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Mesozoic history of the southeastern Tibetan plateau: Sediment provenance, paleoclimate, and surface elevation history

Fei Shang

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MESOZOIC HISTORY OF THE SOUTHEASTERN TIBETAN PLATEAU: SEDIMENT PROVENANCE, PALEOCLIMATE, AND SURFACE ELEVATION HISTORY

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**Dissertation submitted
to the Eberly College of Arts and Sciences at West
Virginia University**

in partial fulfillment of the requirements for the degree of

**Doctor of Philosophy in
Geology**

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**Keywords: Southeastern Tibet; Qamdo Basin; Sediment Provenance;
Paleoclimate; mid-Cretaceous Climate Sensitivity; Paleoaltimetry**

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ABSTRACT

Mesozoic History of the Southeastern Tibetan Plateau: Sediment Provenance, Paleoclimate, and Surface Elevation History

Fei Shang

The region of southeastern Tibetan Plateau is one of the most geologically complicated areas on earth. Numerous continental fragments, volcanic arcs, and oceanic crust were swept together as a collage from Paleozoic to Recent to form the basic tectonic framework of the region. The history of fragmentation, accretion, and deformation is controlled by tectonic activities of several plate boundaries that have been activated and reactivated at numerous times. The multi-phase tectonic history of eastern Tibetan Plateau makes tectonic history reconstruction very difficult due to the tectonic overprints of younger tectonisms (i.e., Cenozoic Indo-Asia collision and Indochina lateral extrusion). The lack of geological knowledge about the pre-Cenozoic southeastern Tibetan Plateau (e.g., pre-collisional upper crustal configuration and initial age of thickening of Tibetan crust) hinders the development of a comprehensive mechanical model that can account for the rise of the vast plateau.

This dissertation presents new results of a multi-disciplinary investigation of the linkages between erosion, uplift and regional climate change associated with the Mesozoic accretionary tectonism in southeastern Tibet. We studied sedimentary record preserved in a robust and strategically-located geologic archive —the Qamdo basin system. Since tectonics can greatly influence the geomorphic and stratigraphic development of sedimentary basins, studying evolution of the Qamdo basin could provide useful insights into the tectonic information of the adjacent areas.

The following questions are addressed in the three chapters of this thesis:

- What are the sediment sources for the Mesozoic Qamdo basin?
- Is there a link between provenance of Qamdo sediments and Mesozoic accretionary tectonism of southeastern Tibet, such as the closure of the Paleo-Tethyan and Meso-Tethyan oceans?
- What impact did Cretaceous global climate change have on local climate of southeastern Tibet?
- What impact did Cretaceous orogenesis have on local climate of southeastern Tibet?
- Can we estimate the amount of Cretaceous surface uplift in southeastern Tibet by distinguishing between effects caused by global climate and local orogenesis?

This dissertation contains three separate chapters focus on sediment provenance of the Qamdo basin, Cretaceous paleoclimate and paleoaltimetry of southeastern Tibet.

Chapter 1 provides an overview of the sedimentology, stratigraphic characterization and sandstone provenance of the Mesozoic Qamdo basin. Study revealed that the Qamdo basin was sourced primarily from erosion of Triassic turbidite of the Yidun Group, metasedimentary rocks of the Zhongza massif, igneous rocks of the Jomda-Weixi arc and

Bangong arc, and recycled Upper Triassic Qamdo strata. Upper Triassic sediments of the Qamdo basin were sourced primarily from the inverted Yidun Group as a result of the closure of the Paleo-Tethyan Ganzi-Litang ocean and development of a widespread fold-thrust belt in northeastern Tibet during Late Triassic time. Three distinct depocenters developed in the Qamdo basin by the beginning of Early Jurassic time, as indicated by a spatial diversification of detrital zircon age signatures. The northern depocenter received detritus derived primarily from the recycled Upper Triassic Qamdo strata; whereas, the central depocenter has a mixed provenance of Zhongza massif metasedimentary rocks and recycled Upper Triassic Qamdo strata. The prominent ~230-245 Ma peak in the southern depocenter signifies sediment influx from the Jomda-Weixi arc. A lack of Jurassic age zircons in the Lower Jurassic sedimentary rocks is interpreted to reflect a temporal geographical segregation between the Bangong arc and the Qamdo basin due to the opening of a short-lived Early Jurassic oceanic backarc basin, which was probably related to slab rollback during Meso-Tethys subduction. The Middle-Upper Jurassic strata were sourced from recycled Upper Triassic Qamdo strata and exhumed Bangong arc plutons that intruded the Tongka basement. Abundant ~165-190 Ma zircon grains in the Middle-Upper Jurassic strata suggest exhumation and erosion of the Bangong arc related to the convergence between Lhasa and Qiangtang blocks during Mid-Late Jurassic time. Cretaceous strata were mostly sourced from the Jomda-Weixi arc combined with sediment input from internally-recycled Qamdo strata. The cause of the reactivated exhumation of the Jomda-Weixi arc was probably related to the northward underthrusting of the Lhasa terrane beneath the Qiangtang block during Early Cretaceous time. This chapter was submitted to GSA bulletin and is currently in review.

Chapter 2 employs carbonate clumped-isotope thermometry in combination with paleosol CO₂ barometry to reconstruct mid-Cretaceous terrestrial climate conditions of southeastern Tibet. The mid-Cretaceous (Cenomanian-Turonian) “super-greenhouse” world of ~100 million years ago was characterized by high levels of atmospheric CO₂ driven by intense volcanic activity, and characterized by high sea surface temperatures (SSTs) and a small equator-to-pole temperature gradient. Although it is commonly accepted that the extreme warmth of the mid-Cretaceous was driven by increased atmospheric *p*CO₂, large discrepancies exist between climate models and proxy data, as models often fail to simulate SSTs unless atmospheric *p*CO₂ reaches implausibly high levels (> 3,500 ppm) when using a standard range of model climate sensitivities. We reconciled this discrepancy using new temperature and *p*CO₂ reconstructions that indicate climate sensitivity may have been underestimated for greenhouse conditions. We use carbonate clumped isotope thermometry to constrain soil temperatures, meteoric water δ₁₈O ratios, and *p*CO₂ levels for the Cenomanian-Turonian interval (~101-90 Ma) from paleosol carbonates in southeastern Tibet, which was located at ~15° N during this time. Minimum summer temperatures were ~30 °C during the Cenomanian, with a warming of ~10 °C associated with major increases in *p*CO₂ and soil water δ₁₈O values during the Cenomanian-Turonian transition. The reconstructed terrestrial temperature trends mirror coeval tropical SST proxies. Our most conservative estimate of climate sensitivity during the Cenomanian-Turonian interval yields a value of ~10 °C per CO₂ doubling under high CO₂ “super-greenhouse” conditions, significantly higher than modern day values of ~1.5-4.5 °C suggested by the Intergovernmental Panel on Climate Change (IPCC). This chapter is under review by coauthors and will be submitted to Earth and Planetary Science Letters.

Chapter 3 reports results from clumped isotope, strontium isotope and sedimentological analyses of the entire succession of Late Cretaceous (Cenomanian-Santonian) strata preserved in the Qamdo basin, accompanied by results of Late

Cretaceous climate modeling. This chapter provides evidence of an episode of surface uplift in southeastern Tibet during Late Cretaceous (Coniacian-Santonian) time. We examined two stratigraphic sections in the Cenomanian-Santonian strata of the Qamdo basin and find that petrography, burial history and elemental composition of paleosol carbonates are consistent with preservation of original isotopic signatures and minimal influence from solid-state reordering. Average soil temperatures (i.e., Δ_{47-T}) decreased from 44 ± 1.5 °C (95% CI) in the Turonian to 38 ± 3.5 °C (95% CI) by the Santonian time, coincident with decrease of average $\delta^{18}\text{O}_{\text{water}}$ values from $-6.3 \pm 0.5\text{‰}$ (95% CI) to $-7.3 \pm 0.5\text{‰}$ (95% CI). Climate model outputs and tropical sea surface temperature (SST) records both show that global climate during the Turonian-Santonian interval were relatively warm and stable, without evidence of significant global cooling. Together, the observed changes in Δ_{47-T} and $\delta^{18}\text{O}_{\text{water}}$ values exhibit a pattern that is consistent with surface elevation changes during Late Cretaceous time. Assuming temperature and isotopic lapse rates of ~ 6 °C/km and $\sim 1\text{‰}/\text{km}$, respectively, under the extreme warm Late Cretaceous climate conditions, the observed Δ_{47-T} and $\delta^{18}\text{O}_{\text{water}}$ shifts correspond to a positive elevation change of ~ 1 km in southeastern Tibet during Late Cretaceous time. This inference is supported by evidence of strontium isotope ratios, structure, and sedimentology. Strontium isotope ratios of pedogenic carbonate increase up-section from 0.7106 to 0.7112, indicating enhanced weathering of continental source areas. Meanwhile, the development of growth strata and occurrence of thick-bedded conglomerate layers within the Late Cretaceous strata all implying active tectonics during deposition. Together, out data suggest that a moderate topography with elevation >1.5 km was probably established by Late Cretaceous time in SE Tibet. This chapter is under review by coauthors and will be submitted to *Earth and Planetary Science Letters* shortly.

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Chapter I

Detrital zircon geochronology from the Mesozoic Qamdo (Changdu) basin southeastern Tibet: Implications for the Paleo- and Meso-Tethys evolution

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ABSTRACT

The Qamdo basin, which is bounded by the Paleo-Tethyan Jinsha suture to the northeast and the Meso-Tethyan Bangong-Nujiang suture zone to the southwest, contains up to 8,000 m of well-preserved Late Triassic-Cretaceous marine-continental sediments that record the evolution of southeastern Tibet prior to the development of the Cenozoic Tibetan plateau. Provenance analysis of Qamdo basin strata using sandstone petrology and detrital zircon geochronology techniques reveals the Mesozoic accretionary tectonism of southeastern Tibet. Quartzose Upper Triassic sandstone samples yield detrital zircon age signatures that form five primary populations at 230-330, 400-500, 700-1200, 1750-1900, and 2400-2600 Ma, consistent with derivation primarily from inversion and erosion of Triassic Yidun Group turbidites in response to the

closure of the Paleo-Tethyan Ganzi-Litang ocean. Lower Jurassic Lower Chaya Group samples are all composed of quartz-rich sandstone but show spatial provenance variation in detrital zircon ages. In the northern Qamdo basin, Lower Chaya Group has the same five primary detrital zircon age populations as the Upper Triassic sandstone; however, in the central Qamdo basin, Lower Jurassic sandstone detrital zircon age signatures are dominated by middle Paleoproterozoic ages (1840-1870 Ma) sourced from the Zhongza massif. The detrital zircon age signature of a Lower Jurassic sandstone sample from southern Qamdo basin is characterized by a prominent Triassic age peak (230-245 Ma), indicating sediment influx from the nearby Jomda-Weixi arc system. The lack of Jurassic-aged zircons in any of the Lower Jurassic strata may reflect a geographical separation between the coeval Bangong arc (~170-190 Ma) and the Qamdo basin, possibly due to development of a back-arc oceanic basin.

Middle-Upper Jurassic Upper Chaya Group sandstone samples in northern and southern parts of the basin display largely similar quartzose framework grain composition and variable proportions of five main age populations: 230-330, 400-500, 700-1200, 1750-1900, and 2400-2600 Ma. In contrast, coeval sandstone samples from the central Qamdo basin show much greater abundance of volcanic lithic fragments and are characterized by a detrital zircon age signature with a prominent peak at ~165-190 Ma. We interpret that Middle-Upper Jurassic strata in the northern and southern Qamdo basin were mainly sourced from recycled Upper Triassic Qamdo strata, whereas Middle-Upper Jurassic deposits in the central Qamdo basin were sourced by erosion of the Bangong arc as a result of exhumation in response to Meso-Tethyan ocean closure and subsequent collision of the Qiangtang and Lhasa blocks. Sandstone samples of the Cretaceous Xiangdui Formation show a volcanic lithic-rich composition and a large detrital zircon age peak at 230-245 Ma, along with a secondary population of 1700-1950 Ma, suggesting

Cretaceous strata were sourced primarily from the exhumed Jomda-Weixi arc to the southeast of the Qamdo basin, combined with locally recycled Upper Triassic Qamdo sediments. The Cretaceous exhumation of the Jomda-Weixi arc may be associated with the northward underthrusting of the Lhasa block beneath the Qiangtang block during Early Cretaceous time. These data depict a prolonged history of collisional tectonism that shaped the tectonic development of southeastern Tibetan plateau lithosphere prior to the initiation of early Cenozoic Himalayan orogenesis.

INTRODUCTION

Southeastern Tibet is a collage of continental fragments, volcanic arcs, and oceanic crustal rocks that were accreted successively to Eurasia during Paleozoic to Cenozoic time (Fig. 1) (e.g., Dewey et al., 1988; Yin and Harrison, 2000; Burchfield and Chen, 2013). Fragmentation, accretion, and deformation resulted from tectonism along several plate boundaries that have been reactivated numerous times (Yin and Harrison, 2000; Guynn et al., 2006; Kapp et al., 2007; Roger et al., 2008; Burchfield and Chen, 2013). The multi-phase tectonic history and overprinting by recent Cenozoic tectonic events makes reconstructing the pre-Himalayan, Mesozoic geologic history of southeastern Tibet very difficult (Tapponnier et al., 2001; Royden et al., 2008; Burchfield and Chen, 2013). Consequently, aspects of the evolution of Tibetan lithosphere prior to Himalayan orogenesis remain subject to debate. One of the outstanding issues is the role and timing of pre-Himalayan crustal shortening and its relationship to overall history of crustal thickening in the region, which is important in accurate geodynamic modeling of Tibetan plateau formation (e.g., Molnar and Tapponnier, 1975; Barazangi and Ni, 1982; Powell, 1986; Dewey et al., 1988; Holt and Wallace, 1990; Molnar et al., 1993; Wu et al.,

1993; Nelson and INDEPTH, 1996; Chang, 1997; Owens and Zandt, 1997; Yin and Harrison, 2000; Kind et al., 2002; Raphine et al. 2003; Kumar et al., 2006; Royden et al., 2008).

Although the modern Tibetan plateau is a topographic feature produced mainly by tectonism associated with the Cenozoic India-Asia collision (e.g., Dewey and Burke, 1973; Powell and Conaghan, 1973; Dewey et al., 1988), many studies across Tibet have documented crustal shortening, crustal thickening, and rapid cooling histories that predate the India-Asia collision (Murphy et al., 1997; Kapp et al., 2007; Studnicki-Gizbert et al., 2008). For example, Murphy et al. (1997) suggested that the Lhasa block underwent only ~40% of north-south crustal shortening during Cenozoic time and that ~60% of its current 70 km crustal thickness was developed during Early Cretaceous time, likely due to the collision of the Lhasa block with the Qiangtang block. Similarly, the Gonjo basin in eastern Tibet rests upon a pre-Cenozoic angular unconformity, which suggests at least some, if not most, of the shortening of Mesozoic rocks in the basin must have occurred prior to the India-Asia collision (Studnicki-Gizbert et al., 2008). In the Nima basin of central Tibet, the upper crust underwent >47% shortening during Cretaceous to mid-Cenozoic time with more than half of this shortening predating the Indo-Asia collision (Kapp et al., 2007). All these studies indicate that a deeper understanding of pre-Cenozoic lithospheric configuration of Tibetan plateau is required in order to fully evaluate the mechanisms that may responsible for the Cenozoic development of the plateau (e.g., lower crustal flow and lithospheric delamination models) (e.g., Clark et al., 2000; Currie et al., 2005).

Estimation of pre-Cenozoic crustal shortening is often difficult in southeastern Tibet due to overprinting by younger, co-axial Cenozoic deformation and/or along-strike displacement of genetically related structures by strike-slip faulting (e.g., Studnicki-Gizbert et al., 2008).

However, sedimentary basins adjacent to major convergent tectonic boundaries preserve a record

of unroofing from adjacent mountains from which geodynamic processes associated with convergent tectonism can be inferred (Ingersoll et al., 2003). Although continuous, pre-Cenozoic sedimentary records are generally rare in southeastern Tibet, the Qamdo basin contains a relatively complete succession of Late Triassic-Cretaceous sedimentary rock >8,000 m thick that is exposed across an area of ~12,000 km² (Fig. 1). The Qamdo basin rests on top of Precambrian and Paleozoic metamorphic basement of the eastern Qiangtang block, bound by the Jinsha suture to the northeast and the Bangong-Nujiang suture zone to the southwest (BNSZ) (Du et al., 1997; Pan et al., 2004). The basin strata preserve an important record of Mesozoic hinterland erosion and basin evolution in southeastern Tibet that was closely related to convergent deformation along both the Jinsha and Bangong-Nujiang sutures bordering the eastern Qiangtang block. Here, we investigate the provenance of Qamdo basin strata, which provides valuable constraints on the Mesozoic tectonic evolution and configuration of southeastern Tibetan lithosphere that was later modified by Cenozoic Himalayan orogenesis to become the modern Tibetan plateau.

We report framework grain compositions for 40 sandstone samples and >1,200 detrital zircon U-Pb ages from 14 sandstone samples of Upper Triassic-Cretaceous Qamdo basin strata in a first attempt to identify sediment source areas for this major basin system and assess the Mesozoic tectonic evolution of southeastern Tibetan lithosphere. In addition to study of Qamdo basin sedimentary rocks, we include 257 detrital zircon ages from 3 Yidun Group samples to further characterize the zircon U-Pb age signatures of potential source rocks. Our results suggest that during Mesozoic time, sediment provenance of the Qamdo basin evolved from uniform, regional, far-field sources toward the east to more local sourcing by erosion of uplifted basin-bounding ranges as paleogeography evolved in response to convergent tectonism associated with the closure of the Paleo- and Meso-Tethyan oceans.

GEOLOGICAL BACKGROUND

Tectonic Setting of the Qamdo Basin

Qamdo basin strata rest unconformably on top of Precambrian and Lower Triassic-Paleozoic metamorphosed sedimentary rocks of the eastern Qiangtang block (EQB) (Chang, 1997; Pan et al., 2004; Burchfield and Chen, 2013). The Jinsha suture delineates the northeastern EQB margin and records the closure of the Paleo-Tethyan ocean due to convergence of the EQB with the Yidun terrane (YDT) during Early-Middle Triassic time (Mo et al., 1993; Wang et al., 2000; Reid et al., 2007). To the southwest, the EQB abuts the Lhasa block (LHB) along the Bangong-Nujiang suture zone (BNSZ), which represents the closure of the Meso-Tethyan ocean between the EQB and LHB during Late Jurassic-Early Cretaceous time (e.g., Guynn et al., 2006, 2012; Kapp et al., 2007; Zhang et al., 2012). The Jitang arc (JTA) lies to the northeast of the BNSZ within the EQB, whereas the Bangong arc (BGA), and the LHB lie to the southwest of the BNSZ (Fig. 1B). These tectonic elements of southeastern Tibet are further described below.

Yidun Terrane (YDT)

The Yidun terrane (YDT) is comprised of the eastern Yidun terrane (EYDT) and Zhongza massif (ZM), which are separated by the north-south trending Xiangcheng-Geza fault (Fig. 1B; BGMRSP, 1991). The EYDT contains widespread, deformed Middle-Upper Triassic turbidite deposits of the Yidun Group (BGMRSP, 1991; Hou, 1993; Wang et al., 2014). These turbidites are intruded by Upper Triassic (216-219 Ma) granitic-dioritic intrusions commonly referred to as the ‘Yidun’ or ‘Litang-Batang’ arc (e.g., Pan et al., 2004). Triassic magmatic rocks of the EYDT are thought to have formed either as the result of the westward subduction of the Ganzi-Litang Ocean (e.g., Hou, 1993; Reid et al., 2007; Weislogel, 2008; Yang et al., 2014) or as

the results of northward subduction along the Jinsha suture (Roger et al., 2008; Burchfield and Chen, 2013); whereas the Cretaceous (95-105 Ma) granitic intrusions are possibly related to subduction of the Meso-Tethyan oceanic crust underneath Asia and corresponding regional extension (Reid et al., 2007). The western YDT contains the ZM, which is intruded by plutons of the Permo-Triassic Jomda-Weixi arc (JWA).

Zhongza Massif (ZM) and Jomda-Weixi Arc (JWA)

The Zhongza massif (ZM) is thought to have originated as a micro-continent separated from the western South China block during Late Permian time based on similar Paleozoic fossil assemblages and sedimentary sequences between the South China craton and the ZM (BGMRSF, 1991; Hou, 1993; Song et al., 2004). The Neoproterozoic granitic gneiss basement and metamorphosed Paleozoic cover strata of the ZM were intruded by Permo-Triassic granitic plutons of the Jomda-Weixi arc, which yields U-Pb zircon ages of 230-245 Ma (Wang et al., 2010; Yang et al., 2011; Yang et al., 2014). The JWA may have been created by the westward subduction of Jinsha oceanic lithosphere (Mo et al., 1993; Wang et al., 2000; Xiao et al., 2008), although eastward subduction has also been proposed (Chang, 1997; Reid et al., 2005). The eastern side of the ZM is structurally overlapped by tightly folded Middle to Upper Triassic (~245-220 Ma) clastic turbidites of the Yidun Group (Pan et al., 2004; Wang et al., 2013).

Lhasa Block (LHB)

The Lhasa block (LHB) consists mainly of Paleozoic and Mesozoic marine strata and arc-type volcanic rocks that are intruded by Jurassic through Early Tertiary granitoids of the Gangdese batholiths (Yin and Harrison, 2000). The block was accreted to the Eurasian plate along the Bangong-Nujiang suture zone (BNSZ), as the Meso-Tethyan oceanic basin closed

during Late Jurassic to Early Cretaceous time (e.g., Dewey et al., 1988; Yin and Harrison, 2000; Kapp et al., 2007; Zhang et al., 2012).

Bangong Arc (BGA)

The Bangong arc (BGA) is a proposed Jurassic continental magmatic arc system that parallels the length of the Bangong-Nujiang suture zone (BNSZ). BGA magmatism was fueled by the northward subduction of the Meso-Tethyan oceanic lithosphere (e.g., Girardeau et al., 1984; Leeder et al., 1988; Guynn et al., 2006; Zhang et al., 2012), however, much of the BGA was later subducted beneath the Qiangtang continental lithosphere as a result of the closure of the Meso-Tethyan ocean, leaving only scattered ophiolite belts and fragments exposed along the BNSZ (Fig. 1B; Pan et al., 2004; Guynn et al., 2006, 2012; Zhang et al., 2012). West of the Qamdo basin, the BGA intruded Cambrian-Neoproterozoic gneissic/granitic crystalline basement rocks of the Tongka basement (Fig. 1B) (Pan et al., 2004; He et al., 2011).

Stratigraphy and Sedimentology of the Qamdo Basin

From Late Triassic to Cretaceous time, Qamdo basin paleoenvironments evolved from deep-marine, to marginal-marine, to lacustrine, then to alluvial, indicating a continuous shallowing-upward trend over ~100 million years (Fig. 2). The ~700 m thick Upper Triassic (Carnian) Jiapila Formation is the basal unit of the Qamdo succession (Fig. 2). It is exposed only along the western Qamdo basin and disconformably underlain by Lower Triassic and Permian metaquartzite of the EQB (Fig. 2; Du et al., 1997). The Jiapila Formation is comprised of marine black shale with fine-grained turbidite sandstone interbeds (Fig. DR1A) interpreted to deposited in a deep-marine setting (Du et al., 1997; Peng et al., 2005). The overlying bioclastic Upper Triassic (early Norian) Bolila Formation is a 300 m thick carbonate platform deposit. The uppermost Triassic unit is the Upper Triassic (late Norian to Rhaetian) Bagang Group, which is

comprised of ~1,500 m thick siliciclastic strata and interlayered coal beds interpreted to have been deposited in a marginal marine/deltaic setting (Fig. DR1B; Peng et al., 2005).

Conformably overlying the Triassic marine-influenced deposits is the ~4,300 m thick Jurassic Chaya Supergroup which can be further divided into the Lower Chaya (Lower Jurassic) and Upper Chaya (Middle-Upper Jurassic) groups (Fig.2; Du et al., 1997). The ~800 m thick Lower Chaya Group is characterized by abundant well-developed red and purple paleosols interbedded with red lacustrine sandstone, siltstone, and mudstone (Fig. DR1C), representing a major transition of depositional environment from marginal marine to terrestrial. The Upper Chaya Group consists of ~3,500 m thick interbedded red lacustrine fine-grained sandstone, siltstone, and mudstone, with occasional paleosol horizons (Fig. DR1D; Du et al., 1997). The sandstone/mudstone cyclothems occur from centimeter- to meter-scale recording periodic lake level fluctuations (Fig. DR1D, E), which may reflect both regional climatic- and tectonic-driven changes in sediment and runoff flux. Mud-cracks are common throughout the Upper Chaya Group (Fig. DR1F), indicating deposition occurred in a shallow lake setting with periodic desiccation. Erosional sedimentary structures (e.g., rip-up mud clasts; Fig. DR1G) are common, indicating strong currents alternated with periods of desiccations.

The Cretaceous Xiangdui Formation is only present in the southern Qamdo basin (Fig. 1; Du et al., 1997; Pan et al., 2004). It contains ~1,200 m thick alluvial deposits composed of clast-supported cobble conglomerate, cross-stratified pebbly red sandstone, and thick-bedded carbonate nodule-bearing paleosol layers (Fig. DR1H, I, J). Abundant clast-supported conglomerate and calcisol layers suggest that deposition occurred proximal to source areas in an arid to semi-arid alluvial fan setting with periodic flooding.

METHODOLOGY

Field Procedures

A total of 14 sandstone samples were collected from the Upper Triassic Jiapila Formation (N=1) and Bagang Group (N=3), the Jurassic Lower (N=4) and Upper (N=4) Chaya groups, and the Cretaceous Xiangdui Formation (N=2) of the Qamdo basin for paired sandstone petrology and detrital zircon geochronology provenance study. An additional 26 sandstone samples were collected from the same localities for thin-section petrographic study (Fig. 1B). Geologic maps produced by the Chengdu Institute of Geology & Mineral Resources and China Geological Survey (Pan et al., 2004) were used to determine the locations and stratigraphic ages of the samples. Three additional sedimentary rock samples were collected from the Upper Triassic Yidun Group turbidites of the western YDT, a potential source area, with the purpose of characterizing its detrital zircon age signature. Paleocurrent orientation data include measurement of 31 cross-bedding structures from the Cretaceous Xiangdui Formation in the south Qamdo basin (Fig. 1B). Unfortunately, statistically robust (e.g., $n > 10$ for a signal site; Potter and Pettijohn, 2012) paleocurrent measurements cannot be obtained from Upper Triassic-Jurassic strata, as paleocurrent indicators (i.e., ripple marks and cross-bedding) are rare.

Sandstone Petrography

Standard petrographic thin-sections were made from 40 medium- to fine-grained Upper Triassic to Cretaceous sandstone samples. The thin-sections were stained for potassium feldspar and calcic plagioclase and a total of 400 point-counts were obtained using a step size of 0.5 mm for each thin-section to achieve a maximum error of 5% at 2σ (Van der Plas and Tobi, 1965). Gazzi-Dickenson method was used to minimize grain-size effects (Dickinson, 1970; Ingersoll et al., 1984). The resulting framework grain estimations from point-counting were then plotted on

provenance discrimination ternary plots after Dickinson et al. (1985) to aid in determining the tectonic setting of the sediment source area. Modal sandstone compositions are listed in Table DR1* and graphically displayed in Figure 3.

Detrital Zircon Geochronology

Sandstone samples were disaggregated and the dense mineral fraction was separated using hydraulic separation and heavy liquid separation; magnetic fraction was removed using a Frantz magnetic separator. Detrital zircon grains were mounted in epoxy without handpicking to prevent sampling bias. Mounts were then polished and imaged using reflected light and cathodoluminescence (CL) to reveal internal heterogeneity of each zircon grain so that analysis mixing two different age domains could be avoided. U-Pb isotopic composition of zircon grains was determined by laser-ablation multi-collector inductively coupled plasma-mass spectrometer (LA-MC-ICP-MS) in the Arizona LaserChron Center at the University of Arizona (Gehrels et al., 2008). The analytical method is described in Gehrels et al. (2008) and briefly summarized here. The mass spectrometer was equipped with a Photon Machines Analyte G2 excimer laser using a spot diameter of 30 or 35 microns. The ablated material is carried in helium into the plasma source of a Nu HR ICPMS, and U, Th, and Pb isotopes can be measured simultaneously. All measurements are made using Faraday detectors for ^{238}U , ^{232}Th , ^{208}Pb - ^{206}Pb , and discrete dynode ion counters for ^{204}Pb and ^{202}Hg (Gehrels et al., 2008). Common-Pb was corrected using the Hg-corrected ^{204}Pb and assuming an initial Pb composition from Stacey and Kramers (1975). Inter-element fractionation of Pb/U was corrected using a standard Sri Lankan zircon with known age of 563.5 ± 3.2 Ma, and another zircon standard R-33 was used as a check standard. Zircon ages that are younger than ca. 1000 Ma are based on common Pb-corrected $^{206}\text{Pb}/^{238}\text{U}$

ratios, whereas ages older than ca. 1000 Ma are based on common Pb–corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ratios.

The resulting interpreted ages were plotted on Pb/U concordia diagrams (Fig. DR2) and relative age-probability diagrams (Fig. 5-9) using Isoplot 3.60 software (Ludwig, 2008).

Uncertainties on individual analyses are given at the 1- σ level and plotted at the 2- σ level.

Detrital zircon ages for each sample are displayed using both probability density plots (PDP), and kernel density estimation plots (KDE), which plots the ages as a set of Gaussian distributions but does not explicitly take into account the analytical uncertainties (Vermeesch, 2012). The KDE produces statistically more robust age distribution curves especially when data quantity and/or analytical precision are high (Vermeesch, 2012; DensityPlotter, version 6.3). Because DensityPlotter 6.3 cannot accept data sets with >1,000 ages due to software limitations, therefore, composite distributions composed of >1,000 ages are plotted only as PDPs. Results from all zircon isotopic measurements are available in Table DR2 in the Data Repository.

The detrital zircon analysis yielded 1275 valid ages from 14 Qamdo sandstone samples and additional 257 valid ages from 3 Yidun Group turbidite sandstone samples (Table DR2). Typically, 100-110 grains were analyzed per sample with 10-20 analyses rejected due to high discordance, high common-lead, or error, leaving ~90 acceptable ages for each sample. According to Vermeesch (2004), 90 ages allows an 82% confidence level that no fraction $\geq 5\%$, or a 100% confidence level that no fraction $\geq 10\%$ was not represented in the final results.

Statistical Methods

Detrital zircon age populations were compared using the Kolmogorov-Smirnov (K-S) test. The K-S test is an unbiased, statistically sound approach for comparing two distributions and determining if there is a significant difference between two distributions (Guynn and

Gehrels, 2010). The P-values returned by the K-S test corresponds to the confidence level that age distributions of two samples are sourced from the same parent population. The K-S test was conducted using an Excel macro program (provided by George Gehrels, 2010). However, the P-values are not only affected by similarity between age distributions, but also factors such as: (1) number of analyses in each sample; (2) proportions of similar ages in samples; and (3) analytical uncertainty (Guynn and Gehrels, 2010). Thus, P-values are used as a secondary indicator for determining if two distributions share a sediment source area or not. Comparison of age distributions between the Qamdo sandstone samples and potential source areas are primarily conducted by determining the presence or absence of prominent age populations along with their relative proportions in different samples.

