**Structural inheritance and selective reactivation in the central Andes:
Cenozoic deformation guided by pre-Andean structures in southern
Peru**

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31 **ABSTRACT**

32

33 Structural, stratigraphic, and geochronologic constraints from the Eastern
34 Cordillera in the central Andean plateau of southern Peru (14–15°S) demonstrate the
35 existence and position of major pre-Andean structures that controlled the accumulation of
36 Triassic synrift fill and guided subsequent Cenozoic deformation. The timing of initial
37 clastic deposition of the Triassic Mitu Group is here constrained to ~242–233 Ma on the
38 basis of detrital and volcanic zircon U-Pb geochronology. Regionally distinct provenance
39 variations, as provided by U-Pb age populations from localized synrift accumulations,
40 demonstrate Triassic erosion of multiple diagnostic sources from diverse rift-flank
41 uplifts. Stratigraphic correlations suggest synchronous initiation of extensional basins
42 containing the Mitu Group, in contrast with previous interpretations of southward rift
43 propagation. Triassic motion along the NE-dipping San Anton normal fault
44 accommodated up to 7 km of throw and hanging-wall deposition of a synrift Mitu
45 succession >2.5 km thick. The contrasting orientation of a non-reactivated Triassic
46 normal fault suggests selective inversion of individual structures in the Eastern Cordillera
47 was dependent on fault dip and strike. Selective preservation of a ~4 km thick succession
48 of Carboniferous–Permian strata in the down-dropped San Anton hanging wall, beneath
49 the synrift Mitu Group, suggests large-scale erosional removal in the uplifted footwall.
50 Field and map observations identify additional pre-Andean thrust faults and folds
51 attributed to poorly understood Paleozoic orogenic events preserved in the San Anton
52 hanging wall. Selective thrust reactivation of normal and reverse faults during later
53 compression largely guided Cenozoic deformation in the Eastern Cordillera. The
54 resulting structural compartmentalization and across-strike variations in kinematics and
55 deformation style highlight the influence of inherited Paleozoic structures and Triassic
56 normal faults on the long-term history of convergent margin deformation in the Andes.

57

2. INTRODUCTION

The effects of inherited crustal architecture on the style, magnitude, and distribution of Andean shortening remain poorly constrained. Many global examples demonstrate that pre-orogenic structural and stratigraphic features define important initial conditions on the evolution of major mountain belts, including the Alpine-Himalayan system and North and South American Cordilleras, among others (e.g., Argand, 1916; Jackson, 1980; Allmendinger et al., 1983; Allègre et al., 1984; Gillcrist et al., 1987; Colletta et al., 1990; Grier et al., 1991; Allmendinger and Gubbels, 1996; Yin and Harrison, 2000; McQuarrie 2002; DeCelles, 2004; Butler et al., 2006; Mescua and Giambiagi, 2012). Regionally, pre-orogenic crustal architecture and lithospheric strength may control shortening and regulate uplift of high topography (e.g., Barragan et al., 2005; Mouthereau et al., 2013). More locally, selective reactivation of individual inherited structures potentially dictates the structural style, degree of basement involvement, and geometries of basin evolution (e.g., Bayona et al., 2008; Giambiagi et al., 2011). Crustal shortening of inherited continental rift systems may result in complex deformational and depositional styles that challenge reconstruction efforts. In the Andes, multiple pre-Andean deformation events of Proterozoic to Mesozoic age have been proposed along various segments of the western margin of South America (e.g., McGroder et al., 2015). Locating these antecedent structures and, in particular, the distribution of normal faults and associated basins, remains a challenge in intensely shortened regions such as the central Andes.

Mesozoic intracontinental rifts and extensional backarc basins have been identified in the northern (Colleta et al., 1990; Mora et al., 2006), central (Manceda and Figueroa, 1995; Sempere et al., 2002), and southern (Dalziel, 1981) segments of the Andean orogenic belt. The age, location, and structural grain of these pre-Andean features vary along strike, potentially influencing subsequent Cenozoic deformation. Inherited crustal and stratigraphic architectures played a key role in developing thick- and thin-skinned deformation, delimiting the along-strike segmentation of structural style, and driving variable subsidence in accompanying foreland basins (Kley et al., 1999; Bayona et al., 2008; Giambiagi et al., 2012). In the northern Andes of Colombia, the

89 main phase of pre-Andean extension began in the Late Jurassic–Early Cretaceous and
90 was accommodated by multiple rapidly subsiding half-graben basins (Casero et al.,
91 1997). Selective inversion of pre-existing basement-involved normal faults strongly
92 influenced the development of Cenozoic fold-thrust structures and modern topography in
93 the Eastern Cordillera (Dengo and Covey, 1993; Mora et al., 2006). In the central Andes
94 of northwestern Argentina, basement-involved Cretaceous rifting occurred at variable
95 orientations relative to later N-S Andean structural trends and Cenozoic inversion of
96 favorably oriented normal faults contributed to thick-skinned deformation in the Eastern
97 Cordillera (Cominquez and Ramos, 1995; Monaldi et al., 2008). Farther south, inversion
98 of Triassic-Jurassic extensional basins within the Neuquén region of Argentina controlled
99 the formation of the Malargüe fold-thrust belt (Manceda and Figueroa, 1995; Giambiagi
100 et al., 2012). These examples demonstrate that the variable timing, orientation, style, and
101 potential basement involvement of pre-existing normal faults may help govern the style,
102 rate, and magnitude of deformation and basin evolution in different segments of the
103 Andes. Despite many documented cases of pre-Andean deformation, few examples of
104 reactivated structures exist in the central Andes of Bolivia and southern Peru. This raises
105 the question whether structural inheritance is important throughout the Andes or is
106 restricted to particular segments.

107 Andean reactivation of pre-existing normal faults often involves thick-skinned
108 deformation (Kley, 1996; Mora et al., 2006; Mescua and Giambiagi, 2012), may
109 contribute to along-strike changes in shortening and orogen width (Kley et al., 1999), and
110 may control deformation transfer to thin-skinned domains (Giambiagi et al., 2009).
111 Basement structures may also compartmentalize discrete structural domains (Rosas et al.,
112 2007). Shortening geometries and kinematics are often guided by the location,
113 orientation, and potential reactivation of pre-existing structures (Sibson, 1985; Sibson,
114 1990; Seeber and Sorlien, 2000; Di Domenica et al., 2014). In northern Peru, Hermoza et
115 al. (2005) demonstrate that Eocene–Pliocene subsidence and shortening in the Huallaga
116 basin was influenced by graben inversion. In Colombia and central Peru, development of
117 passive roof duplexes and hinterland-verging thrusts has been linked to the location and
118 subsurface orientation of normal faults in the Subandean foreland (Mora et al., 2014). In

elevated Andean zones, however, identifying the geometries and kinematics associated with inversion remains challenging.

In southern Peru, the Eastern Cordillera forms the high-elevation, NW-trending range separating the northern Altiplano hinterland basin from the Subandean fold-thrust belt and foreland basin. Sempere et al. (2002) proposed a pre-Andean rift along the modern Eastern Cordillera that initiated in central Peru during the Late Permian and propagated to southern Bolivia by the Middle Jurassic. Although the proposed rift location generally coincides with the preserved synrift Permo-Triassic Mitu Group, the location, orientation, and polarity of major bounding normal faults have not been adequately defined. The geodynamic setting for Mitu deposition remains debated. Early interpretations attributed the Mitu Group to erosion of a western coastal orogen (Newell, 1953), whereas later studies suggested deposition in diverse extensional settings, including a backarc extensional province (Noble et al., 1978; Reitsma, 2012), continental rift (Dalmayrac et al., 1980), or zone of extensional orogenic collapse (Dewey, 1988; Rosas et al., 2007; Ramos, 2009; McGroder et al., 2015). Although some extensional Mitu structures have been imaged in the Subandean Zone of central and northern Peru (Hermoza et al., 2005; Rosas et al., 2007), and hypothesized farther south (McGroder et al., 2015), the location of exposed pre-Andean normal faults in southern Peru has not been established.

