Lateral Heterogeneity in Compressional Mountain Belt Settings

Mary Hubbard[1,1] and Bibek Giri[1,1]

[1]Montana State University

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Abstract

Convergent orogens are typically linear with laterally continuous, orogen-parallel folds and thrusts. Over the years, geoscience research has revealed evidence for important orthogonal/cross structures as well as lateral heterogeneity in deformation style, igneous activity, metamorphic grade, geomorphology, and seismic activity. To assess the occurrence, causal mechanisms, and implications of these lateral heterogeneities, a selection of convergent orogens, with different tectonic settings and history are reviewed. The Appalachians, the North American Cordillera, the Alps, the Himalayas, the Zagros, the Andes, and several other belts all exhibit a degree of lateral heterogeneity. Major factors driving the lateral heterogeneity and/or cross structures include the pre-existing deformational history of the cratonic blocks involved, lateral change in lithology of crustal rocks, variations in crustal/lithospheric rheologic properties, or changes in plate kinematics. The Appalachian orogenic front mimics the Iapetan rift margin. Pre-existing basement structures have control on pre- and syn-orogenic sedimentation, which subsequently impacts how an orogenic wedge evolves. A thicker sedimentary column generally evolves into a salient (as opposed to a recess), which is further enhanced by the presence of weak horizons as seen in the Zagros and the Cordillera. Lateral variation in sedimentary facies also creates changes in thrust-ramp geometry. During orogenic contraction, inherited basement structures can be preferentially reactivated based on their orientation. Several cross faults in the Himalayas spatially coincide with orogen-perpendicular, lower plate, basement structures. In a similar way, oceanic subducting plate physiography can also influence deformation in the overriding plate. Along-strike variations in subduction dynamics have been reflected in the Andean deformation. Orogenic extension in the Alps has been accompanied by a system of orogen-parallel strike slip faults and extensional cross faults. It is evident that lateral heterogeneities can form crucial control on the evolution of orogenic belts and can influence seismic rupture patterns, resource occurrence, and landslide-related natural hazards.

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Bibek Giri and Mary Hubbard
Department of Earth Sciences, Montana State University, Bozeman

Corresponding author: Bibek Giri (bibekgiri@montana.edu)

Key Points

- Lateral heterogeneity, expressed as sharp lateral variations, has been observed in most orogens and often coincides with cross structures.
- Major controls are basement structures, lateral changes in sedimentary facies, varying crustal rheology, and changing plate dynamics.
- Lateral heterogeneities impact orogenic evolution and have geomorphic, seismic, and economic implications.

Abstract

Convergent orogens are typically linear with laterally continuous, orogen-parallel folds and thrusts. Over the years, geoscience research has revealed evidence for important orthogonal/cross structures as well as lateral heterogeneity in deformation style, igneous activity, metamorphic grade, geomorphology, and seismic activity. To assess the occurrence, causal mechanisms, and implications of these lateral heterogeneities, a selection of convergent orogens, with different tectonic settings and history are reviewed. The Appalachians, the North American Cordillera, the Alps, the Himalayas, the Zagros, the Andes, and several other belts all exhibit a degree of lateral heterogeneity. Major factors driving the lateral heterogeneity and/or cross structures include the pre-existing deformational history of the cratonic blocks involved, lateral change in lithology of crustal rocks, variations in crustal/lithospheric rheologic properties, or changes in plate kinematics. The Appalachian orogenic front mimics the Iapetan rift margin. Pre-existing basement structures have control on pre- and syn-orogenic sedimentation, which subsequently impacts how an orogenic wedge evolves. A thicker sedimentary column generally evolves into a salient (as opposed to a recess), which is further enhanced by the presence of weak horizons as seen in the Zagros and the Cordillera. Lateral variation in sedimentary facies also creates changes in thrust-ramp geometry. During orogenic contraction, inherited basement structures can be preferentially reactivated based on their orientation. Several cross faults in the Himalayas spatially coincide with orogen-perpendicular, lower plate, basement structures. In a similar way, oceanic subducting plate physiography can also influence deformation in the overriding plate. Along-strike variations in subduction dynamics have been reflected in the Andean deformation. Orogenic extension in the Alps has been accompanied by a system of orogen-parallel strike slip faults and extensional cross faults. It is evident that lateral heterogeneities can form crucial control on the evolution of orogenic belts and can influence seismic rupture patterns, resource occurrence, and landslide-related natural hazards.

1 Introduction

Many of the world's major mountain belts are found along modern or former convergent plate boundaries and they are typically dominated by major thrust structures (Elliott, 1976; Chapple, 1978; Boyer & Elliott, 1982; Dahlen, 1990). The terms 'continental collision or terrane accretion' have been used for convergent boundaries when both the involved plates are continental, though mountain building also occurs along convergent plate boundaries where an oceanic plate is subducting beneath continental crust (e.g.; Dewey & Bird, 1970). In both tectonic settings, compressive stress is accommodated through imbrication and shortening of upper crustal material as thrust sheets and/or nappes, within a system of fault surfaces, which generally strike perpendicular to the direction of compression (or parallel to the orogen). In general, these faults and related structures are laterally continuous both temporally and spatially. Detailed geological and geophysical research, however, has found that while these convergent mountain belts exhibit overall lateral continuity, there are also lateral heterogeneities that have made an important contribution to the development of the mountain belt. Driven by various mechanisms, these lateral changes serve to induce important structural, geological, and geomorphological lateral heterogeneities along mountains belts that may (or may have) influence(d) igneous or seismic processes or even the occurrence of economic resources. Commonly, significant along-strike changes are known to occur across structures, which strike orthogonal to the overall trend of the orogen. In this article, the term 'cross structure' has been utilized to refer to all generic, syn and post orogenic structural features that bound lateral changes along a mountain belt, though the terms 'cross-strike structural discontinuities' (CSD), 'transverse zones', and 'cross faults' have all been utilized to describe large-scale structures that strike across convergent mountain belts (Drahovzal & Thomas, 1976; Wheeler, 1980; Hubbard et al., 2018; Hubbard et al., 2021). Thomas (1990) suggested that transverse zones (TZ) along fold and thrust belts are domains of several, structurally aligned 'lateral connectors', which encompass transverse/tear faults, lateral ramps, and displacement transfer faults/zones. In general, the term 'transverse zone' has been used in the literature to denote large-scale, inherited basement features with a high-angled orientation to the evolving orogen. Lateral connecters like tear faults, lateral ramps, and transfer faults/zones form during thrust propagation and are rather localized, only spanning across a few thrust sheets. Formed during the propagation of thrust belts, these lateral connectors could be spatially controlled by the inherited structures in the crystalline basement as well as the overlying stratigraphy of the evolving thrust sheet (Thomas, 1990). During orogenesis, inherited basement structures are commonly reactivated or inverted and impart a great control on the tectonic and structural evolution of mountain belts (e.g.; Tavarnelli et al., 2004; Butler et al., 2006). Evidence has also been presented suggesting that oceanic transform faults or seamounts in a subducting oceanic plate can influence the structural development of the mountain belt in the overriding continental plate, thus causing lateral heterogeneity or the presence of cross structures (e.g.; Gutscher et al., 2000; Chapin, 2012).

Mountain belts formed in convergent plate tectonic settings at different time periods in Earth's history all exhibit some degree of lateral heterogeneity. The following review explores a selection of mountain belts from convergent boundaries, examining the proposed driving mechanisms and implications of lateral heterogeneities and the presence of cross structures during and after orogenesis. The Appalachians, the North American Cordillera, the Alps, the Himalayas, the Zagros, and the Andes have been looked at in detail, and several other mountain belts are briefly discussed (Figure 1). This selection represents some of the most well studied and prominent mountain belts around the world and presents a wide array of tectonic settings. Within each mountain belt, presented examples of lateral heterogeneities and cross structures are representative of the lateral variability for that orogen. For each mountain belt, we present a brief introduction of the tectonic setting, observed lateral heterogeneities, and proposed causal factors for those heterogeneities. It is noted that lateral heterogeneities can be expressed in several ways including: 1) along-strike changes in deformation style; 2) variation in igneous activity or metamorphic grade; 3) variation in seismic activity; 4) changes in geomorphology; or 5) abrupt lateral stratigraphic change. Drivers of these along-strike changes can include irregularities along continental margin, inherited basement structures of an overriding plate, features from a subducting plate (oceanic) or footwall block (continental), lateral variation in lithospheric or crustal properties, lateral changes in thickness and facies of the deforming sedimentary sequence, lateral variation in dynamic plate tectonic setting or a combination of these factors. Apart from their influence in the structural/tectonic evolution of the mountain belt, the lateral heterogeneity may also have important seismic, geomorphic, and economic implications.

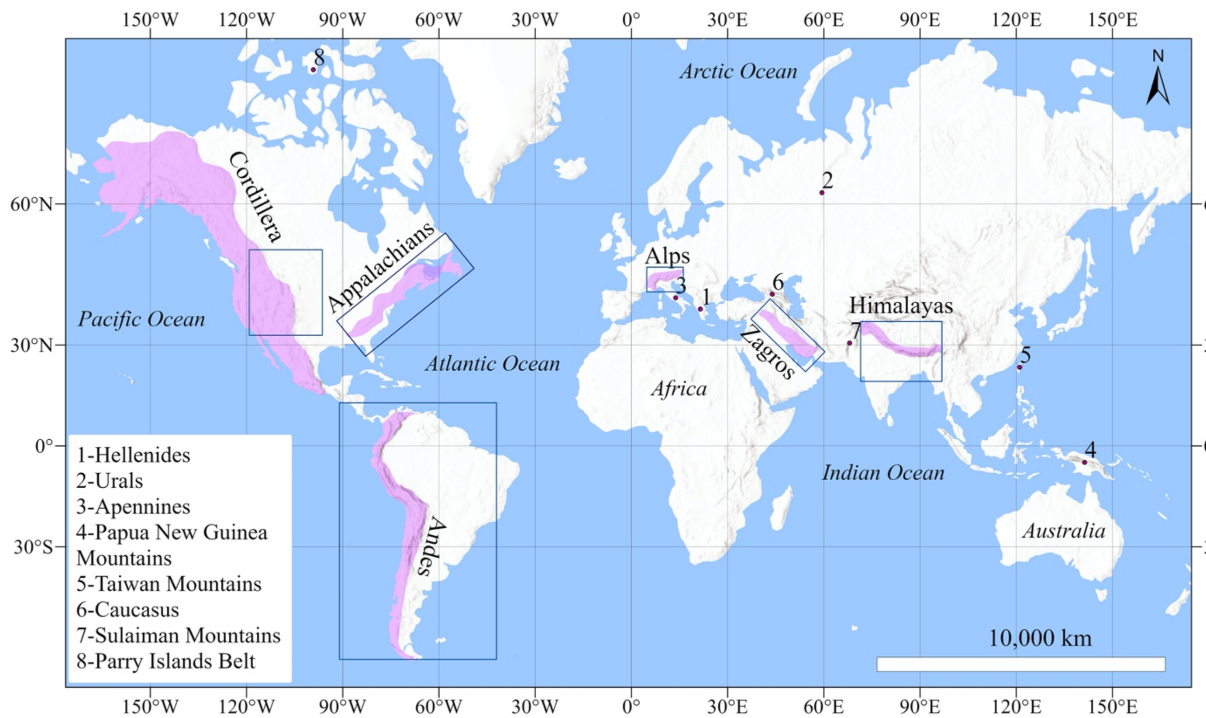


Figure 1: Map of the world showing locations of mountain belts described in this review. The pink shades are spatial extent of respective orogenic belts. Rectangular boxes indicate the extent of tectonic maps presented in the following sections. Numbered areas are other convergent orogens mentioned in the text.

2 Appalachians

2.1 Tectonic Setting and Lateral Heterogeneities

The Appalachian orogenic belt, along the eastern margin of the North American continent, is the structural culmination of a prolonged tectonic history consisting of multiple episodes of subduction, terrain accretion, continental collision, and rifting (Figure 2). The prominent Grenville orogeny (1.3-0.9 Ga; Li et al., 2008; Hynes & Rivers, 2010; Rivers, 2012)) was the coalescence of the Laurentian (modern day North American continent) and Amazonian cratons, which formed the Rodinia supercontinent (Li et al., 2008; Rivers, 2008, 2015). Following the Grenville orogeny, the Laurentian craton remained tectonically inactive for about 250 Ma before being detached from the Rodinia craton in various episodes from 759±12 Ma to 530 Ma (Thomas, 1991; Aleinikoff et al., 1995; Hogan & Gilbert, 1998). The Iapetan rifting, during which the Iapetus ocean opened between the Laurentian and Amazonian (Gondwana) cratons (Domeier, 2016), marked the end of this series of tectonic events (Rankin, 1976; Thomas, 1977; O'Brien & van der Pluijm, 2012). The Middle Ordovician to Late Silurian Taconian orogeny represents the aggregate contractional tectonic activity along the boundary between the Laurentian and Iapetan plates (Hatcher, 1972; Drake et al., 1989; van Staal et al., 2007; Hatcher, 2010). The Late Silurian to Devonian Acadian orogeny has been interpreted to have formed along an Andean-type margin (flat slab subduction) following the accretion of Avalonia to the Laurentia margin (Keppie & Krogh, 1999; Murphy & Keppie, 2005). The youngest event in the building of the Appalachians was the Late Mississippian to Late Permian Alleghenian orogeny (Laurentia and Gondwana) (Hatcher et al., 1989), which also marked the formation of the Pangean supercontinent. In the Mesozoic, the modern day North American plate rifted westward from the African continent. Dominant structures from all of these convergent and divergent tectonic events are range-parallel faults and fold axes. Close inspection, however, reveals lateral heterogeneity that includes changes in the mountain belt orientation, structural style, metamorphic grade, and the presence of cross structures.

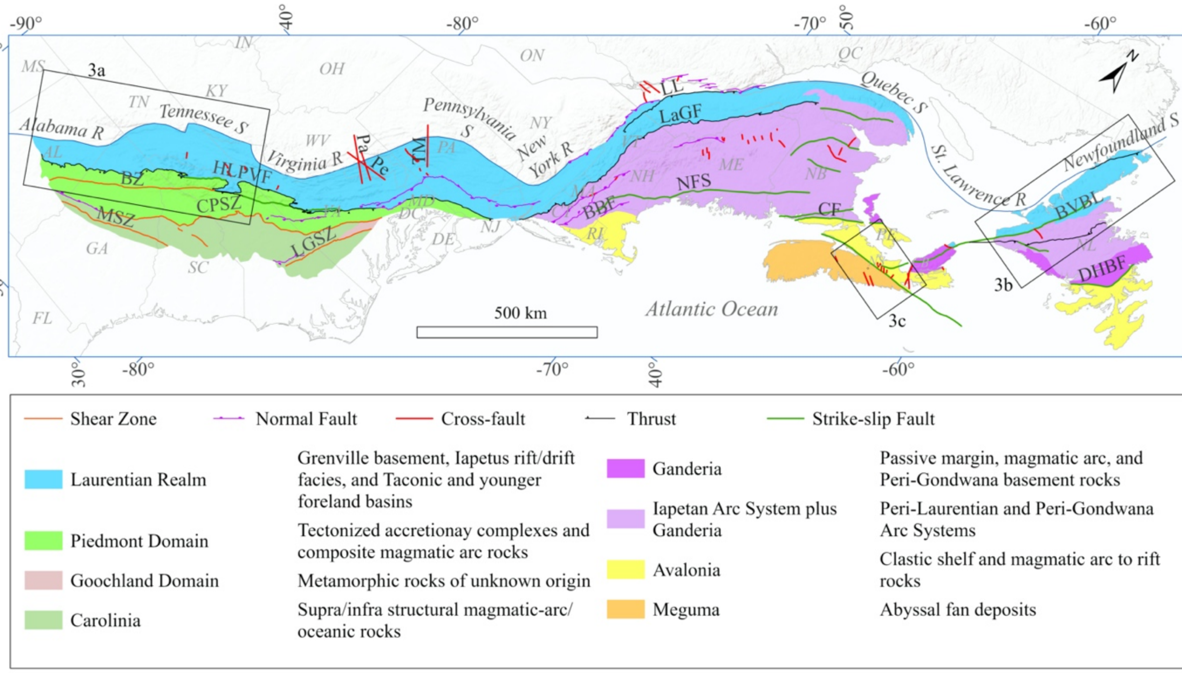


Figure 2: Simplified tectonic map of the Appalachians belt showing the major faults, the salient (S) and recesses (R), and the extent of the Appalachian deformation front. Insets show the locations of figures 3a, 3b, and 3c (simplified from: Southworth, 1986; Hibbard et al., 2006; Waldron et al., 2015).