We conducted 3 K-S tests with results shown in Tables 2, 3, and 4. Results in Table 2 were generated from comparing all Qamdo detrital zircon samples with potential sources. In contrast, results in Table 3 were from comparison of only Mesozoic grains (<252 Ma) in Qamdo samples and potential Mesozoic magmatic arc sources compiled from previous studies (Guynn et al., 2006; Reid et al., 2007; Wang et al., 2010; Yang et al., 2011; Peng et al., 2014; Tao et al., 2014; Wang et al., 2014). Most individual Qamdo detrital zircon samples contain <20 Mesozoic grains, which is insufficient to yield statistically valid K-S analysis results (Guynn and Gehrels, 2010). Thus, the comparisons were conducted between the Mesozoic grains combined from the Upper Triassic, Lower Jurassic, Middle-Upper Jurassic, and Cretaceous samples from the Qamdo basin and the zircon ages from potential magmatic arc sources. K-S test results in Table 4 are from comparison of pre-Mesozoic ages (>252 Ma) in Qamdo samples and pre-Mesozoic (>252 Ma) ages of potential sources compiled from published literature (Reid et al., 2007; Gehrels et al., 2011), as well as from this study.

RESULTS

Sandstone Composition

Upper Triassic Jiapila Formation and Bagang Group

Three Jiapila Formation and eight Bagang Group sandstone samples have very similar detrital framework compositions (Table DR1) with an average quartz abundance of 91% (Table DR1). Monocrystalline quartz is the most abundant component (mean abundance =86%), while polycrystalline quartz is minor (mean abundance =5%) (Fig. 4A, Table DR1). Feldspar grains comprise an average of 5% of the samples and are dominated by plagioclase with minor albitization. Lithic fragments are rare and mainly sedimentary in origin (Table DR1).

Lower Jurassic Lower Chaya Group

Nine Lower Chaya Group samples were point-counted. Monocrystalline quartz is the dominant framework grain type in the Lower Chaya samples (mean abundance =67%). The samples contain 16-24% polycrystalline quartz, and ~2% feldspar dominated by plagioclase (Fig. 4B; Table DR1). Lithic component in the samples tend to vary spatially. The six samples analyzed from the northern and central Qamdo basin contain $\leq 10\%$ lithic fragments, including sedimentary lithic fragments (~6%) and minor volcanic lithic grains (<2%) (Fig. 4C). In contrast, the three samples analyzed from the southern Qamdo basin yield a greater abundance of lithic fragments ($\leq 20\%$), and show much higher content of volcanic lithic fragments (~16%) and lower abundances of sedimentary lithic grains (< 2%) (Figure. 3F; Table DR1).

Middle-Upper Jurassic Upper Chaya Group

In the twelve Upper Chaya Group samples, monocrystalline quartz is the most abundant component (mean abundance =73%; Fig. 4D). Polycrystalline quartz abundances average 11%

(Table DR1). Feldspar makes up $\leq 3\%$ of the total framework grains and lithic grains comprise another $\sim 16\%$. Volcanic and sedimentary lithic fragments are the dominant lithic grain types, with average abundances of $\sim 11\%$ and $\sim 5\%$, respectively (Fig. 4E; Table DR1). The abundance of volcanic lithic grains varies significantly, with the highest volcanic lithic content ($\sim 21\%$) found in samples from the central Qamdo basin. In contrast, volcanic lithic fragments form just $\sim 4\%$ of northern and southern Qamdo basin samples (Fig. 3I).

Cretaceous Xiangdui Formation

Nine Xiangdui Formation sandstone samples were point-counted. Monocrystalline and polycrystalline quartz abundances average 76% and 11%, respectively (Table DR1). Lithic fragments comprise an average abundance of 15%, and most lithic grains ($>80\%$) are volcanic fragments (Fig. 3L; Table DR1). The volcanic lithic fragments display both vitriclastic (Fig. 4F) and microlitic textures, suggesting derivation from both pyroclastic and plutonic igneous rocks. Sedimentary lithic fragments are rare (average $\sim 3\%$). Similar to Upper Triassic-Jurassic samples, feldspars are rare in Cretaceous sandstone samples (Table DR1).

Detrital Zircon Age Analysis Results

Upper Triassic Jiapila Formation and Bagang Group

One Jiapila (i.e., UT-2) and three Bagang (i.e., UT-1, UT-2, UT-4) detrital zircon samples were analyzed: two from north Qamdo (UT-1 and UT-2), and two from south Qamdo (UT-3 and UT-4) (Fig. 1B), which altogether yielded 349 valid ages. Over 80% of the ages form five primary age populations, including 15%-27% that form an Early Carboniferous-Middle Triassic (330-230 Ma) population, 8%-17% that form a Late Cambrian-Early Devonian (500-400 Ma) population, 6%-17% that form a Meso- to Neoproterozoic (1200-700 Ma) population, and

18%-51% and 8%-14% that form a late Paleoproterozoic (1950-1700 Ma) and a late Archean-early Paleoproterozoic (2600-2400 Ma) populations, respectively (Fig. 6A; Table 1). Sample UT-1 differed from the others in that it also contains a population of ca. 220 Ma zircon grains that comprises 7% of the sample. K-S comparison between Upper Triassic samples results in P-values range from 0 to 0.61. There are 4 out of 12 sample pairs yielded P-values >0.05 and 8 out of 12 sample pairs yielded P-values >0.01 (Table 2).

Lower Jurassic Lower Chaya Group

Four Lower Chaya Group samples were analyzed: sample LJ-1 from north Qamdo, samples LJ-2 and LJ-4 from central Qamdo, and sample LJ-3 from south Qamdo (Fig. 1B). Together, the samples yielded 364 valid ages. In north Qamdo, sample LJ-1 has a detrital zircon signature similar to that of the Upper Triassic Qamdo samples with the same five major age populations comprising 81% of total ages, including 17% that form an Early Carboniferous-Middle Triassic (330-230 Ma) population, 20% that form a Late Cambrian-Early Devonian (500-400 Ma) population, 15% that form a Meso- to Neoproterozoic (1200-700 Ma) population, and 25% and 5% that form a late Paleoproterozoic (1950-1700 Ma) and a late Archean-early Paleoproterozoic (2600-2400 Ma) populations, respectively (Fig. 6B; Table 1). In addition, 7% of ages comprise a minor young peak at ca. 220 Ma (Table 1).

In contrast, samples LJ-2 and LJ-4 from central Qamdo display largely similar detrital zircon age populations as LJ-1, but with significantly more middle Paleoproterozoic (1870-1840 Ma) ages (Fig. 6B). Samples LJ-2 and LJ-4 contains 22% and 15% of zircon ages between 1870-1840 Ma, respectively, whereas sample LJ-1 only contains 9% ages of 1870-1840 Ma (Fig. 6B; Table 1).

Sample from south Qamdo (i.e., LJ-3) is distinguishable from the north and central Qamdo samples by a prominent Middle Triassic age peak at 245-230 Ma, which comprises 20% of total zircon grain ages (Fig. 6B; Table 1). In comparison, other coeval sandstone samples from north and central Qamdo (i.e., LJ-1, LJ-2 and LJ-4) contain <6% of 245-230 Ma zircon grains (Table 1). Other primary age populations of sample LJ-3 include a Meso- to Neoproterozoic (1200-700 Ma) population (19% of total grains), and a middle Paleoproterozoic (1870-1840 Ma) population (14% of total grains, Fig. 6B; Table 1).

K-S comparison between Lower Jurassic samples yield P-values range from 0 to 0.46, while half of the sample pairs display P-values <0.05 (Table 2). The relatively low P-values suggest sediment provenance varied spatially during deposition of Lower Jurassic strata.

Middle-Upper Jurassic Upper Chaya Group

Four Upper Chaya Group detrital zircon samples were analyzed: MUJ-1 and MUJ-4 from north Qamdo; MUJ-2 from central Qamdo; and MUJ-3 from south Qamdo (Fig. 1B). Together, the samples yielded 363 valid ages. Samples from north (i.e., MUJ-1, MUJ-4) and south (i.e., MUJ-3) Qamdo display largely similar detrital zircon age distributions with 75%-84% of the ages forming the same five primary age populations observed in most older (i.e., Upper Triassic-Lower Jurassic) Qamdo samples, including 14%-27% that form an Early Carboniferous-Middle Triassic (330-230 Ma) population, 14%-16% that form a Late Cambrian-Early Devonian (500-400 Ma) population, 10%-25% that form a Meso- to Neoproterozoic (1200-700 Ma) population, and 19%-23% and 3%-10% that form a late Paleoproterozoic (1950-1700 Ma) and a late Archean-early Paleoproterozoic (2600-2400 Ma) populations, respectively, along with few Early-Middle Jurassic grains at ~165, ~185, and ~188 Ma (Fig. 7A; Table 1). In contrast, 33% of zircon grains in sample MUJ-2 from central Qamdo yielded ages between 165-190 Ma, forming

a prominent Early-Middle Jurassic peak (Fig. 7A). In MUJ-2, subordinate age populations include Early Carboniferous-Middle Triassic (330-230 Ma; 32% of total grains), Meso- to Neoproterozoic (1200-700 Ma; 5% of total grains), late Paleoproterozoic (1950-1700 Ma; 7% of total grains), and late Archean-early Paleoproterozoic (2600-2400 Ma; 5% of total grains), as were observed in other Qamdo samples (Table 1). K-S comparison between MUJ-1, MUJ-3 and MUJ-4 results in P-values range from 0.12 to 0.74, whereas the comparison of MUJ-2 with other Middle-Upper Jurassic samples all yields P-values =0 (Table 2). These results indicate that MUJ-1, MUJ-3 and MUJ-4 have a similar provenance, whereas MUJ-2 incorporated detritus derived from additional source/sources.

Cretaceous Xiangdui Formation

Two Cretaceous detrital zircon samples, K-1 and K-2, yielded 199 valid ages. Samples were collected from the southern Qamdo basin where Cretaceous strata are present (Fig. 1B). The detrital zircon signature of sample K-1 is characterized by a bimodal age distribution with a prominent Middle Triassic (245-230 Ma) population that comprises 33% of the total ages, along with a late Paleoproterozoic (1950-1700 Ma) population that makes up of 30% of total ages (Fig. 7B; Table 1). In contrast, for sample K-2, the 245-230 and 1950-1700 Ma age populations are much less abundant, and both populations comprise only 10% of total ages. However, the Early Carboniferous-Early Triassic (330-250 Ma; 18% of total ages), Late Cambrian-Early Devonian (500-400 Ma; 14% of total ages), and Meso- to Neoproterozoic (1200-700 Ma; 21% of total ages) ages are more abundant in K-2 compared to K-1 (Fig. 7B; Table 1). K-S comparison between K-1 and K-2 results in P-values =0.07, suggesting a similar provenance between the two samples (Table 2).

Summary of K-S Comparison of Qamdo Zircon Age Distributions

K-S test results are shown in Tables 2, 3, and 4. Results of K-S comparison suggest that most Qamdo samples have similar provenance. P-values that result from comparing all Qamdo detrital zircon samples to each other suggest that 10 out of 14 samples (i.e., UT-1, 2, 3; LJ-1, 3, 4; MUJ-1, 3, 4; K-2) have age distributions similar to each other, with P-values >0.05 for 71% of sample pairs (Table 2). Only 4 samples (i.e., UT-4; LJ-2; MUJ-2; K-1) show relatively low similarity compared to the other samples, as reflected by only 11% of sample pairs with P-values >0.05 (Table 2). This is largely because these 4 samples contain more abundant Mesozoic grains.

Comparison of Mesozoic (<252 Ma) ages among Qamdo composite samples show high P-values (>0.05) for the Upper Triassic, Lower Jurassic, and Cretaceous composite sample pairs (Table 3), indicating the Mesozoic zircon grains in Qamdo basin were likely sourced from similar source/sources. Note that comparisons among the Middle-Upper Jurassic composite sample and the Upper Triassic/Lower Jurassic/Cretaceous composite samples all returned low P-values ($P < 0.05$). We attribute these low values to the presence of 165-190 Ma population in the Middle-Upper Jurassic composite sample that is missing in other composite samples (Fig. 8). K-S comparison of pre-Mesozoic (>252 Ma) ages for all Qamdo samples returned P-values >0.05 for 53% of the sample pairs (Table 4), suggesting the pre-Mesozoic grains have a general similarity in provenance, but with locally varied sediment influx.

COMPARISON WITH POTENTIAL SOURCES OF QAMDO BASIN FILL

Given regional Mesozoic paleogeographic reconstructions and estimated strike-slip movements along Qamdo basin boundary faults (e.g., Jinsha and Bangong-Nujiang sutures) (Yin and Harrison, 2000; Metcalfe, 2009; Gehrels et al., 2011), possible source terranes for the Qamdo basin include the neighboring Zhongza massif (ZM), Jomda-Weixi arc (JWA), eastern

Yidun terrane (EYDT), Qiangtang block (QTB), Lhasa block (LHB), Jitang arc (JTA), and Bangong arc (BGA) (Fig. 1B). In addition, older sedimentary rock of the Qamdo basin can be a potential source for its younger strata because structural estimations suggest shortening of the Qamdo basin beginning by Early Jurassic time (Du et al., 1997; Pan et al., 2004). Thus, folding of pre-Early Jurassic strata generated uplifted fold-thrust belts (e.g., Zhongza thrust system, Xu et al., 2015) that could have served as a sediment source. Published U-Pb zircon ages from detrital, metamorphic, and magmatic rocks of the previously mentioned source terranes, along with detrital zircon ages of the Yidun Group from this study are used to generate reference zircon age signatures of the eight potential sources. Zircon age probability density plots and kernel density plots of all the source areas are shown and compared in Figures 5, 8, and 9.

Eastern Yidun Terrane (EYDT) Signature (Sources to the N-NE of Qamdo Basin)

Basement rocks of the eastern Yidun terrane (EYDT) are not widely exposed, and are limited to the Neoproterozoic-Paleozoic rocks of the Zhongza massif (ZM) exposed along the Jinsha suture (discussed separately). The detrital zircon signature of EYDT is characterized using 257 detrital zircon ages from three Yidun Group turbidite sandstone samples dated in this study (i.e., 13UTSG-1, 13UTSG-2, and 13UTSG-427; Fig. 1B). Detrital zircon age signature of the Yidun Group is characterized by five major age populations of Early Carboniferous-Middle Triassic (330-230 Ma; 18% of total grains), Late Cambrian-Early Devonian (500-400 Ma; 14% of total grains), Meso- to Neoproterozoic (1200-700 Ma; 16% of total grains), late Paleoproterozoic (1950-1700 Ma; 23% of total grains), and late Archean-early Paleoproterozoic (2600-2400 Ma; 12% of total grains; Fig. 5D; Table DR2). U-Pb analyses of widespread granodiorite intrusions within the EYDT yielded mainly Late Triassic (225-215 Ma) and Cretaceous (105-95 Ma) ages (Fig. 8A; Reid et al., 2007; Peng et al., 2014).

Evidence for EYDT Provenance

The Middle-Upper Triassic (Ladinian-Carnian) Yidun Group is a potential source for the Upper Triassic (Carnian-Rhaetian) and younger strata of Qamdo basin. Comparison of detrital zircon age signatures between the Yidun Group and Upper Triassic Qamdo composites (Figs. 5D-E) suggests they contain almost identical primary age populations (i.e., 230-330; 400-500; 700-1200; 1700-1950, 2400-2600) with similar age percentages. K-S test comparisons of Upper Triassic detrital zircon ages with the Yidun Group composite yield P-values >0.05 (range from 0.11 to 0.72) for 2 out of 4 sample pairs, indicating relatively high probability of a shared parent source of detrital zircons (Figs. 5D-E; Table 2). K-S analyses also reveal high correlations between the Yidun Group samples and the Jurassic and Cretaceous Qamdo samples, with P-values >0.05 (range from 0.18 to 0.87) for 56% of total sample pairs (Table 2).

Zhongza Massif (ZM) and Jomda-Weixi Arc (JWA) (Sources to the N-NE of Qamdo Basin)

Although zircons from basement rocks of the Zhongza massif (ZM) have not been directly dated, U-Pb dating of a pegmatite sample from an Early-Middle Triassic granitic pluton that intruded the ZM revealed inherited zircon cores that are dominantly middle Paleoproterozoic in age (1870-1840 Ma; 32% of total grains), along with a few ~500, ~750, and ~1700 Ma cores (Figs. 5F; Reid et al., 2007). These grains were interpreted to be scavenged from Paleozoic metasedimentary rocks of the ZM that experienced partial melting during the Paleo-Tethyan Jinsha ocean subduction (Reid et al., 2007). The age signature of the inherited ZM zircon grains displays South China block affinity, as the ~1800 Ma zircon grains are commonly reported as inherited cores of the South China block (Qiu et al., 2000; Zhang et al., 2006). The 750 Ma grains are similar in age to extensive Neoproterozoic magmatic rocks within the western South China block (Li et al., 2003). Neoproterozoic basement rocks of the ZM were intruded by the

Permian-Triassic plutons of the JWA, which contains abundant ~230-245 Ma zircons (Wang et al., 2010; Yang et al., 2011; Yang et al., 2014).

Evidence for ZM and JWA Provenance

Although Paleoproterozoic (~1700-1950 Ma) zircon ages are common in most Qamdo samples, abundant 1840-1870 Ma zircon is often coincident with abundant Middle Triassic ages at 230-245 Ma (e.g., in samples LJ-3 and K-1). Together these ages indicate sediment influx from the ZM and the JWA (Reid et al., 2007). K-S tests conducted between the ZM basement and the pre-Mesozoic ages of Qamdo samples returned P-values >0.05 (range from 0.07 to 0.49) for 9 out of 14 Qamdo samples (Table 4) and K-S comparisons between the composite Mesozoic ages of Qamdo samples and the JWA shows P-values >0.05 (range from 0.06 to 0.33) for 3 out of 4 sample pairs (Table 3), suggesting the ZM and JWA were important sources for Qamdo sediments.

Qiangtang Block (QTB) Signature (Local Sources of QTB)

The Qiangtang block (QTB) is composed of the south and north Qiangtang terranes separated by a high-pressure eclogite-blueschist metamorphic belt exposed in central QTB (e.g., Metcalfe, 1988; Zhang, 2001). Composite detrital zircon age plot for the north Qiangtang terrane (Fig. 5B) was compiled from 12 samples from Carboniferous-Triassic strata collected across the north Qiangtang terrane (Fig. 1A; Pullen et al., 2010; Gehrels et al., 2011). The south Qiangtang composite (Fig. 5C) was constructed using 21 Precambrian-Triassic samples collected from various locations of the north Qiangtang terrane (Pullen et al., 2010; Gehrels et al., 2011). Although in different proportions, the south and north Qiangtang terranes yield the same zircon age populations of 210-280, 500-650, 750-1000, 1800-2000, and 2400-2600 Ma (Figs. 5B-C; Pullen et al., 2010; Gehrels et al., 2011).

Evidence for Qiangtang Provenance

Although the Qamdo basin was developed on top of the eastern QTB (Du et al., 1997; Pan et al., 2004), a lack of 500-650 Ma ages in the Qamdo sedimentary rock (<3% of total grain counts; Figs. 6-7), which are prevalent in the Qiangtang terranes (Figs. 5B-C), indicates the QTB is not a major source for the Qamdo sedimentary rock. The relationship is also indicated by K-S comparisons between the Qiangtang terranes and Qamdo samples, which all yield P-values of zero (Tables 2; 4).

Recycled Upper Triassic Qamdo Strata (Local Sources of QTB)

As previously discussed, recycled Upper Triassic Qamdo strata could be a potential source for Jurassic-Cretaceous sediments of the Qamdo basin because all characteristic age populations found in Upper Triassic samples (i.e., 230-330, 400-500, 700-1200, 1700-1950, and 2400-2600 Ma) are present in Jurassic and Cretaceous Qamdo samples (Figs. 6B; 7), even in samples that have dominant volcanic arc provenance (e.g., LJ-3, MUJ-2, and K-1) based on zircon ages and sandstone compositions. Similarity between the Upper Triassic and the Jurassic-Cretaceous samples becomes more pronounced when the Mesozoic arc-related ages are removed from consideration. Comparison of all pre-Mesozoic (>252 Ma) detrital zircon ages from the Qamdo basin, it is clear that except for the abundant middle Paleoproterozoic (1870-1840Ma) grains derived from ZM basement rocks that are present in couple Lower Jurassic samples (i.e., LJ-2 and LJ-4), the Jurassic and Cretaceous samples have almost identical detrital zircon signatures as those of the Upper Triassic samples (Fig. 9). The similarity is also supported by K-S analyses comparing Jurassic and Cretaceous samples with the Upper Triassic composite, which show P-values >0.05 (range from 0.05 to 0.54) for 7 out of 10 sample pairs (Table 4). The

recycled origin of Jurassic-Cretaceous deposits is also supported by sandstone petrology data, which reveal QFL ratios indicative of recycled orogen provenance (Fig 3D-L).

However, since the Upper Triassic Yidun Group of the EYDT has almost identical detrital zircon age signatures as those of the Upper Triassic Qamdo strata based on both visual comparison and K-S tests (Figs. 5D-E; Table 2), it is possible the ‘recycled signature’ was related to sediment influx from the Yidun Group rather than recycling of the Upper Triassic Qamdo strata. However, geologic context of the Qamdo basin suggests that the Yidun Group was likely not a major source area for the Qamdo basin during Jurassic through middle Cretaceous time: (1) tectonic deformation of the EYDT likely ceased by ca. 190-200 Ma (Reid et al., 2005), with a subsequent episode of exhumation during Late Cretaceous (Reid et al., 2005; Roger et al., 2008; Jackson et al., 2014; Wang et al., 2014a); and (2) Significant shrinkage of basin size and transition from marine (Fig. DR1A-B) to arid and semi-arid terrestrial environments (Fig. DR1C-F) in the Qamdo basin by Early Jurassic time (Du et al., 1997; Pan et al., 2004) favors development of an internally-drained basin system (e.g., Carroll and Bohacs, 1999; Sobel et al., 2003; Garcia-Castellanos, 2006; Carroll et al., 2010), which therefore favors reworking of the proximal Upper Triassic strata compared to transport of sediments from distal sources in the EYDT. Continued uplift and exhumation of the Upper Triassic Qamdo strata may have been associated with development of the Zhongza thrust belt along the Jinsha suture in latest Triassic as a result of Paleo-Tethyan closure (Xu et al., 2015), and subsequent Jurassic-middle Cretaceous continental-continental collision between the Qiangtang and Lhasa blocks along the BNSZ (Zhang et al., 2012; Wang et al., 2014a; Liu et al., 2015). However, recent thermochronology study of the Yidun terrane reveals an episode of Late Jurassic-Early Cretaceous crustal exhumation that is interpreted to induced by flat-slab subduction of the Meso-Tethys ocean (Tian

et al., 2014), which indicates that the Yidun rocks may be uplifted during Late Jurassic-Early Cretaceous time to source sediment of the Qamdo basin. Therefore it is possible that the recycled Upper Triassic Qamdo strata and EYDT may have both been sources of Jurassic-Cretaceous Qamdo deposits.

Jitang Arc (JTA) Signature (Local Sources of QTB)

The Jitang arc (JTA) is located along the western side of the Qamdo basin (Fig. 1B) and formed by melts derived from a crustal source with Sr-Nd isotopic compositions similar to those of the Indian cratonic crust (Tao et al., 2014). U-Pb zircon age information available for this arc indicates post-collisional magmatism occurred at 219.1 ± 1.7 Ma (Tao et al., 2014).

Evidence for JTA Provenance

Despite close geographic proximity between the JTA and the Qamdo basin, the absence of ca. 220 Ma zircon grains in most Qamdo sedimentary rocks suggests JTA rocks were not a major sediment source for the basin (Table 1). Exceptions are samples UT-1, LJ-1, and K-2, which contain a minor peak at ~220 Ma (Table 1) and were all collected from areas adjacent to the JTA (Fig. 1B). Although Triassic magmatic arcs of the eastern Yidun terrane (EYDT) also contain a minor ca. 220 Ma population (Fig. 8A; Reid et al., 2007), the very localized occurrence of ca. 220 Ma zircon grains in the Qamdo basin and lack of age signature of Triassic plutons of the EYDT (i.e., 215-225 Ma) suggest these young grains were most likely sourced locally from the nearby JTA instead from the distant EYDT. The paucity of JTA aged zircon grains in Qamdo sediments indicates that the JTA may only have experienced limited magmatism and exhumation during Mesozoic time and detritus derived from the arc did not travel far into the Qamdo basin, or perhaps the Mesozoic drainage pattern did not favor the transportation of JTA detritus into the

Qamdo basin, and the detritus shed from the exhumed JTA was routed elsewhere (e.g., to the S-SW toward the Lhasa block)

Lhasa Block (LHB) Signature (Sources to the S-SW of Qamdo Basin)

Granitic basement of the LHB contains abundant Cambrian zircon grains dated at 480-520 Ma, along with a few inherited zircons with ages of ~613 and ~875 Ma (Leier et al., 2007; Pullen et al., 2010; Gehrels et al., 2011). Detrital zircons in the overlying Paleozoic and Mesozoic cover strata also yield mainly Cambrian ages with a peak at 511 Ma (Fig. 5A), probably derived from the underlying granite basement of the LHB (Leier et al., 2007; Pullen et al., 2010; Gehrels et al., 2011). Here, the composite detrital zircon age plot of LHB (Fig. 5A) was constructed from compiling the samples from a Paleozoic sequence and the overlying Jurassic-Lower Cretaceous strata from central LHB (Leier et al., 2007; Pullen et al., 2010; Gehrels et al., 2011).

Evidence for Lhasa Provenance

Cambrian zircons (~480-540 Ma) that are abundant in the LHB rocks are rare to absent in most Mesozoic Qamdo samples (Fig. 9). At most, the age population of 480-540 Ma forms <5% of Qamdo samples. K-S comparisons of Lhasa composite and Qamdo samples all returned P-values =0 (Tables 2; 4), suggesting that the LHB was probably not contributing sediments into the Qamdo basin during Mesozoic time. The interpretation is consistent with the Mesozoic paleogeographic reconstructions of Tibet (Metcalf, 2009; Gehrels et al., 2011), which suggest the LHB was separated from the Qiangtang block by Late Jurassic time. Subsequent closure of the Meso-Tethyan ocean and collision between the Lhasa and Qiangtang blocks probably formed a bivergent foreland thrust belt along the northern margin of the LHB (Kapp et al., 2003; Leier et

al., 2007). This thrust belt may have functioned as a geographic barrier, further preventing the LHB from shedding sediment into the Qamdo basin.

Bangong Arc (BGA) Signature (Sources to the S-SW of Qamdo Basin)

Magmatism of BGA appears to have developed diachronously along the Bangong-Nujiang suture zone (BNSZ) (e.g., Zhang et al., 2012; Guynn et al., 2013; Liu et al., 2015). In the vicinity of the Qamdo basin, BGA magmatic rocks that intruded the Tongka basement (Fig. 1) (He et al., 2011) yielded ~167 Ma zircon U-Pb ages (Zhang et al., 1994). In contrast, westward in central Tibet BGA magmatism mainly occurred during ~170-185 Ma (Guynn et al., 2006; 2012), and even further to the west Tibet it developed later at ~160-175 Ma (Shi, 2007). However, recent studies of Early Cretaceous (~130-100 Ma) ophiolites found in central Tibet (Zhang et al., 2012; Liu et al., 2015) suggest the Meso-Tethyan ocean may not have fully closed by Early Cretaceous time.

Evidence for BGA Provenance

The ~165-190 Ma zircon grains in the Middle-Upper Jurassic Upper Chaya Group (Figs. 7A; 8D) were most likely sourced from erosion of BGA extrusive rocks (Fig. 1B), as there are no other possible sources of Jurassic zircon near the Qamdo basin, even possible removal of source rocks from the immediate area by strike-slip fault movement is taken into consideration (Pan et al., 2004). The BGA source is also supported by an increase of volcanic lithic content from <4% to >20% in the Upper Chaya sandstones with increasing proximity to the BGA (Fig. 1B; Table DR1)

INTERPRETATION OF PROVENANCE

Provenance of the Upper Triassic Strata: Late Triassic Folding and Erosional Exhumation of the Yidun Terrane (YDT)

Sandstone modal data plotted on provenance discrimination diagrams indicate that Upper Triassic strata of the Qamdo basin were derived from craton block and quartzose recycled orogen source areas (Figs. 3A-C). Regional tectonic studies suggest that by the end of the Norian (~204 Ma) in Late Triassic time, the Bayan Har, Songpan-Ganzi, and Yidun terranes had been deformed into a regional fold-thrust belt due to closure of the Paleo-Tethys to the north of the Qamdo basin (Roger et al., 2008; Weislogel, 2008; Gehrels et al., 2011; Burchfiel and Chen, 2013). We interpret that the Upper Triassic Qamdo deposits were most likely sourced from detritus derived from erosion of deformed EYDT sedimentary rocks to the north (Fig. 1B). This interpretation is consistent with the QpLvmLsm plot (Fig. 3C) which shows mainly collisional suture and fold-thrust belt origins for the Upper Triassic sandstone samples, and high similarities between detrital zircon age signatures of the Upper Triassic Qamdo strata and the Triassic Yidun Group (Figs. 5D-E). An alternative interpretation is that Qamdo basin Upper Triassic strata may represent a continuation of the Yidun Group deposystem in Late Triassic, however, depositional age constraints indicate that the Upper Triassic Qamdo strata are generally younger than Yidun Group deposits. The maximum depositional age inferred from the youngest 4 grains from the Bagang Group (Table DR2) yield a maximum depositional age of 218.4 ± 5.0 Ma (2σ) (Dickinson and Gehrels, 2009). While the maximum depositional age of the underlying Jiapila Formation derived from the youngest single detrital zircon grain gives a maximum depositional age of 219 ± 3.3 Ma (2σ) (Table DR2). Although maximum depositional age derived from the youngest single grain age is statically less robust than estimates made from multiple grains, study of Colorado Plateau sediments indicates the method can still yield reliable estimates of

maximum depositional ages in >90% cases (Dickinson and Gehrels, 2009). The inferred maximum depositional ages for the Bagang Group and Jiapila Formation are both consistent with Carnian (228-216 Ma; Jiapila Formation) and Norian-Rhaetian (216-200 Ma; Bagang Group) ages inferred from previous paleontological work (Du et al., 1997; Peng et al., 2005), and are generally younger than the minimum depositional age of ca. 230-220 Ma of the Yidun Group inferred from ages of dacitic intrusions (Wang et al., 2011), despite some overlap at ~220 Ma for the Jiapila Formation.

We note that there was a phase of widespread carbonate deposition occurred during early Norian (~215 Ma; Peng et al., 2005) which deposited ~300 m thick Bolila Limestone on top of the Jiapila Formation, succeed by ~1500 m thick marginal marine/deltaic deposition of the Bagang Group (Fig. 2). Whether the termination of carbonate deposition marks the beginning of inversion and erosion of the EYDT due to eastern Paleo-Tethyan (i.e., Ganzi-Liang ocean) closure, or just reflects a redirection of drainages into the Qamdo basin, however, is not clear.

Together, the combined evidence of sandstone provenance, depositional age and tectonic synthesis suggest that Upper Triassic deposits of the Qamdo basin were most likely derived from erosion of EYDT sedimentary rocks during the Late Triassic orogeny of northeastern Tibet (e.g., Roger et al., 2008; Weislogel, 2008).