Recent changes to the Mesozoic chronostratigraphic framework for central and southern Peru highlight key uncertainties in the timing and style of extension (Carlotto et al., 2010). Originally attributed to Permian deposition (Newell, 1953), subsequent studies of the Mitu Group have suggested depositional ages of Late Permian to Early Triassic (Sempere et al., 2002), exclusively Triassic (Reitsma, 2012), or potentially Late Triassic to Early Jurassic (Carlotto et al., 2010). Additionally, other workers have suggested a lower member present locally in the Eastern Cordillera originally attributed to the Mitu Group may be correlative with marine Middle-Upper Permian Ene Formation deposits of the Subandean Zone in central and northern Peru (Peña Guimas, 2008; Carlotto et al., 2010). Although the widespread occurrence of the Mitu Group in the Eastern Cordillera throughout Peru suggests a single integrated basin, rapid lateral variations in thickness, facies, and depositional environments suggest the potential for localized subsidence in

discrete nonmarine subbasins with minor marine transgressions (Newell, 1953; Carlotto et al., 2010; Reitsma, 2012). Questions regarding the potential connection, distribution, and age of Mitu basins have hindered reconstruction of Permian-Triassic basin geometries and regional subsidence. The chronostratigraphic framework, spatio-temporal provenance history, and distribution of synrift Mitu facies are poorly constrained, yet critical to establishing the pre-Andean structural and stratigraphic architecture.

In this paper we demonstrate the location of major pre-Andean structures within the central Andean plateau using regional subcrop, map, thickness, and facies relationships across the Eastern Cordillera of southern Peru. Cross-cutting structural relationships define several structural domains that delimit these pre-Andean structures, suggesting that potential basement-involved deformation guided structural compartmentalization. We further refine the chronostratigraphic and provenance framework of the Triassic Mitu Group and assess the timing of basin initiation and coalescence along strike. Finally, we show how selective reactivation of inherited structural features guided deformation from deeper basement levels to shallower cover strata, influencing the kinematic evolution of the Eastern Cordillera.

3. GEOLOGIC FRAMEWORK

The Andes of southern Peru define a NW-trending orogen consisting of the subduction trench, onshore forearc slope, Western Cordillera magmatic arc, Altiplano plateau, Eastern Cordillera fold-thrust belt, Subandean Zone deformation front, and modern foreland basin (Figure 1). The Eastern Cordillera preserves two major features that are expressed for hundreds of kilometers along strike in Peru and Bolivia. First, the NW-trending Cordillera de Carabaya (and Cordillera Real of Bolivia) is defined by Triassic plutonic rocks intruded into Ordovician-Devonian metasediments that support a high-elevation (>5–6 km) geomorphologic boundary (drainage divide) between the 4-km-high Altiplano and <1-km-high Subandean Zone (Kontak et al., 1990; Gillis et al., 2006). Second, the NW-trending central Andean backthrust belt (CABB; Figure 1) is defined by dominantly SW-verging fold-thrust structures involving Cretaceous marine sandstones, shales, and limestones. The CABB continues in Bolivia (including the Huarina fold-

thrust belt), where it exposes dominantly Paleozoic strata (McQuarrie and DeCelles, 2001; Murray et al., 2010). The Altiplano-Eastern Cordillera structural boundary in Peru is defined by the Ayaviri fault, part of a larger Ayaviri fault system (AFS), one of the major backthrust structures within the CABB that was previously referred to by the Spanish acronym SFUACC (Sempere et al., 1990; Ibarra et al., 2004; Perez and Horton, 2014). Between the Cordillera de Carabaya and the CABB is the Macusani Structural Zone (MSZ), also known as the Precordillera de Carabaya (Kontak et al., 1990; Sandeman et al., 1997) or domain of "Late Hercynian" folds (Laubacher, 1978) where exposures are dominated by Carboniferous, Permian, and Triassic Mitu Group rocks. Most faults and folds within the MSZ are not parallel with regional NW-SE trends of southern Peru, preserving N-S, NE-SW and NNW-SSE orientations (Figure 2).

The tectonic setting of the Andean margin during late Paleozoic-early Mesozoic time remains debated. In Argentina, Carboniferous–Permian rocks folded during the Permian San Rafael orogeny are unconformably overlain by bimodal volcanics and nonmarine siliciclastics of the Triassic synrift Choiyoi Group. Triassic extension has been interpreted as collapse of the late Paleozoic orogen (Dewey, 1988; Mpodozis and Kay, 1992; Kleiman and Japas, 2009; Ramos, 2009; McGroder et al., 2015). In southern Peru, similar relationships between folded Carboniferous to Permian rocks and the overlying Mitu Group suggest post-orogenic collapse may have been widespread (Mpodozis and Kay, 1992; Rosas et al., 2007). Other workers have suggested Triassic extension occurred within an intracontinental rift system (Dalmayrac et al., 1980; Kontak et al., 1990; Sempere et al., 2002). Alternatively, Noble et al. (1978) and Reitsma (2012) proposed that this Triassic extension occurred in a backarc setting, where Triassic plutons emplaced in the Cordillera de Carabaya represent magmatism inboard of a volcanic arc that was largely destroyed by later subduction erosion.

Subcrop relationships preserved in Ordovician to Cretaceous strata of the Eastern Cordillera provide insights into the pre-Andean structural and stratigraphic framework. We identify these subcrop relationships from map data to refer to the variable substrate upon which a particular unit was deposited, thus revealing spatial patterns in unconformable relationships. The following is a summary of the major patterns, although local variations are common. A thick (up to 10 km) succession of Ordovician–Devonian

slates, phyllites, and metasedimentary rocks defines the Altiplano, Eastern Cordillera, and Subandean Zone (Laubacher and Mégard, 1985). No lowermost Ordovician, Cambrian or older rocks are exposed. A major unconformity attributed to a middle Paleozoic orogenic event (previously referred to as "Early Hercynian") separates the Ordovician–Devonian strata from overlying Carboniferous, Permian, Triassic, or Cretaceous rocks (Steinmann and Hoek, 1912; Newell, 1949; Newell et al., 1953; Ahlfeld and Branisa, 1960; Rivas, 1971; Isaacson, 1975; Suarez Soruco, 1976, 1992, 2000; Castaños and Rodrigo, 1978; Dalmayrac, 1978; Laubacher, 1978; Marocco, 1978; Martinez, 1980; Dalmayrac et al., 1980; Sempere et al., 2004). Up to 4 km (Barros and Carneiro, 1991) of Carboniferous shales and quartzites to Permian limestones with 3–5 km wavelength folds are locally preserved beneath another angular unconformity. Cross-cutting relationships suggest that these folds formed during late Paleozoic deformation, previously referred to as the Juruá or "Late Hercynian" orogeny (Audebaud and Laubacher, 1969; Laubacher, 1978; Soler and Bonhomme, 1987; Barros and Carneiro, 1991; Tankard, 2001; Rosas et al., 2007). Where preserved, the ~4 km thick Carboniferous–Permian section is capped by the Mitu Group in angular unconformity (Newell, 1953). Elsewhere, the Mitu Group overlies Silurian–Devonian rocks. Upper Jurassic–Cretaceous marine sandstones, shales and limestones often overlie the Mitu Group in disconformable or angular unconformable contacts, but also overlie Silurian–Devonian or Carboniferous units where the Mitu is absent. The widespread carbonates of the Jurassic Pucará Group overlying the Mitu Group in central and northern Peru (Rosas et al., 2007) and not observed in southern Peru.