Acronyms: BBF: Bloody Bluff Fault; BVBL: Baie Verte-Brompton Line; BZ: Brevard Zone; CF: Caledonia Fault; CPSZ: Central Piedmont Shear Zone; DHBF: Dover Hermitage Bay Fault; HLPVF: Hollins Line-Pleasant Valley Fault System; LaGF: La Gaudaloupe Fault; LGSZ: Lake Gordon Shear Zone; LL: Logan's Line; MSZ: Modoc Shear Zone; NFS: Norumbega Fault System; Pa: Parson's Lineament; Pe: Petersburg Lineament; TM: Tyrone-Mount Union lineament.

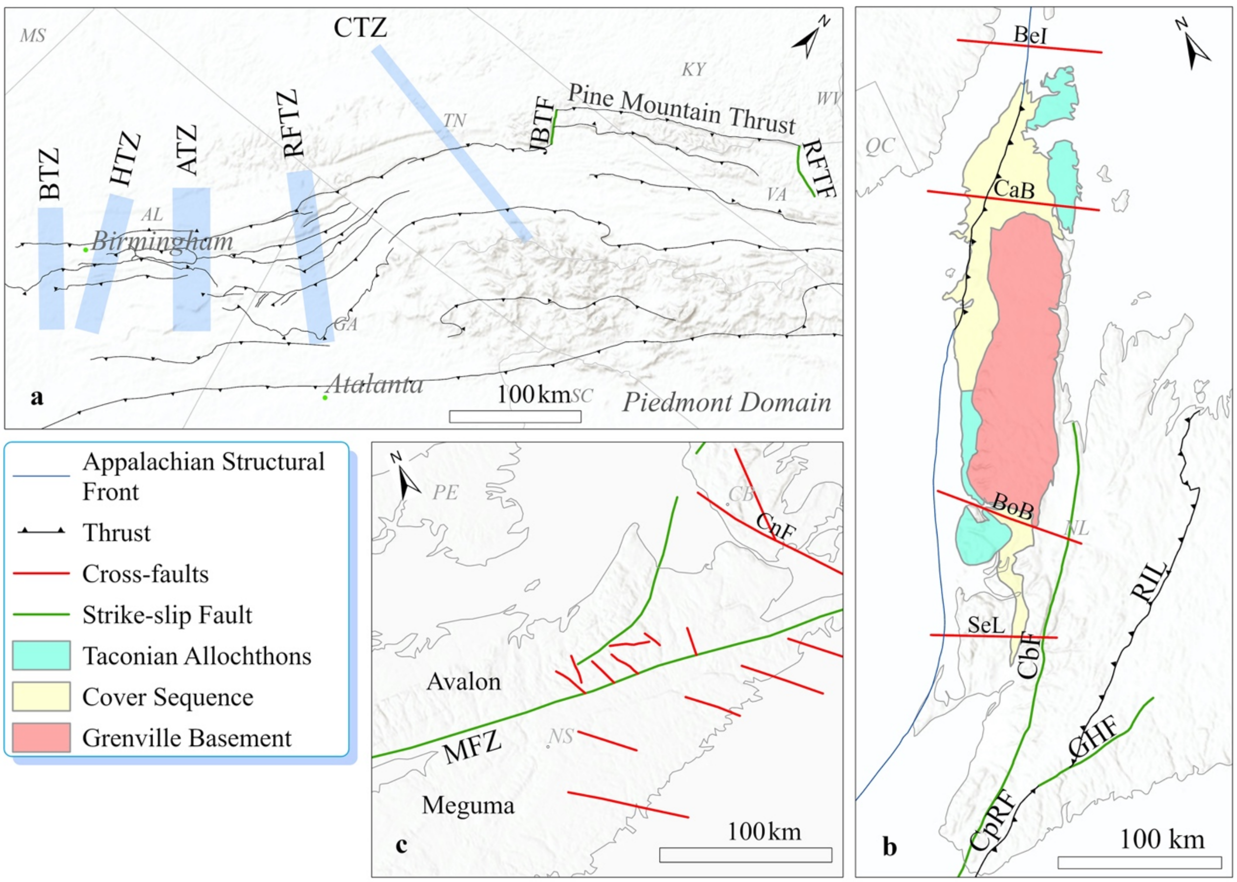


Figure 3: Cross structures along the Appalachians. a) Map of the Alleghenian fold and thrust belt in the Alabama recess and the Tennessee salient showing the major transverse zones (light blue rectangles) and the tear faults on either side of the Pine Mountain (source:Mitra, 1988; Cook & Thomas, 2009). b) Tectonic-sketch map of the Newfoundland Island showing the major cross-faults. Note the lateral termination of the Grenville basement inliers at the Bonne Bay cross-strike discontinuity (BoB) and the Canada Bay (CaB) cross-strike discontinuity. (source: Cawood & Botsford, 1991; Hibbard et al., 2006; Waldron et al., 2015). c) The Minas Fault Zone (MFZ)

between the Avalon and Meguma terranes and its conjugate faults including the Canso Fault (CnF) (source: Hibbard et al., 2006; Waldron et al., 2015).

Acronyms: ATZ: Anniston Transverse Zone; BeI: Belle Isle cross-strike discontinuity; BoB: Bonne Bay cross-strike discontinuity; BTZ: Bessemer Transverse Zone; CaB: Canada Bay cross-strike discontinuity; CbF: Cabot Fault; CnF: Canso Fault; CpRF: Cape Ray Fault; CTZ: Cartersville Transverse Zone, GHF: Gunflap Hills Fault; HTZ: Happersville Transverse Zone; JBTF: Jacksboro Tear Fault; MFZ: Minas Fault Zone; RFTF: Russel Fork Tear Fault; RFTZ: Rising Fawn Transverse Zone; RIL: Red Indian Line; SeL: Serpentine Lake cross-strike discontinuity.

One of the obvious heterogeneities in topography is the fact that the NE-trending Appalachian mountain front curves around a series of recesses and salients along the length of its western margin (Figure 2). These recesses and salient are proposed to be bound by broad transverse structures that offset or truncate thrust structures (Thomas, 1990). In the southern Appalachians, the large-scale thrust faults and related structures are predominantly striking northeast and verging towards the northwest (Bayona et al., 2003). Abrupt along-strike terminations of these structures are known to spatially coincide with these transverse zones (TZ). In the Alabama recess (Figure 3a), major TZs include the Bessemer Transfer Zone (TZ), the Happersville TZ, the Anniston TZ, and the Rising Fawn TZ from southwest to northeast respectively (Thomas, 1990; Bayona et al., 2003; Brewer, 2004). In general, these transfer zones are tens of kilometers wide, hundreds of kilometers long, and cut into the Proterozoic basement rocks (Thomas & Bayona, 2002; Brewer, 2004). Thrust-related structures plunge into these TZs, abruptly terminate at lateral or oblique ramps, show offsets along transverse faults (tear faults), or break into en-echelon segments of transfer zones (Thomas & Bayona, 2002; Thomas, 2007; Cook & Thomas, 2009). Structural strike deviates within these TZs, while propagating over oblique/lateral ramp and irregular basement and at junctions between lateral/oblique and frontal ramps (Bayona et al., 2003). Above a lateral/oblique ramp, hanging wall rocks are generally folded into upright, isoclinal folds driven by local-scale, ramp perpendicular contraction in the hanging wall due to crowding of materials along that trend (Cook and Thomas 2009). Across the Cartersville TZ, deformation style, metamorphic grade, and thrust sheet stratigraphy show sharp contrast between the Alabama recess and the Tennessee salient (Tull & Holm, 2005). In the salient, passive margin and foreland sedimentary sequences are much thicker and have been deformed and metamorphosed (biotite grade of the Barrovian metamorphic series) more strongly than in the recess (Tull & Holm, 2005). Towards the adjoining Virginia recess, a section of the Pine Mountain belt, bounded by the left-lateral Jacksboro tear fault and the right lateral Russel Fork tear fault (Figure 3a), demonstrates a double ramp geometry in the middle section, while a single ramp geometry around the edges (Mitra, 1988). Presence of the orogen-parallel, large-scaled (about 35 km wide, (Brewer, 2004)) Birmingham graben beneath the southern Alleghenian thrust belt also had major controls on the foreland sedimentation and subsequent propagation of the thrust sheet (Thomas, 2007). Transverse zones in the southern Appalachians have been interpreted as inherited transform fault related to the Iapeton rifting along the continental margin of Laurentia (e.g.; Figure 4) (Rankin, 1976; Thomas, 1977, 2014).

Cross-structures in the Appalachians of Pennsylvania and West Virginia (central Appalachians) such as the Parsons, Petersburg, and Tyrone-Mount Union lineaments (Figure 2) also have similar effects on the orogen-parallel structures as is seen in the southern Appalachians (Wheeler, 1980; Southworth, 1986). In the northern Appalachians, however, genesis of cross-structures presents much more diversity. The Serpentine Lake cross-strike discontinuity (CSD), the Bonne Bay CSD, the Canada Bay CSD, and Belle Isle CSD in Newfoundland mark the

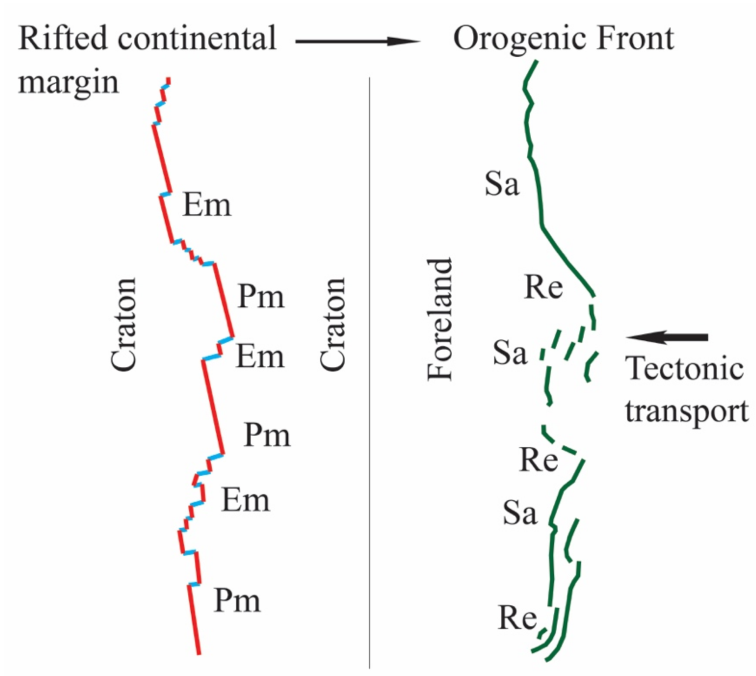


Figure 4: Impact of the continental margin geometry on the geometry of an orogenic front. Embayments (Em) evolve into salients (Sa) and Promontories (Pm) evolve into recesses (Re).

boundaries of second-order salients and recesses and had controls on the Taconic and younger orogenic events (Figure 3b) (Cawood & Botsford, 1991). Across these structures, the depositional facies and structural style of the rock units overlying the Grenville basement change abruptly suggesting these faults have Precambrian roots. The Grenville basement inlier terminates southward at the Bonne Bay CSD (Figure 3b), which also delineates the southern limit of the Acadian uplift of the Grenville Inlier (Cawood & Botsford, 1991). North of the Bonne Bay CSD, thrusts and associated structures are west verging, while towards the south the vergence reverses (Williams & Cawood, 1986). In southern Newfoundland and the Cape Breton islands, the Silurian tectonic wedging of the Avalon craton into the Laurentian craton has resulted in a very narrow Appalachian central mobile belt characterized by stronger syncollisional deformation, high-grade Barrovian metamorphism (‘Kyanite and Sillimanite grade, 700˚-750˚C at 6-10 Kbar (Plint & Jamieson, 1989), 5 kbar elsewhere’), and a greater volume of Silurian magmatism than elsewhere (Lin et al., 1994). This intensely deformed and metamorphosed narrow band is bound by the NW-striking Canso fault (Figure 3c) to the southwest and the E-striking Gunflap Hills fault to the northwest, interpreted as transverse, right lateral wrench faults that accommodated the differential movement on either side of the wedging (Lin et al., 1994). In its western end, the Gunflap Hills fault (Colman-Sadd et al., 1992) merges with the NW-striking Cape Ray fault zone (Figure 3b). In eastern Nova Scotia, several NW-striking, transverse, left-lateral strike-slip faults (e.g. the Country Harbour Fault) have been described as conjugates of the Minas Fault Zone (Figure 3c) (Waldron et al., 2015), which is an E-W striking 300 km long, continental transform fault in New Brunswick and Nova Scotia that marks the boundary between the Avalon and Meguma terranes (Murphy et al., 2011). Towards the west, the Paleozoic Norumbega Fault System (Figure 2) demonstrates the effect of the subducting plate physiography on the upper plate structures. This NE-striking, 300 km long, transpressive, dextral fault system (West Jr & Hubbard, 1997; Hubbard, 1999) has recently been interpreted to be a result of subduction of an oceanic ridge and related transform fault (Kuiper, 2016; Kuiper & Wakabayashi, 2018). Around the orogenic front in Quebec and northern Vermont, cross structures were formed due to reactivation of earlier-formed transverse, normal faults with an oblique thrust-sense or served as an oblique ramp during the westward thrusting of the Lower to Middle Ordovician shelf-carbonates (Séjourné & Malo, 2007).

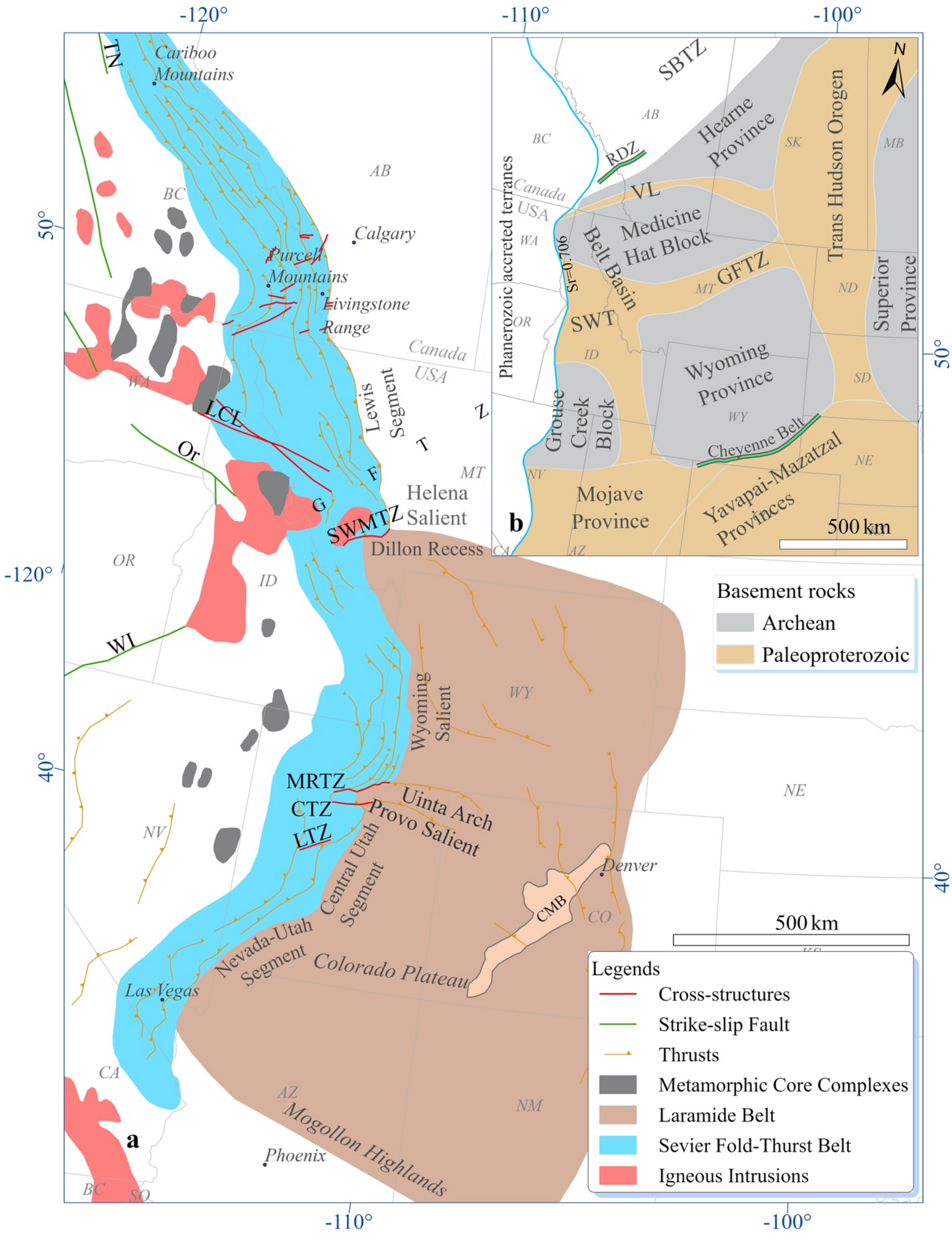
2.2 Proposed Factors Controlling Lateral Heterogeneities

Lateral heterogeneity along the Appalachians has largely been controlled by the irregularities along pre-collisional continental margins and by inherited basement structures. The eastern continental margin of Laurentia went through a series of rifting events including the Iapetan rifting, all of which acted in concert to develop a series of embayments (concave oceanward) and promontories (convex oceanward) separated by transform faults (Rankin, 1976; Thomas, 1977, 2014). During subsequent orogenic events (Taconic, Acadian, and Alleghenian orogenies), embayments evolved into salients (convex towards the foreland), promontories evolved into recesses (concave towards the foreland), and the transform fault boundaries evolved into transverse zones (Figure 4). As Tull and Holm (2005) noted, foreland sedimentary sections along salients are much thicker than along recesses, which has a great control on the geometry and deformation style of the evolving thrust sheets (discussed in detail in the North American Cordillera section of this article). Irregularity along continental margins also has another important control, which results in a sharp lateral heterogeneity along an orogen. A collisional event between two promontories (of two different landmasses) results in a narrower but stronger orogenic deformation and a higher grade of metamorphism that a promontory-embayment collision (Lin et al., 1994). The evolution of rift-related transform faults into transverse zones also had a profound impact on the evolution of the Appalachians mountains by creating barriers or zones that truncated or offset major thrust faults. Along the range front, transverse zones have served as nucleation sites for cross structures in the form of lateral ramps or tear faults.