Provenance of the Lower Jurassic Strata: Localized Exhumation along the Jinsha Suture and Development of Three Separate Depocenters

At the beginning of Early Jurassic time, Qamdo depositional system shifted from marginal marine/deltaic deposition of the Upper Triassic Bagang Group to terrestrial/lacustrine deposition of the Jurassic Chaya Group. Similar to the Upper Triassic Qamdo samples, most of

the Lower Jurassic Qamdo samples are derived from quartzose recycled orogen sources (Figs. 3D-E), as indicated by high quartz content and common presence of siliciclastic sedimentary lithics fragments (Fig. 4C). However, spatial variability of detrital zircon ages in the Lower Jurassic samples reveals development of at least three distinct depocenters in the Qamdo basin during Early Jurassic time. The southern depocenter samples contain abundant volcanic lithic fragments (i.e., LJ-3, LJ-8, LJ-9) (Fig. 3F; Table DR1) and detrital zircon ages at ~230-245 Ma (Fig. 6B), suggesting local sediment influx from erosion of the JWA (Fig. 8A). The occurrence of abundant middle Paleoproterozoic (1870-1840 Ma) detrital zircon ages in the central depocenter samples (Fig. 6B) indicates erosion of Zhongza basement rocks. The presence of JWA and ZM signatures in Lower Jurassic samples suggests erosional exhumation along the Jinsha suture occurred by Early Jurassic time, probably related to development of the Zhongza thrust system and continued convergence along the Paleo-Tethyan margin (Jolivet et al., 2015; Xu et al, 2015). Sediments in the northern Qamdo depocenter mainly received sediments derived from recycled Upper Triassic Qamdo basin strata, as indicated by quartzose recycled provenance (Fig 3D-F), and high similarity of detrital zircon ages between the northern depocenter Lower Jurassic sample and Upper Triassic samples indicated from comparison of common age populations (i.e., 230-330, 400-500, 700-1200, 1700-1950, 2400-2600 Ma) (Table 1) and high P-values (>0.05) returned from K-S test for 2 out of 4 sample pairs (Table 2).

Our provenance interpretation suggests that during Early Jurassic time sediment routing system was more restricted, transporting detritus derived only from nearby sources (Fig. 10B). The inference is also broadly consistent with fine-grained alluvial and lacustrine deposition of the Lower Chaya Group (Fig. DR2C-F), indicating more stagnant basin hydrology during Early Jurassic.

Provenance of the Middle-Upper Jurassic Strata: Exhumation of the BGA and Internally-recycled Sources

The Middle-Upper Jurassic strata were sourced primarily from erosion of uplifted Upper Triassic Qamdo strata and BGA volcanic rocks along the BNSZ. Spatial variability of detrital zircon ages in the Middle-Upper Jurassic samples still exist and indicates that at least three distinct depocenters that formed during Early Jurassic persisted until Middle-Late Jurassic time (Figs. 10B-C). The samples from the northern and southern Qamdo basin have a quartzose recycled provenance (Figs. 3G-H) and are interpreted to sourced from recycled Upper Triassic strata as supported by high similarities of pre-Mesozoic detrital zircon ages between the Upper Triassic and Middle-Upper Jurassic Qamdo samples (Fig. 9), with K-S test of pre-Mesozoic ages yielded P-values >0.05 for 9 out of 12 sample pairs (Table 4). In contrast, samples from the central Qamdo basin show greater magmatic arc component as indicated by a high volcanic lithic content (Fig. 3I), suggesting significant sediment influx from magmatic arc source/sources. The presence of abundant Jurassic ages (190-165 Ma; 33% of total grains) in the sample (i.e., MUJ-2) from central Qamdo depocenter (Fig. 7A) further suggest localized influx of volcanoclastic detritus derived from the BGA exposed to the west of central Qamdo basin (Fig. 1). It has been proposed that the Jurassic BGA magmatism was fueled by northward subduction of the Mesotethys beneath the Qiangtang block (Girardeau et al., 1984; Yin and Harrison; 2000; Kapp et al., 2007; Zhang et al., 2012; Liu et al., 2015), with subsequent collision between Lhasa and Qiangtang blocks likely causing erosional exhumation of the BGA to source Qamdo basin Middle-Upper Jurassic deposits. In comparison, Jurassic zircon grains (165-190 Ma) are rare in coeval samples from the northern and southern Qamdo depocenters (Fig. 7A), indicating the

BGA was not a widespread source of sediments for the Qamdo basin. However, it is unclear that why exhumation of the BGA only occurred locally close to the central Qamdo basin during Middle-Late Jurassic time. Recent studies have suggested that collision between the Qiangtang and Lhasa blocks occurred highly diachronously along the BNSZ, with the proposed time of collision ranging from Middle Jurassic (Girardeau et al., 1984; Leeder et al., 1988; Guynn et al., 2006; Shi, 2007) to Early Cretaceous (Zhang et al., 2012; Liu et al., 2015). The diachronous nature of Lhasa-Qiangtang collision may responsible for the heterogeneous exhumation of the BGA along the BNSZ, which only exposed the arc to surface level in limited regions (e.g., Guynn et al., 2006; 2012).

Provenance of the Cretaceous Xiangdui Formation: Reactivation of Exhumation of the JWA and Internally-recycled Sources

The Cretaceous samples have mixed recycled orogen and magmatic arc provenance sandstone petrology signature (Fig. 3J-L), as indicated by high quartz content and common presence of volcanic lithic grains. High volcanic lithic content (Fig. 3L; Table DR1) and abundant ~230-245 Ma detrital zircon grains (Fig. 7B) most likely reflect sediment influx from the JWA located to the southeast of the Qamdo basin (Fig. 1). The interpretation is consistent with northwestward paleocurrent directions recorded in Cretaceous alluvial strata (Fig. 2). Another important sediment source came from the recycled Upper Triassic Qamdo strata, as indicated by high similarity of pre-Mesozoic zircon ages between the Cretaceous and Upper Triassic samples (Fig. 9) and K-S test which returned P-values >0.05 (range from 0.07 to 0.95) for 5 out of 8 sample pairs (Table 4).

During Cretaceous, significant shrinkage of basin size (Fig. 1) accompanied by a major shift of sedimentary facies from fine-grained Jurassic lacustrine deposition (Fig. DR1D-F) to Cretaceous alluvial deposition characterized by coarse-grained sandstone and thick-bedded conglomerate layers (Fig. DR1H-I) signifies a major transition in regional tectonic setting of the Qamdo area. Localized uplift and exhumation of the JWA can be inferred from northwestward paleocurrent directions (Fig. 2), abundant granite clasts (Fig. DR1I), and presence of abundant ~230-245 Ma detrital zircon ages in Cretaceous samples (Fig. 7B).

TECTONIC CONTROL ON QAMDO BASIN MESOZOIC PROVENANCE EVOLUTION

Late Triassic Closure of the Eastern Paleo-Tethys

Provenance data indicate Upper Triassic deposits of the Qamdo basin were sourced from erosion of the exhumed Yidun terrane as a result of the closure of the Paleo-Tethyan Ganzi-Litang ocean and development of a widespread fold-thrust belt in northeastern Tibet during Late Triassic time (Fig. 10A; Roger et al., 2008; Weislogel, 2008; Gehrels et al., 2011; Burchfiel and Chen, 2013; Xu et al., 2015). Closure of the Paleo-Tethyan Ganzi-Liang ocean and subsequent collision between the Songpan-Ganzi and Yidun terranes generated orogen-perpendicular compression that propagated to the southeastern Tibet (Yang et al., 2014). The widespread Late Triassic fold-thrust belt (composed of the Bayan Har, Songpan-Ganzi, and Yidun terranes; Roger et al., 2008) not only provided sediments into the Qamdo basin, but may have also contributed detritus into the Sichuan (Luo et al., 2014) and even Ordos (Xie and Heller, 2013) basins.

Inside the Qamdo basin, marine shale and thin-bedded turbidite sandstone of the Jiapila Formation (Carnian in age) (Fig. DR1A) and the overlying Bolila Limestone (early Norian in

age) (Fig. 2) reflect a long period of low clastic input into the Qamdo basin. The reduced clastic influx from the folded Yidun terrane suggests the locus of max subsidence during Carnian-early Norian time was probably closer to the fold belt and the Qamdo basin was probably located in a more distal part of the foreland basin. In contrast, ~1500 m thick marginal marine/deltaic deposition of the Bagang Group (late Norian-Rhaetian in age) (Fig. DR1B) signifies a period of enhanced clastic input, which was probably resulted from southwestward fold-thrust belt growth related to the widespread Late Triassic orogenesis in northeastern Tibet (~220-200 Ma) (Roger et al., 2008; Weislogel, 2008; Burchfiel and Chen, 2013; Wang et al., 2014; Yang et al., 2014; Xu et al., 2015). Alternatively, the change of deposition settings in the Late Cretaceous Qamdo basin from black shale/carbonate platform to sandstone deltaic depositions could have also been associated with the onset of syn-collisional orogenesis in the Yidun terrane followed the closure of the Paleo-Tethyan Ganzi-Litang ocean, though a better constraint on the timing of Yidun orogenesis is needed to test this interpretation (e.g., Wang et al., 2014).

Early Jurassic collisional tectonism of Eastern Paleo-Tethyan closure and Middle-Late Jurassic Closure of the Eastern Meso-Tethys

Transition from Late Triassic marginal marine (e.g., Bagang Group) to Early Jurassic alluvial and lacustrine (e.g., Chaya Group) deposition (Fig. DR1A-D) signifies the Qamdo basin was overfilled by earliest Jurassic time. Growth of the Zhongza fold belt along the northeastern Qamdo margin continued during Early Jurassic, driven by compressional stress created by the continued collision between the Songpan-Ganzi and Yidun terranes (Jolivet et al., 2015; Xu et al., 2015). Significant spatial variations of detrital zircon ages (Figs. 6B, 7A) and sandstone detrital modes (Figs. 3D-I) in Jurassic samples reveal development of three distinct depocenters

in the Qamdo basin during Jurassic time. During Early Jurassic, abundant ~1840-1870 and ~230-245 Ma grains found in the central and southern depocenters, respectively, suggest erosion of the ZM, JWA and associated rocks along the northeastern margin of Qamdo basin (Fig. 10B). In contrast, sample from the northern depocenter lacks the JW or JWA zircon age signature, and has detrital zircon signature similar to the Upper Triassic Qamdo samples (Fig. 7B). The spatial variation of detrital zircon signatures in Lower Jurassic samples suggest during Early Jurassic time, sediment routing system was more restricted, only tapping sediments derived from nearby sources.

The occurrence of ~165-190 Ma zircon grains (Fig. 7A) and abundant volcanic lithic fragments (Fig. 3L) in Middle-Upper Jurassic strata suggest a change of drainage influx which started sourcing sediments from the BGA located to the southwest of Qamdo basin (Fig. 10C), although paleocurrent indicators are rare in lacustrine facies (Fig. DR1D-F). Erosion of the BGA was likely resulted from the collision between the Lhasa and Qiangtang blocks initiated in Middle Jurassic time (e.g., Girardeau et al., 1984; Yin and Harrison; 2000; Kapp et al., 2007; Zhang et al., 2012; Liu et al., 2015). During this period, the Qamdo basin accumulated ~3,500 m thick of sediments that comprise about half of the total thickness of the Qamdo Mesozoic sequence (Fig. 2), reflecting an episode of enhanced basin subsidence. We suggest that the Middle-Late Jurassic subsidence of the Qamdo basin was probably related to crustal loading of the fold-thrust belt developed along the BNSZ, generated by the Lhasa-Qiangtang collision (e.g., Guynn et al., 2013). The interpretation is consistent with distribution of Middle-Late Jurassic sedimentation in the Qamdo basin which was concentrated in south half of the basin along the BNSZ (Fig. 1). The distribution of Middle-Upper Jurassic strata suggests location of max

subsidence center was close to the BNSZ during Middle-Late Jurassic time, reflecting influence of orogenic crustal loading existed to the southwest of the basin (i.e., the BNSZ).

No Jurassic zircon grains have been found in the Lower Jurassic strata, as compared to abundant ~190-180 Ma grains (up to 30%) found within the Middle-Upper Jurassic sediments (Fig. 8C-D). Onset of the northward subduction of the Meso-Tethyan ocean beneath the Qiangtang block by Early Jurassic time (~190 Ma) is indicated by active BGA magmatism during Early Jurassic (e.g., Dewey et al., 1988; Yin and Harrison, 2000; Kapp et al., 2007; Zhang et al., 2012; Guynn et al., 2006, 2013; Zhang et al., 2014b; Liu et al., 2015). The lack of Jurassic zircons in the Lower Jurassic sedimentary record (Fig. 6B) suggests the BGA contributed no detritus into the Qamdo basin during Early Jurassic time, despite their close geographic proximity (Fig. 1B). It is possible that sediment transport direction during Early Jurassic was from northeast to southwest which prevented the BGA from sourcing the Qamdo basin, however, given the close proximity of the sampling sites of Lower Jurassic sandstones and the BGA (Fig. 1) and the low-energy deposition of Lower Jurassic lacustrine and paleosol deposits (Fig DR1C), the Lower Jurassic sediments should at least record a trace of BGA signal if the arc was actively developed on the southern margin of Qiangtang block (e.g., Guynn et al., 2013). Another possibility is that the Qamdo basin and the BGA may have originally been far away from each other during Early Jurassic time, and later been moved closer during Middle Jurassic by strike slip faulting, however, this is not likely given the lack of evidence for Mesozoic strike slip faulting in southeastern Tibet region (Yin and Harris, 2000).

Here, we propose that the missing of BGA signature in the Lower Jurassic Qamdo basin was probably resulted from opening of an Early Jurassic oceanic back-arc basin between the Tongka basement and the southeastern Qiangtang block (Fig. 10B) based on: 1) the occurrence

of widespread Jurassic ophiolitic mélange to the west of Qamdo margin and ophiolite belt within the Tongka basement (Fig. 1); and 2) similar tectonic models suggested from the BGA that intruded the Amdo basement along BNSZ in central Tibet (Guynn et al., 2006; 2013).

Guynn et al. (2006) studied the Amdo granite gneiss basement exposed along the central BNSZ and suggest a short-lived Early Jurassic oceanic back-arc basin may have existed between the Amdo basement and Qiangtang block, possibly due to slab rollback during Meso-Tethyan subduction. Existence of the back-arc basin is further supported by ophiolitic fragments north of the Amdo basement (Pan et al., 2004), along with extensive Early-Middle Jurassic marine sedimentation in the southern Qiangtang block and BNSZ, which is lacking in the Lhasa block (Leeder et al., 1988; Yin et al., 1988; Schneider et al., 2003; Guynn et al., 2006).

In comparison, geologic data indicate that the Tongka orthogneiss basement exposed to the west of the Qamdo basin along the southern BNSZ may have a similar origin as the Amdo gneiss. The Tongka basement is bounded by an ophiolite belt to the southwest and Jurassic ophiolitic mélange deposits to the northeast (Fig. 1B; Pan et al., 2004), similar to the Amdo gneiss (Guynn et al., 2006; 2012). U-Pb crystalline ages of zircon from the Tongka orthogneiss range from 450 to 550 Ma (He et al., 2011), identical to the Cambro-Ordovician (540-460 Ma) crystallization ages of the Amdo basement (Guynn et al., 2006; 2012). Meanwhile, both the Amdo and Tongka basements were intruded by extensive Jurassic igneous rocks related to the subduction of the Meso-Tethyan ocean. Therefore, the Amdo and Tongka basements may have once belonged to an integrated continental sliver oriented parallel to the BNSZ, which had rifted from the Qiangtang block during the opening of an Early Jurassic backarc basin on the western Qiangtang margin due to the Meso-Tethyan subduction. We speculate that now the continental fragment is largely ‘missing’ along the BNSZ because it was either buried by younger cover

strata or subducted northward beneath the Qiangtang block (e.g., Guynn et al., 2006; 2012; 2013). Opening of an Early Jurassic back-arc basin between the Tongka basement and Qiangtang block could explain the lack of Jurassic-age zircons in the Lower Jurassic Qamdo strata, as the backarc basin would have functioned as a geological barrier, preventing the BGA-derived sediments from reaching the Qamdo basin (Fig. 10B). The Early Jurassic ophiolitic mélange deposits within the BNSZ may contain sediments derived from the Tongka basement and associated BGA, but the sediments were not transported further to the Qamdo basin.

Cessation of the Meso-Tethyan subduction and subsequent collision between the Lhasa and Qiangtang blocks during Middle to Late Jurassic time in our study area is marked by an ophiolite belt exposed to the west of the Tongka basement (Fig. 1B), representing the southward obduction of Meso-Tethyan oceanic crust onto the northern margin of the Lhasa block (e.g., Girardeau et al., 1984; Leeder et al., 1988; Zhou et al., 1997; Guynn et al., 2006; 2013). As a result, abundant Early-Middle Jurassic (~165-190 Ma) BGA zircon grains were transported into the Qamdo basin during Middle-Late Jurassic time, recording erosion of the BGA-related rocks formed during convergence between the Lhasa and Qiangtang blocks (Fig. 10C). The presence of BGA zircon but lack of Tongka basement zircon (450-550 Ma) in Qamdo samples suggest erosion of BGA extrusive igneous/volcanic rocks that covered up the Tongka basement. Post-collisional orogenesis associated with the collision between the Lhasa and Qiangtang blocks probably caused deformation and development of foreland fold-thrust belt along the southern margin of the Qamdo basin, causing subsidence and deposition of ~3,500 m thickness Middle-Upper Jurassic lacustrine strata in the Qamdo basin (Fig. DR1D-G).

Cretaceous Intraplate Tectonism

Cretaceous sediments in the Qamdo basin were derived from the JWA combined with recycled sources from Upper Triassic Qamdo strata interpreted to reflect an episode of uplift and exhumation of the JWA along the Jinsha suture. Across the Tibetan plateau, ample evidence of Cretaceous crustal deformation and shortening (e.g., Murphy et al., 1997; Kapp et al., 2005; 2007; DeCelles et al., 2007; Leier et al., 2007; Sun et al., 2015) and rapid rock cooling histories (e.g., Hetzel et al., 2011; Wilson and Fowler, 2011; Deng et al., 2012; Rohrmann et al., 2012; Zhang et al., 2012; Wang et al., 2014) have been documented, with associated mechanisms commonly attributed to continued shortening related to the collision between the Lhasa and Qiangtang blocks and/or the Neo-Tethyan flat-slab subduction beneath the Lhasa block causing crustal thickening. Recent thermochronology data from the Yidun and eastern Songpan-Ganzi terranes reveals an episode of extensive Late Jurassic-Early Cretaceous regional cooling that is interpreted to reflect widespread rock uplift in this area, resulting from compressional strain field related to Lhasa-Qiangtang collision (Tian et al., 2014). The JWA was probably been uplifted during this period of time and became a sediment source for the Cretaceous Qamdo basin. If that's the case, it is possible that the uplifted Yidun and even eastern Songpan-Ganzi terranes may had contributed sediments into the Qamdo basin, however, occurrence of thick-bedded conglomerates (Fig. DR1I) and a paucity of ~220 Ma zircon grains in Cretaceous strata (Fig. 8E) suggest: 1) sediment transport distance was likely short during Cretaceous; and 2) EDYT was probably not a major source for the Cretaceous Qamdo basin as indicated by the missing of eastern Yidun arc signatures (Fig. 8A-E).

IMPLICATION FOR MESOZOIC CLIMATE

Provenance data of the Qamdo basin suggest that except for the Upper Triassic strata which were sourced primarily from the distant YDT, other Mesozoic sediments were mostly derived from the basin bounding ranges (e.g., ZM and BGA) combined with internally recycled sources. This source to sink pattern suggests that for ~100 million years from Early Jurassic to Late Cretaceous time, the Qamdo basin was dominated by a persistent internal drainage system. However, why this internally-drained basin system persisted for such a long time without being defeated by drainage network expansion is unclear. Studies of modern internally drained basins in China (Sobel et al., 2003; Carroll et al., 2010) found that internally-drained basins are usually surrounded by active contractional orogenic belts and restricted to arid or semiarid regions. Fluvial incision modeling also suggests that high bounding range uplift rates, arid regional climatic conditions, and exposure of resistant lithologies of bounding ranges are the main causes of establishment of an internally drained basin system (e.g., Sobel et al., 2003).

We speculate that maintenance of the long-standing internal drainage system of the Jurassic to Cretaceous Qamdo basin was caused by active contractional tectonics related the subduction of Meso-Tethyan ocean and subsequent collision between the Qiangtang and Lhasa blocks (e.g., Zhou et al., 1997; Yin and Harrison, 2000; Kapp et al., 2005; 2007; Zhang et al., 2012; Liu et al., 2015); combined with arid tropical-subtropical greenhouse climatic conditions of southeastern Tibet under the Mesozoic greenhouse world (e.g., Takashima et al., 2006). This interpretation is supported by Jurassic and Cretaceous sedimentary records which show abundant desiccation features (Fig. DR1F) and calcic paleosols (Fig. DR1C; 1J) indicating arid to semi-arid climate condition (e.g., Mack et al., 1993).

CONCLUSIONS

Sediment provenance study of the Mesozoic Qamdo basin revealed that the basin was sourced primarily from erosion of Triassic turbidite of the Yidun Group, metasedimentary rocks of the Zhongza massif, igneous rocks of the Jomda-Weixi arc and Bangong arc, and recycled Upper Triassic Qamdo strata. Upper Triassic sediments of the Qamdo basin were sourced primarily from the inverted Yidun Group as a result of the closure of the Paleo-Tethyan Ganzi-Litang ocean and development of a widespread fold-thrust belt in northeastern Tibet during Late Triassic time (e.g., Roger et al., 2008). Three distinct depocenters developed in the Qamdo basin by the beginning of Early Jurassic time, as indicated by a spatial diversification of detrital zircon age signatures. The northern depocenter received detritus derived primarily from the recycled Upper Triassic Qamdo strata; whereas, the central depocenter has a mixed provenance of Zhongza massif metasedimentary rocks and recycled Upper Triassic Qamdo strata. The prominent ~230-245 Ma peak in the southern depocenter signifies sediment influx from the Jomda-Weixi arc. A lack of Jurassic age zircons in the Lower Jurassic sedimentary rocks is interpreted to reflect a temporal geographical segregation between the Bangong arc and the Qamdo basin due to the opening of a short-lived Early Jurassic oceanic backarc basin, which was probably related to slab rollback during Meso-Tethys subduction (Guynn et al., 2006; 2013).

The Middle-Upper Jurassic strata were sourced from recycled Upper Triassic Qamdo strata and exhumed Bangong arc plutons that intruded the Tongka basement. Abundant ~165-190 Ma zircon grains in the Middle-Upper Jurassic strata suggest exhumation and erosion of the Bangong arc related to the convergence between Lhasa and Qiangtang blocks during Mid-Late Jurassic time.

Cretaceous strata were mostly sourced from the Jomda-Weixi arc combined with sediment input from internally-recycled Qamdo strata. The cause of the reactivated exhumation

of the Jomda-Weix arc was probably related to the northward underthrusting of the Lhasa terrane beneath the Qiangtang block during Early Cretaceous time (e.g., Kapp et al., 2003; 2005; Zhang et al., 2012; Tian et al., 2014; Liu et al., 2015).

The provenance data also reveal that the Qamdo basin was dominated by a long-standing internally drained basin system by Early Jurassic time as most Jurassic and Cretaceous sediments were derived from the basin bounding ranges and internally recycled sources. This internally drained basin system persisted for about 100 million years without being defeated by orogen-traversing channels which most likely due to topographic drainage barriers created by active contractional tectonism and maintained by arid climate conditions in southeastern Tibet during middle to late Mesozoic time.

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REFERENCES CITED

- Barazangi, M., and Ni, J., 1982, Velocities and propagation characteristics of Pn and Sn beneath the Himalayan arc and Tibetan Plateau: possible evidence for underthrusting of Indian continental lithosphere beneath Tibet: *Geology*, v. 10, p. 179–185.
- Burchfield, B.C., and Chen, Z.L., 2013, Tectonics of the southeastern Tibetan Plateau and its adjacent foreland: *GSA Memoirs* 210, Geological Society of America, 164 p.
- Bureau of Geology and Mineral Resources of Sichuan Province (B.G.M.R.S.P.), 1991, Regional Geology of Sichuan Province: Geological Memoir 23: Beijing, Geological Publishing House, 707 p.
- Carroll, A.R., Graham, S.A., and Smith, M.E., 2010, Walled sedimentary basins of China: *Basin Research*, v. 22, p. 17–32, doi:10.1111/j.1365-2117.2009.00458.x.
- Carroll, A.R., Bohacs, K.M., 1999, Stratigraphic classification of Ancient Lakes: balancing tectonic and climatic controls: *Geology*, 27, 99-102.
- Chang, C.F., 1997, *Geology and Tectonics of Qinghai-Xizang Plateau*: Science Press, Beijing, 153 p.
- Chinese Geological Society (C.G.S.), 2004, *Geological Map of the People's Republic of China Explanatory Notes*: Beijing, SinoMap Press, 198 p.
- Clark, M. K., and Royden, L. H., 2000, Topographic ooze: Building the eastern margin of Tibet by lower crustal flow: *Geology*, v. 28, no. 8, p. 703-706.
- Currie, B. S., Rowley, D. B., and Tabor, N. J., 2005, Middle Miocene paleoaltimetry of southern Tibet: implications for the role of mantle thickening and delamination in the Himalayan orogen: *Geology*, v. 33, no. 3, p. 181-184.

- DeCelles, P.G., Kapp, P., Ding, L., and Gehrels, G.E., 2007, Late Cretaceous to mid-Tertiary basin evolution in the central Tibetan Plateau: Changing environments in response to tectonic partitioning, aridification, and regional elevation gain: *Geological Society of America Bulletin*, v. 119, p. 654–680, doi: 10.1130/B26074.1.
- DeGraaff-Surpless, K., Graham, S.A., Wooden, J.L., and McWilliams, M.O., 2002, Detrital zircon provenance analysis of the Great Valley Group, California: evolution of an arc forearc system: *Geological Society of America Bulletin*, v. 114, no. 12, p. 1564–1580, doi: 10.1130/0016-7606(2002)114<1564:DZPAOT>2.0.CO;2.
- Deng, B., Liu, S.G., Li, Z.W., Cao, J.X., Sun, W., 2012. Late Cretaceous tectonic change of the Eastern Margin of Tibetan Plateau: results from multi-system thermochronology. *Journal of the Geological Society of India* 80, 241–254.
- Dewey, J.F., Shackleton, R.M., Chang, C., and Yiyin, S., 1988, The tectonic evolution of the Tibetan Plateau: *Philosophical Transactions of the Royal Society of London, series A*, v. 327, p. 379–413.
- Dickinson, W.R., 1985, Interpreting provenance relations from detrital modes of sandstones, in Zuffa, G.G., ed., *Provenance of Arenites*: Boston, D. Reidel Publishing Company, p. 333–361.
- Dickinson, W.R., and Gehrels, G.E., 2009, Use of U-Pb ages of detrital zircons to infer maximum depositional ages of strata: A test against a Colorado Plateau Mesozoic database: *Earth and Planetary Science Letters*, v. 288, p. 115–125, doi:10.1016/j.epsl.2009.09.013.
- Du, D., Li, J., and Li, X., 1997, Sedimentary evolution and palaeogeography of the Qamdo block in Xizang: *Sedimentary Facies and Palaeogeography*, v 17, no. 4, p. 1-17.

- England, P., and Searle, M., 1986, The Cretaceous - Tertiary deformation of the Lhasa block and its implications for crustal thickening in Tibet: *Tectonics*, v. 5, no. 1, p. 1-14.
- Folk, R. L., 1974, *The petrology of sedimentary rocks*: Hemphill, Austin, Texas 182 p.
- Garcia-Castellanos, D. , 2006, Long-term evolution of Tectonic Lakes: climatic controls on the development of internally drained basins. In: *Tectonics, Climate, and Landscape Evolution* (Ed. by S.D.Willet, N. Hovius, M.T. Brandon & D. Fisher), *Penrose Conf. Ser., Geol. Soc. Am.* 283-294.
- Gehrels, G.E., Valencia, V., Ruiz, J., 2008, Enhanced precision, accuracy, efficiency, and spatial resolution of U-Pb ages by laser ablation-multicollector-inductively coupled plasma-mass spectrometry: *Geochemistry, Geophysics, Geosystems*, v. 9, Q03017, doi:10.1029/2007GC001805.
- Gehrels, G.E., 2010, K-S Test: An Excel macro program for doing statistic K-S comparison between age distributions: [http://laserchron.org/laserchron/home/K-S Test.xls](http://laserchron.org/laserchron/home/K-S%20Test.xls) (accessed December 2012).
- Gehrels, G.E., Kapp, P., DeCelles, P., Pullen, A., Blakey, R., Weislogel, A., Ding, L., Guynn, J., Martin, A., McQuarrie, N., Yin, A., 2011, Detrital zircon geochronology of pre-Tertiary strata in the Tibetan-Himalayan orogen: *Tectonics*, v. 30, TC5016, 27 p. doi:10.1029/2011TC002868
- Girardeau, J., Marcoux, J., Allègre, C.J., Bassoullet, J.P., Youking, T., Xuchang, X., Yougong, Z., and Xibin, W., 1984, Tectonic environment and geodynamic significance of the Neo-Cimmerian Donqiao ophiolite, Bangong-Nujiang suture zone, Tibet: *Nature*, v. 307, p. 27–31.

- Gradstein, F.M., Ogg, J.G., Smith, A.G., Bleeker, W., and Lourens, L.J., 2004, A new geologic time scale, with special reference to Precambrian and Neogene: *Episodes*, v. 27, no. 2, p. 83–100.
- Guynn, J.H., Kapp, K., Pullen, A., Heizler, M., Gehrels, G., and Ding, L., 2006, Tibetan basement rocks near Amdo reveal “missing” Mesozoic tectonism along the Bangong suture, central Tibet: *Geology*, v. 34, p. 505–508, doi: 10.1130/G22453.1.
- Guynn, J., and Gehrels, G.E., 2010, Comparison of detrital zircon age distributions using the K-S test: [http://laserchron.org/laserchron/home/K-S Test Instructions.pdf](http://laserchron.org/laserchron/home/K-S%20Test%20Instructions.pdf) (accessed December 2012).
- Guynn, J., Kapp, P., Gehrels, G.E., and Ding, L., 2012, U-Pb geochronology of basement rocks in central Tibet and paleogeographic implications: *Journal of Asian Earth Sciences*, v. 43, p. 23-50, doi:10.1016/j.jseaes.2011.09.003.
- Guynn, J., Tropper, P., Kapp, P., Gehrels, G.E., 2013. Metamorphism of the Amdo metamorphic complex, Tibet: implications for the Jurassic tectonic evolution of the Bangong suture zone. *Journal of Metamorphic Geology* 31, 705–727.
- Hearn, P., Jr., Hare, T., Schruben, P., Sheriff, D., LaMar, C., and Tsushima, P., 2001, Global GIS Database: Digital Atlas of South Asia: U.S. Geological Survey Digital Data Series DDS-62-C.
- Hetzel, R., Dunkl, I., Haider, V., Strobl, M., von Eynatten, H., Ding, L., and Frei, D., 2011, Peneplain formation in southern Tibet predates the India-Asia collision and plateau uplift: *Geology*, v. 39, p. 983–986, doi: 10.1130/G32069.1.
- Holt, W.E., and Wallace, T.C., 1990, Crustal thickness and upper mantle velocities in the Tibetan plateau region from the inversion of regional Pnl waveforms: Evidence for a thick

- upper mantle lid beneath southern Tibet: *Journal of Geophysical Research*, v. 95, p. 499–525.
- Hou, Z.Q., 1993, Tectono-magmatic evolution of the Yidun island-arc and geodynamic setting of Kuroko-type sulfide deposits in Sanjiang region, SW China: *Resource Geology Special Issue 17*, p. 336–350.
- Ingersoll, R.V., Bullard, T.P., Ford, R.L., Grimm, P., Pickle, L.D., and Sares, S.W., 1984, The effect of grain size on detrital modes: A test of the Gazzi-Dickinson point-counting method: *Sedimentary Petrology*, v. 54, p. 103–116.
- Ingersoll, R.V., Dickinson, W.R., and Graham, S.A., 2003, Remnant-ocean submarine fans; largest sedimentary systems on Earth, in Chan, M.A., and Archer, A.W., eds., *Extreme depositional environments: Mega end members in geologic time*: Geological Society of America Special Paper 370, p. 191–208.
- Jackson, T.W., Robinson, D., Weislogel, A., Jian, X., Shang, F., 2014, Mesozoic Deformation Recorded in Sedimentary Basin on top of the Northern Yidun Arc, Eastern Tibetan Plateau: *Geological Society of America Abstracts with Programs*. Vol. 46, No. 6, p.791
- Jian, X., Guan, P., Zhang, D-W., Zhang, W., Feng, F., Liu, R-J., Lin, S-D., 2013, Provenance of Tertiary sandstone in the northern Qaidam basin, northeastern Tibetan Plateau: Integration of framework petrography, heavy mineral analysis and mineral chemistry. *Sedimentary Geology*, v. 290, p. 109-125, doi: 10.1016/j.sedgeo.2013.03.010.
- Jolivet, M., Roger, F., Xu, Z.Q., Paquette, J.L., and Cao, H., 2015, Mesozoic–Cenozoic evolution of the Danba dome (Songpan Garzê, east Tibet) as inferred from LA-ICPMS U-Pb and fission-track data: *Journal of Asian Earth Sciences*, v. 102, p. 180–204, doi: 10.1016/j.jseaes.2015.02.009 .