The Mitu Group has been defined as the intercalated nonmarine red sandstones, shales, conglomerates, volcanoclastics, and alkali volcanic rocks above Permian Copacabana Formation carbonates and below Jurassic–Cretaceous marine siliciclastic and limestone rocks (Newell et al., 1953; Laubacher, 1978; Kontak et al., 1990). Basal Mitu exposures have been reinterpreted as marine deposits equivalent to the Middle to Upper Permian Ene Formation ubiquitous throughout the Subandean Zone of central and northern Peru (Carlotto et al., 2010). In southern Peru, Mitu deposits are preserved between the Ayaviri fault and the Cordillera de Carabaya. The thickest and best exposures of the Mitu Group are in the MSZ, where rapid lateral thickness, lithology, and

facies changes are observed (Figure 3). Previous workers have linked Mitu deposition of the bimodal calc-alkaline volcanic and nonmarine siliciclastic deposits with the intrusion of monzogranitoids along the Eastern Cordillera, and suggested synrift deposition occurred in multiple half grabens with associated magmatism (Mégard, 1978; Laubacher, 1978; Dalmayrac et al., 1980; Kontak et al., 1990; Sempere et al., 2002; Mišković et al., 2009). Although the Mitu Group is commonly attributed to Permian through Triassic deposition on the basis of regional stratigraphic relationships and correlations (Newell, 1953; Dalmayrac et al., 1980; Kontak et al., 1990; Sempere et al., 2002), a refined chronostratigraphic framework for the Altiplano near Cusco and Sicuani suggests solely Triassic deposition (Reitsma et al., 2010; Reitsma, 2012).

Because deep structural levels exposed at high elevations within the Eastern Cordillera coincide with synrift Mitu exposures, some authors suggest the Triassic extensional basin was inverted during Cenozoic Andean shortening (Mégard, 1984; Barros and Carneiro, 1991; Rosas et al., 2007; Scherrenberg et al., 2012). Nevertheless, the locations of major extensional structures along the Eastern Cordillera have not been defined. Reconstructing the Mitu basin and identifying the major basin-bounding faults is necessary to define the upper crustal structural and stratigraphic architecture prior to Andean orogenesis.

4. DEPOSITIONAL AGE AND PROVENANCE OF THE SYNRIFT MITU GROUP

4.1. U-Pb zircon geochronology methods

Detrital zircon U-Pb geochronology of five medium to coarse-grained sandstones from various levels of the Mitu Group facilitate assessments of depositional (stratigraphic) ages and sediment provenance (Figure 4). Samples were collected from three localities (1, 2, and 5; Figure 1). Our simplified numbering scheme corresponds to sample locations (i.e., samples 1a, 1b, 1c, 1d were all collected at locality 1), as presented in the supplementary materials. Sample locations displayed variable facies including medium to coarse-grained sandstone, pebble to cobble conglomerate, volcaniclastic, and

thin extrusive basalt deposits. We interpret coarser Mitu deposits as evidence for proximal rather than distal sources.

At localities 1 and 2, detrital samples were collected near the base (samples 1b and 2a) and top (samples 1c and 2b) of the Mitu Group (Figure 1). Samples were prepared using standard crushing, density, and magnetic separation techniques (Gehrels, 2000; Gehrels et al., 2008). High purity zircon separates were mounted into epoxy pucks, polished to reveal zircon interiors, and analyzed by laser ablation-multicollector-inductively coupled plasma-mass spectrometry at the Arizona LaserChron Center at the University of Arizona following standard procedures (Gehrels et al., 2008). Back-scattered electron imaging was used to positively identify zircon grains. Approximately 120 zircons of varying size, shape and quality were randomly analyzed from each sample. Zircons with obvious inclusions were avoided.

Results from LA-MC-ICP-MS analyses are plotted as probability density plots with age distribution histograms of 50 Myr bins (Figure 4). Populations of zircon ages are depicted as peaks in the probability density plot and reflect contribution from source regions with variable zircon ages. Differences among zircon populations between samples have been interpreted as variable contributions from distinct source areas.

4.2. U-Pb detrital zircon results

We identify six populations of detrital zircon ages from the samples. Following is a discussion of populations from oldest to youngest.

Detrital zircon grains exhibiting 1500–3000 Ma ages are observed in all five samples. The original sources of these Mesoproterozoic zircons potentially include South American cratonic blocks, the Arequipa terrane in the Western Cordillera (Loewy et al., 2004; Ramos, 2009; Bahlburg et al., 2011), or recycled Paleozoic strata of the Eastern and Western Cordilleras (Reimann et al., 2010; Bahlburg et al., 2011; Perez and Horton, 2014). Detritus sourced directly from cratonic blocks would likely reflect dominantly Mesoproterozoic populations and lack Paleozoic zircons. The minor presence of Mesoproterozoic zircons and higher population of Paleozoic populations observed in all five detrital samples from the Mitu Group is similar to detrital populations observed in

Eastern and Western Cordillera units. We suggest this population represents recycled detritus from either the Western or Eastern Cordillera, rather than far traveled first cycle zircons shed directly from the craton.

A 1300–1500 Ma age population ($n=13$) is observed in sample 1b (lowest detrital sample at locality 1), but is much reduced ($n \leq 4$) in all other samples. This age signature has three potential sources. The first is the Rondonia/San Ignacio cratonic rocks of Bolivia (Teixeira et al., 1989; Cordani and Teixeira, 2007; Bettencourt et al., 2010). The second potential source includes isolated orthogneiss exposures in central to northern Peru (Chew et al., 2008). This population is relatively rare in Paleozoic sedimentary and intrusive rocks from the Western and Eastern Cordillera of southern Peru (Mišković et al., 2009; Reimann et al., 2010; Bahlburg et al., 2011; Perez and Horton, 2014). The third potential source is Ordovician clastic deposits from the Precordillera terrane of Argentina, where zircons of these ages possibly originated in the Granite-Rhyolite province of Laurentia (Thomas et al., 2015).

A broad 850–1300 Ma population is observed in all five samples of the Mitu Group. This population has multiple potential sources, including first cycle zircons shed directly from the Sunsas orogen to the east along the Brazilian craton-Andean interface (Bahlburg et al., 2011), or the Arequipa terrane of the Western Cordillera (Loewy et al., 2004). Alternatively, this age population may represent recycled zircons from Paleozoic strata throughout Peru. The omnipresence of this population from eastern and western sources and lack of paleocurrent data present a challenge to delineate a unique sediment source.

A 450–750 Ma age population is present in all five samples. This population is common in Paleozoic strata of the Eastern and Western Cordilleras throughout Peru. Zircons of these ages are attributed to formation during the Paleozoic Famatinian (~450–520 Ma) and Pampean/Brasiliano orogenic cycles (~550–750 Ma) (Bahlburg et al., 2009) and are likely recycled from Paleozoic rocks. Zircon ages alone are insufficient to identify the source of this population.

Carboniferous zircons (300–350 Ma) are present in all five samples. In samples 1b, 1c, 2a, and 2b, this population is insignificant ($n \leq 2$). In sample 5, this population is more prominent ($n=6$). Potential sources are limited to Carboniferous plutons and Ambo

Group rocks of the Eastern Cordillera near Cusco (Mišković and Schaltegger, 2009; Reitsma, 2012). This population was not observed in Ambo Group rocks in this study area situated SE of Cusco (n=0; Perez and Horton, 2014).

A population of 200–300 Ma zircons is present in all five samples, and is a large proportion of samples 2a and 2b (>10%). Sample 5, the westernmost sample, has the fewest (n=1) zircons from this age population. Late Paleozoic to early Mesozoic zircons observed in Mitu deposits are likely first cycle volcanic zircons. Potential sources are limited to synrift volcanic or plutonic rocks of the Eastern Cordillera, probably situated close to the Cordillera de Carabaya plutons.