3 Cordillera

3.1 Tectonic Setting and Lateral Heterogeneities

The North American Cordilleran orogenic system records the Jurassic (and possibly earlier) to Paleogene history of terrane accretion and the eastward convergence of the Farallon and Kula plates beneath the western margin of Laurentia (North American continent) (Oldow et al., 1989; Burchfiel et al., 1992; Saleeby et al., 1992; DeCelles, 2004; Dickinson, 2004). This convergence


-120°
-110°
-100°
50°
40°
-120°
TN
Cariboo Mountains
AB
BC
Calgary
Purcell Mountains
Livingstone Range
Canada
USA
WA
LCL
Lewis Segment
Z
T
Or
F
G
MT
Helena Salient
SWMTZ
Dillon Recess
OR
ID
WI
WY
Wyoming Salient
MRTZ
CTZ
LTZ
Uinta Arch
Provo Salient
Central Utah Segment
NV
NE
Denver
CMB
CO
Nevada-Utah Segment
Colorado Plateau
Las Vegas
CA
AZ
NM
Mogollon Highlands
Phoenix
a
SBTZ
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AB
BC
Hearne Province
SK
MB
Trans Hudson Orogen
RDZ
Phanerozoic accreted terranes
Canada
USA
VL
WA
Sr=0.706
Belt Basin
Medicine Hat Block
GFTZ
MT
ND
Superior Province
OR
SWT
ID
Wyoming Province
SD
WY
Grouse Creek Block
Cheyenne Belt
Yavapai-Mazatzal Provinces
NE
NV
Mojave Province
b
500 km
Basement rocks
Archean
Paleoproterozoic
500 km
Legends
Cross-structures
Strike-slip Fault
Thrusts
Metamorphic Core Complexes
Laramide Belt
Sevier Fold-Thurst Belt
Igneous Intrusions
-110°
-100°

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Figure 5: Simplified tectonic and basement map around the western North American Cordilleran belt. a) Tectonic map of the US and Southern Canadian Cordillera showing the major orogenic curvature and the major cross-structures (modified from: Moffat & Spang, 1984; McGugan, 1987; Kwon & Mitra, 2006; Cooley et al., 2011; McMechan, 2012; Whisner et al., 2014; Yonkee & Weil, 2015) b) Generalized basement map of the western North America. The Sr=0.706 line represents the eastern boundary of the accreted terranes (source: Foster et al., 2006; Whitmeyer & Karlstrom, 2007; McMechan, 2012; Yonkee & Weil, 2015; Ronemus et al., 2020).

Acronyms: CMB: Colorado Mineral Belt; CTZ: Charleston Transverse Zone; GFTZ: Great Falls Tectonic Zone; LCL: Lewis and Clark Line; LTZ: Leamington Transverse Zone; MRTZ: Mount Raymond Transverse Zone; Or: Orfino Shear Zone; RDZ: Red Deer Zone; SBTZ: Snowbird Tectonic Zone (the Thorsby Low is located within the SBTZ); SWMTZ: Southwestern Montana Transverse Zone; SWT: Selway Terrance; TN: Tintina-Rocky Mountain Shear Zone; VL: Vulcan Low; WI: West-Idaho Fault.

resulted in a number of different ranges including the Sierra Nevada, the Rocky Mountains, and the various mountain systems of Alaska (Figure 5a). Changing plate dynamics has resulted in varied deformation style and distribution throughout the Cordilleran evolution. The Late Cretaceous (Beck et al., 1988; Decelles et al., 1991) transition of deformation style from the thin-skinned, taper-wedge in the US Sevier fold-thrust belt to predominantly thick-skinned, basement-involved Laramide orogenesis is one of the remarkable changes. Initiation of the Laramide system uplifts has generally been correlated to the subduction of a flat-slab segment (e.g.; Coney & Reynolds, 1977; Dickinson & Snyder, 1978; Saleeby, 2003; Chapin, 2012). North of the Canadian border, the Laramide-style deformation is absent. Orogenic collapse of the Cordilleran system, focused on the Sevier belt, initiated during Middle to Late Eocene due to the change in relative plate kinematics and has resulted in the active Basin and Range province of extensional deformation (e.g.; Constenius, 1996; Yonkee & Weil, 2015). The contractional phase of deformation resulted in spectacular range-parallel fold and thrust belts (Yonkee & Weil, 2015), yet there are important lateral heterogeneities and a number of cross structures along the Cordillera.

Basement provinces and their suture zones and major faults along the western margin of the North American continent include the Snowbird Tectonic Zone (Ross et al., 1991; Hope &

Eaton, 2002), the Archean Hearne Province (Hope & Eaton, 2002), the Archean Medicine Hat Block (Ross et al., 1991; Lemieux et al., 2000); the Archean Wyoming Province (Wooden & Mueller, 1988; Mogk et al., 1992), and the Paleoproterozoic Yavapai and Mazatzal Provinces (Foster et al., 2006; Whitmeyer & Karlstrom, 2007), from north to the south (Figure 5b). The Trans-Hudson Orogen and the Superior Craton lie towards the east, around the central region of the North America. Towards the west, the basement block include the Archean Grouse Creek Block and the Selway Terrane (Foster et al., 2006).

Transverse crustal boundaries between these provinces and the other Archean/Paleoproterozoic transverse basement structures are one of the primary causes of lateral heterogeneity in the Cordillera due to their influence in the subsequent sedimentary and deformational history (e.g.; Paulsen & Marshak, 1999; Sears & Hendrix, 2004; McMechan, 2012). Significant crustal-scale, transverse boundaries around the southeastern Canadian Cordilleran front (Alberta) include the Thorsby Low, a crustal suture within the Snowbird tectonic zone (Ross et al., 1991; Hope & Eaton, 2002), the Red Deer zone, the northern boundary of the Hearne Province (Hope & Eaton, 2002), and the Vulcan Low, a crustal suture between the Medicine Hat Block and Hearne Province (Figure 5b) (Hoffman, 1988; Ross et al., 1991). These structures and terrane boundaries served as loci for activation of transverse zones during the deposition of the Mesoproterozoic Belt-Purcell sequence (Benvenuto & Price, 1979; Foo, 1979; Root, 1987; Anderson & Davis, 1995; McMechan, 2012). Multiple other transverse structures were also initiated in Mississippian and Triassic as normal faults above these transverse boundaries (Cooley et al., 2011; McMechan, 2012). Segments of some of these transverse zones were reactivated as dextral, oblique reverse faults or as tear faults during Jurassic to Eocene in response to compressional stresses (Bielenstein, 1969; Foo, 1979; Price, 1981; McMechan & Price, 1982; McMechan, 2000; Cooley et al., 2011). Similar impacts of basement structures on the evolution of thrust belts have been reported from other domains in the Canadian Cordillera (e.g.; Pride et al., 1986; Thompson, 1989; Pilkington et al., 2000; Berger et al., 2008). In northern British Columbia, a geometric inversion of the northern end of a NW-trending “basement uplift” during the Sevier contraction resulted in the formation of NE to E-trending (transverse) contractional structures, which were superimposed on the regional NW-striking structures (McMechan, 2007). Aside from their influence in the sedimentary and deformational history, these basement structures and terrane boundaries also formed suitable conduits for melt transport and served as loci for igneous intrusions and mineralization (McMechan, 2012).

The Sevier FTB is generally classified into the Main Ranges, the Front Ranges, and the Foothills, from the hinterland to the foreland, in the northern US and the southern Canada. Cross structures have been observed in all these sub-divisions at different scales. In the Main Ranges of British Columbia (Purcell Mountains), Mesoproterozoic and Eocambrian synsedimentary faults have been reactivated as dextral, east-striking, transverse, thrust faults (Figure 5a) during Mesozoic and Cenozoic (Höy & Heyden, 1988; Larson et al., 2006; McMechan, 2012). Several transverse faults have also been identified in the Front Ranges (e.g.; Moffat & Spang, 1984; McGugan, 1987; Fermor, 1999; McMechan, 2001; Norris, 2001). In the foothills, the north to NW-trending Livingstone Range anticlinorium consists of en-echelon, chevron-style fault-propagation folds, which have been compartmentalized by multiple tear faults and transverse zones of intense fracturing (Cooley et al., 2011). Anticlines plunge into these tear faults and form domal culminations. Tear faults also mark along-strike changes in fold-style, from chevron to concentric and changes in the presence or absence of back-thrusts. These tears are thought to be the results of syn-thrusting reactivation of preexisting NE-striking Paleoproterozoic basement faults, some of which were active during Mississippian (Cooley et al., 2011).

The Great Falls Tectonic Zone (GFTZ; Figure 5) in Montana is a NE-trending, transverse (in relation to the Cordillera) crustal suture between the Medicine Hat Block and the Wyoming Craton (O'Neill & Lopez, 1985). This deformation zone was initiated prior to ~1.9 Ga as a convergent boundary and subsequently evolved into a transpressive boundary in response to the eastward movement of the Wyoming Craton (Dahl et al., 1999; Mueller et al., 2002; Gifford et al., 2014). Cenozoic alkali magmatism, belonging to the Montana Alkali Province, occurred within this zone (Gifford et al., 2014). In central Montana and Idaho, the Mesoproterozoic Belt rifting most likely led to the initiation of the Lewis and Clark fault zone from the southern margin of the GFTZ (Harrison et al., 1974; Reynolds, 1979; Wallace et al., 1990; Sears & Hendrix, 2004; Foster et al., 2006). The Lewis and Clark fault zone (Figure 5a) is a WNW-trending, >800 km long, system of 'steeply dipping strike-slip, oblique-slip, and dip-slip faults', which obliquely cuts across the Sevier belt (Foster et al., 2007). During the Mesozoic-Cenozoic Cordilleran orogeny, this fault zone deformed as a ca. 40 km wide, sinistral shear zone with 'transpressional flower structures' (Hyndman et al., 1988; Sears et al., 2000; Sears & Hendrix, 2004). With the onset of the Basin and

Range extension in the Eocene, polarity of the fault movement reversed, and this zone served as a dextral, extensional, TZ that facilitated the exhumation of metamorphic core complexes (Figure 5a) (Foster et al., 2007).

In the Sevier FTB (USA), the geometry of basement structures had first-order control on the formation of orogenic curvatures and on the evolution of their transverse boundaries (e.g.; Paulsen & Marshak, 1999). A deeper basin generally corresponds to a thicker sedimentary column and more material available to be incorporated into the deforming taper, which results in a wider wedge (salient) (e.g.; Marshak & Wilkerson, 1992; Boyer, 1995). The Helena salient likely formed over an E-trending, asymmetrical depositional trough of the Middle Proterozoic Belt basin called the Helena embayment (Figure 5b) (Harrison et al., 1974). A north-dipping Middle Proterozoic normal fault on the southern edge of the embayment probably evolved into the Southwest Montana Transverse Zone (SWMTZ), which forms the southern boundary of the salient with the Dillon recess (Whisner et al., 2014). Thrusts and related structures in the southern domain of this salient are strongly converging into the right-lateral, reverse faults within the SWMTZ, due to a gradual clockwise rotation of the shortening direction during their evolution (Whisner et al., 2014).

Further south, controls of the basin boundary geometry on the evolution of transverse zones have been exemplified by the Mount Raymond Transverse Zone (MRTZ) and the Charleston TZ. The MRTZ and the Charleston TZ form the northern and the southern boundaries of the Uinta/Cottonwood arch (recess) with the Wyoming Salient and the Provo Salient respectively (Figure 5a) (Paulsen & Marshak, 1997, 1998). Paulsen and Marshak (1999) noted contrasting structural styles between these two zones and explained this contrast in light of corresponding basement structures. An east-west trending asymmetric basement high, with a gentle northern flank and a steep southern flank, existed just north of the present Uinta/Cottonwood arch, along the boundary between the Archean Wyoming province (north) and Proterozoic terranes (south) (Paulsen & Marshak, 1999). A gentler northern flank meant that the sedimentary thickness gradually increased northwards from the Uinta recess into the Wyoming Salient. The MRTZ initiated above this flank as NNE-trending thrusts along the southern margin of the Wyoming salient, which were later tilted northward creating an E-W strike during the uplift of the Uinta/Cottonwood arch (Paulsen & Marshak, 1997). The steeper southern flank, however, marked an abrupt increase in sedimentary thickness towards the south and thereby formed the boundary between two contrasting taper wedges. The Charleston TZ (Figure 5a) served as zone of lateral ramp between the two contrasting tapers and gradually evolved into a left-lateral strike slip zone, which accommodated the differential motion between the Uinta recess and the Provo salient (Paulsen & Marshak, 1998 and references therein). The southern boundary of the Provo salient with the central Utah segment is the Leamington TZ, which is an ENE trending, >50 m long, complex, cross structure (Lawton et al., 1997; Kwon & Mitra, 2006). In the salient, an initial E-directed vergence over the TZ rotated clockwise during subsequent deformational phases, which likely reflects the interaction between a deforming wedge and an oblique ramp (Lawton et al., 1997; Paulsen & Marshak, 1999; Kwon & Mitra, 2006).

In southern Wyoming, the E- to NE-trending Cheyenne belt (Figure 5b) represents the transverse, crustal suture/ transpressional shear zone between the Wyoming and Yavapai-Mazatzal Provinces, across which the Precambrian geology, metamorphism, and metallogenesis abruptly change between adjacent blocks (Karlstrom & Houston, 1984). During the Laramide orogeny, this weak crustal zone was reactivated as a left-lateral transpressional structure and subsequently as a right-lateral transtensional zone during the Tertiary extension (Bader, 2008). Just to the south, east-west trending Precambrian basement fault zones, genetically linked to the Cheyenne belt (Sims et al., 2001; Whitmeyer & Karlstrom, 2007), have been identified to influence Laramide uplift resulting in accumulation of oil and gas resources (Bader, 2009).

A peculiar feature in the Laramide belt of Colorado is a 500-km long, 25-50 km-wide, linear zone of numerous magmatic intrusions, known as the Colorado Mineral Belt (Figure 5a). This zone marks an abrupt along-strike change in: 1) the structural trend of the Laramide uplifts (north-trending in the southern region versus NW-trending northwards); 2) the thickness of the Late Cretaceous and Paleogene sedimentary sequences; and 3) the composition of the Laramide plutons (Chapin, 2012). This economically valuable belt has been interpreted to have formed over an extensional boundary between two adjacent segments of the underlying Farallon flat slab (Chapin, 2012), which serves to demonstrate the effects of the down-going plate features and processes on the upper plate structures. In the Laramide belt, preexisting basement structures and lateral heterogeneities are suggested to have a control on the stress-field and thereby led to the development of structures with an orientation different from the regional trend (Weil et al., 2016).