- Kapp, P., Yin, A., Manning, C.E., Murphy, M., Harrison, T.M., Spurlin, M., Ding, L., Deng, X.G., and Wu, C.M., 2000, Blueschist-bearing metamorphic core complexes in the Qiangtang block reveal deep crustal structure of northern Tibet: *Geology*, v. 28, p. 19–22.
- Kapp, P., Murphy, M.A., Yin, A., Harrison, T.M., Ding, L., and Guo, J., 2003, Mesozoic and Cenozoic tectonic evolution of the Shiquanhe area of western Tibet: *Tectonics*, v. 22, p. 1029, doi: 10.1029/2001TC001332.
- Kapp, P., Yin, A., Harrison, T.M., and Ding, L., 2005, Cretaceous-Tertiary shortening, basin development, and volcanism in central Tibet: *Geological Society of America Bulletin*, v. 117, p. 865–878, doi: 10.1130/B25595.1.
- Kapp, P., DeCelles, P.G., Gehrels, G.E., Heizler, M., and Ding, L., 2007, Geological records of the Lhasa-Qiangtang and Indo-Asian collisions in the Nima area of central Tibet: *Geological Society of America Bulletin*, v. 119, p. 917–933, doi:10.1130/B26033.1.
- Kind, R., Yuan, X., Saul, J., Nelson, D., Sobolev, S.V., Mechie, J., Zhao, W., Kosarev, G., Ni, J., Achauer, U., and Jiang, M., 2002, Seismic images of crust and upper mantle beneath Tibet: evidence for Eurasian plate subduction: *Science*, v. 298, p. 1219–1221.
- Kumar, P., Yuan, X., Kind, R., and Ni, J., 2006, Imaging the colliding Indian and Asian lithospheric plates beneath Tibet: *Journal of Geophysical Research*, v. 111, doi:10.1029/2005JB003930.
- Leeder, M.R., Smith, A.B., and Yin, J., 1988, Sedimentology, palaeoecology and palaeoenvironmental evolution of the 1985 Lhasa to Golmud Geotraverse: *Royal Society of London Philosophical Transaction, series A*, v. 327, p. 107-143, doi:10.1098/rsta.1988.0123.

- Leier, A. L., P. Kapp, G. E. Gehrels, and P. G. DeCelles, 2007, Detrital zircon geochronology of Carboniferous–Cretaceous strata in the Lhasa terrane, Southern Tibet: *Basin Research*, v. 19, p. 361–378, doi:10.1111/j.1365-2117.2007.00330.x.
- Li, Z.X., Li, X.H., Kinny, P.D., Wang, J., Zhang, S., and Zhou, H., 2003, Geochronology of Neoproterozoic syn-rift magmatism in the Yangtze craton, South China, and correlations with other continents: Evidence for a mantle superplume that broke up Rodinia: *Precambrian Research*, v. 122, p. 85–109, doi:10.1016/S0301-9268(02)00208-5.
- Liu, D.L., Shi, R.D., Ding, L., Huang, Q.S., Zhang, X.R., Yue, Y.H., Zhang, L.Y., 2015. Zircon U–Pb age and Hf isotopic compositions of Mesozoic granitoids in southern Qiangtang, Tibet: implications for the subduction of the Bangong–Nujiang Tethyan Ocean. *Gondwana Research*. <http://dx.doi.org/10.1016/j.gr.2015.04.007>.
- Ludwig, K., 2008, Isoplot 3.6: Berkeley Geochronology Center Special Publication 4, 77 p.
- Luo, L., Qi, J.F., Zhang, M.Z., Wang, K., Han, Y.Z., 2014, Detrital zircon U–Pb ages of Late Triassic–Late Jurassic deposits in the western and northern Sichuan Basin margin: constraints on the foreland basin provenance and tectonic implications: *Int J Earth Sci (Geol Rundsch)* v. 103 p.1553–1568, DOI 10.1007/s00531-014-1032-7
- Mack, G.H., James, W.C., Monger, H.C., 1993, Classification of plaesols: *Geol. Soc. Am. Bull.* 105, 129–136.
- Metcalf, I., 1988, Origin and assembly of Southeast Asian continental terranes: *Gondwana and Tethys*, edited by M. G. Audley-Charles and A. Hallam, Geological Society Special Publication, v. 37, p. 101–118.
- Metcalf, I., 2009, Late Palaeozoic and Mesozoic tectonic and palaeogeographical evolution of SE Asia: Geological Society, London, Special Publications 315, pp. 7–23.

- Meng, Q., and Zhang, G., 2000, Geologic framework and tectonic evolution of the Qinling orogen, central China: *Tectonophysics*, v. 323, no. 3–4, p. 183–196, doi:10.1016/S0040-1951(00)00106-2.
- Mo, X.X., 1993, Sanjiang Tethyan volcanism and related mineralization: Geological Memoirs of PRC Ministry of Geology and Mineral Resources, Series 3, 20, Geological Publishing House, Beijing, 267 p.
- Molnar, P., and Tapponnier, P., 1975, Cenozoic tectonics of Asia: effects of a continental Collision: *Science*, v. 189, p. 419–426.
- Molnar, P., England, P., and Martinod, J., 1993, Mantle dynamics, uplift of the Tibetan Plateau, and the Indian monsoon: *Review of Geophysics*, v. 31, p. 357–396.
- Nelson, K.D., and INDEPTH seismic team, 1996, Partially molten middle crust beneath southern Tibet: synthesis of project INDEPTH results: *Science*, v. 274, p. 1684–1688.
- Owens, T.J., and Zandt, G., 1997, Implications of crustal property variations for models of Tibetan plateau evolution: *Nature*, v. 387, p. 37–43.
- Pan, G.T., Ding, J., Yao, D.S., and Wang, L.Q., 2004, Geological map of Qinghai-Xiang (Tibet) Plateau and adjacent areas: Chengdu, Chengdu Institute of Geology and Mineral Resources, China Geological Survey, scale 1:1,500,000, 6 sheets.
- Peng, Yongmin, Liu, Jiaoduo, Luo, Jianning, 2005, Triassic sequence stratigraphy and evolution of sedimentary basin in Qamdo area: Geological publishing house: Beijing, 82 p.
- Peng, T.P., Zhao, G.C., Fan, W.M., Peng, B.X., and Mao, Y.S., 2014, Zircon geochronology and Hf isotopes of Mesozoic intrusive rocks from the Yidun terrane, Eastern Tibetan Plateau: Petrogenesis and their bearings with Cu mineralization: *Journal of Asian Earth Sciences*, v. 80, p. 18–33, doi: 10.1016/j.jseaes.2013.10.028.

- Potter, P. E., and Pettijohn, F. J., 2012, Paleocurrents and basin analysis, Springer Science & Business Media.
- Powell, C.McA., 1986, Continental underplating model for the rise of the Tibetan Plateau: Earth and Planetary Science Letter, v. 81, p. 79–94.
- Pullen, A., Kapp, P., Gehrels, G.E., Ding, L., and Zhang, Q.H., 2010, Metamorphic rocks in central Tibet: Lateral variations and implications for crustal structure: Geological Society of America Bulletin, v. 123, p. 585–600, doi:10.1130/B30154.1.
- Qiu, Y.M., Shan, G., McNaughton, N.J., Groves, D.I., and Ling, W., 2000, First evidence of >3.2 Ga continental crust in the Yangtze craton of South China and its implications for Archean crustal evolution and Phanerozoic tectonics: Geology, v. 28, no. 1, p. 11–14, doi:10.1130/0091-7613(2000)028<0011:FEOGCC>2.0.CO;2.
- Raphine, R., Tilmann, F., West, M., Ni, J., Rodgers, A., 2003, Crustal structure of northern and southern Tibet from surface wave dispersion analysis: Journal Of Geophysics Research, v. 108 (B2), p. 2120, doi:10.1029/2001JB000445.
- Reid, A.J., Wilson, C.J.L., and Liu, S., 2005, Structural evidence for the Permo-Triassic tectonic evolution of the Yidun arc, eastern Tibetan Plateau: Journal of Structural Geology, v. 27, no. 1, p. 119–137, doi:10.1016/j.jsg.2004.06.011.
- Reid, A., Wilson, C.J.L., Shun, L., Pearson, N., Belousova, E., 2007, Mesozoic plutons of the Yidun Arc, SW China: U/Pb geochronology and Hf isotopic signature: Ore Geology Reviews 31, p. 88–106, doi:10.1016/j.oregeorev.2004.11.003.
- Roger, F., Jolivet, M., & Malavieille, J. 2008. Tectonic Evolution of the Triassic Fold Belts of Tibet. Comptes Rendus Geoscience, Académie des Sciences, Paris, 340, 180–189.

- Rohrmann, A., Kapp, P., Carrapa, B., Reiners, P.W., Guynn, J., Ding, L., and Heizler, M., 2012, Thermochronologic evidence for plateau formation in central Tibet by 45 Ma: *Geology*, v. 40, p. 187–190, doi: 10.1130/G32530.1.
- Royden, L., Burchfiel, B.C., and Van der Hilst, R., 2008, The geological evolution of the Tibetan Plateau: *Science*, v. 321, no. 5892, p. 1054–1058, doi: 10.1126/science.1155371.
- Schneider, W., Mattern, F., Wang, P., and Li, C., 2003, Tectonic and sedimentary basin evolution of the eastern Bangong-Nujiang zone (Tibet): A Reading cycle: *Geologische Rundschau*, v. 92, p. 228–254.
- Shi, R., 2007, SHRIMP dating of the Bangong Lake SSZ-type ophiolite: Constraints on the closure time of ocean in the Bangong Lake-Nujiang River, northwestern Tibet: *Chinese Science Bulletin*, v. 52, p. 936–941.
- Sobel, E.R., Hilley, G.E., and Strecker, M.R., 2003, Formation of internally drained contractional basins by aridity-limited bedrock incision: *Journal of Geophysical Research*, v. 108, p. 2344, doi: 10.1029/2002JB001883.
- Song, X.Y., Zhou, M.F., Cao, Z.M., Robinson, P.T., 2004, Late permian rifting of the South China Craton caused by the Emeishan mantle plume?: *Journal of the Geological Society*, v. 161, no. 5, p. 773–781.
- Stacey, J.S., and Kramers, J.D., 1975, Approximation of terrestrial lead isotope evolution by a two stage model: *Earth and Planetary Science Letters*, v. 26, p. 207–221.
- Studnicki-Gizbert, C., Burchfiel, B., Li, Z., and Chen, Z., 2008, Early Tertiary Gonjo basin, eastern Tibet: Sedimentary and structural record of the early history of India-Asia collision: *Geosphere*, v. 4, no. 4, p. 713–735.

- Sun, G.Y., Hu, X.M., Sinclair, D.H., BouDagher-Fadel, K.M., and Wang, J.G., 2015, Late Cretaceous evolution of the Coqen Basin (Lhasa terrane) and implications for early topographic growth on the Tibetan Plateau, *Geological Society of America Bulletin*, doi: 10.1130/B31137.1
- Takashima, R., Nishi, H., Huber, B. T., and Leckie, R. M., 2006, Greenhouse: *Oceanography*, v. 19, no. 4, p. 64.
- Tao, Y., Bi, X.W., Li, C.S., Hu, R.Z., Li, Y.B., Liao, M.Y., 2014, Geochronology, petrogenesis and tectonic significance of the Jitang granitic pluton in eastern Tibet, SW China: *Lithos*, v. 184-187, p. 314-323, doi: 10.1016/j.lithos.2013.10.031.
- Tapponnier, P., Xu, Z., Roger, F., Meyer, B., Arnaud, N., Wittlinger, G., and Yang, J., 2001, Oblique stepwise rise and growth of the Tibet Plateau: *Science*, v. 294, p. 1671–1677.
- Tian, Y., Kohn, B.P., Hu, Sh., Gleadow, A.J.W., 2014, Postorogenic rigid behavior of the eastern Songpan-Ganze terrane: insights from low-temperature thermochronology and implications for intracontinental deformation in central Asia. *Geochem. Geophys. Geosyst.* 15, 453–474.
- Van der Plas, L., and Tobin, A.C., 1965, A chart for judging the reliability of point counting results: *American Journal of Science*, v. 263, p. 87–90, doi: 10.2475 /ajs .263.1 .87 .
- Vermeesch, P., 2004, How many grains are needed for a provenance study?: *Earth and Planetary Science Letters*, v. 224, no. 3–4, p. 441–451, doi: 10.1016/j.epsl.2004.05.037.
- Vermeesch, P., 2012, On the visualisation of detrital age distributions: *Chemical Geology*, v. 312-313, p. 190-194.
- Wang, Y.B., Han, J., Zeng, P.S., Wang, D.H., Yin, G.H., Li, W.C., 2010, U–Pb dating and Hf isotopic characteristics of zircons from granodiorite in Yangla copper deposit, Denqin

- County, Yunnan, southwest China: *Acta Petrologica Sinica* v. 26, p. 1833–1844 (in Chinese with English abstract).
- Wang, X., Metcalfe, I., Jian, P., He, L., and Wang, C., 2000, The Jinshajiang-Ailaoshan suture zone, China; tectonostratigraphy, age and evolution: *Journal of Asian Earth Sciences*, v. 18, no. 6, p. 675–690, doi: 10.1016/S1367-9120(00)00039-0.
- Wang, B.Q., Zhou, M.F., Li, J.W., Yan, D.P., 2011. Late Triassic porphyritic intrusions and associated volcanic rocks from the Shangri-La region, Yidun terrane, Eastern Tibetan Plateau: Adakitic magmatism and porphyry copper mineralization. *Lithos* 127, 24–38.
- Wang, B.Q., Zhou, M.F., Chen, W.T., Gao, J.F., 2012, Petrogenesis and tectonic implications of the Triassic volcanic rocks in the northern Yidun Terrane, Eastern Tibet: *Lithos*, v. 175–176, p. 285–301, doi:10.1016/j.lithos.2013.05.013.
- Wang, B.Q., Zhou, M.F., Chen, W.T., Gao, J.F., and Yan, D.P., 2013, Constraints of detrital zircon U–Pb ages and Hf isotopes on the provenance of the Triassic Yidun Group and tectonic evolution of the Yidun Terrane, Eastern Tibet: *Sedimentary Geology*, v. 289, p. 74–98, doi: 10.1016/j.sedgeo.2013.02.005.
- Wang, X.S., Hu, R.Z., Bi, X.W., Leng, C.B., Pan, L.C., Zhu, J.J., Chen, Y.W., 2014a, Petrogenesis of Late Cretaceous I-type granites in the southern Yidun Terrane: New constraints on the Late Mesozoic tectonic evolution of the eastern Tibetan Plateau, *Lithos*, Volumes 208–209, November 2014, Pages 202–219, ISSN 0024-4937, <http://dx.doi.org/10.1016/j.lithos.2014.08.016>.
- Wang, B.D., Wang, L.Q., Chen, J.L., Yin, F.G., Wang, D.B., Zhang, W.P., Chen, L.K., Liu, H., 2014b. Triassic three-stage collision in the Paleo-Tethys: constraints from magmatism in the

Jiangda–Deqen–Weixi continental margin arc, SW China. *Gondwana Research*.

<http://dx.doi.org/10.1016/j.gr.2013.07.023>

Weislogel, A.L., 2008, The Songpan-Ganzi complex: Record of basin initiation, evolution and collapse for the northeastern-most Tethys: *Tectonophysics*, v. 451, p. 331–345, doi: 10.1016/j.tecto.2007.11.053.

Weislogel, A.L., Graham, S.A., Chang, E.Z., Wooden, J.L., and Gehrels, G.E., 2010, Collisional orogenesis and spatial distribution of erosional exhumation: detrital zircon provenance and depocenter discrimination of the Middle-Late Triassic Songpan-Ganzi complex, central China: *Geological Society of America Bulletin*, v. 122; no. 11-12; p. 2041-2062; doi: 10.1130/B26606.1

Wilson, C.J.L., Fowler, A.P., 2011. Denudational response to surface uplift in east Tibet: evidence from apatite fission-track thermochronology. *Geol. Soc. Am. Bull.* 123 (9/10), 1966–1987.

Wu, G.J., Xiao, X.C., and Li, T.D., 1993, Lithospheric structure and evolution of the Tibetan Plateau: the Yadong–Golmud Geoscience transect: *Tectonophysics*, v. 219, p. 213–221.

Xiao, L., He, Q., Pirajno, F., Ni, P., Du, J.X., Wei, Q.R., 2008, Possible correlation between a mantle plume and the evolution of Paleo-Tethys Jinshajiang Ocean: Evidence from a volcanic rifted margin in the Xiaru-Tuoding area, Yunnan, SW China: *Lithos*, v. 100, i. 1-4, p. 112-126.

Xie, X. Y., and Heller, P. L., 2013, U-Pb detrital zircon geochronology and its implications: the early Late Triassic Yanchang Formation, south Ordos basin, China: *Journal of Asian Earth Sciences*, v. 64, p. 86–98, <http://dx.doi.org/10.1016/j.jseaes.2012.11.045>

- Xu, Z.Q., Dilek, Y., Cao, H., Yang, J.S., Robinson, P., Ma, C.Q., Li, H.Q., Jolivet, M., Roger, F., Chen, X.J., 2015, Paleo-Tethyan evolution of Tibet as recorded in the East Cimmerides and West Cathaysides: *Journal of Asian Earth Sciences*, v. 105, n. 1 , p. 320-337, ISSN 1367-9120, <http://dx.doi.org/10.1016/j.jseaes.2015.01.021>.
- Yang, X.A., Liu, J.J., Han, S.Y., Zhang, H.Y., Luo, C., Wang, H., Chen, S.Y., 2011, U–Pb dating of zircon from granodiorite, Re–Os dating of molybdenite from the ore body and their geological significances in Yangla copper deposit, Yunnan: *Acta Petrologica Sinica* v. 27, p. 2567–2576 (in Chinese with English abstract).
- Yang, T.N., Ding, Y., Zhang, H.R., Fan, J.W., Liang, M.J., Wang, X.H., 2014, Two-phase subduction and subsequent collision defines the Paleotethyan tectonics of the southeastern Tibetan Plateau: evidence from zircon U–Pb dating, geochemistry, and structural geology of the Sanjiang orogenic belt, southwest China: *Geol. Soc. Am. Bull.* v.126, n.1654–1682.
- Yin, J., Xu, J., Liu, C., and Li, H., 1988, The Tibetan plateau: Regional stratigraphic context and previous work: *Philosophical Transactions of the Royal Society of London, Series A*, v. 327, p. 5–52.
- Yin, A., and Harrison, T.M., 2000, Geologic evolution of the Himalayan-Tibetan orogen: *Annual Review of Earth and Planetary Sciences*, v. 28, p. 211–280, doi: 10.1146/annurev.earth.28.1.211.
- Zhang, D.K., Li, Q.X., Luo, K.Y., Luo, J., 1994, Geological report and map of the Basu region (1: 200,000). Geological Survey of Xizang Province, China, 332 P.
- Zhang, K.J., 2001, Blueschist-bearing metamorphic core complexes in the Qiangtang block reveal deep crustal structure of northern Tibet: *Comment: Geology*, v. 29, no. 90, doi:10.1130/0091-7613(2001)029<0090:BBMCCI>2.0.CO;2.

- Zhang, S.B., Zheng, Y.F., Wu, Y.B., Zhao, Z.F., Gao, S., and Wu, F.Y., 2006, Zircon isotope evidence for ≥ 3.5 Ga continental crust in the Yangtze craton of China: *Precambrian Research*, v. 146, p. 16–34, doi: 10.1016/j.precamres.2006.01.002.
- Zhang, K.J., Zhang, Y.X., Tang, X.C., and Xia, B., 2012, Late Mesozoic tectonic evolution and growth of the Tibetan plateau prior to the Indo-Asian collision: *Earth-Science Reviews*, v.114, no.3-4, p. 236-249, doi:10.1016/j.earscirev.2012.06.001.
- Zhang, X.R., Shi, R.D., Huang, Q.S., Liu, D.L., Gong, X.H., Chen, S.S., Wu, K., Yi, G.D., Sun, Y.L., Ding, L., 2014a. Early Jurassic high-pressure metamorphism of the Amdo terrane, Tibet: constraints from zircon U–Pb geochronology of mafic granulites. *Gondwana Research* 26, 975–985.
- Zhang, L.-Y., Ding, L., Pullen, A., Xu, Q., Liu, D.-L., Cai, F.-L., Yue, Y.-H., Lai, Q.-Z., Shi, R.-D., Ducea, M.N., Kapp, P., Chapman, A., 2014b. Age and geochemistry of western Hoh-Xil–Songpan-Ganzi granitoids, northern Tibet: implications for the Mesozoic closure of the Paleo-Tethys ocean. *Lithos* 190–191, 328–348.
- Zhou, M.F., Malpas, J., Robinson, P.T., and Reynolds, P.H., 1997, The dynamothermal aureole of the Donqiao ophiolite (northern Tibet): *Canadian Journal of Earth Sciences*, v. 34, p. 59-65.

FIGURE CAPTIONS

Figure 1. (A) Sketch tectonic map of Tibet, western China showing location of the Qamdo basin. Black box delineates extent of the study area and adjacent geological units in southeastern Tibet (after Meng and Zhang, 2000; C.G.S., 2004; Weislogel et al., 2010). General localities of reference detrital zircon samples from the Qiangtang and Lhasa blocks are shown in dashed areas. (B) Simplified geological map of Qamdo region showing major tectonic units and distribution of Mesozoic rocks of the Qamdo basin (adapted from Pan et al., 2004). Locations of reference detrital zircon samples (marked by white number in black cycle): 1 – 13UTSG-1; 2 – 13UTSG-2; 3 – 13UTSG-427. Locations of reference magmatic arc zircon samples (marked by black number in triangle): 1 – Cretaceous pluton of the eastern Yidun terrane; 2 – Bangong arc; 3 – Triassic plutons of the eastern Yidun terrane; 4 – Jomda-Weixi arc.

Figure 2. Generalized stratigraphic column for the Qamdo Basin, modified from Du et al. (1997), with stratigraphic location of detrital zircon samples, timing of regional tectonic events (from Kapp et al., 2005; 2007; Roger et al., 2008), paleo-flow directions, and interpretation of depositional environments. Time scale is from Gradstein et al. (2004). Where multiple detrital zircon samples of a formation have been analyzed, the total number is in parentheses.

Figure 3. Detrital modes of Upper Triassic-Cretaceous sandstone samples from the Qamdo Basin. Data are listed in Table DR1. Provenience fields of QtFL and QmFLt ternary plots are after Dickinson (1985). Adapted from Xing et al. (2013). QtFL - total quartz-feldspar-lithic fragment; QmFLt - monocrystalline quartz-feldspar-total lithic fragment.

Figure 4. Photomicrographs of the sandstone thin section samples from the Qamdo basin. (A) Upper Triassic sandstone characterized by high monocrystalline quartz content with few sedimentary lithic fragments (Sample UT-3). (B) Lower Chaya sandstone with high polycrystalline quartz abundance (Sample LJ-2). (C) Sedimentary lithic grains in the Lower Chaya sandstone (Sample LJ-2). (D) Sedimentary lithic grains in the Middle-Upper Jurassic

strata. (E) Upper Chaya sample (MUJ-5) shows high quartz content with few volcanic and sedimentary lithic grains. (F) Volcanic lithic grains found in the Cretaceous sandstone show vitriclastic textures (Sample LK-2). All photomicrographs were taken with cross-polarized light. Qm—monocrystalline quartz; Qp—polycrystalline quartz; P—plagioclase; Lsm—sedimentary and metasedimentary lithic fragments; Lvm—volcanic and metavolcanic lithic fragments.

Figure 5. Histograms and zircon age plots of potential source area zircon data (black solid line — probability density plot; black dash line — kernel density plot). White number in black circle represents sample locations of reference detrital zircon samples from the Yidun Group shown on Figure 1B.

Figure 6. Detrital zircon age distribution probability density plots (black solid line), kernel density plots (black dash line) and histograms for (A) upper Triassic samples and (B) Lower Chaya samples from the Qamdo basin. Sample locations are shown on Figure 1B.

Figure 7. Detrital zircon age distribution probability density plots (black solid line), kernel density plots (black dash line) and histograms for (A) Upper Chaya samples and (B) Cretaceous Xiangdui samples from the Qamdo basin. Sample locations are shown on Figure 1B.

Figure 8. Histograms, probability density plots (black solid line) and kernel density plots (black dash line) of (A) potential arc sources; (B) zircon grains <252 Ma from the Qamdo basin. Gray bars represent prominent arc-signatures observed in detrital record. Sample locations are shown on Figure 1B.

Figure 9. Probability density plots (black solid line) and kernel density plots (black dash line) of zircon ages of the Zhongza massif and the zircon ages >252 Ma from the Qamdo basin. Gray bars represent commonly observed age groups in the Qamdo basin.

Figure 10. A generalized depositional model for evolution of the Qamdo basin, showing separate depocenters and associated drainages tapping different source areas. (A) Late Triassic (ca.220-200 Ma) Qamdo basin was primarily sourced from synorogenic detritus derived from folding and exhumation of the Yidun terrane. (B) Early Jurassic (ca.200-175 Ma), three distinct depocenters were developed in the basin due to basin contraction and localized uplift along the Jinsha suture.

The basin was sourced by the Zhongza massif, the Jomda-Weixi arc, and recycled Upper Triassic Qamdo strata. A short-lived backarc basin to the west of the Qamdo basin prevented the Bangong arc from sourcing the basin. (C) Middle to Late Jurassic (ca.175-145 Ma), abundant Bangong arc-derived sediments started to reach the basin as a result of collision between the Qiangtang and Lhasa blocks, causing uplift and exhumation of the arc. The basin was sourced from both the Bangong arc and internally recycled sources. (D) Cretaceous (ca.145-65 Ma), the basin shrunk significantly toward the south and sediments were derived from the Jomda-Weixi arc and recycled Upper Triassic strata of the Qamdo basin. Cretaceous uplift and exhumation of the Jomda-Weixi arc may related to the underthrusting of the Lhasa block beneath the Qiangtang block (Kapp et al., 2003; 2005; Tian et al., 2014).

¹GSA Data Repository item xxx, **[[Data Repository items include one item list, two figures, two tables and a method description]]**, is available online at www.geosociety.org/pubs/, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

Table 1. U-Pb AGE POPULATIONS AND THEIR PRECENTAGES IN EACH DETRITAL ZIRCON SAMPLES FROM THE QAMDO BASIN

Sample Name	Sam- ple	Age range (Ma)									
	Loca- -tion	165- 190	ca. 220	230- 330	(230- 245)*	(250- 330)*	400- 500	700- 1200	1750- 1900	(1840- 1870)*	2400- 2600
<u>Upper Triassic</u>											
UT-1 (n=88)	NQB	NA	7%	27%	5%	20%	17%	7%	18%	10%	11%
UT-2 (n=78)	NQB	NA	NA	18%	5%	12%	12%	12%	29%	13%	14%
UT-3 (n=84)	SQB	NA	NA	27%	4%	21%	14%	17%	20%	10%	8%
UT-4 (n=99)	SQB	NA	NA	15%	NA	13%	8%	6%	51%	13%	10%
<u>Lower Jurassic</u>											
LJ-1 (n=85)	NQB	NA	7%	18%	5%	12%	20%	15%	25%	9%	5%
LJ-2 (n=97)	CQB	NA	NA	23%	4%	17%	4%	5%	39%	22%	13%
LJ-4 (n=96)	CQB	NA	NA	13%	3%	9%	10%	23%	23%	15%	5%
LJ-3 (n=86)	SQB	NA	NA	29%	19%	8%	7%	19%	18%	14%	2%
<u>Middle-Upper Jurassic</u>											
MUJ-1 (n=93)	NQB	3%	NA	14%	1%	12%	14%	25%	19%	8%	3%
MUJ-4 (n=81)	NQB	NA	NA	27%	7%	19%	16%	10%	23%	4%	7%
MUJ-2 (n=92)	CQB	33%	NA	32%	6%	24%	3%	5%	7%	2%	5%
MUJ-3 (n=97)	SQB	NA	NA	16%	2%	12%	14%	20%	20%	8%	10%
<u>Cretaceous</u>											
K-1 (n=94)	SQB	NA	2%	43%	33%	10%	6%	5%	30%	14%	4%
K-2 (n=105)	SQB	NA	6%	28%	10%	18%	14%	21%	10%	3%	6%

Note: NA—The particular age population is either absent or comprises less than 2% of total grains.

* Indicating a U-Pb age sub-population.

Abbreviations: NQB—Northern Qamdo Basin; CQB—Central Qamdo Basin; SQB—Southern Qamdo Basin.

Table 2. P-VALUES FROM K-S STATISTICAL TEST OF TOTAL DETRITAL ZIRCON POPULATIONS

	UT -1	UT -2	UT -3	UT -4	LJ- 1	LJ- 2	LJ- 3	LJ- 4	MUJ -1	MUJ -2	MUJ -3	MUJ -4	K-1	K-2
UT-1	x	0.0 12	0.6 14	0.0 00	0.5 71	0.0 00	0.4 30	0.0 02	0.01 4	0.00 0	0.02 0	0.93 5	0.0 29	0.0 92
UT-2	0.0 12	x	0.0 25	0.2 10	0.0 15	0.5 66	0.0 12	0.5 90	0.09 0	0.00 0	0.25 7	0.05 8	0.0 02	0.0 00
UT-3	0.6 14	0.0 25	x	0.0 00	0.9 48	0.0 00	0.0 61	0.0 53	0.04 7	0.00 0	0.21 0	0.70 3	0.0 00	0.2 78
UT-4	0.0 00	0.2 10	0.0 00	x	0.0 00	0.2 31	0.0 00	0.0 02	0.00 0	0.00 0	0.00 0	0.00 0	0.0 00	0.0 00
LJ-1	0.5 71	0.0 15	0.9 48	0.0 00	x	0.0 00	0.4 57	0.0 62	0.16 6	0.00 0	0.37 7	0.91 2	0.0 06	0.2 10
LJ-2	0.0 00	0.5 66	0.0 00	0.2 31	0.0 00	x	0.0 00	0.0 29	0.00 0	0.00 0	0.00 4	0.00 1	0.0 00	0.0 00
LJ-3	0.4 30	0.0 12	0.0 61	0.0 00	0.4 57	0.0 00	x	0.0 61	0.12 7	0.00 0	0.15 3	0.79 3	0.3 38	0.3 23
LJ-4	0.0 02	0.5 90	0.0 53	0.0 02	0.0 62	0.0 29	0.0 61	x	0.75 5	0.00 0	0.76 2	0.03 3	0.0 00	0.0 03
MUJ -1	0.0 14	0.0 90	0.0 47	0.0 00	0.1 66	0.0 00	0.1 27	0.7 55	x	0.00 0	0.73 6	0.13 0	0.0 01	0.0 11
MUJ -2	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.00 0	x	0.00 0	0.00 0	0.0 00	0.0 00
MUJ -3	0.0 20	0.2 57	0.2 10	0.0 00	0.3 77	0.0 04	0.1 53	0.7 62	0.73 6	0.00 0	x	0.17 5	0.0 01	0.0 17
MUJ -4	0.9 35	0.0 58	0.7 03	0.0 00	0.9 12	0.0 01	0.7 93	0.0 33	0.13 0	0.00 0	0.17 5	x	0.0 37	0.2 26
K-1	0.0 29	0.0 02	0.0 00	0.0 00	0.0 06	0.0 00	0.3 38	0.0 00	0.00 1	0.00 0	0.00 1	0.03 7	x	0.0 66
K-2	0.0 92	0.0 00	0.2 78	0.0 00	0.2 10	0.0 00	0.3 23	0.0 03	0.01 1	0.00 0	0.01 7	0.22 6	0.0 66	x
LH	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.00 0	0.00 0	0.00 0	0.00 0	0.0 00	0.0 00
QT- N	0.0 00	0.0 00	0.0 00	0.0 00	0.0 01	0.0 00	0.0 03	0.0 00	0.00 0	0.00 0	0.00 0	0.00 0	0.0 00	0.0 00
QT- S	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 02	0.00 0	0.00 0	0.00 0	0.00 0	0.0 00	0.0 00
YDG	0.0 13	0.7 18	0.1 14	0.0 00	0.0 86	0.0 07	0.0 46	0.5 81	0.33 4	0.00 0	0.87 2	0.17 6	0.0 00	0.0 00
UT- QB	0.0 44	0.6 41	0.0 29	0.0 02	0.0 14	0.0 19	0.0 15	0.2 88	0.09 2	0.00 0	0.39 4	0.08 1	0.0 00	0.0 00

Note: Abbreviations: LH—Lhasa Terrane Composite; QT-N—North Qiangtang Terrane Composite; QT-S—South Qiangtang Terrane Composite;

YDG—Yidun Group Composite; UT-QB—Qamdo Basin Upper Triassic Composite.