4.3. U-Pb depositional age constraints

Zircon U-Pb ages of interbedded volcanic horizons and the youngest populations from detrital zircon samples provide chronostratigraphic constraints for the Mitu Group (Figure 4). Geochronologic results from volcanic horizons constrain the true depositional age, whereas the youngest population of detrital zircons (defined as a calculated weighted mean age using Isoplot; Ludwig, 2008) in a sample constrains the maximum depositional age, recognizing that true depositional age may be younger. Locality 1 represents the thickest preserved Mitu succession (>2 km) in the study area. Sample 1a, the stratigraphically lowest Mitu sample, is an extrusive basaltic-andesite interbedded with coarse-grained clastic and volcanoclastic deposits; the sample yielded few zircons. The youngest zircon is 441.2 ± 7.2 Ma, older than the Late Permian to Triassic stratigraphic age for the basal Mitu Group. Detrital zircon sample 1b is ~1000 m upsection, with a maximum depositional age of 259.1 ± 6.4 Ma (n=2), although the true depositional age could be younger. Reworked tuff sample 1d and detrital sample 1c are ~1000 m further upsection and dated as 242.0 ± 11.0 Ma (n=3) and 232.9 ± 4.0 Ma (n=3), respectively. Reworked tuff sample 1d is ~3 m below detrital sample 1c. From these results, we interpret Late Permian to early Late Triassic deposition of the Mitu Group at locality 1.

Detrital zircon samples at locality 2 constrain depositional ages for the base and top of the Mitu Group (2a and 2b, respectively). Here, within 7 km NE of the major San Anton map boundary (Figure 1), the preserved Mitu succession is <1.5 km thick. Basal sample 2a was collected ~20 m above the basal angular unconformity capping Permian

Copacabana limestones. The maximum depositional age is 234.8 ± 2.6 Ma ($n=17$). Sample 2b, from <150 m below the depositional contact with the overlying Cretaceous Muni Formation, has a maximum depositional age of 235.8 ± 2.9 Ma ($n=30$). Within error, both samples from locality 2 yield the same age, and are slightly older than the youngest age at locality 1. If these results approximate true depositional ages, the 1.5 km thick Mitu section at locality 2 must have accumulated rapidly. Alternatively, the ~235 Ma zircon age population observed in both samples may have derived from an earlier pulse of synrift volcanism at the onset of Mitu extensional subsidence. Therefore, the entire Mitu succession at locality 2 may record ~235 Ma volcanic and pluton sources, resulting in uniform pre-depositional ages throughout the succession. In either case, we consider locality 2 to preserve generally early Late Triassic depositional ages.

Sample 3 was collected from a local Mitu exposure within the CABB beneath an angular unconformity with the overlying Cretaceous Muni Formation. The sample is from a fine-grained purple andesitic tuff with penetrative cleavage and small-amplitude folds that do not affect overlying Cretaceous rocks. The sample yielded few zircons, and many were inherited older zircons. The three youngest euhedral, zoned zircons yield an age of 239.0 ± 10.0 Ma, consistent with a Middle to Late Triassic age of Mitu deposition in the CABB.

Samples 4a and 4b were collected from locality 4 near the base and top, respectively, of a 300–400 m thick succession of intercalated medium-grained tuffs and volcanoclastic deposits. Here, the basal contact of the Mitu Group is defined by an angular unconformity with the Silurian/Devonian Chagrapi Formation. The upper contact is defined by an angular unconformity with the Cretaceous Muni Formation. Samples 4a and 4b give ages of 236.1 ± 2.4 Ma and 236.1 ± 2.9 Ma, respectively. Locality 4 preserves Middle to Late Triassic Mitu Group deposition and volcanism.

Locality 5, the westernmost locality, has one detrital zircon sample. The youngest detrital zircon grain has an age of 289.2 ± 6.7 Ma. This single grain is older than the Leonardian age (~270–280 Ma) suggested for the top of the underlying Copacabana Formation, based on biostratigraphic constraints (Carlotto, 2010). Hence, the true depositional age of the Mitu Group is younger than the single detrital zircon grain age, but remains poorly constrained at this location.

Zircon U-Pb geochronologic constraints for the Mitu Group broadly support previous interpretations of Late Permian to early Late Triassic deposition (Newell et al., 1953; Laubacher, 1978; Kontak et al., 1990). However, we suggest that the youngest age populations for samples 1a, 1b, and 5 do not reflect true depositional ages. The young populations for sandstone samples 1b and 5 are defined by very few grains ($n=2$ and $n=1$, respectively). Sample 1a is from an extrusive basaltic-andesite within the Mitu Group, yet contains multiple zircons older than 440 Ma, suggesting entrainment of inherited zircons from surrounding Paleozoic host rock. A lack of eruptive zircons may be attributable to the relatively mafic composition of these igneous materials. In contrast, other detrital zircon samples, tuffs, and volcaniclastic samples have significant ($n \geq 3$) populations of potentially syndepositional zircons. These samples (1c, 1d, 2a, 2b, 3, 4a, 4b) yield ages between 242 ± 11 Ma and 232.9 ± 4.0 Ma, suggesting a Triassic depositional age of the Mitu Group in southern Peru. Reitsma (2012) demonstrated initial Mitu deposition between 241.5 ± 6.0 Ma and 234.3 ± 0.3 for areas up to ~700 km NW of this study area, consistent with our new depositional age constraints.

We interpret two potential scenarios for the U-Pb results. First, synrift Mitu deposition may have initiated in the Late Permian but lacked the felsic volcanic sources with significant zircon yields. We consider this unlikely, as the upper Copacabana Formation is well constrained as Late Permian in age. Moreover, ages of intrusions that cross cut folded Permian Copacabana Formation have been dated as ~245–233 Ma (Soler and Bonhomme, 1987). The Mitu Group is often unconformably overlies folded Permian Copacabana Formation strata (Figure 2) and therefore postdates deposition, folding, and Early Triassic intrusive activity that affected the Copacabana Formation. Second, in our preferred scenario, zircon U-Pb constraints from the Abancay-Cusco region (Reitsma, 2012) and this study suggest nearly uniform Middle to Late Triassic initiation of Mitu Group deposition over a ~700 km distance along strike. Samples lacking Triassic zircons (e.g., sample 1b) do not record true depositional ages. Rather these samples likely record the maximum possible depositional age and were deposited some time after the youngest zircon ages.

5. PRE-ANDEAN NORMAL FAULTS

5.1. Outcrop example

Few outcrop examples of pre-Andean normal faults have been identified in the Eastern Cordillera of the central Andes. Here we present a previously undocumented normal fault situated at ~4800 m elevation ~20 km east of the town of Macusani in southern Peru (Figure 5). The fault is exposed along the west side of a glacially carved ~500 m deep valley. It strikes ~110° and is exposed for <2 km along strike. Cross-cutting relationships demonstrate that the ~40° SSW-dipping normal fault post-dated harmonic folding of the Permian Copacabana Formation limestones and older rocks. No overlap assemblage is preserved that post-dates normal fault motion.

Mitu Group rocks preserved in the normal fault hanging wall dip ~20° N. Restoration of Mitu beds to their original horizontal orientation suggests that the original normal fault dipped ~60° SSW. Although we observed no evidence for folding of the Mitu Group at this outcrop, the normal fault and associated fault blocks underwent regional tilting during later Andean deformation. The normal fault is interpreted as a probable Triassic structure, although uncertainty remains due to a lack of growth strata and a preserved overlap assemblage.

5.2. Subcrop relationships and lateral thickness variations

Depositional, unconformable, and fault relationships were evaluated across the Eastern Cordillera for the Carboniferous (Ambo Group and Tarma Formation), Permian (Copacabana Formation), Triassic (Mitu Group), and Cretaceous units based on existing geologic maps (INGEMMET, 1999), our own observations, and previous studies (Audebaud and Laubacher, 1973; Mégard, 1978; Dalmayrac et al., 1980). We agree with previous workers on the presence of three major unconformities. Well-constrained sections at seven key columns A–G (Figure 1, Figure 6a) illustrate lateral changes in thickness, subcrop relationships, and these unconformities across the Eastern Cordillera.