3.2 Proposed Factors Controlling Lateral Heterogeneities

Along the Cordillera (primarily along the Sevier FTB), pre-existing basement features/structures likely had the greatest impact on creating lateral heterogeneity, which has generally manifested as curvatures along the orogenic front and as variation in the geometry of cross structures. Prior to the Cordilleran orogeny, transverse crustal boundaries served as loci for activation of transverse zones during various rift-related extensional features (McMechan, 2012). Subsequently, these transverse zones were reactivated as cross structures during the orogenic contraction/extension and served to partition or distort deformation of the evolving crustal wedge. Both transverse structures and irregular basement topography further added lateral heterogeneity through their profound control on the lateral continuity of facies and thickness of the pre and syn orogenic sedimentary succession. Abrupt lateral changes in the facies and thickness of the sedimentary column caused the deforming wedge to partition into segments, which are separated by cross structures of various geometries and genesis. As Paulsen and Marshak (1999) explained, a thicker sedimentary column corresponds to a farther propagation of the deforming wedge (salients) than a thinner column (recesses). Further the geometry of the basement irregularities and transverse structures dictate the geometry of cross structures. A near vertical transverse structure is more likely to evolve into a tear fault, whereas, an inclined structure evolves into a lateral ramp (Paulsen & Marshak, 1999). And finally, features of the subducting plate (Farallon) may have influenced the development of cross structures and lateral heterogeneity along the range (e.g.; Chapin, 2012).

4 Alps

4.1 Tectonic Setting and Lateral Heterogeneities

The European Alps formed during a Late Cretaceous collision between the European and African plates following the closure of the Alpine Tethys, which consisted of the northern Valais ocean and the southern Piemont-Liguria ocean, separated by continental crust in the middle known as the Brianconnais (Tricart, 1984; Stampfli & Borel, 2002; Schmid et al., 2004; Handy et al., 2010). The European side of the collision was the lower plate with the Apulian/Adriatic blocks of the African plate forming the upper plate (Dewey et al., 1998; Handy et al., 2010). The Apulian plate refers to all continental domains located south of the Alpine Tethys. The Adriatic micro-plate or Adriatic indenter, a part of the Apulian plate, is situated south of the modern-day Periadriatic Fault System (PA; Stampfli & Borel, 2002; Schmid et al., 2004; Handy et al., 2010). Today this orogen trends approximately east-west in the Eastern and Central Alps and has a NE-SW orientation in the Western Alps (Figure 6). Major range-parallel lithotectonic units of the Alpine orogen are from north to south, the European foreland, Jura Mountains, Molasse Basin (Oligocene-Miocene Foreland), Helvetics, Penninic zone, Austroalpine, Southern Alpine, and Po Basin (retro-arc basin for the Alps and foreland basin for the Apennines) (Pfiffner, 2014). The Helvetic domains are derived from the Mesozoic and Cenozoic shelf and upper slope deposits along the southern European plate margin, and in places include the Pre-Triassic basement (Zerlauth et al., 2014). For simplicity, the classic term 'Penninic nappes/zone' has been largely used in the literature to denote the tectonic units derived from the subducted European margin, the Valais ocean, the Piemont-Liguria ocean, and the Brianconnais continental crust (e.g.; Schmid et al., 2004; Pfiffner, 2014). Parts of the Apulian plate to the north and south of the PA are represented by the Austoalpine and Southern Alpine units respectively (Polinski & Eisbacher, 1992; Schmid et al., 2004). While the Austroalpine unit dominates the eastern Alps, this unit has fully eroded away in the Western Alps, exposing up to blueschist to eclogite facies rocks of the Penninic unit and sub-greenschist facies rocks of the Helvetics (Pfiffner, 2014). The rheologically strong Dolomites, the Adriatic indenter, lie within the Southern Alpine unit.

Following the initial collision, the eastern Alps underwent E-W directed orogen parallel extension in the Miocene (Oligocene; Ring, 1994; Steck, 2008). This extension has been referred to as "lateral extrusion" (Ratschbacher et al., 1991). The lateral extrusion has been interpreted as; i) coupling of compression and gravitational collapse (Ratschbacher et al., 1991), ii) upper plate extension due to roll back of a subduction zone beneath the Carpathian orogen (Royden et al., 1983; Royden, 1993; Horváth & Cloetingh, 1996; Sperner et al., 2002), iii) northward indentation of the southern Dolomites (Rosenberg et al., 2004; Rosenberg & Garcia, 2011; Reiter et al., 2018), and iv) an extension related to the roll back of the Mediterranean plate in the west (Ring & Gerdes, 2016). This Miocene extension has been accommodated along a series of orogen-parallel, strike-slip faults and orogen-perpendicular, normal faults. Modern-day topographic evolution was largely controlled by these faults as much of the lateral heterogeneity we see along the range (Bartosch et al., 2017). Major orogen-parallel strike-slip faults in the Eastern and Central Alps are the Periadriatic Fault System (PA), the Defreggen-Antholz-Vals Fult (DAV), Salzach-Ennstal-Mariazell-Puchberg Fault (SEPM), Inntal Fault (IN), and Mur-Murz Valley Fault (MM) (Figure 6). The PA trends E-W for ~700 km and forms a rheological boundary between the weaker Eastern

Alps and the relatively stronger Southern Alps (Robl & Stüwe, 2005). From the west to east, the PA consists of the Tonale (or Insubric) Line, Giudicarie Fault System, Mauls, Puresetral, and Gailtal segments (Figure 6). Different segments of the PA were active during 32-29 Ma and 22-16 Ma (Müller et al., 2001). North of the PA (Pustertal and Mauls segments), the sinistral, normal DAV runs E-W, for ~80 km and forms the southern boundary of alpine metamorphism (Müller et al., 2000; Bartosch et al., 2017). Within the Austroalpine domain, the EW-striking, ~400 km long, SEPM has a cumulative left-lateral slip of about 60 km (Urbanek et al., 2002) and separates the Mesozoic Northern Calcareous Alps (NCA) from the Middle Austroalpine basement rocks (Bartosch et al., 2017). The NE-striking, sinistral MM was active during 17-13 Ma (Dunkl et al., 2005) as a conjugate of the NNW-striking, dextral Pols-Lavanttal fault system (PL) (Bartosch et al., 2017).

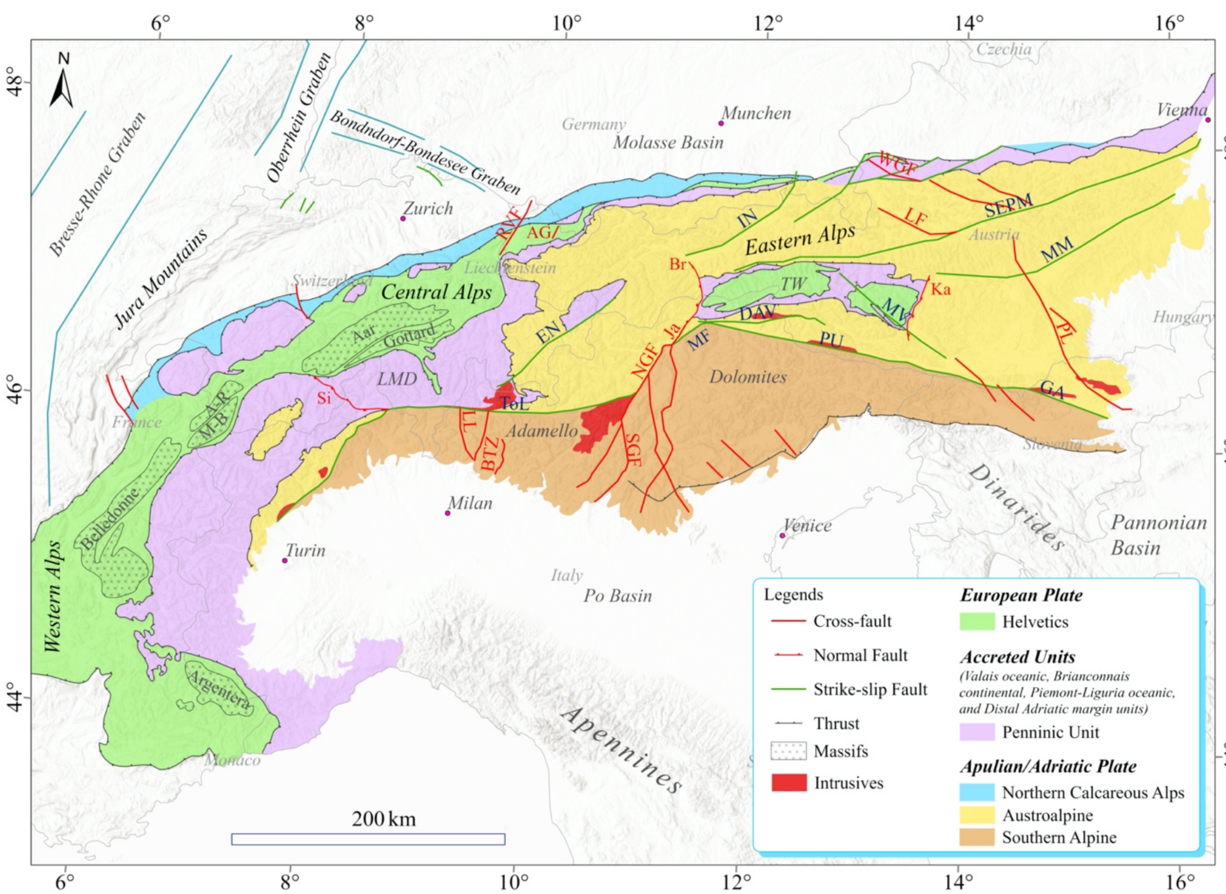

Figure 6: Geological map of the Alps showing the major lithotectonic units, major strike-slip faults (golden lines), and major cross-faults (red lines) (after Laubscher, 1985; Polinski & Eisbacher,

1992; Schönborn, 1992; Linzer et al., 1995; Castellarin et al., 2006; Pfiffner, 2014; Zerlauth et al., 2014; Ring & Gerdes, 2016; Bartosch et al., 2017).

Acronyms: AG: Alpenrhein Graben; A-R: Aiguilles Rouges Massif; Br: Brenner Fault; BTZ: Ballabio-Barzio Transfer Zone; EN: Enagdin Fault; GA: Gailtal Fault; IN: Inntal Fault; Ja: Jaufen Fault; Ka: Katschberg Faults; LF: Lammertal Fault; LL: Lecco Line; LMD: Lepontine Metamorphic Dome; M-B: Mont-Blanc Massif; MF: Meran-Mauls Fault; MM: Murz-Valley Fault; MV: Moll Valley Fault; NGF: North Giudicarie Fault System; PL: Pols-Lavanttal fault system; PU: Purestal Fault; RVF: Rhine Valley Fault; SEPM: Salzach-Ennstal-Mariazell-Puchberg Fault; SGF: South Giudicarie Fault System; Si: Simplon Fault Zone; ToL: Tonale (Insurbic) Line; TW: Tauern Window; WGF: Wolfgangsee Fault.

4.1.1 Cross Structures in the Eastern and Central Alps

Major transverse, normal faults in the eastern and central Alps are the Simplon Fault Zone (Si), the Pols-Lavanttal fault system (PL), the Brenner Fault (Br), the Moll Valley Fault (MV), and the Katschberg Faults (Ka) (Figure 6). The Brenner and Katschberg Faults mark the western and eastern edges of the Tauern Window respectively (Behrmann, 1988; Selverstone, 1988; Fiigenschuh et al., 1997), where blue-schist and eclogitic facies rocks of the Penninic zone have been exposed (Pfiffner, 2014). Exhumation of the Tauern Window has been widely linked to Miocene extension. Rosenberg and Garcia (2011) argue that localized intensive folding deformation due to an irregular geometry of the Dolomite indenter coupled with erosion can also exhume deep-seated rocks without a significant crustal extension. The southern edge of the window has been cut by the NW-striking dextral, normal MV fault. Gently west-dipping mylonitic fabric, with top-to-the-west shear along the Brenner Fault (Behrmann, 1988; Selverstone, 1988), has been overprinted by steeply west-dipping cataclastic zones (Prey, 1989). Much like the Brenner Fault, the Simplon Fault zone is a low-angle, SW-dipping, extensional fault that exhumes the Lepontine Metamorphic Dome in its footwall. Mylonitic shearing that formed the Simplon Fault zone initiated at 30 Ma (Campani et al., 2010) and was overprinted by brittle detachment faults of the Simplon Line from 14.5-10 Ma until 3-5 Ma (Hubbard & Mancktelow, 1992; Mancktelow, 1992; Grosjean et al., 2004; Campani et al., 2010). Both the Brenner and Simplon Faults expose a telescoped crustal section via an initial ductile shearing and subsequent brittle faulting (Grosjean et al., 2004; Campani et al., 2010; Mancktelow et al., 2015). It has been proposed that Brenner and Simplon extension may be related to step-overs in the range-parallel dextral strike-slip deformation in the Neogene (Schmid et al., 1989; Hubbard & Mancktelow, 1992).

Around the eastern Alpine periphery, the NW-trending, NE-verging, Paleogene folds in Permo-Mesozoic cover of the Austroalpine are cut by sets of NE-striking, high-angle, cross-faults (Figure 6) (Polinski & Eisbacher, 1992). In the Miocene, both the Austroalpine and Southern Alpine units were folded into NE-trending folds, which are dissected by a system of NW-trending high angle, dextral cross-faults, which includes the 150 km long PL. These Miocene cross faults either offset or merge into the PA and partition deformation between the contractional Alpine domain and the extensional Pannonian basin (Polinski & Eisbacher, 1992). In the western Alps, NE to E-striking transverse faults are represented by the Neogene normal faults with a few faults showing minor left-lateral reactivation (Sue & Tricart, 2003) and the late to post Oligocene right-lateral, strike-slip faults (Malusà, 2004; Perello et al., 2004; Perrone et al., 2011). Anti-clockwise rotation of the Apulian plate has been attributed as the cause of the orogen-parallel dextral displacement and the NE to SW extension (Hubbard & Mancktelow, 1992; Calais et al., 2002).

The Giudicarie Fault System (GFS) is a system of WNW-NW dipping fault-links between the Tonale and Pustertal line of the PA (Castellarin & Cantelli, 2000; Mancktelow et al., 2001; Müller et al., 2001; Viola et al., 2001). The Meran-Mauls (MF) and the Northern Giudicarie Fault (NGF) within the GFS form the trace of the PA, while the Southern Giudicarie Fault (SGS) is entirely located in the Southern Alps. The Jaufen Fault, as a secondary fault within the GFS, has been considered as a potential continuity of the Brenner Fault (Rosenberg & Garcia, 2011). An explanation of the genesis of the GFS assumes evolution of an initially curved section of the PA, which underwent Neogene sinistral transpression over an inherited NE-trending horst and graben structure (Castellarin & Cantelli, 2000; Müller et al., 2001; Viola et al., 2001). A second explanation, however, considers an initially straight PA bent by the Late Oligocene to Early Miocene, northward movement of the Dolomite indenter. The MF represents a section of the bend, which was subsequently cut by the NGS (Laubscher, 1971; Schmid et al., 1996; Frisch et al., 1998; Pomella et al., 2012). The SGF has been interpreted as a Serravallian-Tortonian transfer fault between the Guidicarie belt and the pre-Adamello belt of the Southern Alps (Castellarin et al., 2006). The Adamello batholith is an Upper Eocene and Lower Oligocene batholith, the northern rim of which has been sheared by the Tonale line (Brack, 1981; Laubscher, 1985; Zanchetta et al., 2011). Just to the east of the NGS, a shallow, NNE-striking, sinistral transpressive fault has been mapped in the Southern Alps (Fondriest et al., 2015).