Bold type denotes $P > 0.05$. Shaded areas show P-values from comparisons of Qamdo samples

with potential sources.

Table 3. P-VALUES FROM K-S STATISTICAL TEST OF MESOZOIC MAGMATIC ARC-DERIVED DETRITAL ZIRCON POPULATIONS (U-Pb AGES <252 MA)

	UT	LJ	MUJ	K
UT		0.918	0.001	0.335
LJ	0.918		0.000	0.251
MUJ	0.001	0.000		0.000
K	0.335	0.251	0.000	
EYDT-K	0.000	0.000	0.000	0.000
BGA	0.000	0.000	0.000	0.000
EYDT-TR	0.006	0.000	0.000	0.000
JWA	0.321	0.327	0.000	0.059

Note: Abbreviations: UT—Upper Triassic Composite; LJ—Lower Jurassic Composite; MUJ—Middle-Upper Jurassic Composite; K—Cretaceous Composite; EYDT-K—Cretaceous plutons of the eastern Yidun terrane; EYDT-TR—Triassic plutons of the eastern Yidun terrane; BGA—Bangong Arc; JWA—Jiangda-Weixi Arc.

Bold type denotes $P > 0.05$. Shaded areas show P-values from comparisons of Qamdo samples with potential arc sources.

Table 4. P VALUES FROM K-S STATISTICAL TEST OF PRE-MESOZOIC DETRITAL ZIRCON POPULATIONS (U-Pb AGE > 252 MA)

	UT -1	UT -2	UT -3	UT -4	LJ- 1	LJ- 2	LJ- 3	LJ- 4	MUJ -1	MUJ -2	MUJ -3	MUJ -4	K-1	K-2
UT-1		0.0 29	0.4 21	0.0 01	0.5 75	0.0 01	0.0 38	0.0 06	0.03 6	0.13 6	0.06 1	0.95 0	0.0 70	0.0 79
UT-2	0.0 29		0.0 13	0.4 69	0.0 23	0.6 90	0.1 39	0.5 53	0.07 2	0.00 3	0.19 4	0.10 2	0.5 62	0.0 00
UT-3	0.4 21	0.0 13		0.0 00	0.3 63	0.0 00	0.0 23	0.0 16	0.02 1	0.29 5	0.09 6	0.79 1	0.0 10	0.9 54
UT-4	0.0 01	0.4 69	0.0 00		0.0 00	0.3 82	0.0 01	0.0 09	0.00 0	0.00 0	0.00 1	0.00 1	0.1 10	0.0 00
LJ-1	0.5 75	0.0 23	0.3 63	0.0 00		0.0 00	0.2 91	0.1 33	0.33 3	0.00 4	0.64 1	0.95 8	0.0 78	0.2 19
LJ-2	0.0 01	0.6 90	0.0 00	0.3 82	0.0 00		0.0 02	0.0 45	0.00 0	0.00 0	0.00 5	0.00 4	0.5 92	0.0 00
LJ-3	0.0 38	0.1 39	0.0 23	0.0 01	0.2 91	0.0 02		0.7 23	0.99 5	0.00 2	0.57 2	0.26 4	0.0 81	0.0 55
LJ-4	0.0 06	0.5 53	0.0 16	0.0 09	0.1 33	0.0 45	0.7 23		0.67 2	0.00 0	0.72 4	0.10 3	0.6 34	0.0 15
MUJ -1	0.0 36	0.0 72	0.0 21	0.0 00	0.3 33	0.0 00	0.9 95	0.6 72		0.00 1	0.70 3	0.33 9	0.0 41	0.0 60
MUJ -2	0.1 36	0.0 03	0.2 95	0.0 00	0.0 04	0.0 00	0.0 02	0.0 00	0.00 1		0.00 1	0.03 4	0.0 14	0.0 72
MUJ -3	0.0 61	0.1 94	0.0 96	0.0 01	0.6 41	0.0 05	0.5 72	0.7 24	0.70 3	0.00 1		0.49 7	0.1 62	0.0 49
MUJ -4	0.9 50	0.1 02	0.7 91	0.0 01	0.9 58	0.0 04	0.2 64	0.1 03	0.33 9	0.03 4	0.49 7		0.1 74	0.1 93
K-1	0.0 70	0.5 62	0.0 10	0.1 10	0.0 78	0.5 92	0.0 81	0.6 34	0.04 1	0.01 4	0.16 2	0.17 4		0.0 00
K-2	0.0 79	0.0 00	0.9 54	0.0 00	0.2 19	0.0 00	0.0 55	0.0 15	0.06 0	0.07 2	0.04 9	0.19 3	0.0 00	
LH	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 00	0.0 09	0.0 00	0.00 0	0.00 0	0.00 0	0.00 0	0.0 00	0.0 00
QT- N	0.0 00	0.0 00	0.0 00	0.0 00	0.0 04	0.0 00	0.0 00	0.0 00	0.00 0	0.00 0	0.00 0	0.00 2	0.0 00	0.0 01
QT- S	0.0 00	0.0 04	0.0 00	0.0 00	0.0 00	0.0 00	0.0 34	0.0 24	0.00 0	0.00 0	0.00 0	0.00 0	0.0 03	0.0 00
YDG	0.0 39	0.7 57	0.0 48	0.0 03	0.1 22	0.0 13	0.2 82	0.5 40	0.25 6	0.00 0	0.76 9	0.47 7	0.4 88	0.0 01
UT- QB	0.3 02	0.3 70	0.0 26	0.0 03	0.0 51	0.0 12	0.2 85	0.0 84	0.14 4	0.01 1	0.25 5	0.30 9	0.5 40	0.0 01
ZM	0.0 11	0.3 75	0.0 12	0.1 94	0.0 74	0.3 30	0.1 10	0.4 90	0.08 1	0.00 1	0.21 9	0.03 9	0.2 70	0.0 04

Note: Abbreviations: LH—Lhasa Terrane Composite; QT-N—North Qiangtang Terrane Composite; QT-S—South Qiangtang Terrane Composite;

YDG—Yidun Group Composite; UT-QB—Qamdo Basin Upper Triassic Composite.
ZM—Zhongza Massif.

Bold type denotes $P > 0.05$. Shaded areas show P-values from comparisons of Qamdo samples with potential sources.

Chapter II

Paleosol records of tropical terrestrial temperatures and atmospheric $p\text{CO}_2$ levels during the mid-Cretaceous (Cenomanian-Turonian) ‘super-greenhouse’ world

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Abstract

The mid-Cretaceous ‘super-greenhouse’ world of ~100 million years ago was characterized by high sea surface temperatures (SSTs), greatly reduced equator-to-pole temperature gradient and high levels of atmospheric $p\text{CO}_2$ driven by elevated rates of seafloor spreading. Although many mid-Cretaceous paleoclimate reconstructions have been published for marine environment, little is known about the evolution of terrestrial temperatures, especially during the Cenomanian-Turonian (C-T) interval (~101-90 Ma). To improve our understanding of terrestrial climate response to the mid-Cretaceous ‘super-greenhouse’ world, we apply the carbonate clumped isotope paleothermometry and paleosol CO_2 paleobarometry on paleosol carbonates collected from an alluvial sequence that spans the C-T interval in southeastern Tibet of China. Δ_{47} -derived soil temperature ($\Delta_{47}\text{-T}$) and $p\text{CO}_2$ reconstructions reveal that average summer temperatures in tropics (~15° N) were ~35 °C with estimated $p\text{CO}_2$ levels at ~614 ppmV during the Cenomanian, followed by a warming of ~11 °C coincident with increase of $p\text{CO}_2$ levels by >40% at the C-T boundary. During the Turonian, $\Delta_{47}\text{-T}$ and $p\text{CO}_2$ values dropped to ~44 °C and ~775 ppmV, respectively. The reconstructed terrestrial temperature trend mirrors coeval tropical SST records, and agrees with mid-Cretaceous climate simulation showing high land surface temperatures in tropics. Our estimate of regional climate sensitivity during mid-Cretaceous yields an average value of ~10 °C per CO_2 doubling, consistent with the view that Earth’s climate response to increasing $p\text{CO}_2$ level was significantly enhanced in greenhouse world conditions.

Introduction

The Cenomanian-Turonian (C-T) interval (~101-90 million years ago) of the mid-Cretaceous was characterized by an extraordinary warm ‘hothouse’ climate superimposed on the greenhouse conditions of the Cretaceous (e.g., Bice et al., 2003; Bice et al., 2006; Huber et al., 2002; Jenkyns et al., 2004; Norris et al., 2002; Schouten et al., 2003; Wilson et al., 2002). Proxy data indicate sea surface temperatures (SSTs) may have reached 33-36 °C in the tropics (Norris et al., 2002; Schouten et al., 2003; Wilson et al., 2002), 30-32 °C in the southern Atlantic Ocean (60° S paleolatitude) (Bice et al., 2003), and >20 °C in the Arctic ocean (Jenkyns et al., 2004), implying a substantially reduced equator-to-pole temperature gradient (e.g., Huber et al., 1995). Geochemical study of brachiopod shells from subtropical and mid-latitude shelf-sea environments reveals a long-term warming trend within the C-T interval, with a dramatic temperature rise of 4-5 °C at the C-T boundary (Voigt et al., 2004). Similarly, organic geochemical analyses (the TEX₈₆ proxy) of mudstones recovered in sediment cores from the deep tropical Atlantic Ocean have revealed a gradual warming trend, with SSTs rising from ~31 °C in the early Cenomanian to ~36 °C in the Turonian (Forster et al., 2007). The C-T global warming is coeval with a worldwide oceanic anoxic event (i.e., Bonarelli event or OAE 2), causing widespread deposition of organic-rich black shales (e.g., Jenkyns, 2010) and global extinction of marine biota (Sepkoski Jr, 1986), with extinction rates locally reaching up to ~79% for macro-invertebrates (i.e., in the Western Interior Seaway of the USA) (Harries and Little, 1999). The high temperatures observed during the mid-Cretaceous have been commonly attributed to elevated atmospheric greenhouse gas levels, with increased $p\text{CO}_2$ driven by enhanced volcanism (e.g., Arthur et al., 1985). Atmospheric $p\text{CO}_2$ estimates from geochemical models range from ~1,000 and ~1,750 ppm (Bergman et al., 2004; Berner and Kothavala, 2001).

Values for $p\text{CO}_2$ estimated from various proxies such as paleosol barometry, stomatal density, liverworts, and phytoplankton carbon isotope ratios are largely similar to those of the model outputs, but display a slightly larger variation, with most values falling between ~500 and ~1,500 ppm (Barclay et al., 2010; Bice et al., 2006; Breecker et al., 2010; Fletcher et al., 2008; Freeman and Hayes, 1992; Van Benthum et al., 2012).

However, most of our knowledge about the mid-Cretaceous climate was derived from marine records, and there is a general paucity of continental record indicating an extraordinary mid-Cretaceous warmth (e.g., Chamberlain et al., 2013), as might be expected. The scarcity of data mainly arises from a lack of quantitative methods for extracting temperature information from terrestrial sedimentary records. Recent development of carbonate clumped-isotope thermometry provides an innovative technique that holds potential for reconstruction of carbonate-formation temperatures over a range of timescales (e.g., Dennis et al., 2013; Eagle et al., 2013; Eiler, 2011; Fan et al., 2014; Garzzone et al., 2014; Hough et al., 2014; Huntington et al., 2015; Petryshyn et al., 2015; Quade et al., 2013; Snell et al., 2014; Snell et al., 2013; Suarez and Passey, 2014; Suarez et al., 2011; Tripathi et al., 2010; Tripathi et al., 2015; VanDeVelde et al., 2013; Zaarur et al., 2011). Bonding of ^{13}C and ^{18}O ions in carbonate minerals is temperature-dependent, and thus the abundance of ^{13}C - ^{18}O bonds is closely correlated to the formation temperatures of carbonate minerals (Eiler, 2007; Ghosh et al., 2006). The technique has been applied to a few Cretaceous climate studies, including the study of the Early Cretaceous (Berriasian-late Valanginian) marine temperatures in sub-arctic region using mollusk shells (Price and Passey, 2013), the Late Cretaceous (late Campanian and Maastrichtian) climate of Western Interior Seaway of North America using cephalopod fossils (Dennis et al., 2013), and the late Early Cretaceous (Aptian-Albain) terrestrial climate by studying lacustrine carbonates

from the Xiagou Formation of northwestern China (Suarez et al., 2012), whereas continental temperature records from the C-T interval are still lacking.

Figure 1. Regional map. (A) Global paleogeographic map of mid-Cretaceous (~90 Ma) showing general location of the study area in southeastern Tibet (red square). Figure adapted from Jarvis *et al.* (2011) based on the paleogeographic reconstruction of Blakey (Blakey, 2008) (B) Simplified geological map of southwestern China showing major tectonic units and block boundaries (Pan et al., 2004). (C) Local geologic map of Markham Area, eastern Tibet (Pan et al., 2004).

In order to improve our understanding of terrestrial climate evolution in response to changing $p\text{CO}_2$ levels and associated global warming during mid-Cretaceous, we examined a paleosol-rich Cretaceous alluvial sequence that spans the C-T interval, located in the Markam County of southeastern Tibet (Fig. 1) using carbonate clumped isotope paleothermometer. Soil forms at the Earth's surface, and directly records climatic and environmental conditions over the period of growth, and thus is a powerful tool for paleoclimate reconstructions (Sheldon and Tabor, 2009). For clumped isotope-based paleoclimate studies, paleosol carbonates have been increasingly used as an archive for reconstruction of terrestrial temperatures (e.g., Eagle et al., 2013; Garzione et al., 2014; Huntington et al., 2015; Passey et al., 2010; Quade et al., 2013; Snell et al., 2013; Suarez et al., 2011). Additionally, atmospheric CO_2 concentrations can be modeled using the paleosol carbonate barometer developed by Cerling (1999). If soil carbonate forming temperatures can be constrained by clumped isotope thermometry, then key parameters used for $p\text{CO}_2$ estimates based on this barometer (e.g., $\delta^{13}\text{C}_\text{S}$: $\delta^{13}\text{C}$ of soil CO_2 and S_Z : soil CO_2 concentration) can be more accurately estimated (Breecker and Retallack, 2014; Breecker et al., 2009; Montanez, 2013; Romanek et al., 1992).

Here, we report Δ_{47} -derived soil temperatures ($\Delta_{47}\text{-T}$), $\delta^{18}\text{O}$ values of soil water ($\delta^{18}\text{O}_\text{water}$) and estimated atmospheric $p\text{CO}_2$ levels from paleosol carbonate nodules. Influence of diagenesis on soil carbonates is rigorously screened using isotopic, petrographic, chemical techniques to assess if the soil carbonates preserve original isotopic and Δ_{47} signals. The reconstructed $\Delta_{47}\text{-T}$ and $p\text{CO}_2$ values from soil record are compared with marine temperature datasets and mean annual surface-air temperature (MAAT) from Global Climate Model (GCM) simulation (Kump and Pollard, 2008), and existing $p\text{CO}_2$ estimates (Barclay et al., 2010; Bice et al., 2006; Breecker et al., 2010; Fletcher et al., 2008; Freeman and Hayes, 1992; Van Bentum et

al., 2012), respectively, to improve our understanding of regional continental climate response to mid-Cretaceous global warming and Earth's system climate sensitivity in 'super-greenhouse' world.

Geologic and climate settings

Our study area is located in the Markam County of southeastern Tibet where Cretaceous alluvial strata are present (Fig. 1) (Du et al., 1997; Pan et al., 2004). The Cretaceous sequence was deposited in the southern Qamdo basin developed on top of the eastern Qiangtang block, bounded by the Jinsha suture to the east and the Bangong-Nujiang suture to the west (Fig. 1C). Paleogeographic reconstruction suggests that the study area was located at approximately 15° N in a tropical zone during the mid-Cretaceous (~90 Ma) (Fig. 1A; Blakey, 2008). The Cretaceous strata cover an area of ~2,000 km² and contain ~1,200 m thick alluvial deposits composed of clast-supported conglomerate, cross-stratified pebbly red sandstone, and thick-bedded carbonate nodule-bearing paleosol layers (Fig. 2A) underlain by Triassic-Jurassic marine-lacustrine deposits (Du et al., 1997; Pan et al., 2004). Paleosol profiles range from 1.5 m to up to 8 m thick and were identified in the field based on lithology (i.e., silty clay), lack of stratification, and presence of carbonate nodules, root traces, bioturbation features, color mottling and ped structures (Fig. 2). Paleosol carbonates occur as discrete nodules ranging from 1-5 cm in diameter.

Figure 2. Field photos of Cretaceous section and individual paleosol profiles. (A) Overview of Cretaceous strata composed of clast-supported conglomerate, cross-stratified pebbly red sandstone, and thick-bedded carbonate nodule-bearing paleosol layers. Letters in white triangle indicate positions of individual paleosols shown in Figs. 2B-E. (B-E) Sedimentary logs and photographs of individual paleosol profiles, showing Bk-depth and pedogenic features such as

root traces, bioturbations, color mottling, ped structures and carbonate nodules.

Paleoclimate reconstructions suggest that our study area was located in a widespread evaporite zone near the equatorial region ($\sim 15^{\circ}$ N) during the mid-Cretaceous (Hay and Floegel, 2012). The evaporite zone was characterized by increased aridity and evaporation during the C-T time (Hay and Floegel, 2012). Although C-T terrestrial temperature records from proxies are still lacking, climate model simulations (i.e., GENESIS v. 3.0) show that during mid-Cretaceous (~ 100 Ma) ‘super-greenhouse’ interval, mean annual surface-air temperatures (MAAT) in southeastern Tibet may have reached 35-40 °C (Kump and Pollard, 2008). If the simulated mid-Cretaceous MAATs are representative, it is very likely that warm-month surface-air temperatures during mid-Cretaceous may even higher in this region.

Stratigraphy and age constraint

The Cretaceous sequence exposed near the city of Markam is divided into two major units: the Lower Cretaceous Jingxing Formation and the Upper Cretaceous Nanxin Formation (GSX, 2007) (Fig. 3A). The Jingxing Formation consists of ~ 700 m light grey sandy conglomerate and volcanic lithic arenite in the basal part and transitions upwards to red medium- to coarse-grained sandstone with interbedded Calcisols (Du et al., 1997). The Nanxing Formation consists of ~ 600 m of thick-bedded red to purplish lithic sandstone interbedded with layers of conglomerate, pebbly sandstone, mudstones, and carbonate nodule-bearing Calcisols (Du et al., 1997).

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Stratigraphic boundaries and ages are determined based on paleomagnetic dating (Huang et al., 1992), biostratigraphy and lateral correlations with Cretaceous type sections for the region mapped by Xizang (Tibetan) Geological Survey (GSX, 2007), and palynological analysis conducted in this study (Fig. 3A). The Cretaceous strata in our study area cannot be radiometrically dated since it contains no volcanic ash layers that are suitable for dating, and the youngest detrital zircon age populations of the Cretaceous sediments are all >200 Ma (Shang, 2014). Full list of faunal assemblages and palynological data are included in Supplementary Text 1.

In our studied sections, depositional ages of the Lower Cretaceous Jingxing Formation is constrained at between middle Aptian and Albian (118-100 Ma), whereas the age of the lower Nanxing Formation is constrained at between Cenomanian and Turonian (~100-89 Ma), and that of the upper Nanxing Formation is at between Coniacian and Santonian (~89-83 Ma) (Fig. 3A) (GSX, 2007; Gradstein et al., 2004; Huang et al., 1992). The position of the Cenomanian-Turonian (C-T) boundary in the studied sections is further determined by assuming a constant average sediment accumulation rate, as indicated by relatively continuous stratigraphic deposition with absence of major unconformities and depositional hiatuses (Fig. 2A). Note that when calculating the stratal thickness, thick-bedded conglomerate layers were excluded since they most likely reflect rapid, short-duration depositional events. The C-T boundary was therefore placed at ~200 m height of the composited section (Fig. 3A).

Methodology

A total of 21 individual paleosol horizons from the C-T interval were sampled, with 12 samples from section A-A' and 9 samples from section C-C'. Our rationale for sampling the

same paleosol interval at two different locations (~3 km apart) is that soil temperature can be affected by vegetation shading and local slopes (Quade et al., 2013). Thus, sampling the same paleosol interval at different locations will help us to distinguish between regional paleoclimate signals and local variations. Carbonate nodules were sampled from depth of ~50 cm below top of the preserved paleosol surface to constrain the variation of soil temperature with depth and reduce the potential effects of soil water evaporation (Quade et al., 2013).

Mineralogical analysis, petrographic analysis and isotopic screening

Carbonate nodule samples were characterized using a combination of mineralogy, microscopy, petrology and geochemical analyses of Mg/Ca and Sr/Ca ratios. X-ray diffraction (XRD) analysis was performed on all samples. Diffraction patterns were obtained using a PANalytical X'Pert Pro XRD diffractometer equipped with Cu K α radiation at 40 keV and 40 mA. Samples were first powdered and scanned from 10° to 100° at 2 θ with a step interval of 0.05°. Carbonate mineralogy was determined using the computer-based Highscore Plus software equipped with a Powder Diffraction File 3 database. In addition, Sr/Ca and Mg/Ca ratios of all samples from section A-A' were measured using inductively coupled plasma mass spectrometry (ICP-MS) at Actlabs, Canada. Polished petrographic thin-sections were produced for all samples collected from section A-A' (n=12). The thin-sections were examined under transmitted and cross-polarised light to characterize their mineral compositions, pedogenic textures, and carbonate phases.

Before the more time (~3 hrs per sample) and material (average sample size = ~10 mg) consuming clumped isotope analysis, an stable oxygen and carbon isotopic screening was performed on carbonate nodules collected from 11 paleosol horizons span the ~300 m C-T

interval at the West Virginia University Stable Isotope Lab, using a GasBench device coupled to a Delta Advantage isotope ratio mass spectrometer. Standardization is based on NBS18 and NBS19 for $\delta^{13}\text{C}/\delta^{18}\text{O}$ of carbonate. Analytical precision is better than ± 0.1 for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (1s), based on repeated analysis of internal standards. Nodules with both micritic and sparry textures (Fig. A.1.) were micro-sampled ($n=1$ for sparry calcite; $n=2$ for micritic calcite; average sample size = 0.3mg) using a low-speed dremel drill. Carbonate powders were analyzed for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values to detect isotopic differences between micritic and co-existing sparry phases (Fig. A.1.), which is useful in characterizing diagenetic histories of carbonate cement (e.g., Garzzone et al., 2004).

Clumped-isotope thermometry

Carbonate clumped-isotope thermometry is an innovative technique that is based on the temperature-dependant bonding of heavy isotopes (^{13}C and ^{18}O) into bonds (e.g., $^{13}\text{C}-^{18}\text{O}-^{16}\text{O}$) in carbonate minerals, which is favored at low temperatures (Eiler, 2007; Ghosh et al., 2006). The abundance of $^{13}\text{C}^{18}\text{O}^{16}\text{O}$ produced on acid digestion of carbonate minerals is measured (Ghosh et al., 2006). The relative abundance of the $^{13}\text{C}^{18}\text{O}^{16}\text{O}$ isotopologue is reported using the variable “ Δ_{47} ” (Eiler and Schauble, 2004). Carbonate-formation temperatures can be calculated by applying published calibrations relating Δ_{47} values to carbonate formation temperatures (e.g., Tripathi et al., 2015).

The approach not only constrains the formation temperature of carbonate minerals, but also makes calculation of $\delta^{18}\text{O}$ value of water from which carbonate grew more accurate because fractionation of O^{18} between calcite and water is temperature-dependant (Kim and O'Neil, 1997). For soil carbonate, the relationship between soil forming temperature and $\delta^{18}\text{O}_{\text{water}}$ is very useful

in detecting signs of evaporation and diagenetic effects, as well as discriminating between primary climate signals and diagenetic resetting (Quade et al., 2013).

Carbonate clumped isotope analysis was conducted at University of California, Los Angeles (UCLA), using a specially modified Thermo Fisher 253 gas source mass spectrometer with mass 44-49 ion collectors (“Chewbacca”). Carbonate nodules with homogeneous micritic textures were carefully sampled using a low-speed dremel drill with periodic cooling of the drill pit to avoid potential resetting of the samples. Samples and carbonate standards are first processed using a custom-built automated system for digestion and purification (after Passey et al., 2010) before being transferred to the inlet of the mass spectrometer. For each measurement, 8-10 mg of calcite powder were digested at 90 °C in phosphoric acid, the resultant CO₂ being cryogenically purified before analysis on the mass spectrometer.

The absolute reference frame (ARF) was established by analyzing equilibrium CO₂ samples prepared at 1000 °C and 25 °C. The standard gas are purified and analyzed using the same protocol as gases generated from carbonate samples. For carbonates, the Δ_{47} values are normalized to 25 °C using a 0.092‰ value from Passey et al. (2010) as recalculated in Henkes et al. (2013), which is within error of the fractionation factor reported by Defliese et al. (2015). Clumped-isotope data for pedogenic calcite are reported relative to the ARF of Dennis et al. (2011). Uncertainties in reported Δ_{47} values and calculated temperatures include the propagated uncertainty in equilibrated gas determination and in sample measurement. Δ_{47} -temperatures are calculated using the inorganic calibration of Tripathi et al. (2015), generated on the same instrument that measurements for this study were made on. For completeness, Supplementary Table 1 also shows estimated temperatures using other published calibrations.

During the clumped isotope analysis, three internal carbonate standards (Carrerra Marble,

Carmel Chalk and Veinstrom standard) were analyzed multiple times. The mean Δ_{47} values and standard deviation (SD) of these standards during the period of analysis were $0.390 \pm 0.012\text{‰}$ (n =11), $0.697 \pm 0.014\text{‰}$ (n =8), and $0.725 \pm 0.011\text{‰}$ (n =12), respectively, which comes to an average lab precision during the period of sample analysis of 0.012‰, lower than the long-term lab precision of 0.009‰. The standard errors (1 s.e.) for Δ_{47} values are calculated as $SE=SD/SQRT(N)$ (1), where SD is the observed standard deviation for repeated analyses of samples, and N is the number of repeated analyses. If the observed SD is smaller than the long-term lab precision (0.009‰), SE is calculated using $SD=0.009\text{‰}$ (Tripathi et al., 2015). Error in Δ_{47} -T is calculated by propagating the SE for Δ_{47} through the Tripathi et al. (2015) temperature calibration equation.

Reconstruction of atmospheric pCO₂

We reconstructed atmospheric CO₂ concentrations using the paleosol carbonate barometer developed by Cerling (1999) who reported the following equation:

$$C_a = S(z) * \frac{\delta^{13}C_s - 1.0044 * \delta^{13}C_r - 4.4}{\delta^{13}C_a - \delta^{13}C_s} \quad (2)$$

where C_a is pCO₂ in atmosphere (ppmV). The value of S_z (soil-respired CO₂ concentration, ppmV) was commonly assumed to be 5000 ppmV (e.g., Ekart et al., 1999), however, recent studies show that this value has been strongly overestimated since soil carbonated only form in dry months when soil respiration rates are low (e.g., Breecker et al., 2009; Hough et al., 2014). Currently, S_z value can be estimated using three approaches: 1) correlation between S_z values and depth to B_k horizon (Breecker and Retallack, 2014; Retallack, 2009); 2) soil-order method which assigns specific ranges of S_z values for different soil types (Montanez, 2013); and 3) correlation between mean annual precipitation (which also correlates to depth to B_k horizon)

(Retallack, 2005) and S_z (Cotton and Sheldon, 2012). Bk horizons in our studied paleosol profiles occur at relatively shallow depth (Fig. 2) at ~10-31 cm (average =21 cm), even after burial decompaction using a standard compaction formula for aridisols (Table 4) (Sheldon and Retallack, 2001). The S_z values estimated using this approach yields values range from 776 to 1531 ppmV (average =1157 ppmV) and associated uncertainty for each paleosol profile $\sim \pm 100$ ppmV (Table 4). In cases when paleosol profiles were eroded by overlying alluvial sandstones which can be detected by presence of scour surface (Fig. 2E), S_z values are adopted from the nearest paleosol horizons with better preserved A horizons. The S_z values estimated from Bk-depth method are consistent with estimates based on soil-order method (Montanez, 2013). Paleosol in our studied sections are dominated by Calcisols (Fig. 2) and modern observations suggest that carbonate precipitation in Aridisols (and its paleosol analog Calcisols) occurs when S_z values range from 500-2000 ppmv (Montanez, 2013).

Parameters of $\delta^{13}C_s$, $\delta^{13}C_r$, and $\delta^{13}C_a$, are the stable carbon isotopic compositions of soil CO_2 , soil-respired CO_2 , and atmospheric CO_2 , respectively. $\delta^{13}C_s$ values can be estimated from measured $\delta^{13}C_{\text{carbonate}}$ values and growth temperatures (Δ_{47-T}) of pedogenic carbonate, using the temperature-dependent carbon isotopic fractionation equation for calcite and CO_2 (Romanek et al., 1992). Values for $\delta^{13}C_r$ are normally approximated by the $\delta^{13}C$ value of soil organic matter ($\delta^{13}C_{\text{org}}$) (e.g., Cerling, 1999), however, microbial decay of soil organic matters may cause enrichment of $\delta^{13}C_{\text{org}}$ by several per mil (Wynn, 2007). Here, instead of analyzing the $\delta^{13}C_{\text{org}}$ from bulk paleosol samples, we adopt a slightly different method of analyzing only the organic matters enclosed in the pedogenic carbonate nodules. We argue that organic matter enclosed in the carbonate nodules should have a better chance of preserving their original isotopic signals from aerobic decomposition. More importantly, exposed paleosols are often contaminated by

modern plants, fungi, and soil bacteria, therefore, analyzing only the organic matters enclosed in the carbonate nodules could greatly reduce the risk of sampling the modern organic contaminations.