A cross-section synthesis of these subcrop and thickness variations (Figure 6a) reveals relationships that necessitate pre-Andean structures (Figure 6b). The dip direction of normal faults 1* – 3* are correlated with the modern thrust faults that mark the subcrop and thickness boundaries. We suggest these may be inverted normal faults. Subsurface normal fault 4* is inferred based on stratigraphic and subcrop relationships; it

454 is the conjugate to fault 3*. Fault 4* was cut by the Cordillera de Carabaya backthrust
455 (Figure 1, 5, and 6). One of the most significant subcrop and thickness changes is
456 observed between columns C and D across the major San Anton map boundary (3*,
457 Figure 1). At column C we demonstrate that >2 km of Cretaceous rocks are deposited
458 unconformably on <300 m of Mitu Group rocks, which lay unconformably on Devonian
459 and older strata. Carboniferous to Permian strata are absent. In contrast, column D
460 situated 10–20 km NE preserves ~4 km of folded Carboniferous to Permian rocks
461 beneath up to 700 m of Mitu Group rocks. At the surface, there is no fault relationship at
462 the San Anton boundary. We interpret this as a tilted unconformity. However, we suggest
463 that this boundary was achieved by a NE-dipping normal fault at depth.

464 We present two possible scenarios to explain the observations across the San
465 Anton boundary (Figure 7). Both scenarios use a conservative thickness estimate of 3500
466 m for Carboniferous to Mitu Group, although true thicknesses are often greater. Scenario
467 A recreates the subcrop and thickness relationships with a tapered stratigraphic wedge
468 and does not require a pre-Andean structure. We investigate two cases for stratigraphic
469 wedges with dips of 10 and 3°. With these dips a wedge ~18–61 km wide is required to
470 create a 3500 m thick Carboniferous to Mitu Group succession. To laterally juxtapose
471 this thick package within ~10 km of the <300 m thick Mitu Group would require ~8–51
472 km of shortening of the wedge (~44–84%). We consider this scenario unlikely, as our
473 lowest shortening magnitude is greater than published total shortening estimates for this
474 segment of the Andes is 123 km, or 40% (Gotberg et al., 2010). The shortening required
475 in scenario 1 (~8–51 km) suggests up to ~41% of total shortening across this segment of
476 the Andes was accommodated along a single structure, which is also unlikely.

477 Scenario B recreates the subcrop and thickness variations with a NE-dipping
478 normal fault that preserves Carboniferous to Permian strata and deposition of thick Mitu
479 Group strata in the hanging-wall setting, a steeply dipping normal fault would require
480 minimal extension to accommodate >3500 m of Carboniferous to Mitu Group strata. In
481 the footwall, Carboniferous to Permian strata were not preserved. We suggest these units
482 were eroded from the footwall. Observations of the Mitu Group immediately to the NE of
483 the San Anton boundary at sample localities 2a, b and column D demonstrate a mix of
484 coarse volcanoclastics interbedded with pebble-cobble conglomerates composed primarily

of Permian limestone clasts (Figure 3e). Absence of Permian Copacabana Formation limestones in the proposed footwall and the presence of Copacabana limestone clasts in Mitu Group deposits in the proposed hanging wall are consistent with a normal fault at this position. These fault proximal facies, subcrop relationships, and thickness variations observed across the San Anton boundary suggest it is the result of a pre-Andean normal fault. The thickness of strata preserved in the hanging wall (up to 7 km) suggests the San Anton boundary represents a major normal fault that controlled the Mitu Rift.

5.3. Pre-Andean deformation preserved by structural compartmentalization

Multiple phases of shortening and a phase of extension have been proposed for southern Peru (Mégard et al., 1971; Laubacher, 1978; Mégard, 1978; Dalmayrac et al., 1980; Laubacher and Mégard, 1985; Jimenez et al., 2009). Identifying the pre-Andean San Anton normal fault is critical because Paleozoic rocks preserved in the hanging-wall (MSZ) record multiple pre-Andean deformational events. These pre-Andean deformational events have variable trends, including non-coaxial orientations oblique to typical Andean trends. Outside of the MSZ, Andean structures trend NW-SE. In the CABB and Eastern Cordillera, Cretaceous and older strata are deformed by SW-verging faults and folds. In the MSZ, post-Triassic, potentially Andean age thrust faults have variable orientations that are not parallel to Andean trends, and often reutilized pre-Andean structural fabrics.

The middle Paleozoic phase is expressed as NW-trending, shallowly plunging fold axes in Silurian-Devonian shales and older units in the MSZ (Figure 2, 8a). Isolated exposures of Ordovician-Devonian rocks in the CABB limit interpretation of the distribution and magnitude of middle Paleozoic deformation outside the MSZ. However, these regions were likely affected (Laubacher, 1978), as evidenced by unconformities at sample location 4 and stratigraphic column B. Cross-cutting relationships suggest deformation occurred during Late Devonian to Early Carboniferous time (Dalmayrac et al., 1980; Laubacher and Mégard, 1985). Folding is generally harmonic, with moderately dipping limbs and ~1 km wavelength. Silurian-Devonian units that preserve this

deformation are observed in localized erosional windows beneath an angular unconformity attributed to middle Paleozoic orogenesis.

Above this unconformity are Carboniferous to Permian shales, highly quartz cemented sandstones, and carbonates folded in broad synclines and anticlines at ~3–5 km wavelengths with 6–7 km long axial traces (Figure 8a, b). These folds are upright, with moderately dipping limbs, and rounded hinges, and do not affect overlying Mitu Group rocks. Cross-cutting relationships suggest a Late Permian age of deformation (Mégard et al., 1971, Laubacher, 1978; Marocco, 1978). Fold axes trend NNW-SSE, NE-SW and N-S. Folds are < 20 km long and often terminate into other non-parallel folds. Some folds change orientation along strike, either from NNW-SSE to ENE-WSW or from NE-SW to NNW-SSE. Previous interpretations of these folds have been unclear, but suggest refolding as a mode of curved fold creation (Laubacher, 1978). However, other curved folds from the Andes have been interpreted as the result of complex basement interaction (Jimenez et al., 2012), although the short strike length of these folds are difficult to reconcile with deep basement deformation. Another potential explanation of these folds could invoke deformation over an extremely weak detachment, likely Devonian shales.

A N-trending thrust fault juxtaposes W-dipping hanging-wall flat Carboniferous sandstones and shales on E-dipping footwall ramp Permian carbonates (Figure 8b). Mitu Group rocks dip ~20°E and rest in angular unconformity on the broad, variably oriented folds (Figure 8a, c) and the N-trending thrust fault. NNW-trending fold axes and the main thrust fault are attributed to late Paleozoic deformation prior to Mitu deposition (Figure 8b). Therefore, accumulation of the Mitu Group occurred after fold-thrust deformation, sealing late Paleozoic deformation. Our new geochronologic constraints from Mitu Group samples (Figure 4) paired with Late Permian biostratigraphic constraints on Copacabana Formation deposition (Grader et al., 2008) bracket the late Paleozoic deformation between ~275 and 233 Ma.