4.1.2 Cross Structures in the Southern Alps

Basement rocks in the Southern Alps have undergone extension in the Permian, resulting in ENE-trending faults in Triassic, NS-trending faults, and in Jurassic and Cretaceous EW-trending faults (Gaetani et al., 1986; Schumacher, 1990; Bertotti et al., 1993; Picotti et al., 1997; Festa et al., 2020). During the south-verging tectonic transport, these basement structures acted as lateral/oblique ramps, initiated more lateral ramps, and produced complex geometries like en-echelon ramp folds and back thrusting (Schönborn, 1992). In the central Southern Alps, inherited transfer zones cut through decoupling surfaces, partition thrust sheets into discrete blocks, and thereby serve to generate a laterally heterogenous deformation style (Laubscher, 1985; Schönborn, 1992). Several transverse structures that have been identified in this area include the Lecco Line and Ballabio-Barzio TZ (Figure 6) (Laubscher, 1985; Schönborn, 1992; Zanchi et al., 2012). The central Southern Alpine thrust belt has been compartmentalized by these transverse zones (Schönborn, 1992). Moreover, many upper Triassic to Jurassic (Bernoulli, 2007), orogen-perpendicular (e.g.; Berra & Carminati, 2010), normal faults were reactivated during the Neo-Alpine deformation (Oligocene-Miocene; Castellarin & Cantelli, 2000) along with nucleation of several transverse, sinistral, strike-slip faults (Prosser, 1998; Zanchi et al., 2012).

4.1.3 Cross Structures in the Northern Calcareous Alps (a subdivision of the Austroalpine unit)

The Permo-Mesozoic sedimentary succession (3-5 km thick) of the Northern Calcareous Alps were deformed into roughly NE-trending contractional structures during Early Cretaceous to Late Eocence (Gaupp & Batten, 1983; Kralik et al., 1987). The synchronous, en echelon, WNW-striking dextral, tear/transfer faults cut these contractional structures at all scales (Linzer et al., 1995). Some examples include the Lammertal and Wolfgangsee fault systems in the northeastern Alps (Figure 6). Linzer et al. (1995) observed a peculiar deformation decoupling between the sedimentary cover and basement during the ongoing oblique convergence. The sedimentary cover accommodated the convergence obliquity via deformation partitioning between the contractional and strike-slip structures, while the basement deformed through ‘crystal-plastic flow’ (Linzer et al., 1995). Together these displacement transfer faults and deformation decoupling caused a 30˚ clockwise rotation of the entire NCA belt about a vertical axis (Linzer et al., 1995). Moreover, tear faults related to the synorogenic inversion of a Jurassic rift-related graben shoulder have also been identified in the NCA (Oswald et al., 2019). The Miocene, extensional, NE-trending, sinistral, strike-slip faults, such as the SEPM and IN, cut all the earlier structures and divide the NCA into a number of rhombohedral crustal blocks (Linzer et al., 1995).

4.1.4 Cross Structures in the Helvetics

In the Helvetic units, NW to N-striking tear faults were formed due to lateral variation in shortening of the nappe stack, primarily during 35-30 Ma (Hunziker et al., 1986). Nappe imbrication in the Helvetics changes laterally due to the absence or presence of a decoupling layer (Zerlauth et al., 2014). The Permo-Carboniferous and Jurassic extensional structures along the European margin resulted in these along-strike facies changes, which subsequently controlled the deformation style. During the nappe formation, syn-sedimentary normal faults were also reactivated as lateral ramps (e.g.; the Rhine Valley Fault) and tear faults (Zerlauth et al., 2014). In the Oligocene, the NW-trending, sinistral, transtensional Alpenrhein graben was formed in the Helvetics (Ring & Gerdes, 2016). It's conjugate, the Bonndorf-Bodensee in the Jura Mountains is a NE-trending graben formed due to dextral transtension prior to 18 Ma (Hofmann et al., 2000; Ring & Gerdes, 2016). The Bresse-Rhone and Oberrhien grabens in the Alpine foreland are related to the European Cenozoic Rift System (Ring & Gerdes, 2016).

4.2 Proposed Factors Controlling Lateral Heterogeneities

The Eastern and Western Alps show remarkable heterogeneity in timing of major orogenic events (Late Cretaceous in the Eastern Alps Cenozoic in the Western Alps), timing of the regional metamorphism (older in the Eastern Alps and younger in the Western Alps), and overall direction of tectonic transport (NW- to W-directed in the Eastern Alps vs N to NW directed transport in the Western Alps) (Handy et al., 2010, page 123 and references therein). Within each lateral sub-division of the Alps, several factors have contributed to the development of cross structures and lateral heterogeneity. Extension of the upper plate has given rise to several cross faults such as the Simplon and Brenner Faults, which serve as a large-scale, displacement transfer faults between major, range-parallel, strike-slip faults (e.g.; Selverstone, 1988; Hubbard & Mancktelow, 1992).

Lateral connecters (tear faults, lateral ramps, and displacement transfer faults) are common in most of the tectonic units of the Alps. As observed in the North American Cordillera and the Appalachians, extension related basement structures, in both the European and Apulian plates, had controls on the lateral continuity of facies and thicknesses of sedimentary rocks and thereby on subsequent deformation, primarily in the Southern Alps and Helvetics (e.g.; Laubscher, 1985; Schönborn, 1992; Zanchi et al., 2012; Zerlauth et al., 2014). Lateral variability in the composition of the sedimentary section has influenced deformation style based on the presence or absence of decollement horizons. Oblique convergence coupled with partitioning of deformation between the basement and sedimentary cover has resulted in multiple cross faults in the Northern Calcareous Alps (Linzer et al., 1995).

5 Himalayas

5.1 Tectonic Setting and Lateral Heterogeneities

The Himalaya is our planet's most prominent example of a collisional mountain belt. While the geology of this orogen is generally presented in the context of range-parallel thrust faults that separate lithotectonic units of differing metamorphic grade (Figure 7; Hodges, 2000), there is also lateral heterogeneity in various aspects of the geology along the range and in some areas cross structures have played a role in the segmentation of the Himalaya (Mukul, 2010; Godin & Harris, 2014). Early geological and geophysical studies on the Indo-Gangetic plain presented evidence for lateral variations in sediment thickness and geophysical properties along the Himalayan foreland (Burrard, 1915; Oldham, 1917). In the 1970's new data from the oil and gas industry connected these lateral variations to a series of NE-trending basement ridges (Sastri, 1971; Rao, 1973; Raiverman, 1983). It was also proposed that these transverse basement structures may influence Himalayan deformation (Valdiya, 1976). In recent years, evidence from field surveys and laboratory analyses confirms the lateral heterogeneity and locates structural transfer zones and cross structures (Mugnier et al., 1999; Mukul, 2010; Godin & Harris, 2014; Hubbard et al., 2018; DeCelles et al., 2020; Duvall, 2020).

The Himalaya is characterized by range-parallel zones of differing geologic characteristics separated by north-dipping thrust faults (Figure 7). From south to north, the Indo-Gangetic plain is separated from the Sub-Himalaya by the active Main Frontal Thrust (MFT). The Sub-Himalaya is bound to the north by the Main Boundary Thrust (MBT). The hanging wall to the MBT is the

Lesser Himalayan zone, which is bound to the north by the Main Central Thrust (MCT). The unit that includes the highest peaks of the range is the Greater Himalayan Sequence (GHS). Geophysical data supports the merging of these thrusts into a master fault known as the Main Himalayan Thrust (Zhao, 1993; Avouac, 2003; Nabelek, 2009). There is an extensional fault system bounding the northern GHS, the South Tibetan Detachment System (STDS; Hodges, 2000). The hanging wall of the STDS is the Tibetan or Tethyan Sedimentary Sequence (Gansser, 1964; DeCelles et al., 2020). Each of these zones from the Sub-Himalaya to the Tibetan Sedimentary Sequence and their bounding structures exhibits some degree of lateral heterogeneity along the length of the range in either the expression of structural style, depositional history, geomorphology, or seismicity.

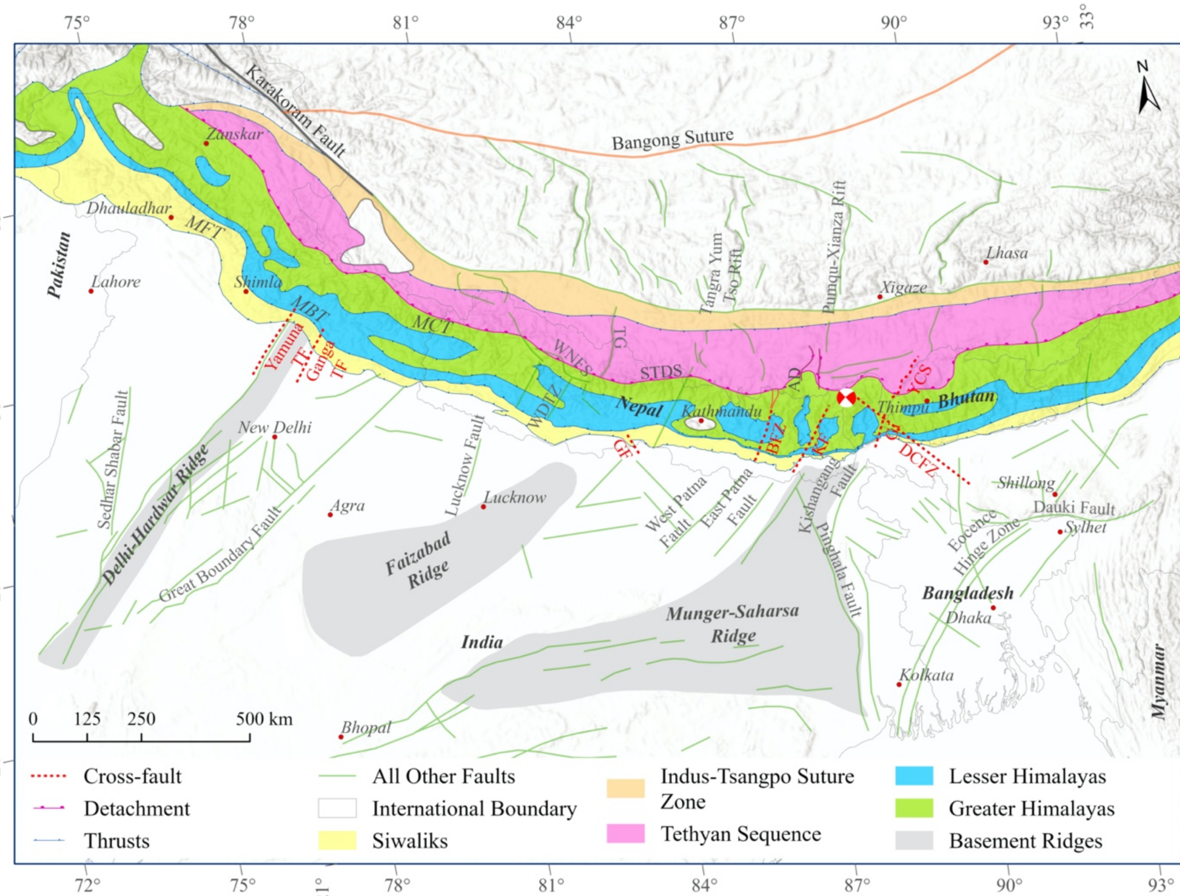

Figure 7: Simplified tectonic map of the Himalaya showing the major cross structures. The beach ball shows focal mechanism of the 2011 M_w 6.9 Sikkim Earthquake (source: the Department of Mines and Geology, Nepal; the Geological Survey of India; Gansser, 1964; Sastri, 1971; Mugnier

et al., 1999; Sahoo, 2000; Searle et al., 2003; Guillot et al., 2008; Jessup et al., 2008; Godin & Harris, 2014; Silver et al., 2015; Diehl, 2017; Mukul, 2018; Divyadarshini, 2019; Seifert, 2019). Acronyms: AD: Ama Drime Detachment; BFZ: Benkar Fault Zone; DCFZ: Dhurbi-Chungthan Fault Zone; GF: Gardi Tear Fault; Gi: Gish Fault; KF: Kosi Fault; MBT: Main Boundary Thrust, MCT: Main Central Thrust, MFT: Main Frontal Thrust, MMT: Main Mantle Thrust, MZB: Main Zanskar Back Thrust, STDS: South Tibetan Detachment System; TF: Tear Fault; WDTZ: Western Dang Transfer Zone; WNFS: Western Nepal Fault System; YCS: Yadong Cross Structure.

In the Sub-Himalayan there is notable variability in the morphology of the range front including the local presence of dun structures and a series of recesses and salients (Yeats, 1991; Mukul, 2010). In some areas the irregular mountain front has been linked to differences in shortening (Dubey, 2001) and in other areas there is a connection with cross faults that mark the transitions from salients to recesses. Examples of cross faults in the Sub-Himalaya include the Yamuna and Ganga Tear Faults in the NW Himalaya and the Kosi and Gish faults in eastern Nepal and Sikkim (India), respectively (Figure 7) (Sahoo, 2000; Srivastava, 2018). In the Lesser Himalaya, lateral heterogeneity is seen in topographic data, seismic data, and cooling history data (Harvey, 2015; van der Beek, 2016; Soucy La Roche, 2019). In western Nepal there is a zone of change in a variety of parameters that has led researchers to conclude that the MHT may have a ramp at that location, possibly coupled with a change in strike (Harvey, 2015; van der Beek, 2016; Soucy La Roche, 2019). This change also aligns with the proposed West Dang Transfer Zone (Figure 7) (Mugnier et al., 1999). Soucy La Roche (2019) propose a connection between this structural change and the NE-striking Lucknow fault in the Indian basement. In several areas, along-strike changes in the geometry of duplex structures in the Lesser Himalaya have also been noted (Hauck, 1998; DeCelles et al., 2001; Grujic, 2002; Long, 2011).

In the Greater Himalaya there is lateral heterogeneity in exhumation rates (Eugster, 2018), topographic profiles (Duncan et al., 2003), the presence of leucogranitic intrusions (Weinberg, 2016), and the presence or absence of discontinuities (Carosi, 2010; Larson, 2014). Hubbard et al. (2018) recently recognized a cross structure, the Benkar Fault zone, in the Greater Himalaya of eastern Nepal (Figure 7). This structure has dextral normal displacement and was active in the past ~12 Ma. The Benkar Fault zone may continue into the Lesser Himalaya, however it is yet to be mapped to the south.

Microseismicity in the Himalaya has occurred along cross-strike trends, has terminated at cross-strike zones, and has made other changes along cross-strike boundaries (Rajaure, 2013; Mugnier et al., 2017; Bilham, 2019; Mendoza, 2019). The aftershock seismicity from the 2015 Gorkha earthquake terminates along a sharp NE-striking boundary east of Kathmandu in Nepal (Hubbard et al., 2016; Mendoza, 2019). In the eastern Himalaya, several seismic events have been consistent with strike-slip displacement on transverse structures (Drukpa, 2006; Paul, 2015; Diehl, 2017). While a number of these events are minor, there have also been major earthquakes such as the 2011 M_w 6.9 Sikkim event that was interpreted to have occurred along a NW-striking (cross-strike) plane with dextral kinematics (Figure 7) (Paul, 2015). The Sikkim event and a number of the smaller events have originated at depths of ~50-60 km suggesting that rupture initiation was below the MHT (Drukpa, 2006; Paul, 2015) but possibly penetrating the hanging wall. Major historic thrust-related earthquake events are known to have had a finite rupture area (Bilham et al., 2001) and tracking of these areas has facilitated the recognition of seismic gaps that may represent regions with higher risk (Bilham, 2019). It has been proposed that cross structures may limit the seismic rupture area, thus contributing to the lateral heterogeneity in seismic activity (Mugnier et al., 2017; Hubbard et al., 2021).

5.2 Proposed Factors Controlling Lateral Heterogeneities

The combination of pre-existing transverse basement structures and the lateral variation in sedimentary history may both contribute to the lateral heterogeneity and cross structures that we see today in the Himalaya. Godin and Harris (2014) proposed a connection between variations in gravity data across a broad region in the Himalaya and Tibet to ridges in the Indian basement detected in the foreland. Soucy La Roche (2019) demonstrated lateral variations in cooling histories in western Nepal and proposed a cross structure in the form of a lateral ramp or tear fault separating the regions with differing cooling histories. This cross structure aligns with the Lucknow basement fault in the Indian foreland and these authors propose that the Himalayan cross structure is linked to the basement fault. Boundaries between salient and recesses in range front geomorphology has also been linked to basement faults (Hubbard et al., 2021). These basement faults in the foreland may also separate down-dropped blocks with thicker sedimentary sequences from higher blocks and these differences in sedimentary thickness may continue into the Himalaya, thus influencing lateral variations in structural style (DeCelles et al., 2020). Understanding the nature of segmentation and segment boundaries may shed light on details of the mountain building process in collisional orogens but may also help us understand how convergence is accommodated and factors that control seismic energy propagation.