Paleosol organic carbon isotope ($\delta^{13}\text{C}_{\text{org}}$) analysis was performed on all 21 samples at the Environmental Isotope Laboratory of the University of Arizona. Pedogenic carbonate nodules were carefully washed using distilled water and dried in room temperature before crushing into powders using mortar and pestle. Carbonates were removed by adding sulfurous acid (H_2SO_3) to the sample in a silver-foil combustion capsule. Every 24 hours two drops of acid were added to the samples and then they were dried overnight at 55 °C. After visible bubbles were no longer generated by the addition of acid, two more drops of acid were added to ensure complete removal of carbonate and the samples were dried. The foil capsule was then closed and placed in a tin capsule for combustion. $\delta^{13}\text{C}_{\text{org}}$ was measured on a continuous-flow gas-ratio mass spectrometer (Finnigan Delta PlusXL). Samples were combusted using an elemental analyzer (Costech) coupled to the mass spectrometer. Standardization is based on acetanilide for elemental concentrations, and NBS-22 and USGS-24 for $\delta^{13}\text{C}$. Precision is better than ± 0.08 for $\delta^{13}\text{C}$ (1s), based on repeated analysis of internal standards.

Alternatively, the $\delta^{13}\text{C}_{\text{org}}$ values can be estimated using an empirical equation relating the $\delta^{13}\text{C}_a$ to the $\delta^{13}\text{C}$ of paleosol organic matter: $\delta^{13}\text{C}_a = (\delta^{13}\text{C}_{\text{org}} + 18.67)/1.1$ (3) (Arens et al., 2000), where the $\delta^{13}\text{C}_a$ can be derived from the coeval $\delta^{13}\text{C}$ of planktonic foraminifera, assuming a -7.5‰ isotopic equilibrium fractionation between ocean and atmospheric CO_2 under the high ocean temperatures (>30 °C) of mid-Cretaceous (Forster et al., 2007; Lynch-Stieglitz et al., 1995). The $\delta^{13}\text{C}$ of planktonic forams can be derived from the Deep Sea Drilling Project (Huber et al., 1995) and the Tanzania Drilling Project (MacLeod et al., 2013), showing $\delta^{13}\text{C}$ of

planktonic forams $\sim 2\text{‰}$. The equation (3) thus yields a $\delta^{13}\text{C}_{\text{org}}$ value of -24.7‰ . This value is consistent with our measured $\delta^{13}\text{C}_{\text{org}}$ from paleosol record (~ -26 to $\sim -24\text{‰}$) and $\delta^{13}\text{C}$ values of ~ -25 to -23‰ measured from the C-T terrestrial organic matter from northern Japan (Hasegawa, 2003). Together, the consistence of $\delta^{13}\text{C}_{\text{org}}$ estimates derived from various proxies suggest the organic matters in the Markam paleosol preserved an original signal of $\delta^{13}\text{C}_{\text{org}}$ without significant ^{13}C -enrichment by microbial decay. Therefore, the coexisting paleosol $\delta^{13}\text{C}_{\text{org}}$ is adopted as $\delta^{13}\text{C}_r$ in the $p\text{CO}_2$ equation (2). Value for $\delta^{13}\text{C}_a$ is further calculated using equation (2): $\delta^{13}\text{C}_a = (\delta^{13}\text{C}_{\text{org}} + 18.67)/1.1$.

Results

Mineralogical analysis, petrographic analysis and isotopic screening

Computer-aided mineral identification using X'Pert High Score software indicates that carbonate nodule samples all consist of calcite, quartz, and minor clay minerals. Calcite is the dominant component of the samples and the only carbonate species present, as inferred from maximum XRD peaks varying between 29.20° and $29.50^\circ 2\theta$, and a d-spacing of 3.03 to 3.06 Å (Table 1), which is almost identical to published laboratory standards for calcium carbonate ($29.40^\circ 2\theta$ and 3.04 Å, respectively) (Swanson et al., 1953). Elemental analysis on carbonate mineral phases reveals low Mg/Ca and Sr/Ca ratios (<0.01) (Table 1) (Graham et al., 2005).

Table 1. Sample mineralogy as determined by XRD, Mg/Ca, and Sr/Ca ratios

Sample ID	Stratigraphic Position (m)	XRD results				Elemental results	
		Number of counts	100% 2°θ	100% d-spacing (Å)	Main component	Mg/Ca	Sr/Ca
Section A-A'							
MK6325	281	72892	29.364	3.04170	Calcite, Quartz	0.0054	0.0008
MK6322	263	83277	29.380	3.04016	Calcite, Quartz	0.0047	0.0006
MK6321	250	78943	29.436	3.03444	Calcite, Quartz	0.0050	0.0012
MK6319	215	61613	29.422	3.03588	Calcite, Quartz	0.0048	0.0009
MK6318	195	85589	29.431	3.03497	Calcite, Quartz	0.0047	0.0004
MK6317	192	52815	29.414	3.03667	Calcite, Quartz	0.0064	0.0006
MK6316	140	37980	29.203	3.05812	Calcite, Quartz	0.0062	0.0007
MK6315	107	78881	29.416	3.03644	Calcite, Quartz	0.0048	0.0004
MK6312	84	66107	29.393	3.03884	Calcite, Quartz	0.0070	0.0006
MK6311	59	49812	29.412	3.03690	Calcite, Quartz	0.0077	0.0010
MK639	29	71606	29.436	3.03444	Calcite, Quartz	0.0066	0.0006
MK638	15	80815	29.440	3.03407	Calcite, Quartz	0.0072	0.0009
Section C-C'							
MK6230	310	11670	29.475	3.03051	Calcite, Quartz	N/A	N/A
MK6228	300	11272	29.477	3.03036	Calcite, Quartz	N/A	N/A
MK6227	234	11663	29.468	3.03125	Calcite, Quartz	N/A	N/A
MK6226	192	10956	29.481	3.02988	Calcite, Quartz	N/A	N/A
MK6223	162	10615	29.502	3.02778	Calcite, Quartz	N/A	N/A
MK6222	112	9994	29.474	3.03059	Calcite, Quartz	N/A	N/A
MK6220	100	11314	29.466	3.03142	Calcite, Quartz	N/A	N/A
MK6221	72	11327	29.485	3.02949	Calcite, Quartz	N/A	N/A
MK6218	35	7719	29.498	3.02824	Calcite, Quartz	N/A	N/A
N/A: data not available							

Thin-section petrographic examination shows that carbonate nodules are dominated by red-brown micritic calcite groundmass occasionally overprinted by dark-brown soil mottling (Fig. 4). Detrital siliciclastic quartz grains are common in thin-sections (Fig. 4) and no detrital carbonate grains are observed. Sparry calcite cement typically occurs in isolated vugs/veins (Figs. 4E, F) and makes up <5% of total carbonate phases. The sparry calcite has distinct crystal

boundaries from the surrounding micritic groundmass and displays rhombic crystal shape and lamellar twinning in plane-polarized light (Figs. 4E, F). Organic matters are preserved in some carbonate nodules, occur as dark brown-colored opaque matrix.

Stable isotopic screening of soil nodules with both micritic and sparry textures from 11 paleosol horizons (sampling position within stratigraphic sections shown in Table 2) reveals that the two types of calcite have very similar isotopic values within same nodules (Table 2). Average differences of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values between micritic and co-existing sparry calcite are 0.25‰ (range =0.01-0.63‰) and 0.24‰ (range =0.03-0.68‰), respectively, whereas the stable isotopic values of both micrite and sparry calcite can vary as much as ~3‰ (for $\delta^{18}\text{O}$) and ~2.5‰ (for $\delta^{13}\text{C}$) among different paleosol horizons (Table 2).

Figure 4. Photomicrographs of studied carbonate nodules. (A-D) Photomicrographs taken under polarized lights showing dominant micritic calcite groundmass with variable amounts of detrital quartz grains and soil color mottling feature. (E, F) Photomicrographs of carbonate cement showing crack-filling spar calcite (under polarized lights). Only the areas with homogeneous micritic textures were sampled for clumped isotopic analyses.

Table 2. Isotopic values of micro-sampled micritic and co-existing sparry calcite

Nodule Sample ID	Nodule 1		Nodule 2		Nodule 3		Nodule 4		Nodule 5		Nodule 6	
Sampling Position	15 m		29 m		84 m		112 m		150 m		162 m	
$\delta^{13}\text{C}, \delta^{18}\text{O}$ (‰, VPDB) micrite 1	-8.5	-10.5	-8.6	-10.9	-8.9	-10.1	-9.0	-10.5	-7.2	-10.0	-7.8	-9.6
$\delta^{13}\text{C}, \delta^{18}\text{O}$ (‰, VPDB) micrite 2	-8.7	-10.3	N/A	N/A	-8.7	-9.5	-9.2	-10.5	-7.6	-9.8	-7.0	-9.8
$\delta^{13}\text{C}, \delta^{18}\text{O}$ (‰, VPDB) spar	-9.0	-10.4	-8.3	-10.8	-9.3	-10.2	-9.1	-10.6	-7.8	-9.5	-7.6	-9.1
Note: Detailed sampling spots within each nodule are shown in Fig. A.1.												
Continue...												
Nodule 7	Nodule 8		Nodule 9		Nodule 10		Nodule 11					
215 m	234 m		250 m		265 m		300 m					
-8.3	-12.1	-9.0	-12.3	-9.0	-12.5	-9.2	-12.3	-8.6	-11.5			
-8.3	-12.0	-9.0	-12.0	-9.1	-12.3	-9.2	-12.2	-8.6	-11.7			
-8.4	-11.8	-9.4	-11.9	-8.8	-12.3	-9.2	-12.4	-8.6	-11.6			

Clumped-isotope thermometry and atmospheric $p\text{CO}_2$ reconstruction

A summary of isotopic results, including $\delta^{18}\text{O}_{\text{carbonate}}$, $\delta^{13}\text{C}_{\text{carbonate}}$, $\delta^{13}\text{C}_{\text{org}}$, Δ_{47} and calculated $\Delta_{47}\text{-T}$, $\delta^{18}\text{O}_{\text{water}}$ and $p\text{CO}_2$ values, are provided in Tables 3 and 4. Details of clumped isotope analysis, including standards measured over the period of sample analysis are listed in Supplementary Table 1. Δ_{47} values for all samples ranged from 0.609‰ to 0.677‰ (absolute reference scale), with external precision (1 standard error for replicates of a sample) varying from 0.009‰ to 0.018‰.

The range of Δ_{47} values translate to carbonate formation temperatures of 27-48 °C, with external errors in $\Delta_{47}\text{-temperature}$ ($\Delta_{47}\text{-T}$) values averaging $\pm \sim 4$ °C and ranging from $\pm \sim 2$ °C to $\pm \sim 5$ °C (Table 3). Carbonate $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values vary between -9.39 ‰ and -7.66 ‰ VPDB, and between -12.65 ‰ and -9.26 ‰ VPDB, respectively, with calculated $\delta^{18}\text{O}_{\text{water}}$ values range from ~ -3 ‰ to ~ -8 ‰ VSMOW (Table 3). Estimated $p\text{CO}_2$ levels range from ~ 500 to ~ 1200 ppmV with parameters used for calculations listed in Table 4. The combined soil temperature, $\delta^{18}\text{O}_{\text{water}}$ and $p\text{CO}_2$ reconstructions for the C-T sequence are characterized by several marked changes that are used to subdivide this succession into 3 intervals: an interval that generally corresponds to the Cenomanian (0-150 m height in the composite section), a transition interval that contains the C-T boundary (150-230 m height in the composite section), and an interval that contains the Turonian (230-320 m height in the composite section) (Fig. 5). Mean values and associated standard error (1 s.e.) of soil temperature, $\delta^{18}\text{O}_{\text{water}}$ and $p\text{CO}_2$ were calculated for each intervals (Tables 3 and 4).

Average $\Delta_{47}\text{-T}$ increases from 35 ± 1.2 °C in the Cenomanian to 46 ± 1.1 °C during the C-T transition, and then decrease slightly back to 44 ± 0.5 °C in the Turonian (Fig. 5A; Table 3). Average $\delta^{18}\text{O}_{\text{water}}$ values show a similar trend, increasing up-section from -6.7 ± 0.3 ‰

Table 3. Stable isotope data including clumped isotope results. Temperatures calculated using the inorganic calibration of Tripati et al. (2015), generated on the same instrument.

Time Interval	Sample ID	Section ID	N	Composite section height (m)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Δ_{47}^{a} (‰, ARF)	SE ^b (‰, analytical)	SE ^b (‰, external)	$\Delta_{47}\text{-T}^{\text{c}}$ (°C)	SE $\Delta_{47}\text{-T}$ (°C)	$\delta^{18}\text{O}_{\text{water}}^{\text{d}}$ (‰, VSMOW)	SE $\delta^{18}\text{O}_{\text{water}}$ (‰, VSMOW)
Cenomanian	MK638	A-A'	1	15	-8.38	-10.68	0.651	0.010	0.010	35	2.9	-6.5	0.6
	MK639	A-A'	1	29	-8.54	-10.89	0.677	0.018	0.018	27	4.7	-8.2	1.0
	MK6218	C-C'	1	35	-8.88	-11.25	0.669	0.010	0.010	30	2.7	-8.1	0.6
	MK6311	A-A'	1	59	-7.93	-11.45	0.642	0.017	0.017	37	5.0	-6.8	1.0
	MK6221 avg	C-C'	2	72	-8.51	-9.69	0.645	0.001	0.009	37	2.6	-5.2	0.5
	MK6221				-8.51	-9.60	0.644						
	MK6221				-8.51	-9.78	0.646						
	MK6312	A-A'	1	84	-9.04	-10.51	0.651	0.010	0.010	35	2.8	-6.4	0.6
	MK6220	C-C'	1	100	-8.36	-11.36	0.656	0.010	0.010	33	2.8	-7.5	0.6
	MK6315	A-A'	1	107	-8.34	-10.96	0.647	0.014	0.014	36	4.0	-6.6	0.8
	MK6222	C-C'	1	112	-8.47	-10.49	0.630	0.015	0.015	41	4.6	-5.2	0.8
	MK6316	A-A'	1	140	-8.15	-11.14	0.649	0.015	0.015	35	4.3	-6.9	0.8
Average all samples										35	±1.2	-6.7	±0.3
Cenomanian-Turonian transition													
	MK6223	C-C'	1	162	-7.66	-9.71	0.629	0.014	0.014	41	4.2	-4.3	0.8
	MK6226	C-C'	1	188	-8.48	-9.26	0.615	0.011	0.011	46	3.6	-3.1	0.6
	MK6317	A-A'	1	192	-8.29	-10.30	0.609	0.010	0.010	48	3.2	-3.8	0.6
	MK6318	A-A'	1	198	-8.70	-11.25	0.613	0.012	0.012	47	3.8	-5.0	0.7
	MK6319 avg	A-A'	2	215	-8.30	-12.07	0.611	0.004	0.009	47	2.9	-5.8	0.5
	MK6319				-8.31	-12.09	0.607						
	MK6319				-8.28	-12.04	0.615						
Average all samples										46	±1.1	-4.4	±0.5
Turonian													
	MK6227	C-C'	1	234	-9.27	-11.71	0.626	0.013	0.013	42	4.0	-6.2	0.7
	MK6321	A-A'	1	250	-9.17	-12.35	0.624	0.015	0.015	43	4.5	-6.8	0.8
	MK6322 avg	A-A'	2	265	-9.21	-12.40	0.617	0.003	0.009	45	2.9	-6.4	0.5
	MK6322				-9.18	-12.37	0.620						
	MK6322				-9.24	-12.43	0.615						
	MK6325	A-A'	1	283	-8.72	-12.65	0.621	0.012	0.012	44	3.7	-6.9	0.7
	MK6228 avg	C-C'	2	300	-8.57	-11.61	0.620	0.001	0.009	44	2.8	-5.8	0.5
	MK6228				-8.56	-11.57	0.621						
	MK6228				-8.57	-11.66	0.619						
	MK6230	C-C'	1	310	-9.39	-12.12	0.615	0.011	0.011	46	3.5	-6.0	0.6
Average all samples										44	±0.5	-6.4	±0.2

^a ARF: Absolute Reference Frame.

^b SE: the standard error of Δ_{47} (SE=analytical error for N=1; SE=SD(Δ_{47})/SQRT(N) for N≥2). When SE of a sample is less than the observed long term SE of lab standards (0.009), the long-term lab average SE of 0.009 is assigned.

^c Temperature is calculated using the inorganic calibration of Tripati et al. (2015): $T = \text{SQRT}(50500/(\Delta_{47}-0.1185)) - 273.15$.

^d Kim and O'Neil (1997): $10^3 \ln \alpha = 18030/(T+273.15) - 32.42$.

Table 4. Parameters used for $p\text{CO}_2$ estimation.

Time Interval	Sample ID	Section ID	Composite section height (m)	$\delta^{13}\text{C}_c^a$ (‰, VPDB)	Carbon isotopic fractionation factor					$\delta^{13}\text{C}_s^c$ (‰, VPDB)	soil-respired CO ² (ppm)			$\delta^{13}\text{C}_{org}^e$ (‰, VPDB)	$\delta^{13}\text{C}_a^f$ (‰, VPDB)	$p\text{CO}_2^g$ (ppm)			
					$\alpha_{\text{Calcite-CO}_2}^b$						Burial depth (km)	Depth to Bk horizon (cm)	S _(z) ^d (ppm)						
Cenomanian																			
	MK638	A-A'	15	-8.38	35	±	2.9	7.8	±	0.4	-16.17	0.7	21—27 ^h	1154—1342	-25.91	-6.58	718	±	125
	MK639	A-A'	29	-8.54	27	±	4.7	8.7	±	0.6	-17.22	0.7	21—27	1154—1342	ⁱ -26.24	-6.88	587	±	147
	MK6218	C-C'	35	-8.88	30	±	2.7	8.4	±	0.3	-17.30	0.7	27—32	1342—1531	-26.24	-6.88	649	±	108
	MK6311	A-A'	59	-7.93	37	±	5.0	7.5	±	0.6	-15.42	0.7	21—27	1154—1342	-24.86	-5.63	675	±	171
	MK6221	C-C'	72	-8.51	37	±	2.6	7.6	±	0.3	-16.11	0.7	21—27 ^h	1154—1342	-26.46	-7.08	849	±	137
	MK6312	A-A'	84	-9.04	35	±	2.8	7.8	±	0.4	-16.86	0.7	21—27	1154—1342	-25.18	-5.92	466	±	89
	MK6220	C-C'	100	-8.36	33	±	2.8	8.0	±	0.3	-16.34	0.7	16—21	965—1154	-25.20	-5.94	472	±	92
	MK6315	A-A'	107	-8.34	36	±	4.0	7.7	±	0.5	-16.02	0.7	21—27	1154—1342	-24.93	-5.69	570	±	129
	MK6222	C-C'	112	-8.47	41	±	4.6	7.0	±	0.6	-15.52	0.7	16—21	965—1154	-25.21	-5.94	613	±	153
	MK6316	A-A'	140	-8.15	35	±	1.2	7.7	±	0.5	-15.88	0.7	11—16	776—965	-25.81	-6.49	539	±	136
Average all samples																	614±37 ^j		
Cenomanian-Turonian transition																			
	MK6223	C-C'	162	-7.66	41	±	4.2	7.0	±	0.5	-14.66	0.6	16—21 ^h	965—1154	ⁱ -25.81	-6.49	907	±	203
	MK6226	C-C'	188	-8.29	46	±	3.6	6.5	±	0.4	-14.78	0.6	16—21	965—1154	-26.22	-6.86	969	±	196
	MK6317	A-A'	192	-8.29	48	±	3.2	6.3	±	0.4	-14.54	0.6	21—27	1140—1326	-26.16	-6.81	###	±	212
	MK6318	A-A'	198	-8.70	47	±	3.8	6.4	±	0.5	-15.09	0.6	16—21	965—1154	-25.50	-6.21	741	±	159
	MK6319	A-A'	215	-8.30	47	±	2.9	6.3	±	0.4	-14.64	0.6	11—16	776—965	-25.92	-6.59	764	±	152
Average all samples																	915±82 ^j		
Turonian																			
	MK6227	C-C'	234	-9.27	42	±	4.0	6.9	±	0.5	-16.17	0.5	11—16	776-965	-26.37	-7.00	569	±	136
	MK6321	A-A'	250	-9.17	43	±	4.5	6.8	±	0.6	-16.00	0.5	21—27	1140 - 1326	-26.05	-6.71	785	±	177
	MK6322	A-A'	265	-9.22	45	±	2.9	6.6	±	0.3	-15.78	0.5	26—31	1326 - 1511	-25.47	-6.18	806	±	132
	MK6325	A-A'	283	-8.72	44	±	3.7	6.7	±	0.5	-15.42	0.5	21—26	1140 - 1326	-25.75	-6.44	845	±	168
	MK6228	C-C'	300	-8.57	44	±	2.8	6.7	±	0.3	-15.25	0.5	26—31	1326 1511	-25.49	-6.20	942	±	150
	MK6230	C-C'	310	-9.39	46	±	3.5	6.5	±	0.4	-15.88	0.5	21—27	1140 - 1326	-25.56	-6.26	703	±	137
Average all samples																	775±52 ^j		

^a $\delta^{13}\text{C}_c$: $\delta^{13}\text{C}$ of soil carbonate.^b $\alpha_{\text{Calcite-}\text{CO}_2}$: Fractionation factor for carbon between calcite and CO_2 . Equation: $\alpha_{\text{Calcite-}\text{CO}_2} = 11.98 - 0.12 T$ (Romanek et al., 1992).^c $\delta^{13}\text{C}_s$: $\delta^{13}\text{C}$ of soil CO_2 .^d $S_{(z)}$: Soil CO_2 concentration. $S_{(z)}$ calculated using a revised Bk-depth correlation equation (Breecker and Retallack, 2014). Reported Bk-depth has been corrected for soil compaction using a standard compaction formula (Sheldon and Retallack, 2001)^e $\delta^{13}\text{C}_{\text{org}}$: $\delta^{13}\text{C}$ of soil organic matters.^f $\delta^{13}\text{C}_a$: $\delta^{13}\text{C}$ of atmospheric CO_2 .^g Errors of estimated $p\text{CO}_2$ reflect uncertainties in soil carbonate forming temperatures (Δ_{47-T}) and $S_{(z)}$ determination.^h Bk-depths were not be able to accurately determined due to erosion and the depth values were adopted from measurement made from the nearest paleosol profiles.ⁱ The coexisting $\delta^{13}\text{C}_{\text{org}}$ values were not be able to determined due to low organic carbon content, and their $\delta^{13}\text{C}_{\text{org}}$ values were adopted from $\delta^{13}\text{C}_{\text{org}}$ measured from the nearest paleosol layers.^j Standard error of average $p\text{CO}_2$ (1.s.e.).

in the Cenomanian to $-4.4 \pm 0.5\text{‰}$ at the C-T transition, and then decreasing back to $-6.4 \pm 0.2\text{‰}$ in the Turonian. Note that within the transition interval, individual $\delta^{18}\text{O}_{\text{water}}$ values fluctuate considerably, ranging from $\sim -3\text{‰}$ to $\sim -6\text{‰}$ (Fig. 5B; Table 3). The estimated atmospheric CO_2 concentrations from paleosols show an average value of 614 ± 37 ppm in the Cenomanian, followed by a large increase by $>40\%$ in the transition stage with the average value reaching 915 ± 82 ppm (Fig. 5C; Table 4). The rise in $p\text{CO}_2$ level is coincident with a large increase of average soil temperatures (from $\sim 35^\circ\text{C}$ to $\sim 46^\circ\text{C}$). During the Turonian, the average value of reconstructed atmospheric CO_2 decreases to 775 ± 52 ppm (Fig. 5C; Table 4).

Figure 5. Δ_{47} -temperature, $\delta^{18}\text{O}_{\text{water}}$, and atmospheric CO_2 ($p\text{CO}_2$) estimated from Markam paleosols. Note that the black and red diamonds represent samples from two different sections (A-A' and section C-C', respectively), with associated error bars. Samples were assigned as

either: Cenomanian (blue shading); Cenomanian-Turonian transition (red shading); or Turonian (green shading). **(A)** Δ_{47} -temperature estimates are based on the inorganic calibration of Tripathi et al. (2015), generated on the same instrument, plotted with simulated mean annual surface-air temperature (MAAT) of southeastern Tibet during mid-Cretaceous (Kump and Pollard, 2008). **(B)** Calculated oxygen isotope values of soil water ($\delta^{18}\text{O}_{\text{water}}$). **(C)** Estimated $p\text{CO}_2$ levels from Markam paleosol record, plotted with $p\text{CO}_2$ estimates from previous studies: 1. $p\text{CO}_2$ estimates from geochemical models (Bergman et al., 2004; Berner and Kothavala, 2001); 2. $p\text{CO}_2$ estimates from stomatal density (Barclay et al., 2010) and carbon isotope of marine sediments (Bice et al., 2006). **(D)** Cenomanian-Turonian (C-T) tropical sea surface temperature estimates from organic geochemical proxy (TEX_{86}) (Bornemann et al., 2008; Forster et al., 2007) and oxygen isotope ratios of planktonic foraminifera based on assumed values for water $\delta^{18}\text{O}$ (Bornemann et al., 2008).

Discussion and Implication

Evaluating potential diagenetic effects

Diagenetic alteration is an important issue that must be examined before meaningful interpretation of any isotopic results. The burial and thermal history of the Late Cretaceous Nanxin Formation derived from fluid inclusion and vitrinite reflectivity studies suggest a maximum burial temperature less than 80 °C, with vitrinite reflectivity values of organic matter less than 0.5 (Chen et al., 2010). This temperature estimate is consistent with geological observation that the Cretaceous red beds are poorly compacted rocks that form the youngest and shallowest sedimentary unit in the Markam region, except some very scattered, localized Eocene deposits (Pan et al., 2004). Henkes et al. (2014) suggest that Δ_{47} value of calcite should not be affected by solid-state C-O bond reordering if temperatures were below 100 °C over timescales of up to 10^8 years. Thus, the low burial temperatures provide the first evidence that carbonate nodules in the Nanxin Formation was unlikely experienced major solid-state resetting during

burial.

Petrographic observations show that the studied soil carbonates are dominated by micritic texture with minor sparry calcite cement filling veins and vugs (Fig. 4). Although materials used for clumped isotope analysis are sampled from micritic calcite, the large amounts of material (~10 mg for each analysis) necessary for clumped isotope measurements makes it likely that some calcite cement was probably included. Therefore, it is important to determine the diagenetic history of cement and evaluate its potential influence on clumped isotope analysis. Carbonate cement can form during early diagenetic, mesodiagenetic and late diagenetic phases (Choquette and Pray, 1970). Early diagenetic cement forms at the surface or shallow subsurface environments and therefore may primarily record the original surface climate and environmental conditions (e.g., surface temperature and $\delta^{18}\text{O}_{\text{water}}$). In contrast, mesodiagenetic and late diagenetic cements occur during deep burial and post-burial uplift and erosion stages, respectively, and are precipitated from fluids that may have different isotopic compositions from the original surface waters (Brand and Veizer, 1981). $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analysis of micrite and co-existing sparry cement of soil nodules from 11 horizons reveals that micritic calcite and sparry cement have very similar isotopic values (Table 2), with average differences of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values between micrite and co-existing sparry cement of 0.25‰ (range =0.01-0.63‰) and of 0.24‰ (range =0.03-0.68‰), respectively (Table 2), whereas isotopic values of nodules (for both micrite and sparry calcite) from different horizons can vary from 6.97‰ to 9.42‰ for $\delta^{13}\text{C}$ and from 9.12‰ to 12.48‰ for $\delta^{18}\text{O}$ (Table 2). The similar isotopic values between micrite and co-existing sparry cement suggests that carbonate cements in our studied paleosols were most likely formed during early diagenesis in phreatic zones, forming rhombic spars filling vugs and cracks within soil nodules (Tucker and Wright, 2009). Although mesodiagenesis during deep

burial could also homogenize isotopic values between micritic and sparry calcite phases, such process was unlikely to happen given relatively shallow burial depth as indicated by basin thermal history study (Chen et al., 2010), and relatively low Δ_{47} -T (~ 30 °C) from the deepest part of the studied section (Fig. 5A), indicating maximum burial temperature was unlikely exceeded 100 °C for a prolonged period of time (Henkes et al., 2014). The carbonate cement could also formed during late diagenesis after erosion and exhumation exposed the Cretaceous rocks to the surface. Regional tectonic study suggests that major deformation in the study area occurred in early Cenozoic during Indo-Asian collision (Studnicki-Gizbert et al., 2008). If the carbonate cements were precipitated during late diagenesis, we would expect to see: 1) difference of isotopic values between micrite and sparry cement as climate conditions during mid-Cretaceous and early Cenozoic were likely very different (e.g., Pagani et al., 2014; Zachos et al., 2008) and 2) similar isotopic values of carbonate cement among different soil horizons because late diagenesis would likely to happen synchronously throughout the C-T strata, given that the sampled sections are only 300 meters long. However, isotopic analysis of micrite and co-existing sparry cement shows that the two phases have very similar isotopic values within in the same soil horizons, whereas isotopic values of calcite cement varies significantly among different paleosol profiles ($>3\%$ for $\delta^{18}\text{O}$ and $>2\%$ for $\delta^{13}\text{C}$) (Table 2). These observations suggest that carbonate cements in our studied nodules were mainly formed during early diagenesis, and mesodiagenesis and late diagenesis have limited influence on soil nodules. Therefore, $\delta^{18}\text{O}$ and Δ_{47} -T values recorded by sparry cement primarily reflect near-surface environmental conditions similar to those recorded by micritic calcite.

Finally, we observed no systematic Δ_{47} -T versus depth relationship in the studied sections. The Δ_{47} -T trends we observed indicate that soil temperatures do not increase with burial

depth. On the contrary, it generally decreases with depth (Fig. 5A). For example, the three samples from the base of the Cenomanian strata (MK6218, MK638, MK629) exhibit Δ_{47} -T ~30-35 °C, cooler than all the other values observed above, whereas, the highest Δ_{47} -T occurs in the middle part of the section. It is also unlikely that the observed temperature trend was due to modification by hydrothermal fluids given the complete absence of major intrusive mineral veins or nearby igneous bodies, relatively simple structure of the study area, and poorly-compacted nature of the Cretaceous sedimentary rocks.

In summary, burial history, thermal maturity, petrography, and isotopic screening all indicate that our carbonate samples did not experience deep burial and associated post-depositional diagenesis, which favors the preservation of primary climate signals of soil formation without solid-state C-O bond reordering.

Evaluation of Δ_{47} -derived soil temperatures

The calculated Δ_{47} -T values show that more than half of the carbonate nodules exhibit growth temperatures in excess of 40 °C (Table 3) that, if interpreted as a primary climate signal, indicate very high ground temperatures in the tropics at a site located at ~15° N during the mid-Cretaceous. Studies have shown that in most cases soil carbonates form under temperatures considerably warmer than mean annual temperature (e.g., Passey et al., 2010; Quade et al., 2013), concentrated in the summer months of the year during drought seasons, when soil temperature exceeds mean annual temperature by more than 15 °C (e.g., Breecker et al., 2009). If representative, that means only hot summer temperatures will be recorded in soil carbonates, especially in arid regions where soil dewatering commonly happens in the summer months (Breecker et al., 2009; Hough et al., 2014). Clumped isotope study of modern soil carbonates

from various places and climate regimes reveal that 40 °C is the upper limit for most modern soil carbonates formed at soil depth >25 cm (Quade et al., 2013), and $\Delta_{47-T} > 40$ °C indicates either substantially warmer climate regimes compared to the present, or diagenetic signals produced by resetting during deep burial (Quade et al., 2013), the latter of which we have been able to preclude based on assessment of the burial history and observations of the sedimentology and geochemistry, as discussed above.