Within the MSZ, Andean age deformation is accommodated differently than in other parts of the Eastern Cordillera. A NNW-trending, ENE-dipping thrust fault juxtaposes E-dipping hanging-wall flat Carboniferous rocks on generally E-dipping footwall flat Mitu Group rocks (Figure 8b, d). This thrust fault cuts the Mitu Group and demonstrates that post-Triassic (potentially Andean) shortening reactivated structures

parallel to the previous late Paleozoic tectonic grain. Thrust timing remains poorly constrained. A SW-verging thrust fault is parallel to Andean trends (Figure 8a). The hanging wall is defined by an overturned anticline with NE-dipping axial plane developed in Carboniferous strata. This fault folds a NE-trending late Paleozoic anticline. In the footwall, Mitu Group and Cretaceous rocks are folded into an upright NW-trending syncline (Figure 8a, 8e, 9) suggesting Andean deformation.

The complete record of middle and late Paleozoic shortening, Triassic Mitu extension, and Andean shortening is present only to the NE of the San Anton boundary (Figure 2). This major pre-Andean normal fault played a key role in preserving pre-Andean deformation and compartmentalized styles of Andean deformation. Pre-Andean normal faults controlled deposition of the Mitu Group in the MSZ and contributed to structural compartmentalization by separating contrasting thin-skinned deformation in the CABB with oblique and potentially thick-skinned deformation in the MSZ. These field-based observations inform the creation of a cross section across the San Anton boundary that illustrates how reactivation of the San Anton fault linked basement deformation to the shallow detachment in cover strata of the SW verging CABB (Figure 10). Oblique late Paleozoic deformation of Carboniferous–Permian rocks is well constrained in the MSZ. However, limited exposures of Ordovician–Devonian rocks in the MSZ and CABB preclude accurate representation of regional structural trends associated with middle or late Paleozoic orogenic events. We depict deformation of Ordovician–Devonian units related only to Andean structures, but note that these Paleozoic units have experienced multiple phases of pre-Andean deformation.

6. DISCUSSION

6.1. Triassic Mitu basin

Spatial and temporal constraints on Triassic accumulation of the Mitu Group provide evidence for individual basins with localized provenance signals that evolved through time. Although the exact boundaries of these extensional subbasins remain poorly constrained, they dictated accumulation within a larger graben system defined by major normal faults at the MSZ margins. Identification of multiple disconnected basins is

575 based on several key provenance observations, summarized here. The easternmost
576 locality 1 shows an upsection disappearance of a cratonic zircon age population sourced
577 from Rondonia/San Ignacio blocks to the east and a corresponding upsection increase in
578 Triassic zircons. The westernmost locality 5 is the only Mitu sample that contains
579 significant Carboniferous ages and lacks the Triassic and cratonic zircon populations.

580 We consider the presence of two unique zircon populations (1300–1500 Ma, and
581 300–350 Ma) restricted to specific localities, and the lack of Triassic and cratonic zircons
582 at locality 5, to reflect a collection of subbasins separated by topographic barriers that
583 inhibited mixing of sediment sources. If the high proportion of 1300–1500 Ma zircon age
584 population at the base of locality 1 was sourced from orthogneiss rocks situated N or NW
585 of our study, or the Precordillera terrane of Argentina situated to the S, sediment transport
586 from north or south would likely result in this population being observed in correlative
587 stratigraphy along strike. However, this population is not observed in Mitu samples near
588 Cusco (Reitsma, 2012), between the northern orthogneiss source and our study area. We
589 suggest that this zircon age population was delivered by west-directed sediment transport
590 from the Rondonia/San Ignacio cratonic block in Bolivia. The upsection disappearance of
591 cratonic zircons at locality 1 may represent reorganization of sediment routing pathways,
592 or development of a significant topographic barrier that effectively blocked cratonic
593 contributions.

594 The accompanying upsection increase in Triassic zircons may reflect sediment
595 sourced from volcanic centers that become more felsic as extension progressed (Panca et
596 al., 2011). Carboniferous zircons restricted to locality 5 were likely transported to the SW
597 along the rift axis, from the Cusco region. The lack of Triassic ages at locality 5 may be
598 attributed to dilution of the distant Eastern Cordillera source of synrift volcanic products.
599 Alternatively, this absence at locality 5 and the absence of Carboniferous zircons at other
600 localities further supports the presence of topographic boundaries that prevented drainage
601 integration across subbasins. These regionally variable provenance signatures are
602 consistent with rapid changes in lithology and thickness of the Mitu Group and support
603 the interpretation of multiple basin depocenters during extension (Newell et al., 1953;
604 Laubacher, 1978; Dalmayrac et al., 1980; Kontak et al., 1990; Sempere et al., 2002;
605 Carlotto et al., 2010; Reitsma et al., 2010; Reitsma, 2012).

Although often assigned a broadly Late Permian to Triassic age, new zircon U-Pb geochronologic constraints demonstrate Triassic deposition of the Mitu Group, from ~242 to 233 Ma, similar to Mitu deposits ~700 km along strike (Reitsma, 2012). These constraints suggest initial Triassic extension was regionally synchronous. In contrast, Sempere et al., (2002) interpreted Eastern Cordillera plutons that become progressively younger from Cusco along strike to the SE as evidence for a southward propagating rift. Reconciling along-strike patterns in synrift plutonism and the surface expression of rifting (basin formation, upper crustal fault motion) may have implications for discriminating among backarc extension, continental rifting, or orogenic collapse as the principal driver of basin genesis (Noble et al., 1978; Dalmayrac et al., 1980; Dewey, 1988; Rosas et al., 2007; Ramos, 2009; Reitsma, 2012; McGroder et al., 2015).

6.2. Selective reactivation of pre-Andean faults

Stratigraphic, structural, and subcrop relationships across the Eastern Cordillera identify pre-Andean faults and folds. In this study we present examples of normal faults that were not reactivated, normal faults that influenced subsequent thrust faults, and inherited shortening-related structures that guided Andean deformation.

The outcrop-scale minor normal fault in the Eastern Cordillera (Figure 5) is an example of a non-reactivated normal fault of probable Triassic age. Typical normal-fault inversion geometries such as footwall shortcuts, hanging-wall anticlines, or buttressing (e.g., Williams et al., 1989; Cooper et al., 1989) are not observed. The NNW fault strike is subparallel to Andean structural trends. We suggest that the key factor inhibiting reactivation was a steep fault dip, rather than strike orientation, consistent with others suggesting normal fault inversion is more likely for faults dipping $< 45^\circ$ (Sibson 1985, 1990).

Surrounding geologic relationships suggest the outcrop-scale normal fault is synthetic to a larger normal fault that accommodated Triassic Mitu accumulation and preserved ~4 km of Carboniferous to Permian strata in its hanging wall. The Cordillera de Carabaya fault is a SW-verging thrust fault that places Carboniferous strata and Triassic plutonic rocks on Permian Copacabana Formation and Triassic Mitu Group

rocks. This major Andean fault dips in the opposite direction and cuts the proposed normal fault. Although pure normal-fault inversion did not occur, the original fault may have helped localize the new thrust fault of different orientation (Figure 11), similar to fault interactions proposed for Alpine fold-thrust structures (e.g., Welbon and Butler, 1992; Butler et al., 2006).

The San Anton boundary separating the CABB and MSZ is interpreted as a tilted unconformity related to thrust reactivation of a major normal fault at depth. We interpret the MSZ block to be elevated above the CABB by thrust reactivation of the Triassic San Anton normal fault at depth (fault 3*) (Figure 6, 10). In our kinematic reconstruction (Figure 10), thick-skinned reactivation of the normal fault transfers slip upsection to a shallow basal Cretaceous detachment and guides thin-skinned deformation in the CABB. We suggest that the inherited orientation of the NE-dipping San Anton fault contributed to SW vergence of the CABB.