6 Zagros

6.1 Tectonic Setting and Lateral Heterogeneities

The Zagros orogenic belt runs for about 2000 km along the northeastern margin of the Arabian plate (Figure 8) and is the product of Miocene collision between the Arabian and Iranian continental plates and subsequent convergence (Allen & Armstrong, 2008; McQuarrie & van Hinsbergen, 2013). The eastern boundary is the dextral Zagros-Makran Transfer Zone (Regard et al., 2005), and it's western boundary is the sinistral East Anatolian Fault (Falcon, 1974; Haynes & McQuillan, 1974). During Permian to lower Cretaceous, the Neo-Tethys Ocean opened between the Arabian and Iranian plates. Closure of this short-lived ocean began in Late Cretaceous along with the tectonic emplacement of ophiolites and trench sediments on the Arabian plate (Haynes & Reynolds, 1980; Berberian & King, 1981). While there are dominant range-parallel thrust faults, there are also several important cross structures and other geologic features that vary along the length of the range.

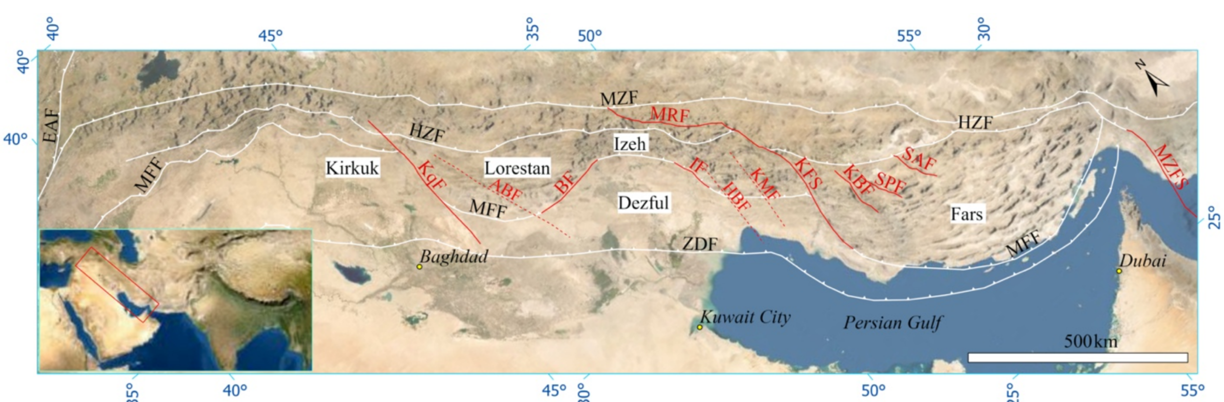

Figure 8: Simplified tectonic map of the Zagros mountains showing the major faults and the recesses and salients. The inset on the bottom-left corner shows the location of the larger map. The Minab-Zendan Fault System (MZFS) is a major fault of the Zagros-Makran Transfer Zone (modified from Authemayou et al., 2006; Farzipour-Saein et al., 2013; Joudaki et al., 2016; Le Garzic et al., 2019). The white lines are range parallel faults and red lines indicate cross structures.

Acronyms: ABF: Anaran Basement Fault; BF: Belarud Fault; EAF: East Anatolian Fault; HBF: Hendijan-Izeh Basement Fault; HZF: High Zagros Fault; IF: Izeh Fault; KBF: Kareh-Bas Fault; KFS: Kazerun Fault System; KMF: Khark-Mish Basement Fault; KqF: Khanequin Fault; MRF: Main Recent Fault; MZFS: Minab-Zendan Fault System; MZF: Main Zagros Reverse Fault; SAF: Saverstan Fault; SPF: Sabz Pushan Fault; ZDF: Zagros Deformation Front (also known as the Zagros Foredeep Fault).

The major, range-parallel tectonic elements in the Zagros have a NW-SE trend (Figure 8) and include the Main Zagros Reverse Fault (the continental suture), the Zagros Imbricate Zone, the High Zagros Faults, the Zagros Folded Belt, Mountain Front Fault, the Zagros Foredeep, and the Zagros Foredeep Fault, from NE to SW (Stocklin, 1968; Berberian, 1995). A peculiar feature in the Zagros is that crustal shortening has been accommodated by deep seated, high-angled reverse faults, which have been interpreted as reactivated, basement normal-faults initiated during the Neotethyan rifting and spreading (Falcon, 1974; Jackson, 1980; Chauvet et al., 2004; Mouthereau et al., 2012). Generally, these faults are segmented with about 85-100 km gaps in between segments and are confined in the Precambrian basement and the lower stratigraphic levels of the overlying sedimentary cover (Berberian, 1995; Bigi et al., 2018). The Cambrian to Pliocene sedimentary cover (~5-13 km thick) overlies the Proterozoic to early Cambrian Hormoz Salt (Stocklin, 1968; Falcon, 1974; Colman-Sadd, 1978). This incompetent salt unit acted as a decollement horizon between the thick-skinned deformation in the basement and the thin-skinned deformation in the sedimentary cover (e.g.; Berberian, 1995; Lacombe et al., 2006). Deformation initiated in the Zagros Imbricate Zone at about 20 Ma or earlier (Fakhari et al., 2008)**,** propagated towards the SW and reached the Zagros Folded belt at 14 Ma (Khadivi et al., 2010).

Since the Neogene, a system of strike-slip faults has also been active in the Zagros (Berberian, 1995). The Main Recent Fault (MRF) was initiated in late Pliocene, which is an active, orogen-parallel, strike-slip fault that follows the trace of the Main Zagros Reverse Fault (Tchalenko & Braud, 1974; Berberian, 1995; Authemayou et al., 2006). Oblique convergence between the Iranian and Arabian plate is partially accommodated along this hinterland fault (Authemayou et al., 2006). At its southeastern end, the MRF diffuses into a series of N to NNW-striking right-lateral, basement-inherited faults, grouped as the Karezun Fault System (Authemayou et al., 2006). Major faults belonging to this zone are the Saverstan, Sabz Pushan, and Kareh-Bas Faults (Figure 8) with an offset of about 100-150 km along each one (Berberian, 1995; Authemayou et al., 2006). During rifting of the Proto-Tethys and initial stages of the Neo-Tethys, these faults were activated as right-lateral transform faults in the basement rocks (Talbot & Alavi, 1996). Neogene reactivation of these transverse faults has produced a dragging effect on the earlier-formed, orogen-parallel folds. These faults serve to transmit and distribute the slip along the MRF towards the southeast into the Zagros Folded Belt and the Foredeep (Berberian, 1995; Authemayou et al., 2006). Salt diapirs have been intruded along these faults at multiple locations, which suggests that these faults cut into the top of the basement (e.g.; Kent, 1979; McQuillan, 1991; Talbot & Alavi, 1996). A certain degree of seismic hazard is associated with these transverse faults (Berberian, 1995; Authemayou et al., 2006). West of the Karezun Fault System, important transverse faults are the Izeh, Balarud Fault (E-W, 130 km left lateral), the Anaran Basement Fault, and the Khanequin Fault (Figure 8) (Hessami et al., 2001; Joudaki et al., 2016; Sadeghi & Yassaghi, 2016). All these transverse faults were active basin-bounding, normal faults, reactivated as strike-slip faults, and had controls on pre and syn-orogenic sedimentation and subsequent deformation in the Zagros (Sepehr & Cosgrove, 2004).

Along its strike, the Zagros curves into a series of salients and recesses bounded by the transverse faults (Figure 8). Generally, these curvatures are grouped into domains and include the Fars salient, the Dezful embayment, the Izeh zone (juxtaposed with the Dezful embayment in the north), the Lorestan salient, and the Kirkuk recess from SE to NW (e.g.; Stocklin, 1968; Falcon, 1974; McQuarrie, 2004). The deformation zone is widest in the Fars salient and narrows towards the NW. The style and distribution of deformation changes notably along these curvatures, mostly driven by the presence or absence of decollement layers (e.g.; Bahroudi & Koyi, 2003).

Lateral heterogeneity in the Zagros is expressed through the transverse faults, the presence of salt diapirs, and the salient and recesses on the mountain front as discussed above. Further heterogeneity is also expressed in changes in fold geometry and the amount of shortening in adjacent regions. The Fars salient is characterized by a very low taper angle, several salt intrusions, and concentric folds with large amplitude (Talbot & Alavi, 1996; Sepehr et al., 2006; Mukherjee et al., 2010). Towards the west, box folds and large concentric folds are dominant in the Izeh zone. In the Dezful embayment, however, exposed folds are concentric folds with small wavelengths and probably overlie large concentric folds beneath a detachment surface (Sepehr et al., 2006). The presence of a paleo-basement high, bounded by the N-striking, Hendijan-Izeh and Khark-

Mish basement faults, has been suggested beneath the Izeh and Dezful domains (Farzipour-Saein et al., 2013 and references therein). In the Lorestan salient, folds are rounded but with much smaller wavelength than in the Fars salient (Sepehr et al., 2006). Several small-scale transverse faults have also been identified around the inflection zone between the Lorestan salient and the Kirkuk recess (e.g.; Sadeghi & Yassaghi, 2016). Amount of shortening in the Fars, Dezful, and Lorestan segments are 67 km, 85 km, and 57 km respectively (McQuarrie, 2004).

Spatial changes in folding style have been linked to the mechanical anisotropy within the deforming sedimentary column and the depth at which those anisotropies occur (e.g.; Sepehr et al., 2006). Mechanical anisotropy (presence of an incompetent layer) has a major control on the shape of the folds (lower anisotropy equals more rounded folds) and the depth of anisotropy controls the fold wavelength (deeper anisotropy equals large folds) (Sepehr et al., 2006). The deep-seated Cambrian Hormuz Salt is about 1-2 km thick beneath the Fars salient, and possibly present in the Izeh zone, but absent beneath the sedimentary columns in the Lorestan and Dezful domains (Bahroudi & Koyi, 2003). In contrast, the Mesozoic Kazhumdi shales or the Dashtak Evaporites in the Lorestan, and the Miocene Gachsaran Formation in the Dezful Embayment act as higher level decoupling surfaces, while the overlying sedimentary cover has a uniform mechanical stratigraphy (Sepehr & Cosgrove, 2004; Sepehr et al., 2006). The presence of decollement horizons at different structural levels in the Dezful and Izeh domains has generated a peculiar ramp and flat geometry in the sedimentary cover in the central Zagros (McQuarrie, 2004; Sepehr et al., 2006).

Stratigraphy and deformation style are also generally different from the recesses to the salients (Sepehr et al., 2006). It has been suggested that the Zagros foreland in the salient deform by both thin-skinned and thick-skinned deformation, whereas, the foreland deformation in recesses is accommodated only by thin-skinned tectonics (Malekzade et al., 2016). Around the central Zagros foreland, both the intensity of deformation and the amount of shortening increases towards the west and the deformation has been partitioned by left-lateral, blind tear faults reactivated on preexisting basement structures (Sarkarinejad et al., 2018; Pash et al., 2020).

6.2 Proposed Factors Controlling Lateral Heterogeneities

Several ideas have been proposed to explain the sinuosity and lateral heterogeneity of deformation style in the Zagros. These ideas include: 1) rotations of crustal blocks (Hessami et al., 2001; Edey et al., 2020); 2) along strike variations in a viscous decollement horizon between the basement and the sedimentary cover (Bahroudi & Koyi, 2003; McQuarrie, 2004); 3) the presence of a lateral buttress that serves to partition deformation (Cotton & Koyi, 2000; Bahroudi & Koyi, 2003); 4) lateral variation in the degree of oblique convergence (McQuarrie et al., 2003; Vernant et al., 2004; Vernant & Chéry, 2006); and 5) heterogenous rigidity along the plate margin (Malekzade et al., 2016). The block rotation model suggests that adjacent crustal blocks rotate with a reverse polarity along a vertical axis, driven by movements along strike-slip faults and the presence of a rigid backstop (e.g.; Hessami et al., 2001; Edey et al., 2020). These strike-slip faults were activated along N-S trending, inherited basement structures attributed to the Pan-African tectonics (Koop & Stoneley, 1982; Husseini, 1988; Hessami et al., 2001) and subsequent Tethyan rifting (Talbot & Alavi, 1996).

Along strike variation in the distribution of the Cambrian salt unit has a major control in the deformation along the Zagros Fold Belt. The presence of this incompetent unit served as a viscous detachment and deformation propagated much further where the salt was present than in the domains where it is missing (Bahroudi & Koyi, 2003; McQuarrie, 2004; Farzipour-Saein et al., 2013). The patchy arrangement of salt deposition has been attributed to the aforementioned, inherited basement structures, which could have also created tear fault-type transverse structures in the hanging walls of the thrust sheets. The style of deformation is also dependent on the nature of detachment horizons. Above a viscous decollement, the overlying units generally deform by ductile thickening, whereas above a frictional detachment, imbrication and folding deformation dominates (Bahroudi & Koyi, 2003). During deformation, the presence of transverse basement faults or a lateral change in facies can act as a lateral buttress and reorient local kinematics (Cotton & Koyi, 2000; Bahroudi & Koyi, 2003). Obliquity of plate convergence and the frictional strength along the MRF (a hinterland fault) together dictate if strike-slip partitioning initiates or not and thereby can have controls on internal deformation in the Zagros belt (McQuarrie et al., 2003; Vernant et al., 2004; Vernant & Chéry, 2006). Moreover, heterogenous plate-margin rigidity along with the presence of embayments and indenters, can cause along-strike changes in deformation patterns by affecting the local obliquity angle and favoring the escape of the upper crustal material into adjacent reentrants (Malekzade et al., 2016).

It is likely that a number of factors have contributed to lateral heterogeneity in the Zagros. The stratigraphy has played a role in terms of thickness changes causing differences in folding patterns and fault geometry. The presence of localized salt layers and salt diapirs further contribute to differences in thrust displacement and local geology. The geometry of the extensional or rifts structures that pre-date collision may have impacted the geometry of post-collisional faults.

7 Andes

7.1 Tectonic Setting and Lateral Heterogeneities

The Andean mountain belt is a classic example of an active subduction margin and likely represents processes that were active in the world's collisional mountain belts prior to collision. Modern-day Andean topography is largely the manifestation of crustal shortening and thickening in response to the ongoing oblique subduction of the Nazca oceanic plate beneath the South American plate (e.g.; Mpodozis & Ramos, 1989). The two plates are currently converging along a N78˚E vector at a rate of about 66 mm/yr (Angermann et al., 1999; Kendrick et al., 2003). The Andean orogen follows a north-south trend along the western margin of the South American plate, except in the Central Andes, where the orogen makes a curve known as the Arica bend (Figure 9a). Along the length of the mountain belt there is lateral heterogeneity exhibited at several scales. Lateral heterogeneity along the range has been observed in terms of the upper and lower plate dynamics, topography and deformation style of the retro-arc belt, nature of the foreland basin, and igneous activity.

At a coarse scale, the Andes exhibit lateral heterogeneity in topographic changes from north to south that has led to the characterization of the Northern Andes, the Central Andes, and the Southern Andes (Figure 9a). The major visible difference is in the width of the mountain belt where the Northern and Southern Andes are narrow while the Central Andes is much wider and includes the Altiplano and Puna plateau regions. These general topographic differences are the product of differences in their terrane accretion history and subduction zone dynamics. Other broad differences seen along the range include the presence or absence of active volcanism which is likely related to changes in the dip angle of the subduction slab (Ramos & Folguera, 2009). During early Paleozoic, the southwestern margin of South America had a history of exotic terrane accretion and subduction (e.g.; Ramos, 1988). The modern-day subduction margin initiated in late Paleozoic (Giambiagi et al., 2012). The Late Permian to Early Jurassic tectonic history was

Figure 9: Tectonic map of the Andes. a) Map of the South American plate and the Nazca oceanic plate (west) showing major transverse features on the subducting Nazca oceanic plate. The red star

shows the epicenter of the M_w 8.4 2001 Peru earthquake. Red shades indicate volcanic zones. Note the large-scale strike slip faults, indicated by yellow lines, in the Northern Andes. The inset shows the extent of the 'Figure 9b' (modified from Gutscher et al., 2000; Robinson et al., 2006; Egbue & Kellogg, 2010; Schepers et al., 2017). b) Geological map of the Central and Southern Andes showing the location of major cross structures. The deformation style and distribution are quite different from the Central to the Southern Andes. (modified from Ghiglione et al., 2009; Stanton-Yonge et al., 2016; Schepers et al., 2017). The red lines in the 'Figure 9b' represent the Andean Transverse Faults.