Cretaceous climate reconstructions indicate southeastern Tibet was located within a widespread evaporite zone near the equatorial region (at ~15° N) during the mid-Cretaceous (Hay and Floegel, 2012). The evaporite zone was characterized by increased aridity and evaporation during the C-T ‘super-greenhouse’ interval (Hay and Floegel, 2012), consistent with our isotopic data showing high soil carbonate forming temperatures and enriched $\delta^{18}\text{O}_{\text{water}}$ values indicating strong soil evaporation at the C-T boundary (Fig. 5B). In the field, the inferred aridity is supported by relatively shallow Bk-depth even after burial decompaction (decompacted Bk-depth ranges from 10 to 31 cm; average =21 cm) (Table 4), indicating mean annual precipitation of ~350 mm (Retallack, 2005). Climate model (i.e., GENESIS v. 3.0) simulation shows that mean annual surface-air temperature (MAAT) in our study area had reached 35-40 °C during mid-Cretaceous (~100 Ma) (Kump and Pollard, 2008), suggesting that the warm-month surface-air temperature (WAAT) may have reached well above 40 °C. Quade et al. (2013) show that soil carbonate typically forms in warm months when soil temperature exceeds MAAT by 10-15 °C due to timing of calcite precipitation based on study of Holocene soils. If representative, that means soil nodules in our study site should record warm-month temperatures of ~45-55 °C, 10-15 °C higher than the MAAT, though the more equable mid-Cretaceous climate (e.g., Pagani et al., 2014) was probably characterized by reduced seasonalities (i.e., smaller difference between

MAAT and WAAT). Furthermore, direct solar heating can elevate deep soil temperatures by more than 5 °C, especially in arid settings where plant shadings are rare (Quade et al., 2013). We also note that even conservative estimate of SSTs during the Turonian in the tropical Atlantic Ocean can reach up to 36 °C (Forster et al., 2007), whereas estimates derived from oxygen isotope and Mg/Ca ratios from well-preserved planktonic foraminifera suggest the tropical Atlantic Ocean SSTs may reach up to 40 °C (Bice et al., 2006). If these high SSTs are representative, then it is very likely that the tropical ground temperature was even higher especially during summer season when solar radiation is at its strongest.

Therefore, we argue that the observed high Δ_{47} -T values (>40 °C) from the late Cenomanian and Turonian paleosol nodules are plausible because these soil nodules formed in an arid environmental setting during one of the warmest periods in Cretaceous, with limited influenced from late-stage diagenesis and solid-state reordering of soil carbonate.

Reconstruction of Cenomanian-Turonian ground temperatures and atmospheric $p\text{CO}_2$ level

Reconstructed Δ_{47} -soil temperature record shows that the Cenomanian interval is the ‘coolest’ part of our studied section and contains some of the lowest observed soil temperatures from the C-T interval (e.g., 27 °C at 29 m and 30 °C at 35 m composite section heights) (Fig. 5A). Within the Cenomanian interval, several additional temperature trends can be observed. One sample from the lowest paleosol horizon displays Δ_{47} -T value of 35 ± 3 °C. The value decreases up-section to ~ 27 -30 °C, and then increases back to ~ 37 °C by the middle of the Cenomanian section (Fig. 5A). Onset of the C-T transition is marked by a rise in soil temperatures from ~ 37 °C to >45 °C; this rise coincides with a 2‰ increase of $\delta^{18}\text{O}_{\text{water}}$ values from -6 ‰ to -4 ‰ (Fig. 5B). Average soil temperature then reached a thermal maximum at $46 \pm$

1.1 °C which was maintained throughout the transition interval. A temperature drop of ~5 °C in the lower Turonian is observed in both sections and marks the termination of the C-T transition thermal maximum (Fig. 5A). Although the Turonian was a little cooler than the C-T transition, average soil temperature is maintained at 44 ± 0.5 °C, consistent with the reconstruction that indicates Turonian climate was remarkably hot and stable (MacLeod et al., 2013).

Average values of $\delta^{18}\text{O}_{\text{water}}$ are relatively constant throughout the section at ~6‰, except at the C-T transition where a large positive excursion is observed. Average $\delta^{18}\text{O}_{\text{water}}$ values increase dramatically from ~-6‰ to ~-4‰ (Fig. 5B). The positive excursion is coincident with the onset of the C-T thermal maximum. A cross-plot of $\Delta_{47}\text{-T}$ and $\delta^{18}\text{O}_{\text{water}}$ also reveals a broad correlation between the two factors (Fig. 6A), suggesting enhanced soil evaporation associated with rising soil temperatures. The very high Δ_{47} -temperatures (average = 46 ± 1.1 °C) and the positive excursion of $\delta^{18}\text{O}_{\text{water}}$ at the C-T transition interval are likely related to two factors: 1) extreme high surface temperature associated with the C-T thermal maximum; and 2) intense soil evaporation associated with increased aridity. In comparison, pedogenic carbonate from the Turonian interval also record relatively high formation temperature (average = 44 ± 0.5 °C), but the $\delta^{18}\text{O}_{\text{water}}$ values are much less enriched as compared to the samples from the C-T transition, which indicates reduced soil evaporation and thus reduced aridity during Turonian.

The Cenomanian to early Turonian warming trend evident in this low-latitude (~15° N) terrestrial soil temperature record (Fig. 5A) can be correlated to contemporaneous paleoclimate reconstructions. Tropical sea surface temperature (SST) estimates derived from the TEX₈₆ proxy show a gradual warming trend during the Cenomanian, followed by a thermal maximum at the C-T boundary (Forster et al., 2007) (Fig. 5D), similar to what we observed. The same trend can also be inferred from $\delta^{18}\text{O}$ values of well-preserved marine fossils from mid- and high-latitude

regions (Huber et al., 1995; Huber et al., 2002; Jenkyns et al., 2004), which both show a stepwise temperature rise through the Cenomanian, succeeded by a temperature maximum at the C-T boundary.

Our reconstructed $p\text{CO}_2$ levels are ~500-1200 ppmV, lower than the values derived from geochemical models such as GEOCARB III (Bernier and Kothavala, 2001) and COPSE (Bergman et al., 2004) (Fig. 5C). The trend of increasing $p\text{CO}_2$ level at the C-T transition seen in our reconstruction is also observed in the COPSE model (Bergman et al., 2004), although it is not seen in the GEOCARB III model (Bernier and Kothavala, 2001) results, which is likely due to its coarse temporal resolution (~10 Ma). Our estimated $p\text{CO}_2$ values generally overlap estimates derived from marine organic matter (Bice et al., 2006), and the short-term record estimated using stomata density (Barclay et al., 2010) (Fig. 5C). Barclay et al. (2010) observed a 20% rise of atmospheric CO_2 concentrations at the C-T boundary, succeeded by a 26% decrease of $p\text{CO}_2$ levels associated with extensive carbon burial during OAE 2 (~93.5 Ma). In our paleosol record, a position excursion of $p\text{CO}_2$ levels at the C-T boundary is also observed, with $p\text{CO}_2$ levels increased to >1000 ppmV at the C-T transition, followed by a ~20% drop right after the transition (Fig. 5C). The decrease of $p\text{CO}_2$ concentration may associated with extensive carbon sequestration during OAE 2 (e.g., Barclay et al., 2010), though a better geochronological constraint is needed (e.g., U-Pb dating of pedogenic carbonate) (Rasbury and Cole, 2009) to further confirm the exact position of the C-T boundary in our studied sections.

Implication for mid-Cretaceous regional climate sensitive over tropics

Proxy data have been used to determine climate sensitivity (ΔT_{2X}) (Hansen et al., 2008;

Royer, 2010), which is the equilibrium temperature response to a doubling of $p\text{CO}_2$. Estimates from proxy data that consider fast feedback processes average $\sim 3^\circ\text{C}$ (Solomon, 2007). In contrast, higher estimates of $\sim 6^\circ\text{C}$ are estimated if we consider “deep-time” records where the response is also impacted by slow feedbacks in the Earth system such as changes in surface albedo associated with vegetation migration and landscape evolution, ice sheet retreat, and the release of greenhouse gases from ocean sediments, tundra, and soils (Joos, 2015; Kump and Pollard, 2008; Stocker et al., 2013; Wang et al., 2014).

A cross-plot of our reconstructed mid-Cretaceous surface temperatures and $p\text{CO}_2$ levels reveals a broad positive correlation ($r^2 = 0.41$) (Fig. 6B; solid black line). The increase of $p\text{CO}_2$ levels from ~ 500 to ~ 1000 ppmV is accompanied by an average soil temperature rise from ~ 35 to $\sim 45^\circ\text{C}$, which corresponds to a $\Delta T_{2X} \sim 10^\circ\text{C}$ (Fig. 6B). We note a similarly high estimate of climate sensitivity derived using our $p\text{CO}_2$ estimates and published tropical Cenomanian and Turonian SST dataset (Bornemann et al., 2008; Forster et al., 2007). The increase of average tropical SSTs from ~ 31 to $\sim 35^\circ\text{C}$ (Bornemann et al., 2008; Forster et al., 2007) is coincident with a rise of average $p\text{CO}_2$ levels from 614 ± 37 to 775 ± 52 ppmV (Fig. 6B, dashed blue line), similar to the best-fit correlation between temperature and $p\text{CO}_2$ values determined from the paleosol record. Our estimated regional climate sensitivity in mid-Cretaceous is significantly higher than modern day values of $\sim 1.5\text{--}4.5^\circ\text{C}$ per $p\text{CO}_2$ doubling as suggested by the Intergovernmental Panel on Climate Change (Solomon, 2007), consistent with the view that Earth’s climate response to increasing atmospheric CO_2 concentration was significantly enhanced in greenhouse world (e.g., Pagani et al., 2014; Royer, 2010).

Substantially higher climate sensitivities in mid-Cretaceous time may reflect carbon cycle feedbacks that operate under ‘super-greenhouse’ conditions, in part arising from the impacts of

enhanced water transport to the high-latitudes. The combined effects of moisture and temperature on terrestrial carbon cycling in high-latitude and tropical regions, and changes in fire regimes, may result in high carbon emissions (Joos, 2015; Landry et al., 2015; Stocker et al., 2013). The production of methane in sediments, and the transfer of carbon from the ocean to the atmosphere, might also be greatly accelerated under ‘super-greenhouse’ conditions (Kort et al., 2012).

Figure 6. Constraints on regional climate sensitivity for the mid-Cretaceous. Black and red dots indicate individual samples from sections A-A' and C-C', respectively. Regression lines are generated based on all paleosol samples. **(A)** Cross-plot of Δ_{47} -T and $\delta^{18}\text{O}_{\text{water}}$. Solid black line represents best-fit correlation between Δ_{47} -T and $p\text{CO}_2$. Points plotted above the line likely represent effects of soil evaporation. **(B)** Cross-plot of Δ_{47} -temperatures (Δ_{47} -T) and estimated $p\text{CO}_2$ levels. Solid black line represents best-fit correlation between Δ_{47} -T and $p\text{CO}_2$, which corresponds to an average climate sensitivity (ΔT_{2X}) of $\sim 10^\circ\text{C}$ per CO_2 doubling. Dashed black lines mark the 95% confidence Intervals. Dashed blue line is based on the correlation between tropical sea surface temperature (Bornemann et al., 2008; Forster et al., 2007) and $p\text{CO}_2$ derived from Markam paleosol record.

References

- Arens, N. C., Hope Jahren, A., and Amundson, R., 2000, Can C3 plants faithfully record the carbon isotopic composition of atmospheric carbon dioxide?: *Paleobiology*, v. 26, no. 1, p. 137-164.
- Arthur, M., Dean, W., and Schlanger, S., 1985, Variations in the global carbon cycle during the Cretaceous related to climate, volcanism, and changes in atmospheric CO₂: *The Carbon Cycle and Atmospheric CO₂: Natural Variations Archean to Present*, p. 504-529.
- Barclay, R. S., McElwain, J. C., and Sageman, B. B., 2010, Carbon sequestration activated by a volcanic CO₂ pulse during Ocean Anoxic Event 2: *Nature Geoscience*, v. 3, no. 3, p. 205-208.
- Bergman, N. M., Lenton, T. M., and Watson, A. J., 2004, COPSE: a new model of biogeochemical cycling over Phanerozoic time: *American Journal of Science*, v. 304, no. 5, p. 397-437.
- Berner, R. A., and Kothavala, Z., 2001, GEOCARB III: a revised model of atmospheric CO₂ over Phanerozoic time: *American Journal of Science*, v. 301, no. 2, p. 182-204.
- Bice, K., Huber, B., and Norris, R., 2003, Extreme polar warmth during the Cretaceous greenhouse? The paradox of the late Turonian $\delta^{18}\text{O}$ record at DSDP Site 511: *Paleoceanography*, v. 18.
- Bice, K. L., Birgel, D., Meyers, P. A., Dahl, K. A., Hinrichs, K. U., and Norris, R. D., 2006, A multiple proxy and model study of Cretaceous upper ocean temperatures and atmospheric CO₂ concentrations: *Paleoceanography*, v. 21, no. 2.
- Blakey, R. C., 2008, Gondwana paleogeography from assembly to breakup—A 500 my odyssey: *Geological Society of America Special Papers*, v. 441, p. 1-28.
- Bornemann, A., Norris, R. D., Friedrich, O., Beckmann, B., Schouten, S., Damsté, J. S. S., Vogel, J., Hofmann, P., and Wagner, T., 2008, Isotopic evidence for glaciation during the Cretaceous supergreenhouse: *Science*, v. 319, no. 5860, p. 189-192.
- Brand, U., and Veizer, J., 1981, Chemical diagenesis of a multicomponent carbonate system-2: stable isotopes: *Journal of Sedimentary Research*, v. 51, no. 3.
- Breecker, D., and Retallack, G., 2014, Refining the pedogenic carbonate atmospheric CO₂ proxy and application to Miocene CO₂: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 406, p. 1-8.
- Breecker, D., Sharp, Z., and McFadden, L., 2010, Atmospheric CO₂ concentrations during ancient greenhouse climates were similar to those predicted for AD 2100: *Proceedings of the National Academy of Sciences*, v. 107, no. 2, p. 576-580.
- Breecker, D., Sharp, Z., and McFadden, L. D., 2009, Seasonal bias in the formation and stable isotopic composition of pedogenic carbonate in modern soils from central New Mexico, USA: *Geological Society of America Bulletin*, v. 121, no. 3-4, p. 630-640.
- Cerling, T., 1999, Stable carbon isotopes in palaeosol carbonates: *Palaeoweathering, palaeosurfaces and related continental deposits*, p. 43-60.
- Chamberlain, C. P., Wan, X., Graham, S. A., Carroll, A. R., Doebbert, A. C., Sageman, B. B., Blisniuk, P., Kent-Corson, M. L., Wang, Z., and Chengshan, W., 2013, Stable isotopic evidence for climate and basin evolution of the Late Cretaceous Songliao basin, China: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 385, p. 106-124.

- Chen, H. H., Wu, Y. & Xiao, Q. J., 2010, Fluid Inclusion Evidence of Paleo-Oil Reservoirs in Changdu Basin, Tibet: *Acta Geologica Sinica*, v. 84, no. 10, p. 1457-1469.
- Choquette, P. W., and Pray, L. C., 1970, Geologic nomenclature and classification of porosity in sedimentary carbonates: *AAPG bulletin*, v. 54, no. 2, p. 207-250.
- Cotton, J. M., and Sheldon, N. D., 2012, New constraints on using paleosols to reconstruct atmospheric pCO₂: *Geological Society of America Bulletin*, v. 124, no. 9-10, p. 1411-1423.
- Defliese, W. F., Hren, M. T., and Lohmann, K. C., 2015, Compositional and temperature effects of phosphoric acid fractionation on $\Delta 47$ analysis and implications for discrepant calibrations: *Chemical Geology*, v. 396, p. 51-60.
- Dennis, K., Cochran, J., Landman, N., and Schrag, D., 2013, The climate of the Late Cretaceous: New insights from the application of the carbonate clumped isotope thermometer to Western Interior Seaway macrofossil: *Earth and Planetary Science Letters*, v. 362, p. 51-65.
- Dennis, K. J., Affek, H. P., Passey, B. H., Schrag, D. P., and Eiler, J. M., 2011, Defining an absolute reference frame for ‘clumped’ isotope studies of CO₂: *Geochimica et Cosmochimica Acta*, v. 75, no. 22, p. 7117-7131.
- Du, D., Luo, J., and Li, X., 1997, Sedimentary evolution and palaeogeography of the Qamdo Block in Xizang: *Lithofacies Palaeogeogr*, v. 17, no. 4, p. 1-17.
- Eagle, R. A., Risi, C., Mitchell, J. L., Eiler, J. M., Seibt, U., Neelin, J. D., Li, G., and Tripathi, A. K., 2013, High regional climate sensitivity over continental China constrained by glacial-recent changes in temperature and the hydrological cycle: *Proceedings of the National Academy of Sciences*, v. 110, no. 22, p. 8813-8818.
- Eiler, J. M., 2007, “Clumped-isotope” geochemistry—The study of naturally-occurring, multiply-substituted isotopologues: *Earth and Planetary Science Letters*, v. 262, no. 3, p. 309-327.
- Eiler, J. M., 2011, Paleoclimate reconstruction using carbonate clumped isotope thermometry: *Quaternary Science Reviews*, v. 30, no. 25, p. 3575-3588.
- Eiler, J. M., and Schauble, E., 2004, 18O 13C 16O in Earth’s atmosphere: *Geochimica et Cosmochimica Acta*, v. 68, no. 23, p. 4767-4777.
- Ekart, D. D., Cerling, T. E., Montanez, I. P., and Tabor, N. J., 1999, A 400 million year carbon isotope record of pedogenic carbonate: implications for paleoatmospheric carbon dioxide: *American Journal of Science*, v. 299, no. 10, p. 805-827.
- Fan, M., Hough, B. G., and Passey, B. H., 2014, Middle to late Cenozoic cooling and high topography in the central Rocky Mountains: Constraints from clumped isotope geochemistry: *Earth and Planetary Science Letters*, v. 408, p. 35-47.
- Fletcher, B. J., Brentnall, S. J., Anderson, C. W., Berner, R. A., and Beerling, D. J., 2008, Atmospheric carbon dioxide linked with Mesozoic and early Cenozoic climate change: *Nature Geoscience*, v. 1, no. 1, p. 43-48.
- Forster, A., Schouten, S., Baas, M., and Damsté, J. S. S., 2007, Mid-Cretaceous (Albian–Santonian) sea surface temperature record of the tropical Atlantic Ocean: *Geology*, v. 35, no. 10, p. 919-922.
- Freeman, K. H., and Hayes, J., 1992, Fractionation of carbon isotopes by phytoplankton and estimates of ancient CO₂ levels: *Global Biogeochemical Cycles*, v. 6, no. 2, p. 185-198.
- Garziane, C. N., Auerbach, D. J., Smith, J. J.-S., Rosario, J. J., Passey, B. H., Jordan, T. E., and Eiler, J. M., 2014, Clumped isotope evidence for diachronous surface cooling of the

- Altiplano and pulsed surface uplift of the Central Andes: *Earth and Planetary Science Letters*, v. 393, p. 173-181.
- Garzione, C. N., Dettman, D. L., and Horton, B. K., 2004, Carbonate oxygen isotope paleoaltimetry: evaluating the effect of diagenesis on paleoelevation estimates for the Tibetan plateau: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 212, no. 1, p. 119-140.
- Ghosh, P., Adkins, J., Affek, H., Balta, B., Guo, W., Schauble, E. A., Schrag, D., and Eiler, J. M., 2006, ^{13}C – ^{18}O bonds in carbonate minerals: A new kind of paleothermometer: *Geochimica et Cosmochimica Acta*, v. 70, no. 6, p. 1439-1456.
- Gradstein, F. M., Ogg, J. G., and Smith, A. G., 2004, *A geologic time scale 2004*, Cambridge University Press.
- Graham, S. A., Chamberlain, C. P., Yue, Y., Ritts, B. D., Hanson, A. D., Horton, T. W., Waldbauer, J. R., Poage, M. A., and Feng, X., 2005, Stable isotope records of Cenozoic climate and topography, Tibetan plateau and Tarim basin: *American Journal of Science*, v. 305, no. 2, p. 101-118.
- GSX (Geological Survey of Xizang), 2007, *Regional Geology Report of Basu, Gonji, Ranwu and Markam counties*.
- Hansen, J., Sato, M., Kharecha, P., Beerling, D., Berner, R., Masson-Delmotte, V., Pagani, M., Raymo, M., Royer, D. L., and Zachos, J. C., 2008, Target atmospheric CO_2 : Where should humanity aim?: arXiv preprint arXiv:0804.1126.
- Harries, P. J., and Little, C. T., 1999, The early Toarcian (Early Jurassic) and the Cenomanian–Turonian (Late Cretaceous) mass extinctions: similarities and contrasts: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 154, no. 1, p. 39-66.
- Hasegawa, T., 2003, Cretaceous terrestrial paleoenvironments of northeastern Asia suggested from carbon isotope stratigraphy: Increased atmospheric pCO_2 -induced climate: *Journal of Asian Earth Sciences*, v. 21, no. 8, p. 849-859.
- Hay, W. W., and Floegel, S., 2012, New thoughts about the Cretaceous climate and oceans: *Earth-Science Reviews*, v. 115, no. 4, p. 262-272.
- Henkes, G. A., Passey, B. H., Grossman, E. L., Shenton, B. J., Pérez-Huerta, A., and Yancey, T. E., 2014, Temperature limits for preservation of primary calcite clumped isotope paleotemperatures: *Geochimica et cosmochimica acta*, v. 139, p. 362-382.
- Henkes, G. A., Passey, B. H., Wanamaker, A. D., Grossman, E. L., Ambrose, W. G., and Carroll, M. L., 2013, Carbonate clumped isotope compositions of modern marine mollusk and brachiopod shells: *Geochimica et Cosmochimica Acta*, v. 106, p. 307-325.
- Hough, B. G., Fan, M., and Passey, B. H., 2014, Calibration of the clumped isotope geothermometer in soil carbonate in Wyoming and Nebraska, USA: Implications for paleoelevation and paleoclimate reconstruction: *Earth and Planetary Science Letters*, v. 391, p. 110-120.
- Huang, K., Opdyke, N. D., Li, J., and Peng, X., 1992, Paleomagnetism of Cretaceous rocks from eastern Qiangtang terrane of Tibet: *Journal of Geophysical Research: Solid Earth* (1978–2012), v. 97, no. B2, p. 1789-1799.
- Huber, B. T., Hodell, D. A., and Hamilton, C. P., 1995, Middle–Late Cretaceous climate of the southern high latitudes: stable isotopic evidence for minimal equator-to-pole thermal gradients: *Geological Society of America Bulletin*, v. 107, no. 10, p. 1164-1191.
- Huber, B. T., Norris, R. D., and MacLeod, K. G., 2002, Deep-sea paleotemperature record of extreme warmth during the Cretaceous: *Geology*, v. 30, no. 2, p. 123-126.

- Huntington, K. W., Saylor, J., Quade, J., and Hudson, A. M., 2015, High late Miocene–Pliocene elevation of the Zhada Basin, southwestern Tibetan Plateau, from carbonate clumped isotope thermometry: *Geological Society of America Bulletin*, v. 127, no. 1-2, p. 181-199.
- Jarvis, I., Lignum, J. S., Gröcke, D. R., Jenkyns, H. C., and Pearce, M. A., 2011, Black shale deposition, atmospheric CO₂ drawdown, and cooling during the Cenomanian-Turonian Oceanic Anoxic Event: *Paleoceanography*, v. 26, no. 3.
- Jenkyns, H. C., 2010, Geochemistry of oceanic anoxic events: *Geochemistry, Geophysics, Geosystems*, v. 11, no. 3.
- Jenkyns, H. C., Forster, A., Schouten, S., and Damsté, J. S. S., 2004, High temperatures in the late Cretaceous Arctic Ocean: *Nature*, v. 432, no. 7019, p. 888-892.
- Joos, F., 2015, Global warming: Growing feedback from ocean carbon to climate: *Nature*, v. 522, no. 7556, p. 295-296.
- Kim, S.-T., and O'Neil, J. R., 1997, Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates: *Geochimica et Cosmochimica Acta*, v. 61, no. 16, p. 3461-3475.
- Kort, E., Wofsy, S., Daube, B., Diao, M., Elkins, J., Gao, R., Hintsa, E., Hurst, D., Jimenez, R., and Moore, F., 2012, Atmospheric observations of Arctic Ocean methane emissions up to 82 [deg] north: *Nature Geoscience*, v. 5, no. 5, p. 318-321.
- Kump, L. R., and Pollard, D., 2008, Amplification of Cretaceous warmth by biological cloud feedbacks: *Science*, v. 320, no. 5873, p. 195-195.
- Landry, J.-S., Matthews, H. D., and Ramankutty, N., 2015, A global assessment of the carbon cycle and temperature responses to major changes in future fire regime: *Climatic Change*, p. 1-14.
- Lynch-Stieglitz, J., Stocker, T. F., Broecker, W. S., and Fairbanks, R. G., 1995, The influence of air-sea exchange on the isotopic composition of oceanic carbon: Observations and modeling: *Global Biogeochemical Cycles*, v. 9, no. 4, p. 653-665.
- MacLeod, K. G., Huber, B. T., Berrocoso, Á. J., and Wendler, I., 2013, A stable and hot Turonian without glacial $\delta^{18}\text{O}$ excursions is indicated by exquisitely preserved Tanzanian foraminifera: *Geology*, v. 41, no. 10, p. 1083-1086.
- Montanez, I. P., 2013, Modern soil system constraints on reconstructing deep-time atmospheric CO₂: *Geochimica et Cosmochimica Acta*, v. 101, p. 57-75.
- Norris, R. D., Bice, K. L., Magno, E. A., and Wilson, P. A., 2002, Jiggling the tropical thermostat in the Cretaceous hothouse: *Geology*, v. 30, no. 4, p. 299-302.
- Pagani, M., Huber, M., and Sageman, B., 2014, Greenhouse climates, *Treatise on Geochemistry*, Volume 6, Elsevier Oxford, p. 281-304.
- Pan, G. T., Ding, J., Yao, D. S., and Wang, L. Q., 2004, Geological map of Qinghai-Xiang (Tibet) Plateau and adjacent areas: Chengdu Institute of Geology and Mineral Resources, China Geological Survey, scale 1:1,500,000, 6 sheets.
- Passey, B. H., Levin, N. E., Cerling, T. E., Brown, F. H., and Eiler, J. M., 2010, High-temperature environments of human evolution in East Africa based on bond ordering in paleosol carbonates: *Proceedings of the National Academy of Sciences*, v. 107, no. 25, p. 11245-11249.
- Petryshyn, V., Lim, D., Laval, B., Brady, A., Slater, G., and Tripathi, A., 2015, Reconstruction of limnology and microbialite formation conditions from carbonate clumped isotope thermometry: *Geobiology*, v. 13, no. 1, p. 53-67.

- Price, G. D., and Passey, B. H., 2013, Dynamic polar climates in a greenhouse world: Evidence from clumped isotope thermometry of Early Cretaceous belemnites: *Geology*, v. 41, no. 8, p. 923-926.
- Quade, J., Eiler, J., Daeron, M., and Achyuthan, H., 2013, The clumped isotope geothermometer in soil and paleosol carbonate: *Geochimica et Cosmochimica Acta*, v. 105, p. 92-107.
- Rasbury, E. T., and Cole, J. M., 2009, Directly dating geologic events: U-Pb dating of carbonates: *Reviews of Geophysics*, v. 47, no. 3.
- Retallack, G. J., 2005, Pedogenic carbonate proxies for amount and seasonality of precipitation in paleosols: *Geology*, v. 33, no. 4, p. 333-336.
- Retallack, G. J., 2009, Refining a pedogenic-carbonate CO₂ paleobarometer to quantify a middle Miocene greenhouse spike: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 281, no. 1, p. 57-65.
- Romanek, C. S., Grossman, E. L., and Morse, J. W., 1992, Carbon isotopic fractionation in synthetic aragonite and calcite: effects of temperature and precipitation rate: *Geochimica et Cosmochimica Acta*, v. 56, no. 1, p. 419-430.
- Royer, D. L., 2010, Fossil soils constrain ancient climate sensitivity: *Proceedings of the National Academy of Sciences*, v. 107, no. 2, p. 517-518.
- Schouten, S., Hopmans, E. C., Forster, A., van Breugel, Y., Kuypers, M. M., and Damsté, J. S. S., 2003, Extremely high sea-surface temperatures at low latitudes during the middle Cretaceous as revealed by archaeal membrane lipids: *Geology*, v. 31, no. 12, p. 1069-1072.
- Sepkoski Jr, J., 1986, Phanerozoic overview of mass extinction, *Patterns and Processes in the History of Life*, Springer, p. 277-295.
- Shang, F., 2014, Detrital zircon geochronology from the Mesozoic Qamdo (Changdu) basin, southeastern Tibet: Implications for the Paleo- and Meso-Tethys evolution, in *2014 GSA Annual Meeting in Vancouver, British Columbia*.
- Sheldon, N. D., and Retallack, G. J., 2001, Equation for compaction of paleosols due to burial: *Geology*, v. 29, no. 3, p. 247-250.
- Sheldon, N. D., and Tabor, N. J., 2009, Quantitative paleoenvironmental and paleoclimatic reconstruction using paleosols: *Earth-Science Reviews*, v. 95, no. 1, p. 1-52.
- Snell, K. E., Koch, P. L., Druschke, P., Foreman, B. Z., and Eiler, J. M., 2014, High elevation of the 'Nevadaplano' during the Late Cretaceous: *Earth and Planetary Science Letters*, v. 386, p. 52-63.
- Snell, K. E., Thrasher, B. L., Eiler, J. M., Koch, P. L., Sloan, L. C., and Tabor, N. J., 2013, Hot summers in the Bighorn Basin during the early Paleogene: *Geology*, v. 41, no. 1, p. 55-58.
- Solomon, S., 2007, *Climate change 2007-the physical science basis: Working group I contribution to the fourth assessment report of the IPCC*, Cambridge University Press.
- Stocker, B. D., Roth, R., Joos, F., Spahni, R., Steinacher, M., Zaehle, S., Bouwman, L., and Prentice, I. C., 2013, Multiple greenhouse-gas feedbacks from the land biosphere under future climate change scenarios: *Nature Climate Change*, v. 3, no. 7, p. 666-672.
- Studnicki-Gizbert, C., Burchfiel, B., Li, Z., and Chen, Z., 2008, Early Tertiary Gonjo basin, eastern Tibet: Sedimentary and structural record of the early history of India-Asia collision: *Geosphere*, v. 4, no. 4, p. 713-735.