Previous workers have suggested the Ayaviri fault, a key feature in the Cenozoic evolution of the Altiplano, represents a major pre-Andean structure linked to a long-lived, potentially crustal-scale boundary persisting for >300 km along strike (Sempere et al., 1990; Carlier et al., 2005; Carlotto, 2013; Perez and Horton, 2014). We interpret stratigraphic and subcrop relationships across this thrust as evidence for a pre-Andean history, potentially as a normal fault. However, the corresponding footwall cutoff for the Ayaviri fault hanging wall is offset laterally by significant thrust translation. Linking the geochemical variations observed at the surface across the Ayaviri fault (Carlier et al., 2005) with a proposed major subsurface crustal boundary will need to be compared with kinematic reconstructions. Although the exact nature of its pre-Andean history remains unclear, Oligocene thrust motion of the Ayaviri fault accommodated deformation propagation from the CABB toward the Altiplano. Crustal loading from cover and basement thrust deformation induced rapid subsidence in the Ayaviri basin, controlled sediment provenance, and was key to partitioning the Andean hinterland basin (Perez and Horton, 2014; Horton et al., 2015).

Inherited geometries from middle and late Paleozoic deformation events likely influenced the style and trends of Andean deformation. The MSZ preserves evidence of Triassic and Cretaceous rocks deformed by Andean thrust faults that are parallel to the

orientation of folds attributed to middle Paleozoic deformation (Figure 8a). Other Andean structures, including the broad folds, non-coaxial structural patterns, and deeper structural levels observed in the MSZ, suggest potentially thick-skinned thrusts reutilized oblique structural orientations inherited from late Paleozoic orogenesis (Figure 8b).

6.3. The role of inherited structures on structural compartmentalization and Andean deformation

The complete record of middle and late Paleozoic deformation is preserved only beneath Mitu Group deposits in the MSZ. Middle Paleozoic deformation likely affected much of the Eastern Cordillera and potentially Altiplano, as evidenced by schistosity, foliation, and angular unconformities among lower Paleozoic rocks (Laubacher, 1978; Dalmayrac et al., 1980). Beyond the MSZ, structures in lower Paleozoic rocks from the CABB and much of the Eastern Cordillera are parallel to Andean NW-SE trends. Similarly, fold-thrust structures deforming Cretaceous rocks in the CABB are aligned with Andean trends. In some cases, contacts between Paleozoic, Triassic, and Cretaceous rocks are disconformable, suggesting regional variations in middle Paleozoic deformation magnitudes. Observed bedding orientations from the Altiplano/CABB and frontal Eastern Cordillera/Subandean Zone show that Cenozoic rocks deformed during Andean deformation follow the orientation of Paleozoic deformation trends (Figure 12). Likewise, the variety of pre-Andean structural trends is mimicked by bedding orientations of rocks deformed during Andean deformation. This suggests a component of Andean deformation across the Eastern Cordillera follows structural geometries inherited from Paleozoic and Triassic deformation.

Selectively preservation of oblique late Paleozoic deformation trends in Carboniferous–Permian rocks of the MSZ can be linked to Triassic extension accommodated by the San Anton normal fault and a proposed conjugate normal fault at depth below the Cordillera de Carabaya (fault 4*; Figure 1, 6). Erosional removal of Carboniferous–Permian rocks from the adjacent uplifted footwalls inhibits definition of the distribution of late Paleozoic deformation outside the MSZ, and is less constrained than middle Paleozoic deformation. We interpret the MSZ as a graben that accommodated the majority of Mitu deposition during Triassic extension, preserving

upper Paleozoic rocks and late Paleozoic deformation trends. Later Andean deformation of Triassic Mitu and Cretaceous rocks in the MSZ followed a mix of complex structural orientations inherited from orogenic and rifting events. Some instances are parallel to middle Paleozoic shortening and Triassic extensional trends (Figure 8a). Others reutilize late Paleozoic deformation orientations and remain oblique to typical Andean orientations (Figure 8b).

The paucity of observed oblique late Paleozoic deformation trends outside the MSZ (Figure 12) raises two potential scenarios. First, late Paleozoic deformation affected regions outside the MSZ but did not influence later Andean deformation, or was not oblique to middle Paleozoic and Andean trends and thus not easily detected. Alternatively, oblique late Paleozoic deformation was restricted to the MSZ. Although the San Anton boundary is interpreted here as a Triassic normal fault, if it contributed to compartmentalization of late Paleozoic shortening, it may represent a complex and long-lived orogenic deformation front reutilized during successive deformation events that ultimately guided Andean kinematics in the CABB and MSZ.

7. CONCLUSIONS

1) Detrital zircon U-Pb geochronological results from the Triassic Mitu Group in the Eastern Cordillera of southern Peru reveal compartmentalized extensional basins with unique provenance characteristics. The stratigraphically lowest sample from the easternmost Mitu exposure preserves a significant Rondonia/San Ignacio cratonic source that disappears upsection, likely due to the development of an eastern rift shoulder that blocked cratonic contributions. The westernmost sample has the highest proportion of Carboniferous zircons, sourced from the Cusco region, suggesting axial SW-directed transport. The Triassic extensional system is interpreted as a mosaic of subbasins with discrete provenance signatures, as opposed to a single large extensional basin with homogenous provenance.

2) Zircon U-Pb ages constrain the timing of Mitu Group deposition between ~242 and 233 Ma. This Triassic deposition is consistent with other constraints from up to ~700 km NW along strike, suggesting nearly synchronous Triassic initiation of Mitu accumulation. This finding refines previous chronostratigraphy, which assumed a Late Permian to Triassic age of the Mitu Group. Regionally synchronous initiation of the Mitu extensional system contrasts with interpretations of a SW-propagating rift.

3) The locations of pre-Andean normal faults active during Triassic extension and Mitu accumulation are identified based on outcrop, regional subcrop relationships, and thickness variations. Results show a mix of low- and high-stratigraphic throw associated with identified normal faults. These normal faults parallel the modern Andean structural grain, although fault dip varies between roughly 40° and 60°.

4) Reactivation and interaction with pre-Andean normal faults exerted considerable influence on Andean deformation. The MSZ represents a Triassic graben bounded by two major pre-Andean normal faults. The San Anton normal fault was inverted and guided deformation in the Central Andean Backthrust belt. The NE fault nucleated a backthrust of opposite dip that established the boundary between the Cordillera de Carabaya and the MSZ.

5) Selective reactivation of pre-Andean structures attributed to Paleozoic deformation and Triassic extension contributed to compartmentalization of deformation style across the Eastern Cordillera. Multiple phases of non-coaxial deformation are preserved in the MSZ and result in Cenozoic deformation that reutilized oblique structural trends. In other parts of the Eastern Cordillera, Cenozoic fold-thrust deformation is parallel to Andean trends.

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Figure Captions

Figure 1. Geologic map after INGEMMET (1999). Note typical NW-SE Andean structural grain, and the variation of structural orientations in the Macusani Structural Zone. Major map boundaries are indicated by 1*, 2*, etc. Stratigraphic columns marked by letters. Samples from the Mitu Group are marked by numbers. Map boundaries, stratigraphic columns, and samples are synthesized in figure 6.

Figure 2. Simplified geologic map of Macusani Structure Zone (location, stratigraphic key from Figure 1). Typical NW-SE Andean deformation trends in the Central Andean Backthrust Belt are contrasted by NNW-SSE and NE-SW deformation trends in the Macusani Structural Zone. Note various ages of structures determined from field relationships. Andean thrust faults in the MSZ often reactivate late Paleozoic deformation trends, have longer fault traces than non-reactivated faults, and are oblique to typical Andean trends observed outside the MSZ. Heavy dashed boxes outline figure locations.

Figure 3. Facies variation within the Triassic Mitu Group from Macusani Structural Zone outcrops. A) Cross stratified coarse sandstone. B) Volcaniclastic. C) Breccia with volcaniclastic matrix and clasts of extrusive rock. D) Poorly sorted matrix supported quartzite pebble conglomerate. E) Poorly sorted clast supported limestone cobble conglomerate. F) Bedding surface of planar and ripple cross-laminated siltstone.