Acronyms: CCM: Calliqui-Copahue-Mandolegue Transfer Zone; CCVC: Cordon Caulle Volcanic Complex; ChC: Chillan-Cortaderas Lineament; GFZ: Grijalva Fracture Zone; LATF: Lago Argentino Transfer Fault; LOFS: Liquine-Ofqui Fault System; LVTF: Lago Videma Transfer Fault; MFZ: Mendana Fracture Zone; MVFZ: Mocha-Villarrica Fault Zone; NFZ: Nazca Fracture Zone; TPTF: Torres del Paine Transfer Fault; TSPP: Tarta-San Pedro-Pellado Volcanic Alignment.

characterized by crustal extension and the associated volcanism (Llambías et al., 1993). In the Southern Andean front, there was backarc extension from late Triassic to Early Jurassic (Vergani et al., 1995; Giambiagi et al., 2012) while the Central Andes experienced extension from late Jurassic to Early Cretaceous (Galliski & Viramonte, 1988; Salfity & Marquillas, 1994). The resulting orogen-parallel, rift-related, normal faults and/or transfer faults were reactivated as reverse faults with strike-slip components in Cretaceous to Paleogene (Kley et al., 2005; Mescua & Giambiagi, 2012). In the Southern Andes, this Neogene inversion of the extensional basins led to lateral variations in the thickness and facies of the sedimentary sequences, which were subsequently reflected in heterogeneities of the deforming fold and thrust belt (e.g.; Ghiglione et al., 2009; Likerman et al., 2013).

The style of subduction of the down-going Nazca Plate in the Central Andean margin differs greatly from the adjacent margins to the north and south. The Central segment has had a steep subducting plate while to the north and south the oceanic plate has been subducting at a low angle since 12 Ma in the respective trench segments namely, the northern Peruvian and southern Pampean flat slabs (Schepers et al., 2017). The trench retreat has greatly outpaced the slab roll back and this difference has generated the 200-300 km long flat slab segments. The slab is thought to have been anchored at the 660 km discontinuity, which is preventing the roll back (Schepers et al., 2017; Chen et al., 2019). The crustal shortening along the Arica bend in the Central Andes is 420 km, which is much greater than shortening values of 160 km and 150 km in the Southern and Northern respectively (Arriagada et al., 2008; Schepers et al., 2017). There are also differences in the style of deformation related to flat slabs including the inland development of thick-skinned deformation east of the crest of the Andes. Another lateral variation related to the flat slab zones is the absence of active volcanism. Several causes for the flattening of the slab have included subduction of buoyant oceanic crust around the Nazca and Juan Fernandez Ridges or possible changes in the thermal thickness of the overriding lithosphere (Pilger, 1981; Manea et al., 2012; Schepers et al., 2017).

At a finer scale, there are lateral differences in the timing and nature of deformation along the Andes. In the Northern Andes, crustal blocks are "escaping" northeastwards towards the free Caribbean-North Andes boundary and away from the rigid South American Plate along the system of large-scaled, strike-slip faults (Figure 9a) (Audemard et al., 2005; Audemard, 2009; Egbue & Kellogg, 2010; Monod et al., 2010). The trench-parallel component of the oblique plate convergence has been driving this escape for the last 1.8 Ma. This escape was likely triggered by an increase in coupling between the upper and lower plate when the Carnegie Ridge entered the subduction zone (Egbue & Kellogg, 2010).

The Central Andes segment is dominated by anomalous topography of the Altiplano-Puna plateau, which was uplifted due to thermal softening of the lithosphere followed by crustal shortening and accompanying arc magmatism during the orogeny (Allmendinger et al., 1997; Coutand et al., 2001). Within the plateau region researchers have documented significant differences in the timing of the surface uplift and crustal structure of adjacent blocks (e.g.; Bianchi et al., 2013; Leier et al., 2013; Canavan et al., 2014). The fault boundaries of these blocks are thought to be related to inherited basement faults, possibly bounding older accreted terranes (Jordan et al., 1983) Towards the foreland, spatial location of the Cenozoic fault systems were likely controlled by the presence of Paleozoic and Mesozoic basement structures (Coutand et al., 2001; Gillis et al., 2006). The plateaus and the adjacent ranges in the Central Andes, define the highest topographic features in the world within a non-collisional setting (Isacks, 1988).

Another feature unique to the southern Central Andes is the fragmentation and exhumation of the retro-arc foreland basin (Sierras-Pampeanas) along inherited structures (Figure 9b) (Jordan et al., 1983; Japas et al., 2016). Stronger coupling between the upper and lower plates in the flat-slab segments has been described as the driver behind inland propagation of deformation and subsequently the basin uplift (Ramos et al., 2002; Oriolo et al., 2014). Just to the south of the flat-slab segment, Giambiagi et al. (2012) documented an abrupt southward decrease in topographic uplift and crustal shortening. This sharp, lateral variation is driven by a strong lateral change in the upper plate rheology, which controls the degree of coupling between the upper and lower crust, such that a thick and more felsic crust has strongly coupled upper and lower slabs (Giambiagi et al., 2012).

Thick-skin deformation in the Central Andes terminates towards the south at the northern segment of the Southern Andes, which is characterized by trench-parallel, thin-skinned, fold and thrust deformation. The southern segment, however, is dominated by a system of strike-slip fault systems. The ENE-striking, transverse, dextral Callaqui-Copahue Mandolegue (CCM) fault is thought to decouple the contrasting deformation styles between the retro-arc thrust belt (north) and the Liquine-Ofqui Fault System (LOFS) in south (Folguera et al., 2004).The LOFS (Figure 9b) is a ~1200 km long, NNE-striking, right-lateral, reverse, intra-arc fault in the southern segment (Cembrano et al., 1996; Thomson, 2002; Vargas et al., 2013). Here, a series of NE-striking, en echelon, dextral, normal faults splay off from the LOFS and form a strike slip duplex between two sub-parallel fault branches of the LOFS (Cembrano et al., 1996). Another set of strike-slip faults cut the LOFS and associated faults and are called the Andean Transverse Faults (ATF). The ATF (Figure 9b) are NW-striking, reverse, sinistral faults, primarily formed on inherited basement structures (e.g.; Roquer et al., 2017; Sielfeld et al., 2019). Deformation partitioning occurs within these faults, whereby the trench-parallel component of the oblique convergence is accommodated by the LOFS and its splay faults and the ATFs accommodate the trench-perpendicular component (e.g.; Stanton-Yonge et al., 2016). In the northern termination of the LOFS, the strike-slip dominated domain sharply transitions into margin-parallel fold and thrust deformation (Figure 9b) (Stanton-Yonge et al., 2016). These strike-slip fault systems also form excellent conduits as well as reservoirs for magma and hydrothermal fluids and thereby control the locus of volcanic complexes (e.g.; Petrinovic et al., 2006; Pérez-Flores et al., 2016; Roquer et al., 2017; Sielfeld et al., 2019; Lupi et al., 2020; Piquer et al., 2020).

Effects of the subduction of oceanic fracture zones on seismic activity has also been observed in the central Andes. In the subducted Peruvian slab, generally deeper earthquakes are generated around a subducted segment boundary, the Mendana Fracture Zone (Figure 9a) (Gutscher et al., 2000). Towards the south, a fracture zone in the down-going Nazca plate (most likely the Nazca Fracture zone) is thought to have induced a fracture in the over-riding plate that acted as a temporary lateral barrier during the initial seismic rupture propagation but subsequently allowed energy to pass through this vertical plane releasing the energy of the M_w 8.4 2001 Peru earthquake (Robinson et al., 2006). These examples indicate that transverse discontinuities in the lower plate (or subducting plate) can influence structures in the upper plate and can have great seismic implications.

7.2 Proposed Factors Controlling Lateral Heterogeneities

Along the Andean belt, lateral heterogeneities have been primarily expressed as variation in crustal shortening and topographic uplift (e.g.; Allmendinger et al., 1997; Arriagada et al., 2008; Giambiagi et al., 2012; Schepers et al., 2017), along-strike changes in the styles and timing of deformation (e.g.; Ramos et al., 2002; Bianchi et al., 2013; Leier et al., 2013; Canavan et al., 2014; Oriolo et al., 2014; Stanton-Yonge et al., 2016), laterally contrasting seismicity (Gutscher et al., 2000; Robinson et al., 2006), and as changes in volcanic activity (e.g.; Pérez-Flores et al., 2016; Roquer et al., 2017; Sielfeld et al., 2019; Lupi et al., 2020; Piquer et al., 2020). Researchers interpret that these heterogeneities are governed by the angle of lower plate subduction (flat versus steep slab), physiography of the lower plate, obliquity of the subduction vector, preexisting basement structures in the upper plate, and the rheology of the upper plate. The angle of the lower plate subduction controls the degree of coupling between the lower and upper plates (lower angle equals stronger coupling), which in turn impacts the nature and distribution of the upper plate deformation (e.g.; Ramos et al., 2002; Oriolo et al., 2014). Subduction of physiographical features in the lower plate also can change the amount of plate coupling and generate respective topographic and seismic signatures in the upper plate or it can influence the development of cross structures (Gutscher et al., 2000; Robinson et al., 2006). Oblique plate convergence in the Southern Andes has been accommodated by a system of strike-slip and transverse faults (e.g.; Stanton-Yonge et al., 2016). In the Northern Andes, however, the dynamic plate setting has favored crustal escape (e.g.; Egbue & Kellogg, 2010). Rheology of the upper plate dictates the amount of plate coupling and therefore, has controls on the rate and amount of crustal shortening and the overall topographic uplift (Giambiagi et al., 2012).

8 Other Orogens

Lateral heterogeneities have been documented in several other mountain belts around the world (Figure 1). While there are groups of along-strike variations common across multiple mountain belts, some of the heterogeneities are the result of the broad tectonic setting and the nature of the sedimentary cover, and therefore can be unique to only a few mountain belts. Comparable to the Appalachians, , the Mesozoic rift structures in the External Hellenides thrust belt in Greece were reactivated as transverse zones (e.g.; the Corinth Gulf, Ierapetra, and Omalos transverse zones), which partitioned the belt into a series of salients, recesses, and linear segments (Skourlis & Doutsos, 2003; Kokkalas & Doutsos, 2004; Chatzaras et al., 2013). These inherited structures also disrupted the lateral continuity of the foreland sedimentary sequences and served as crustal-scale lateral/oblique ramps during thrust propagation and thereby had a significant control on the nature and deformation style of the evolving taper wedge (Robertson et al., 1991; Doutsos et al., 2006; Chatzaras et al., 2013). Similarly, in the Urals, Precambrian aulacogens (failed rift arms) served as transverse basement structures, above which tear-faults and lateral ramps were developed during fold-thrust propagation (Rodgers, 1990; Brown et al., 1997; Perez-Estaun et al., 1997). Cross faults are common in the Apennines and are generally called anti-Apennine Faults (e.g.; Coltorti et al., 1996; Sorgi et al., 1998; Butler et al., 2006; Elter et al., 2011). The NNW-striking Apennine belt is underlain by the NNE-trending crustal tectonic lineaments (Valnerina Line, Ancona-Anzio Line, Ortona-Rocca Monfina Line). These large-scale lineaments act as structural barriers (zones of abrupt lateral changes in tectonic and structural style, wedge stratigraphy, and topography) during tectonic transport and laterally limit the propagation of seismic fault rupture (Pizzi & Galadini, 2009; Satolli et al., 2014). There are also smaller-scaled transverse basement structures, that may either serve as seismic segment boundaries/seismic loci or get reactivated as transfer zones (Valensise & Pantosti, 2001; Pizzi & Galadini, 2009). Furthermore, transfer zones in the Apennines are also known to form suitable loci for magma emplacement (Dini et al., 2008). On a much larger scale, the differential retreat along the adjacent segments of the Adriatic plate during the last 5 Ma has been manifested as a lithospheric tear/transfer zone across the Appennines (Scrocca, 2006), which may share a similar genesis to the Colorado Mineral Belt in the Cordilleran belt.

Cross structures have also been interpreted to have great economic and seismic impacts besides their influence in tectonic, stratigraphic, and structural evolution (e.g.; Mahoney et al., 2017). In the Papua New Guinea fold-thrust belt, the Jurassic, extensional, transverse, crustal structures evolved as zones of economically significant copper-gold mineralization during their Late Miocene-Pliocene inversion (Davies, 1991; Corbett, 1994; Hill et al., 2002). Like in the Apennines and the Himalayas, active cross faults in the Taiwan mountains are known to serve as earthquake nucleation sites and as rupture segment boundaries (Deffontaines et al., 1997; Lacombe et al., 2001; Ching et al., 2011). These cross faults have genetically been linked to changing deformation style (thick-skinned vs thin-skinned) and to a lateral change in thickness of the deforming sedimentary sequences (Lacombe et al., 2001; Mouthereau et al., 2002; Mouthereau & Lacombe, 2006; Ching et al., 2007). Much similar to the Zagros belt, controls of the presence or absence of a decoupling layer on the style of the evolving taper wedge (narrow wedge with a high-taper angle, when decoupling layer is absent and vice-versa) have been noted in the Caucasus belt (Upper Jurassic salt and Middle-Lower Jurassic shales) in Russia (Sobornov, 1996), the Sulaiman belt (Paleozoic, Lower Cretaceous, and Eocene strata) in Pakistan (Jadoon et al., 1994), and the Parry Island belts (Ordovician salt) in the Canadian Arctic (Harrison & Bally, 1988).

9 Discussion

Despite the general temporal and spatial continuity of crustal deformation along convergent mountain belts, significant lateral heterogeneities have been observed in several orogenic belts from around the world. There are certain lateral variations, which are unique to only a few orogens such as the lithospheric tear in the Apennines and the asynchronous orogenic events in the Alps. More frequently, however, similar groups of lateral variations have been observed across multiple mountain belts. Generally, lateral heterogeneity in convergent mountain belt settings have been expressed as: 1) an along-strike change in deformation style (thick-skinned vs thin-skinned, imbrication vs duplexing/changes in ramp geometries); 2) variation in igneous activity and metamorphic grade; 3) variation in seismic activity; 4) differential topographic uplift/features along the strike; and/or 5) abrupt changes in thickness and facies of sedimentary sequences in both foreland basins and within the fold-thrust belt. Oftentimes such lateral variations are abrupt rather than gradual and are marked by geological structures, mostly faults, that are nearly orthogonal to the strike of the orogen, generically referred to as cross structures in this review. The causal factors/mechanisms behind the observed lateral heterogeneities are discussed in this section.

9.1 Irregular Continental Margins

The geometry of the continental margin(s), prior to the collision/convergence, has a profound effect on the geometry of the evolving orogenic belt. The orogenic front may mimic the continental margin geometry (Figure 4) such that any irregularities along the margin are reflected along the deformational font. This phenomenon has been proposed in the Appalachians. During the Iapetan rifting (plus the other rifting events) along the Laurentian margin, a series of embayments (concave oceanward) and promontories (convex oceanward) separated by transform faults were produced (Thomas, 2014). Along such an irregular margin, the depositional environment is bound to vary laterally. During subsequent orogenic events, promontories evolved into recesses (concave towards the foreland), embayments evolved into salients (convex towards the foreland), and the transform fault boundaries evolved into transverse zones or cross structures (Thomas, 2014). Kwon and Mitra (2004) have compiled five end-member models for the kinematic development of salients as a function of principal transport direction, amount of shortening, degree of vertical axis rotation, thrust displacement, and presence or absence of tear faults or lateral/oblique ramp boundaries. These models, however, can be viewed as second-order, structural controls on formation of orogenic curvatures.

Lin et al. (1994) further noted that a collisional event between two promontories (of two different landmasses) results in a narrower but stronger orogenic deformation and a higher grade of metamorphism than a promontory-embayment collision. It can be concluded that an irregularity along a continental margin may serve as a nucleation site for lateral heterogeneity.

9.2 Inherited Basement Structures

Inherited basement structures form a first-order control on formation of cross structures. Basement structures are primarily related to continental rifting, hyper-extension of rifted margins (e.g.; Ribes et al., 2019) , crustal sutures, or back-arc extension (Figure 10). Furthermore, it has been noted that a failed rift-arm (aulacogen) can also serves as an inherited basement structure and control structural evolution (Rodgers, 1990; Brown et al., 1997; Perez-Estaun et al., 1997). Crustal sutures mark boundaries between two continental blocks and form as a result of collisional tectonics or terrane accretion. Generally, suture zones consist of highly deformed rock units and form weaker zones within the crust (Hope & Eaton, 2002; Whitmeyer & Karlstrom, 2007). These zones are known to be intruded by igneous bodies during the subsequent crustal stabilization

(Whitmeyer & Karlstrom, 2007), and these intrusion boundaries can also behave as zone of weaknesses during various tectonic events (Simony & Carr, 1997; Bader, 2009).

