Morari; AD: Ama Drime /Arun; SK: Sikkim; BN: Bhutan).

Figure 2. Measured versus predicted Pressure-Temperature-time (P-T-t) paths for the upper (A), middle (B) and lower (C) Himalayan Metamorphic core. Data from Jamieson et al., (2004); Kellett et al., (2010); Jamieson et al., (2006); Larson et al., (2013); Kohn, (2008); Kohn et al., (2001); Caddick et al., (2007), Jessup et al., (2008).

Figure 3. (A) Schematic representation of the critical taper model as applied to the Himalaya. In this model, displacement and convergence accommodation largely occurs along discrete, forward propagating thrust structures. (B) Schematic representation of the channel flow ‘pipe to the surface’ model wherein convergence is accommodated through climatically enhanced lateral extrusion of the orogenic mid-crust between the MCT and STDS. Later deformation propagates toward the foreland as thrusts. (C) Schematic

representation of early stage channel tunneling or tectonic wedging accommodating convergence in the mid-crust (left) followed by incorporation into the frontal wedge and the downward migration of the MCT or subcretion of material to the HMC.

Abbreviations as for Figure 1.

Figure 4. Schematic model illustrating differing styles of distortion and displacement during southward extrusion of the Himalayan-mid-crust in the Early Miocene. The hinterland is dominated by extending flow - vertical thinning and lateral extruding of hot, ductile midcrustal rock from beneath the Tibetan Plateau. The extruded rock cools as it moves laterally along the Main Central thrust and intrudes rocks to the south. Tectonic wedging of the cooled mid-crust into the foreland strata is manifest as compressing flow - vertical thickening and horizontal shortening. Slip lines and streamlines within the Prandtl cell for extending flow (right) and for compressing flow (left) are illustrated. Abbreviations as for Figure 1. Modified from Larson et al., (2010), Kanizay, (1962) and Price, (1972).

Figure 5. Simplified, wireframe version of the geologic map in Figure 1A. White numbers in black circles mark known tectonometamorphic discontinuities: 1) Garhwal India (Spencer et al. 2010); 2) NW Nepal (Yakymchuk and Godin, 2012); 3) Mangri Shear Zone, Mugu Karnali (Montomoli et al., 2013); 4) Toijem Shear Zone, Lower Dolpo (Carosi et al., 2010); 5) Kalopani Shear Zone (Vannay and Hodges, 1996; Godin et al., 2001); 6) Bhanuwa Thrust, Sinuwa Thrust; Modi Khola (Martin et al., 2010; Corrie and Kohn, 2011; Nadin & Martin 2012); 7) Manaslu-Himal Chuli (Larson et al., 2010, 2011); 8) Langtang Thrust (Reddy et al., 1993; Fraser et al., 2000; Kohn et al., 2005;

Kohn, 2008); 9) Nylam thrust, Bhote Kosi (Wang et al., 2013); 10) Tama Kosi (Larson et al., 2012; Larson and Cottle 2014); 11) Hidden Discontinuity, Arun Window, Nepal (Groppo et al., 2009); 12) Khumbu thrust, Everest Region (Searle et al., 1997; 1999); 13) High Himalayan Thrust, Kanchenjunga (Goscombe et al., 2006; Imayama et al., 2010, 2012; Ambrose et al., In Press); 14) Northern Sikkim (Rubatto et al., 2013); 15) Laya-Kakhtang Thrust, Bhutan (Davidson et al., 1997; Grujic et al., 1996, 2002, 2011; Daniel et al., 2003; Long & McQuarrie 2010; Chambers et al., 2011; Warren et al., 2011); Bhutan Thrust (Swapp and Hollister, 1991); 17) Zimithang Thrust NE India (Warren et al., 2014).

Figure 6. Schematic cross-sections illustrating the development of inverted metamorphic isograds at the base of the Himalayan Metamorphic Core. In this model, the inverted metamorphic sequence is assembled via repeated foreland-directed thrusting. Once the thrust propagated towards the foreland, or the underplated material is added to the hanging wall, the process begins again at a higher structural level and farther toward the foreland juxtaposing progressively lower metamorphic grade samples across the active fault.

Figure 7. Schematic reconstructions of the evolution of the Himalayan mid-crust through time; oldest (A) through youngest (E). See text for explanation. Active structures are depicted in bold black lines whereas formerly active structures are shown in grey. Thin gray lines with long dashed indicate the future position of incipient structures. Thin gray lines with short dashes indicate the potential geometry of thrust-stacked material

1405 structurally below that exposed in the mapped area. The location of a representative
1406 tectonometamorphic discontinuity within the HMC is labeled in (d). White dashed lines
1407 indicate thickening of the Himalayan metamorphic core that occurred at different times in
1408 different locations along the length of the orogen. Areas of white fill are undifferentiated
1409 rock of Indian affinity. The vertical scale maintains approximately the same relative
1410 position within the overthrusting material throughout the diagram.

Figure

Cottle et al., Fig. 1

Cottle et al., Fig. 2

Cottle et al., Fig. 3

Cottle et al., Fig. 4

Cottle et al., Fig. 5

Cottle et al., Fig. 6

A) Early crustal thickening and (U)HP metamorphism 55-45 Ma

B) Continued crustal thickening, incipient lateral flow, local overthrusting in the HMC 43-28 Ma

C) Widespread lateral flow (hinterland) and tectonic wedging (foreland) 24-23 Ma

D) Duplication of the HMC in many locations underplating of material below HMC initiates 22-14 Ma

E) HMC & underplated material exhumed by duplexes in the LHS 10-3 Ma

Cottle et al., Fig. 7

Cottle et al., Graphical Abstract

A) Early crustal thickening and (U)HP metamorphism 55-45 Ma

B) Continued crustal thickening, incipient lateral flow, local overthrusting in the HMC 43-28 Ma

C) Widespread lateral flow (hinterland) and tectonic wedging (foreland) 24-23 Ma

D) Duplication of the HMC in many locations underplating of material below HMC initiates 22-14 Ma

E) HMC & underplated material exhumed by duplexes in the LHS 10-3 Ma