Figure 4. U-Pb zircon geochronology from Mitu Group samples (locations shown in Figure 1). X axis for all probability density plots from 0–3000 Ma. Y axis for all plots is number of zircons. Sample name, relative stratigraphic position, zircons analyzed, and youngest age populations listed next to sample plots. A) Samples from tuffs interbedded in Mitu Group deposits. B) Detrital zircon samples. Age populations highlighted by colored rectangles.

Figure 5. Outcrop example of Triassic normal fault. Fault strikes $\sim 110^\circ$ and dips $\sim 40^\circ$ SSW. Location from Figure 1. Schematic location in Figure 6. Note

802 folded Permian Copacabana Fm. rocks are unconformably overlain by
803 unfolded, tilted Triassic Mitu Group rocks.

804 Figure 6. A) Subcrop and thickness variations across the Eastern Cordillera. Column
805 locations, sample locations from Figure 1. Distance between columns
806 measured perpendicular to NW-SE Andean strike. Subcrop, thickness, and
807 bedding orientation relationships from our field and map data. Schematic
808 bedding shown in Ordovician–Devonian rocks based on our observations.
809 Stations A, B, and G preserve angular unconformities between Paleozoic
810 rocks and overlying Cretaceous rocks. Column C preserves a disconformity
811 between Ordovician–Devonian rocks and overlying Mitu Group rocks.
812 Columns D, E, and F preserve folds and angular unconformities between
813 Ordovician–Devonian rocks and overlying Mitu Group rocks. B) Schematic
814 proposed pre-Cenozoic structural framework. Subcrop and thickness
815 variations recreated with normal faults 1*–4*. Fault numbers from Figure 1.

Figure 7. Two potential scenarios to explain subcrop and thickness changes between columns C and D (Figure 1, 6) across the San Anton boundary (3*). Field relationships suggest Cretaceous rocks were deposited across the San Anton fault (3*). At column C, thin Mitu Group rocks underlie Cretaceous rocks, and are deposited on early Paleozoic rocks. At column D, approximately 10 km NE, Mitu Group rocks are up to 3 km thick and deposited on up to 4 km of upper Paleozoic rocks. Although the Paleozoic strata were deformed by middle and late Paleozoic orogenies before deposition of the Mitu group, they are depicted as undeformed here for simplicity. Scenario A achieves lateral changes in stratigraphic thickness and subcrop with a tapered stratigraphic wedge (3–10°) over distances of 18–61 km. To juxtapose thick Carboniferous to Triassic deposits within 10 km of the San Anton boundary would require 8–51 km shortening (44–84%). Total shortening for the region is 123 km (41%) (Gotberg et al., 2009). We consider scenario B more likely. Rapid lateral change in stratigraphic thickness is accommodated by a normal fault (fault 3*, San Anton fault). Reactivation of this fault elevates Carboniferous to Triassic rocks in the MSZ above Cretaceous rocks in the CABB.

Figure 8. Structural relationships documenting pre-Andean deformation preserved in the Macusani Structural Zone. Map and photo locations from Figure 1. A) map showing relationship between "Early Hercynian" folds (NW-SE trends in Silurian/Devonian rocks below unconformity, late Paleozoic folds (NNW-SSE syncline in Carboniferous to Permian rocks), and Andean thrust fault, fold deforming Triassic and Cretaceous rocks, parallel to middle Paleozoic deformation trends. B) Late Paleozoic thrust fault (N-S) and folds (NNW-SSE) sealed by Triassic Mitu Group deposits. Andean thrust fault (NNW-SSE) follows late Paleozoic deformation trends and thrust Carboniferous rocks on Mitu Group rocks. C) Outcrop photo of unconformity between folded Permian Copacabana Formations rocks below and Triassic Mitu Group rocks above. Location for B. D) Outcrop photograph of Andean thrust fault (from map B) parallel to late Paleozoic deformation trends. Carboniferous Ambo Group hanging-wall flat thrust on Triassic Mitu Group footwall flat. E) Outcrop photograph of Andean thrust fault (from map B) parallel to middle Paleozoic and typical Andean trends. Thrust fault juxtaposes Carboniferous Ambo Group rocks on Triassic Mitu Group and Cretaceous Muni and Huancane Formation rocks.

Figure 9. Schematic block diagram from Figure 8A based on bedding orientations measured in the field. Colors same as figure 8. Note middle Paleozoic folds beneath angular unconformity and Carboniferous Ambo Group. Late Paleozoic deformation trends are oblique to middle Paleozoic deformation trends. Paleozoic rocks are thrust over Triassic and Cretaceous rocks by Andean thrust fault that is parallel to the middle Paleozoic deformation trends.

Figure 10. Part of a line balanced cross section across the Central Andean Backthrust belt and Macusani Structural Zone. Note preservation of Carboniferous to Permian rocks in MSZ deformed by late Paleozoic orogenic event. Correlative strata are not preserved across the San Anton boundary in the CABB. In restored section, late Paleozoic deformation is schematically depicted. Carboniferous Tarma Group and Permian Copacabana Formation are not drawn on the restored section. However, the thickness between base of the Triassic Mitu Group and top of the Silurian/Devonian Annanea Formation is consistent with the deformed section. Late Paleozoic deformation is constrained within this stratigraphic interval. We do not attempt to show structures associated with middle Paleozoic deformation, although it is observed in field relationships in Ordovician–Devonian rocks. At depth, San Anton boundary (depositional contact between Cretaceous and Triassic rocks) is depicted as a Triassic normal fault reactivated as an Andean thrust fault. Thrust fault inversion feeds slip upsection to a shallow detachment at the base of Cretaceous rocks in the CABB.

Figure 11. Modified after Butler et al., (2006). Schematic illustration of how normal faults may nucleate new thrust faults that dip opposite direction. Note schematic of location of normal fault from Figure 6 that is synthetic to major normal fault. We suggest fault 4* (figure 1, 6) is analogous to this schematic normal fault. The Cordillera de Carabaya backthrust is analogous to the younger thrust fault that cuts the older normal fault.

Figure 12. Map of dominant structural trends highlight structural blocks within the Eastern Cordillera. Same map area and major fault/subcrop boundaries as Figure 1. Stereonets show poles to bedding from three regions: Altiplano/Central Andean Backthrust Belt (gray background), Macusani Structural Zone (white background), and frontal Eastern Cordillera/Subandean Zone (gray background). Black poles represent bedding measurements from Triassic to Cenozoic rocks. Gray poles represent bedding measurements from Paleozoic rocks. Thin lines are fold

and fault traces. Note consistent NW-SE Andean trends in Altiplano/Central Andean Backthrust Belt and Subandean Zone. Corresponding stereonet show girdles and are consistent with NW-trending Andean fault and fold strikes. In contrast, the Macusani Structural Zone, where multiple structural trends are oblique to typical Andean orientation, the corresponding stereonet lacks a well-defined girdle. This reflects the influence of oblique deformation trends of Carboniferous to Permian rocks during the late Paleozoic orogeny preserved in the Macusani Structural Zone. In this structural block, Andean age structures follow these pre-existing oblique orientations.

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Figure 3

B) Proposed pre-Cenozoic structural framework

Figure 7

**From
subcrop
relationships:**

Scenario A:

~18-61 km

To place 3500 m thick U. Pz - Triassic Mitu Gp. next to 300 m thick Mitu Gp.:

- 3° dip: minimum ~ 51 km lateral translation (84% shortening)

- 10° dip: minimum ~8 km lateral translation (44% shortening)

Preferred shortening estimate in southern Peru: 123 km (41%; Gotberg et al., 2009)

Scenario B:

CABB

*Lower Pz. encompasses Ordovician through Devonian units

*Upper Pz. encompasses Permian Copacabana Fm., Carboniferous Tarma Fm., and Ambo Gp.

NE **Figure 9**

Figure 12