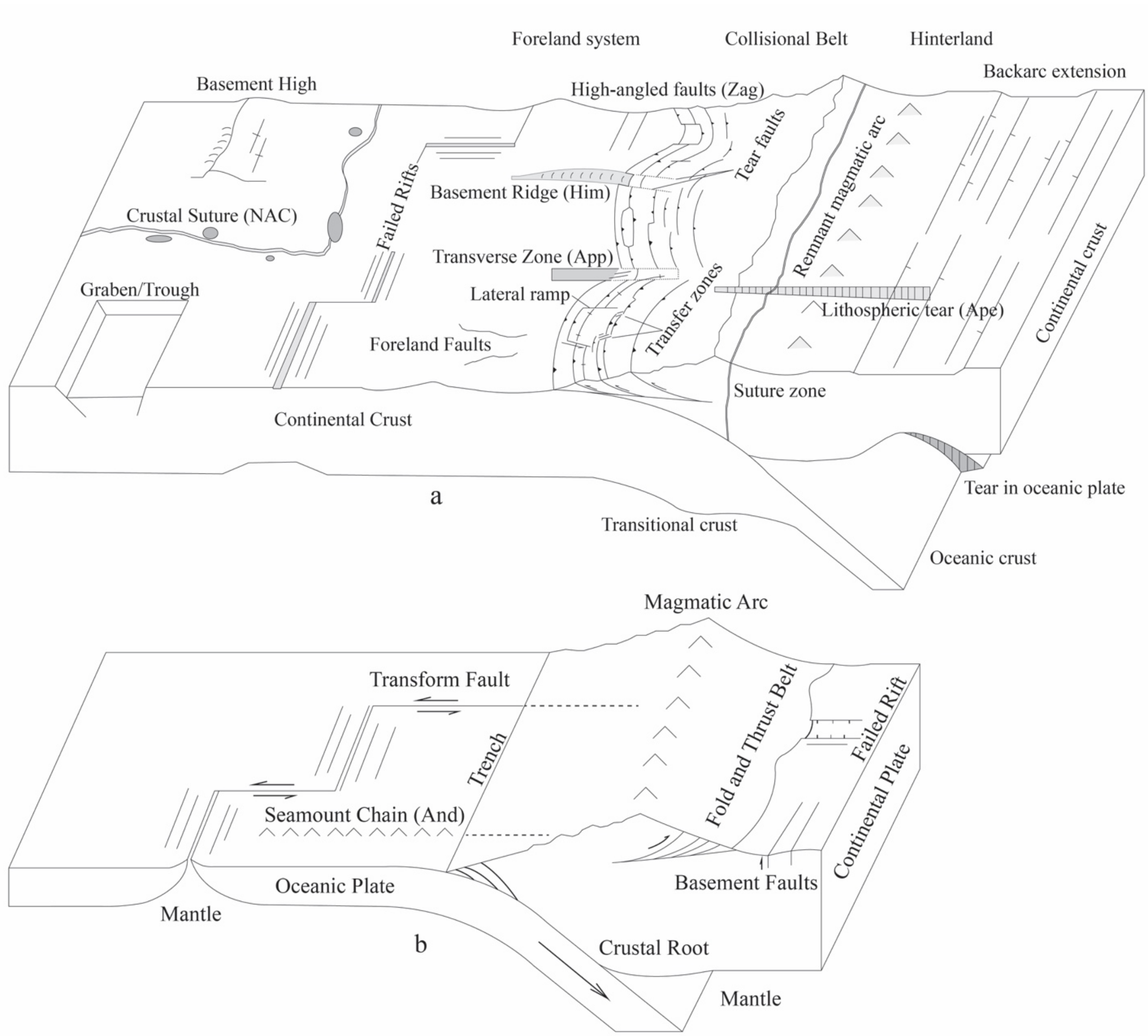

Figure 10: Various inherited basement structures and oceanic plate physiography in contractional settings. The gray ellipses in the 'Figure10a' represent igneous intrusions. The mountain belt that, in general, represents a certain structure are denoted in parenthesis.

And: Andes; Ape: Apennines; App: Appalachians; Him: Himalayas; NAC: North American Cordillera; Zag: Zagros.

9.2.1 Impacts of Basement Structures on Subsequent Sedimentation and Deformation

The presence of basement structures can greatly affect the foreland sedimentation and thereby have a strong control on the evolution of mountain belts as is seen in the Zagros, the Cordillera, and the Himalaya. Basement structures can isolate deposition centers in the foreland, which induces sharp lateral variations in both the thickness and facies of the sedimentary strata. A thicker foreland sedimentary sequence, when incorporated into the evolving taper wedge, will propagate much farther (salient) than a wedge consisting of a thinner sequence (recess) (e.g.; Paulsen & Marshak, 1999). Differential tectonic transport between the adjacent segments (salients and recesses) is often accommodated by cross structures such as tear faults, lateral ramps, and displacement transfer zones (Paulsen & Marshak, 1999). Orogen-parallel faults and folds are generally truncated at, or dragged into cross structures as seen in the Appalachians and the Cordillera (e.g.; Thomas, 2007; Whisner et al., 2014).

Similarly, lateral variations in sedimentary facies will also have a significant impact on the wedge evolution. An abrupt change in the structural elevation of decoupling layers results in lateral ramps (Thomas, 1990). Since thrust-ramp geometry is governed by the nature of the sedimentary column, any lateral variation in sedimentary facies is likely to be reflected in the thrust geometry (e.g.; Mitra, 1988). In a broader stratigraphic framework, if the presence of regional decollement horizons (units of salts, evaporites, and shales) varies laterally, then the deformation style in adjacent segments of the mountain belts will be very different. If a decollement horizon is present, then the taper angle will be low and propagate further than adjacent areas. Likewise, in the absence of such a horizon, the taper will build up to a larger angle in a narrower belt as in the Zagros (e,g,; Jadoon et al., 1994; Sobornov, 1996; Bahroudi & Koyi, 2003). The depth of these decollement horizons will also have a first order control on the fold geometry, such that a deeper horizon will result in folds with larger amplitude (Sepehr et al., 2006). Furthermore, the stratigraphic variation boundaries will also act as a lateral buttress to locally deflect the transport trajectories (Bahroudi & Koyi, 2003). If basement structures are reactivated during deposition, then syn-sedimentary faults and/or drape folds can form in the overlying sedimentary column (Thomas, 1990). These structures will also impact the fold-thrust belt evolution in a similar fashion to other cross-structures (e.g.; Zerlauth et al., 2014).

9.2.2 Reactivation of Basement Structures

Regardless of their origin, inherited basement structures can be broadly categorized into three groups: orogen-parallel, oblique, and transverse/cross basement structures. Various styles of preferential reactivation and inversion of basement structures have been discussed in the Apennines (Tavarnelli et al., 2004; Butler et al., 2006), and are also seen in other mountain belts. When orogen-parallel basement structures/faults are present, it is possible for them to be reactivated as reverse faults during orogenic compression (Figure 10a). It has, however, been noted that basement faults can be deformed under compression without inversion or reactivation (Pantet et al., 2020). In a reactivation scenario, the deformation between the basement and the overlying sedimentary cover is partially decoupled, whereby the basement deforms in thick-skinned style and the cover deforms as thin-skinned, fold and thrust belts (e.g.; Berberian, 1995; Lacombe et al., 2006). Discontinuities along the orogen-parallel basement structures will eventually result in laterally heterogeneous deformation style along the range. If the basement faults are steeply-dipping, then they often cut into the thin-skinned decollement faults of the sedimentary cover and act as frontal ramps during the upper, brittle deformation (e.g.; Bahroudi & Koyi, 2003). These orogen-parallel basement structures also accommodate convergence via strike-slip faulting under a favorable plate kinematic setting (e.g.; Authemayou et al., 2006). Margin perpendicular and oblique structures, however, either serve as lateral/oblique ramps or get reactivated as strike-slip faults with a dip-slip component based on their orientation in the stress field. Based on their aerial extent, these inherited structures can affect a few thrust sheets, or they can partition the entire mountain belt (Scrocca, 2006; Pizzi & Galadini, 2009; Satolli et al., 2014). Cross faults in the Himalayas may have had an origin in reactivated, margin-perpendicular basement faults and basement highs such as the Delhi-Hardwar ridge (Sahoo, 2000; Godin & Harris, 2014; Hubbard et al., 2018; Hubbard et al., 2021).

9.3 Rheology of the Crust

Variation in crustal rheology is another significant element in inducing lateral heterogeneities in orogenic belts, which itself is a function of its composition and thermal structure. In general, a weaker crust is more likely to buckle under contraction resulting in a high crustal relief, whereas a stronger crust will distribute deformation in more brittle manner (Allmendinger et al., 1997; Coutand et al., 2001). It has also been proposed that rheological heterogeneities along a continental margin front could result in orogenic curvatures, such that a stronger segment better resists deformation and evolves as a salient (e.g.; Malekzade et al., 2016). As seen in the Zagros, frictional strength in hinterland faults, which is governed by crustal rheology, can influence the mode of deformation partitioning in the orogenic belt and thereby have a control on the style and distribution of deformation (e.g.; Vernant & Chéry, 2006). Thermal structure and rheology of the crust can vary laterally due to variations in arc magmatism. Therefore, previous tectonic events will also have an indirect impact on lateral heterogeneity of an orogen.

Giambiagi et al. (2012) proposed that a stronger coupling between the upper and lower crust deformation corresponds to a higher, surficial topography and greater crustal shortening. Crustal coupling is thought to be governed by crustal composition such that a thick and more felsic crust has a stronger coupling between the upper and lower units than a thin and mafic crust. Thus, if there is strong lateral variation in crustal properties due to tectonic events such as terrane accretion or rifting then such variation will be manifested as laterally heterogenous deformation segments.

In the Alps, intensive folding and uplift of the Tauern window has been linked to the northward encroachment of the Dolomite acting as an indenter (Rosenberg & Garcia, 2011), which indicates that rheology of both converging crustal blocks have an impact on the morpho-tectonic evolution of an orogen.. As Butler et al. (2006) pointed out, the correlation between precise lithospheric strength profiles and orogenic deformation would further elucidate our understanding about the role of upper crustal rheology during orogenesis.

9.4 Plate Dynamics and Physiography of the Lower Plate

Plate dynamics between the upper and lower plate as well as the lower plate structures can disrupt the lateral homogeneity of orogenic belt (Figure 10b). Effects of lateral change in the style of subduction in the deforming orogen have contributed to lateral heterogeneity in the Andes. Any transverse structure in the lower plate is likely to be inherited in the upper plate, such as the Colorado Mineral Belt and the lithospheric tear in the Apennines (e.g.; Figure 10a) (Scrocca, 2006; Chapin, 2012). These structures in the lower plate impartially dissect the orogen rather than controlling the deformation style. The dynamics of the lower plate, however, generally differs across these transverse structures, which impacts the foreland sedimentation and fold-thrust belt evolution (Chapin, 2012). A similar phenomenon has been observed in cases where oceanic fracture zones (Figure 10b), orthogonal to the subduction zone, get subducted and result in the occurrence of cross faults in the over-riding plate. This scenario has been proposed on the Juan de Fuca plate subduction (Goldfinger, 1997), in Sumatra (Graindorge, 2008), and in the Andes (Robinson et al., 2006). Although not a cross structure, the Norumbega Fault System in the Appalachians has been regarded as a surface manifestation of a subducted oceanic ridge-transform system (Kuiper, 2016; Kuiper & Wakabayashi, 2018).

In orogens associated with oceanic subduction, variations in the relative rates of subduction versus convergence can create lateral heterogeneity in the over-riding plates as is seen in the Andes. When subduction outpaces convergence such that the convergent boundary retreats, back-arc extension occurs. These areas may have contrasting evolution when compared to adjacent areas where there is a better balance of the subduction versus convergence rates.

9.5 Obliquity of Plate Convergence

Lateral heterogeneity induced by oblique convergence is scale dependent. Obliquity of the plate convergence is generally accommodated by arrays of transtensional, transpressional, and strike-slip faults. In the Andes, transpressional cross faults are associated with mafic volcanism, while transtensional faults are linked with felsic magmatism as driven by magmatic viscosity difference (Petrinovic et al., 2006). At the scale of individual faults, these cross-faults disrupt the continuity of the fold-thrust belt as noted in the southern Andes and the Northern Calcareous Alps (Zerlauth et al., 2014; Stanton-Yonge et al., 2016). Lateral variation in the degree of oblique convergence itself, however, will also induce lateral heterogeneity in the orogenic belt due to an overall change in the stress field (McQuarrie et al., 2003; Vernant et al., 2004; Vernant & Chéry, 2006).

9.6 Further Implications of Cross Structures

Besides their role in foreland sedimentation and tectonic evolution of orogens, cross faults also have implications in the seismicity, mineralization, igneous activity, numerical modelling, and in topography of mountain belts. In the Zagros, a small seismic hazard has been linked to cross-faults (Berberian, 1995; Authemayou et al., 2006). Cross-faults in the Apennines are known to either serve as seismic segment boundaries or seismic loci (Valensise & Pantosti, 2001; Pizzi & Galadini, 2009). Similarly, in Taiwan, cross-faults are known to act as earthquake nucleation sites and as rupture segment boundaries (Deffontaines et al., 1997; Ching et al., 2011). Microseismicity in the

Himalayas has occurred along cross-strike trends, has terminated at cross-strike zones, and has made other spatial changes along cross-strike boundaries (Rajaure, 2013; Mugnier et al., 2017; Bilham, 2019; Mendoza, 2019). A transverse fault in the down-going Nazca plate had an enormous impact during the 2001 Peru earthquake (Robinson et al., 2003). From the economic standpoint, cross-faults form excellent conduits as well as reservoirs for magma and hydrothermal fluids (McMechan, 2012), which makes them target sites for economic mineral deposits. In the Papua New Guinea fold-thrust belt, transverse zones form sites of economically significant copper-gold deposits (Davies, 1991; Corbett, 1994; Hill et al., 2002). The Colorado Mineral Belt is another example of valuable mineral accumulation along a cross-structure (Chapin, 2012). Moreover, cross-structures can form potential traps for oil and gas accumulation, which, in some areas makes them an important site for hydrocarbon exploration (Wheeler, 1980; Séjourné & Malo, 2007; Bader, 2009). Fault systems can form excellent conduits and reservoirs for magmatic fluids and therefore can control the locus of volcanic complexes in convergent settings (e.g.; Petrinovic et al., 2006; Pérez-Flores et al., 2016; Roquer et al., 2017; Sielfeld et al., 2019; Lupi et al., 2020; Piquer et al., 2020). Thermo-mechanical tectonic models could potentially be rectified by changing the lithospheric parametric values along the strike such that they agree with the lateral variation in crustal deformation, as can be observed on the surface. Finally, cross-structures also have a major control on the geomorphologic evolution of mountain belts. In the Himalayan front, cross faults spatially coincide with river channels (Sahoo, 2000; Srivastava, 2018). The anti-Apennine faults have distinct topographic signatures (Coltorti et al., 1996). Cross-faults can be zones of weaknesses, and therefore, can also amplify erosional hazards, especially in active mountain belts.

10 Conclusions

While convergent mountain belts are dominated by spatially and temporally continuous orogen-parallel structures, geological and geophysical data shows that various forms of lateral heterogeneity, often marked by cross structures, are ubiquitous in most orogenic settings. In general, lateral heterogeneities along orogens have been manifested as: 1) along-strike changes in deformation style; 2) variation in igneous activity or metamorphic grade; 3) variation in seismic activity; 4) changes in topography and geomorphology; and 5) abrupt lateral stratigraphic changes. Common drivers behind these lateral heterogeneities include the geometry of the continental margin, inherited basement structures, lateral variation in stratigraphy of deforming sedimentary sequences, variation in crustal rheology, along-strike changes in plate tectonic setting, physiography of the lower plate, and obliquity of plate convergence. In most settings, these factors are interrelated and simultaneously influence the morpho-tectonic evolution of an orogen. Apart from their influence on foreland sedimentation and orogenic evolution, lateral heterogeneity and cross structures can have an impact on patterns of seismicity, natural resource occurrence, and natural hazards. We therefore stress the importance of documenting heterogeneity, mapping cross structures, and understanding the role these lateral changes play in mountain belt development along convergent margins.

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