- Suarez, M., Ludvigson, G., Gonzalez, L., and You, H., Linking Cretaceous Continental Climate Change and Ocean Anoxic Events: Evidence from the Xiagou Formation of Gansu Province, China, *in* Proceedings AGU Fall Meeting Abstracts 2012, Volume 1, p. 01.
- Suarez, M. B., and Passey, B. H., 2014, Assessment of the clumped isotope composition of fossil bone carbonate as a recorder of subsurface temperatures: *Geochimica et Cosmochimica Acta*, v. 140, p. 142-159.
- Suarez, M. B., Passey, B. H., and Kaakinen, A., 2011, Paleosol carbonate multiple isotopologue signature of active East Asian summer monsoons during the late Miocene and Pliocene: *Geology*, v. 39, no. 12, p. 1151-1154.
- Swanson, H. E., Tatge, E., and Fuyat, R. K., 1953, Standard X-ray diffraction powder patterns.
- Tripathi, A. K., Eagle, R. A., Thiagarajan, N., Gagnon, A. C., Bauch, H., Halloran, P. R., and Eiler, J. M., 2010, $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ isotope signatures and ‘clumped isotope’ thermometry in foraminifera and coccoliths: *Geochimica et Cosmochimica Acta*, v. 74, no. 20, p. 5697-5717.
- Tripathi, A. K., Hill, P. S., Eagle, R. A., Mosenfelder, J. L., Tang, J., Schauble, E. A., Eiler, J. M., Zeebe, R. E., Uchikawa, J., and Coplen, T. B., 2015, Beyond temperature: Clumped isotope signatures in dissolved inorganic carbon species and the influence of solution chemistry on carbonate mineral composition: *Geochimica et Cosmochimica Acta*, v. 166, p. 344-371.
- Tucker, M. E., and Wright, V. P., 2009, *Carbonate sedimentology*, John Wiley & Sons.
- Van Bentum, E., Reichert, G.-J., Forster, A., and Sinninghe Damsté, J. S., 2012, Latitudinal differences in the amplitude of the OAE-2 carbon isotopic excursion: pCO_2 and paleo productivity: *Biogeosciences*, v. 9, no. 2, p. 717-731.
- VanDeVelde, J. H., Bowen, G. J., Passey, B. H., and Bowen, B. B., 2013, Climatic and diagenetic signals in the stable isotope geochemistry of dolomitic paleosols spanning the Paleocene–Eocene boundary: *Geochimica et Cosmochimica Acta*, v. 109, p. 254-267.
- Voigt, S., Gale, A. S., and Flögel, S., 2004, Midlatitude shelf seas in the Cenomanian-Turonian greenhouse world: Temperature evolution and North Atlantic circulation: *Paleoceanography*, v. 19, no. 4.
- Wang, X., Piao, S., Ciais, P., Friedlingstein, P., Myneni, R. B., Cox, P., Heimann, M., Miller, J., Peng, S., and Wang, T., 2014, A two-fold increase of carbon cycle sensitivity to tropical temperature variations: *Nature*, v. 506, no. 7487, p. 212-215.
- Wilson, P. A., Norris, R. D., and Cooper, M. J., 2002, Testing the Cretaceous greenhouse hypothesis using glassy foraminiferal calcite from the core of the Turonian tropics on Demerara Rise: *Geology*, v. 30, no. 7, p. 607-610.
- Wynn, J. G., 2007, Carbon isotope fractionation during decomposition of organic matter in soils and paleosols: implications for paleoecological interpretations of paleosols: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 251, no. 3, p. 437-448.
- Zaarur, S., Olack, G., and Affek, H. P., 2011, Paleo-environmental implication of clumped isotopes in land snail shells: *Geochimica et Cosmochimica Acta*, v. 75, no. 22, p. 6859-6869.
- Zachos, J. C., Dickens, G. R., and Zeebe, R. E., 2008, An early Cenozoic perspective on greenhouse warming and carbon-cycle dynamics: *Nature*, v. 451, no. 7176, p. 279-283.

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Author contributions

F.S. wrote the paper; A.L.W., F.S. and A.K.T. conceived the project. Isotopic analysis was carried out by F.S., A.K.T., S.S. and R.C. All authors were involved in discussions and commented on the manuscript.

Chapter III

Low-latitude terrestrial climate of southeastern Tibet revealed through clumped isotope thermometry: Implications for Late Cretaceous elevation history of southeastern Tibet

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Abstract

Views range widely on the timing of surface uplift of the Tibetan Plateau to its current height of ~5 km, specifically, whether the current elevation of Tibet developed solely in response to the Cenozoic Indo-Asia collision, or was partially inherited from high topography developed during Mesozoic is unclear. Here, we evaluate the Cretaceous elevation history of southeastern Tibet by studying pedogenic carbonates from an alluvial sequence in the Qamdo basin using clumped isotopes as a paleoaltimetry tool. Climate model simulations using the Cretaceous paleogeographic boundary conditions are also conducted to investigate the impact of global climate change on soil temperatures in our study area.

We examined two stratigraphic sections through the Cenomanian-Santonian strata of the Qamdo basin and find that petrography, burial history and elemental composition of paleosol carbonates are consistent with preservation of original isotopic signatures and minimal influence from solid-state reordering. Average soil temperatures (i.e., Δ_{47-T}) decreased from 44 ± 1.5 °C (95% CI) in the Turonian to 38 ± 3.5 °C (95% CI) by the Santonian time, coincident with decrease of average $\delta^{18}\text{O}_{\text{water}}$ values from $-6.3 \pm 0.5\text{‰}$ (95% CI) to $-7.3 \pm 0.5\text{‰}$ (95% CI). Climate model outputs and tropical sea surface temperature (SST) records both show that global

climate during the Turonian-Santonian interval were relatively warm and stable, without evidence of significant global cooling. Together, the observed changes in Δ_{47-T} and $\delta^{18}\text{O}_{\text{water}}$ values exhibit a pattern that is consistent with surface elevation changes during Late Cretaceous time. Assuming temperature and isotopic lapse rates of $\sim 6^\circ\text{C}/\text{km}$ and $\sim 1\text{‰}/\text{km}$, respectively, under the extreme warm Late Cretaceous climate conditions, the observed Δ_{47-T} and $\delta^{18}\text{O}_{\text{water}}$ shifts correspond to a positive elevation change of $\sim 1\text{ km}$ in southeastern Tibet during Late Cretaceous time. This inference is supported by evidence of strontium isotope ratios, structure, and sedimentology. Strontium isotope ratios of pedogenic carbonate increase up-section from 0.7106 to 0.7112, indicating enhanced weathering of continental source areas. Meanwhile, the development of growth strata and occurrence of thick-bedded conglomerate layers within the Late Cretaceous strata all implying active tectonics during deposition. Together, our data suggest that a moderate topography with elevation $>1.5\text{ km}$ was probably established by Late Cretaceous time in SE Tibet.

Introduction

Although it has long been recognized that the Tibetan Plateau is a Cenozoic feature produced by the India-Asia collision initiated ~ 55 million years ago (e.g., Dewey and Burke, 1973; Dewey et al., 1988; Garzanti, 2008; Powell and Conaghan, 1973; Rowley, 1996) ample evidence of pre-Cenozoic crustal shortening and thickening (and by inference, topographic growth) have been documented throughout the Tibet, particularly from Cretaceous geological records (Decelles et al., 2007; England and Searle, 1986; Kapp et al., 2007; Kapp et al., 2003; Kapp et al., 2005; Murphy et al., 1997; Studnicki-Gizbert et al., 2008; Sun et al., 2015; Volkmer et al., 2007). Cretaceous crustal deformation and shortening (Decelles et al., 2007; Kapp et al.,

2007; Kapp et al., 2005; Leier et al., 2007; Murphy et al., 1997; Wang et al., 2008), as well as rapid Cretaceous rock cooling histories (Deng et al., 2012; Hetzel et al., 2011; Reid et al., 2005; Rohrmann et al., 2012; Wang et al., 2014; Wilson and Fowler, 2011; Zhang et al., 2012) are mainly attributed to post-collisional orogenesis associated with the Qiangtang-Lhasa collision and/or crustal thickening due to flat-slab subduction of the Neotethyan ocean beneath the Lhasa block. It has been proposed that high elevations in at least southern Tibet were inherited from a pre-Eocene proto-plateau referred to as the 'Lhasaplano' (Kapp et al., 2007). The hypothesis is supported by sedimentological evidence, where Late Cretaceous deformation has been documented in the northern Lhasa block reflecting early topographic growth in that area (Sun et al., 2015). However, thermochronology data from southeastern (SE) Tibet (e.g., Yidun and Songpan-Ganzi terranes) also reveal episodes of extensive Late Jurassic-Early Cretaceous regional rock cooling, which may related to widespread rock uplift resulted from compressional strain field created by the Lhasa-Qiangtang collision (Tian et al., 2014). If the later study is representative, it suggests SE Tibet may had been an elevated area during Late Cretaceous, and therefore the proposed Cretaceous Tibet 'proto-plateau' or 'Lhasaplano' may had been spatially more extensive. Recent oxygen isotope-based paleoaltimetry study shows that SE margin of the Tibetan Plateau was at its present elevation of ~3 km as early as late Eocene time (Hoke et al., 2014). Whereas, whether this high late Eocene elevation was entirely obtained through rapid uplift after the initial collision between India and Asia, or was partially inherited from pre-collisional topography of a Late Cretaceous 'proto-plateau' is unclear. The puzzle could only be resolved if pre-Cenozoic elevation history of SE Tibet is constrained.

Oxygen isotope-based paleoaltimetry

Oxygen isotope-based paleoaltimetry (e.g., Currie et al., 2005; Graham et al., 2005; Hren et al., 2009; Kent-Corson et al., 2009; Mulch et al., 2010; Polissar et al., 2009; Quade et al., 2011; Quade et al., 2007; Rowley and Garzione, 2007; Saylor et al., 2009) is a widely used technique for evaluating surface elevation history that is independent from other surface uplift indicators (e.g., crustal deformation and rock cooling history). The method is based on the fractionation of oxygen isotopes from atmospheric water vapor to rainfall, in a process known as ‘Rayleigh distillation’ (Blisniuk and Stern, 2005; Craig and Gordon, 1965; Dansgaard, 1964; Poage and Chamberlain, 2001; Rowley and Garzione, 2007). In general, $\delta^{18}\text{O}$ values of meteoric waters tend to decrease with increasing surface elevations at certain isotopic-elevation gradient (e.g., global average = $\sim -2.8\text{‰/km}$) (Poage and Chamberlain, 2001; Rowley, 2007). Therefore, paleo-elevations can be estimated if $\delta^{18}\text{O}$ signals of paleo-meteoric waters are preserved in geologic records (e.g., lacustrine carbonate, soil carbonate, fossil shells) and proper isotopic lapse rates can be assumed. Soil carbonates are one of the most widely used proxies for $\delta^{18}\text{O}$ values of paleo-meteoric waters (e.g., Graham et al., 2005; Quade et al., 2007; Saylor et al., 2009; Mulch et al., 2010; Quade et al., 2011) and appear to provide the best record of paleo-precipitation as they grew from soil waters mostly sourced from rain waters fall directly on the paleo-ground (Rowley and Garzione, 2007). However, applicability of oxygen isotope-based paleoaltimetry is often complicated by: 1) uncertainty of oxygen isotopic fractionation factor between minerals (e.g., pedogenic carbonate) and the waters from which they grew as the fractionation is temperature-dependent (Kim and O’Neil, 1997); 2) enrichment of heavy oxygen isotopes in surface waters by evaporation (e.g., Quade et al., 2007); and 3) climatic influences (e.g., moisture source, atmospheric circulation patterns and seasonality of carbonate

precipitation) on $\delta^{18}\text{O}$ values of surface waters and therefore precipitated carbonate minerals (e.g., Garzzone et al., 2000).

Clumped isotope-based paleoaltimetry

Carbonate clumped isotope thermometry is an innovative technique that is capable of measuring the formation temperature of carbonate minerals (e.g., Eiler, 2007; Ghosh et al., 2006). Different from conventional stable isotope techniques, clumped isotope thermometry measures the abundance of mass-47 (which reflects the abundance of ^{13}C - ^{18}O bond) in CO_2 produced by acid-digestion of carbonate minerals (e.g., Ghosh et al., 2006; Eiler, 2007). Since “clumping” of ^{13}C and ^{18}O ions in carbonate minerals is temperature-dependent, abundance of ^{13}C - ^{18}O bonds (reported as “ Δ_{47} ”) (Eiler and Schauble, 2004) can be correlated to formation temperature of carbonate minerals (e.g., Eiler and Schauble, 2004; Ghosh et al., 2006; Tripathi et al., 2015). The Δ_{47} -derived temperature (Δ_{47} -T) of carbonate minerals (e.g., soil carbonate, lacustrine carbonate, fossil shell) can then be correlated to surface air temperatures (e.g., Hough et al., 2014; Hren and Sheldon, 2012; Quade et al., 2013; Tripathi et al., 2010). For example, it has been suggested that Δ_{47} -Ts recorded by soil carbonate usually reflect warm season surface air temperatures (e.g., Quade et al., 2013; Hough et al., 2014). After surface temperatures are constrained, paleo-elevation histories can be assessed by applying appropriate temperature-elevation lapse rates (e.g., global average = $\sim 5.5^\circ\text{C}/\text{km}$) (Meyer, 1992; Meyer, 2007), if other temperature-influencing factors (e.g., climate change or seasonality of carbonate growth) can be ruled out. Furthermore, constraint on carbonate forming temperature also makes estimation of $\delta^{18}\text{O}$ values of meteoric water ($\delta^{18}\text{O}_{\text{water}}$) from which carbonate grew more accurate as fractionation of O^{18} between carbonate and water is temperature-dependant (Kim and O'Neil,

1997). The combination of $\delta^{18}\text{O}$ - and Δ_{47} -based paleoaltimetry techniques provide two independent controls on paleo-elevation histories and have so far yielded some of the most robust paleo-elevation estimates (e.g., Carrapa et al., 2014; Fan et al., 2014; Garzione et al., 2014; Huntington et al., 2015; Leier et al., 2013; Snell et al., 2014).

Purpose of the study

Here, we present estimates of Late Cretaceous (Coniacian-Santonian) soil temperatures and $\delta^{18}\text{O}_{\text{water}}$ values for SE Tibet from carbonate clumped isotope thermometry on paleosol carbonates collected from the Markam area (Fig. 1). Combined with soil temperature and $\delta^{18}\text{O}_{\text{water}}$ data from the underlying Cenomanian-Turonian strata (Shang et al., in review), we first reconstruct the Cenomanian-Santonian (~100 to 83 Ma) surface temperature history of SE Tibet. Then, the soil temperature records are compared with contemporaneous global climate inferred from marine temperature datasets (Bornemann et al., 2008; Forster et al., 2007; Huber et al., 2002; Voigt et al., 2004) in terms of magnitude and temporal evolution. Soil temperature variations that are inconsistent with global climate patterns are used to determine local paleo-elevation changes for SE Tibet during Late Cretaceous time. The method can be potentially complicated by several factors: 1) terrestrial climate change may or may not response consistently with marine temperature change (e.g., Lunt et al., 2010); 2) changes of seasonality/timing of soil carbonate growth through time; and 3) uncertainty of temperature-elevation gradient during Late Cretaceous. To better constrain how global climate change affects soil temperatures during Late Cretaceous time, we used a general circulation model HadCM3L (Lunt et al., 2015) to simulate soil temperatures (at depth of 25-65 cm) for East Asia. High resolution ($0.5^\circ \times 0.5^\circ$) paleogeographic reconstructions for each studied geologic stages

adapted from Jarvis et al. (2011) based on the paleogeographic reconstruction of Blakey (2008). Red square marks the estimated location of SE Tibet. (C) Local geologic (Pan et al., 2004) and satellite (Digital Global, 215) maps of the Markam area, showing local lithology and structural features as well as locations of the two studied sections (after Shang et al., in review).

Geologic Background

Paleogeography and stratigraphy

Our study site is located in the Markam County of SE Tibet where Cretaceous red beds are present (Fig. 1). Previous paleomagnetic work in the Markam area suggests the region was located at between $\sim 20\text{--}40^\circ\text{N}$ during Late Cretaceous time (Huang et al., 1992). Recent high-resolution paleomagnetic data from the southern Lhasa terrane reveals that the region maintained relatively stable at $\sim 20\text{--}25^\circ\text{N}$ paleo-latitude during Late Cretaceous before drifted northward to its current latitude at $\sim 30^\circ\text{N}$ since Late Eocene (Lippert et al., 2014). Here, we suggest that paleo-latitude of our study site in SE Tibet was similar to that of suggested by Lippert et al. (2014), as indicated by relatively short distance ($\sim 700\text{ km}$) between our study site and the place in the southern Lhasa terrane where paleomagnetic samples were collected (Lippert et al., 2014), as well as almost identical current day latitudes between the two places (Markam: $29^\circ 31'\text{N}$; southern Lhasa terrane: $\sim 30^\circ\text{N}$).

Cretaceous strata in the Markam area of SE Tibet cover an area of $\sim 2,000\text{ km}^2$ and contain $\sim 1,200\text{ m}$ thick terrestrial deposits composed of clast-supported conglomerate, cross-stratified pebbly red sandstone, and thick-bedded carbonate nodule-bearing paleosol layers underlain by Triassic-Jurassic marine-lacustrine deposits (Du et al., 1997; Pan et al., 2004). The Cretaceous sequence is further divided into two units: the Early Cretaceous (late Aptian-Albian)

Jingxing Formation and the Late Cretaceous (Cenomanian-Santonian) Nanxing Formation (GSX, 2007; Huang et al., 1992; Pan et al., 2004). The Late Cretaceous Nanxing Formation consists of ~600 m of red to purplish lithic sandstone, interbedded with layers of conglomerate, pebbly sandstone, mudstone, and carbonate nodule-bearing paleosol horizons (Du et al., 1997; GSX, 2007). Stage boundaries and ages within the Nanxing Formation are determined based on magnetostratigraphy (Huang et al., 1992; Otofujii et al., 1990), terrestrial macrofossil record (GSX, 2007), and palynological data from Shang et al. (in review) (Fig. A.1.). Depositional ages of the lower Nanxing Formation is constrained at between Cenomanian and Turonian (~100-89 Ma), and that of the upper Nanxing Formation is constrained at between Coniacian and Santonian (~89-93 Ma) (Fig. A.1.).

Figure A.1. (A) Stratigraphic column and age control of the Late Cretaceous sequence (Shang et al., in review). (B) Local satellite map of studied sections exposed near the city of Markham.

Existing Late Cretaceous (Cenomanian-Turonian) soil temperature record

Previous study (Shang et al., in review) has reported soil temperatures from the lower

Nanxing Formation (i.e., Cenomanian-Turonian interval) in the Markam area based on clumped isotope thermometry on soil carbonate. The study reveals a terrestrial warming trend through the Cenomanian-Turonian interval, with average soil temperatures increased from ~35 °C in the Cenomanian to ~44 °C in the Turonian (Shang et al., in review). The observed soil temperature trend closely mirrors coeval global climate inferred from marine dataset. Subtropical sea surface temperatures (SSTs) indicated by geochemistry of brachiopod shell show a long-term warming trend through the Cenomanian-Turonian interval (Voigt et al., 2004). Similarly, tropical SSTs derived from TEX₈₆ proxy show that SSTs raised from ~31 °C in the Cenomanian to ~35 °C in the Turonian (Bornemann et al., 2008; Forster et al., 2007). Together, continental and marine records both reveal an episode of global warming during the Cenomanian-Turonian time (e.g., Clarke and Jenkyns, 1999; Voigt et al., 2004).

Analytical method

Sampling

We sampled 8 carbonate nodule-bearing paleosol horizons in the upper Nanxing Formation (Coniacian-Santonian in age) from two measured sections (i.e., sections: A-A', C-C') (Fig. 1C). Five samples were taken from section A-A' and the rest three samples were collected from section C-C'. The sampled paleosol layers are all classified as Calcisol according to the paleosol classification scheme of Mack et al. (1993). Carbonate nodules were collected at depth of ~50 cm below top of the preserved paleosol surface to constrain variation of soil temperature with depth (e.g., Quade et al., 2013). Four additional paleosol carbonate samples collected in previous study from the lower Nanxing Formation (Cenomanian-Turonian in age) (Shang et al., in review) are also included in this study for strontium isotope (⁸⁷Sr/⁸⁶Sr) analysis.

Carbonate nodules were first cut in halves, polished and examined under transmitted light using microscope. For Δ_{47} analysis, only the areas with homogeneous micritic textures were carefully sampled using a low-speed dremel drill with periodic cooling of the drill pit to avoid potential resetting of the samples. Carbonate powders were drilled from multiple places within a nodule and mixed together; therefore, the resultant soil temperature most likely represents averaged climate conditions during the course of soil nodule formation. For X-ray diffraction (XRD), Mg/Ca, Sr/Ca ratios and strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) analyses, since the required sample sizes are larger (>2 gram), nodule samples were first broken into fine pieces (<0.5 cm in diameter) and examined under microscope to avoid sampling the pieces that contain visible carbonate veins and spars. Selected portions were then grounded into powders using a ceramic mortar and pestle.

Petrographic, mineralogical and isotopic screening for diagenesis

Carbonate diagenesis is assessed using a combination of petrology, mineralogy, and isotopic approaches. Polished petrographic thin sections were produced for all carbonate nodule samples (n=8) and examined under transmitted and cross-polarized lights to characterize carbonate textures (i.e., micrite versus sparry cement). X-ray diffraction (XRD) analysis was performed on all 8 samples using a PANalytical X'Pert Pro XRD diffractometer equipped with Cu K α radiation at 40 keV and 40 mA. Samples were scanned from 10° to 100° at 2 θ with a step interval of 0.05°. Interpretation of mineralogy was aided by Highscore Plus software equipped with the Powder Diffraction File 3 database. Sr/Ca and Mg/Ca ratios of the 8 carbonate samples were measured at Actlabs, Canada. Carbonate phases were first dissolved in 0.5 M acetic acid follow the method of Ray et al. (1957), and then analyzed by inductively coupled plasma mass

spectrometry (ICP-MS) for Ca, Mg and Sr concentrations. In addition, isotopic (i.e., $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) screening was performed on carbonate samples (n=6) that contain both micritic and sparry textures, with the purpose of characterizing carbonate diagenetic history (e.g., Garziane et al., 2004). Micrite and co-existing sparry cement were micro-sampled (n=1 for cement; n=2 for micrite; average sample size =0.3mg) using a hand drill. Isotopic analysis was performed at West Virginia University Stable Isotope Lab. Samples were measured using a GasBench device coupled to a Delta Advantage isotope ratio mass spectrometer. Results of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ were standardized using NBS18 and NBS19 carbonate standards. Analytical precision is better than ± 0.1 for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (1s), based on repeated analysis of internal standards.

Strontium isotope analysis

Strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) analysis of paleosol carbonate samples (n=8) was conducted at the Center for Applied Isotope Studies, University of Georgia. Carbonate fraction was first extracted with 0.5 M acetic acid. $^{87}\text{Sr}/^{86}\text{Sr}$ measurement was conducted by dynamic multi-collection on the Sector 54. Extracted Sr is loaded onto Re filaments with H_3PO_4 and TaCl_5 activator, and produces a stable beam for several hours. The external reproducibility of $^{87}\text{Sr}/^{86}\text{Sr}$ analyses is better than 25 ppm (2σ) based on analyses of past 10 years, with $^{87}\text{Sr}/^{86}\text{Sr}$ of NBS 987 measured over that time period averages 0.710246 ± 0.000011 (2σ).

Clumped isotope analysis

Carbonate clumped isotope was measured at University of California, Los Angeles (UCLA) stable isotope lab, using a specially modified Thermo Fisher 253 gas source mass

spectrometer with mass 44-49 ion collectors (“Chewbacca”). Samples and carbonate standards were processed using an automated system for acid digestion and purification (after Passey et al., 2010) before transferred to the inlet of the mass spectrometer. For each measurement, 10-12 mg of sample powders were digested at 90 °C in phosphoric acid, the resultant CO₂ being cryogenically purified before analysis on the mass spectrometer. Absolute reference frame (ARF) was established by analyzing equilibrium CO₂ samples prepared at 1000 °C and 25 °C. The standard gas are purified and analyzed using the same protocol as gases generated from carbonate samples. For carbonates, the Δ_{47} values are normalized to 25 °C using a 0.092‰ value from Passey et al. (2010) as recalculated in Henkes et al., 2013. Clumped-isotope data of pedogenic carbonate are reported relative to the ARF of Dennis et al. (2011). Δ_{47} -temperatures are calculated using the inorganic calibration of Tripathi et al. (2015), generated on the same instrument. Uncertainties in reported Δ_{47} values and calculated temperatures include the propagated uncertainty in equilibrated gas determination and in sample measurement. Three internal carbonate standards (i.e., Carrerra Marble, Carmel Chalk and Veinstorm Limestone) were analyzed multiple times during the course of the clumped isotope analysis, which yield mean Δ_{47} values of $0.390 \pm 0.012\text{‰}$ (n =11), $0.697 \pm 0.014\text{‰}$ (n =8), and $0.725 \pm 0.011\text{‰}$ (n =12), respectively (Table A.1.). The standard errors (SE) for Δ_{47} values reported in Table 2 are calculated as $SE=SD/\text{SQRT}(N)$, where SD is the observed standard deviation for repeated analyses of samples, and N is the number of repeated analyses. If the observed SE is smaller than the long-term lab precision (0.009‰), SE is calculated using $SD=0.009\text{‰}$ (Tripathi et al., 2015). Error in Δ_{47} -T is calculated by propagating the SE for Δ_{47} through the Tripathi et al. (2015) temperature calibration equation.

The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ ratios were measured during the course of clumped isotope analysis

and reported relative to VPDB standard, with analytical precisions all $>0.1\text{‰}$. The $\delta^{18}\text{O}$ values of soil water ($\delta^{18}\text{O}_{\text{water}}$) were then calculated using the Δ_{47} -T and temperature-dependant oxygen fractionation equation between calcite and water (Kim and O'Neil, 1997). Errors of the calculated $\delta^{18}\text{O}_{\text{water}}$ values reflect uncertainties of Δ_{47} -T measurement.

Cretaceous climate simulation

To investigate the impact of global climate change on soil temperatures in SE Tibet, we used the General Circulation Model (GCM) HadCM3L ($3.75^\circ \times 2.5^\circ$ atmosphere and ocean resolution) (Lunt et al., 2015), a fully coupled Atmosphere-Ocean General Circulation model modified from the HadCM3 model (Gordon et al., 2000), to generate soil temperature at depth of 25-65 cm, sea surface temperature (SST), precipitation and atmospheric circulation pattern data for East Asia ($5\text{--}60^\circ\text{N}$ paleo-latitude, $50\text{--}160^\circ\text{E}$ paleo-longitude) during summer months (i.e., June, July, August and September). It has been suggested that soil carbonates typically form in warm summer months (Hough et al., 2014; Passey et al., 2010; Quade et al., 2013) and therefore record warm season climatic conditions. Four time-slice simulations were constructed utilizing high-resolution ($0.5^\circ \times 0.5^\circ$) paleogeographic boundary conditions created by Getch Plc using the method of Markwick (2007) (Fig. 2), including the Cenomanian (100.5-93.9 Ma), Turonian (93.9-89.8 Ma), Coniacian (89.8-86.3 Ma) and Santonian (86.3-83.6 Ma) geological stages (Gradstein et al., 2004). Note that each time-slice does not represent specific date in geologic record due to age uncertainty of proxy data used in the simulation. Instead, the reconstructions reflect time-averaged mid-point condition for each geologic stage (Lunt et al., 2015).

Unlike marine temperatures that mostly reflect global climate conditions, land surface temperatures can be affected by both global climate and local topography (e.g., the modern day

Tibetan Plateau). Our model simulations limit the effects of topography on soil temperatures by prescribing a relatively constant elevation values (1000 ± 200 m) throughout the modeled time interval (Fig. 2). Therefore, any variation of soil temperature between simulated geologic stages is likely due to changes of global climate rather than local topography.

The climate model provides the most temporally and spatially comprehensive Cretaceous climate simulations that are currently available (Lunt et al., 2015). A more detailed model description including parameters used during simulation (e.g., solar luminosity, paleogeography, $p\text{CO}_2$) can be found in Lunt et al. (2015).

Figure 2. Paleogeographic boundary conditions for the Cenomanian, Turonian, Coniacian and Santonian time-slices used in climate simulation, showing elevation and bathymetry (Getech Plc). Note that the area corresponds to current day SE Tibet (marked by black square) is assumed

to have relatively constant paleoelevation (1000 ± 200 m) during the Cenomanian-Santonian time.

Results

Petrographic, mineralogical and isotopic screening for diagenesis

Thin-section petrographic studies show that the pedogenic carbonate nodules are dominated by red-brown micrite texture with occasional crack-filling sparry cement (Fig. 3). The spars have distinct crystal boundaries from the surrounding micritic groundmass (Fig. 3D). ‘Floating’ angular quartz grains are common (Fig. 3) and no detrital carbonate grains are found in thin sections. Results of XRD mineralogy, Sr/Ca and Mg/Ca elemental ratios are presented in Table 1. Computer-aided mineral identification suggests that calcite is the major component and only carbonate mineral present. Secondary components include quartz and minor clay minerals. Elemental analysis of carbonates shows low Mg/Ca and Sr/Ca ratios (<0.01) (Table 1).

Figure 3. Thin-section photos of studied carbonate nodules. (A-C) Thin-section photos of paleosol carbonate nodules taken under polarized lights showing dominant micritic texture. Detrital quartz grains are ‘floating’ in micritic groundmass. (D) Photomicrograph taken under polarized lights showing crack-filling sparry cement. Only the areas with homogeneous micritic texture were sampled for clumped isotopic analyses.

Table 1. Results of XRD, Mg/Ca and Sr/Ca ratios, and strontium isotope analyses

Sample ID	Stratigraphic Position (m)	XRD results				Elemental results		Sr isotope
		Number of counts	100% 2°θ	100% d-spacing (Å)	Main component	Mg/Ca	Sr/Ca	⁸⁷ Sr/ ⁸⁶ Sr
Section A-A'								
MK6331	582	77387	29.256	3.05275	Calcite, Quartz	0.0046	0.0006	N/A
MK6330	573	73436	29.320	3.04619	Calcite, Quartz	0.0050	0.0005	0.7112
MK6329	560	71706	29.437	3.03666	Calcite, Quartz	0.0062	0.0007	N/A
MK6328	418	10113	29.485	3.02950	Calcite, Quartz	0.0048	0.0006	N/A
MK6327	390	75915	29.333	3.04488	Calcite, Quartz	0.0052	0.0008	0.7107
MK6325*	281							0.7106
MK6319*	215							0.7104
MK6317*	192							0.7106
MK6315*	107							0.7104
Section C-C'								
MK6234	455	37780	29.203	3.05812	Calcite, Quartz	0.0062	0.0007	0.7110
MK6233	410	7687	29.483	3.02969	Calcite, Quartz	0.0238	0.0008	0.7106
MK6232	395	14724	29.487	3.02933	Calcite, Quartz	0.0154	0.0009	N/A
Note: *Sample from Shang et al., in review.								
N/A: data not available								

The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of micrite and co-existing sparry cement show that the two types of carbonate phases have very similar isotopic values (Table 2). Variation of $\delta^{18}\text{O}$ ratios between micrite and co-existing spars averages 0.21‰ and ranges from 0.04‰ to 0.41‰, while the same variation for $\delta^{13}\text{C}$ ratios averages 0.14‰ and ranges from 0.01‰ to 0.32‰ (Table 2). In comparison, the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of both micrite and spars can vary as much as ~1.2‰ (for $\delta^{18}\text{O}$) and ~0.6‰ (for $\delta^{13}\text{C}$) among different paleosol horizons (Table 2).

Table 2. Isotopic values of micro-sampled micrite and co-existing sparry cement

Nodule Sample ID	Nodule 1		Nodule 2		Nodule 3		Nodule 4		Nodule 5		Nodule 6	
Sampling Height	390 m		395 m		410 m		455 m		573 m		582 m	
$\delta^{13}\text{C}$, $\delta^{18}\text{O}$ (‰, VPDB) micrite 1	-9.03	-12.28	-9.14	-12.14	-9.25	-12.10	-8.73	-11.12	-8.99	-12.31	-8.50	-11.97
$\delta^{13}\text{C}$, $\delta^{18}\text{O}$ (‰, VPDB) micrite 2	-8.95	-11.97	-9.02	-11.76	-9.41	-11.86	-8.89	-11.56	-8.73	-12.03	-8.54	-11.54
$\delta^{13}\text{C}$, $\delta^{18}\text{O}$ (‰, VPDB) spar	-9.0	-12.2	-9.34	-11.8	-9.54	-12.16	-8.71	-11.19	-8.8	-11.9	-8.41	-11.70

Strontium isotope values

Strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) ratios of paleosol carbonates from the Nanxing Formation range from 0.7104 to 0.7112 (Table 1). The $^{87}\text{Sr}/^{86}\text{Sr}$ values are all greater than 0.708 (i.e., $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of Cretaceous sea water) (Gradstein et al., 2004), indicating a continental origin of soil carbonates free from marine influence. The $^{87}\text{Sr}/^{86}\text{Sr}$ values display an increasing upward trend within the Nanxing Formation, with $^{87}\text{Sr}/^{86}\text{Sr}$ ratios increase from ~0.7106 to ~0.7112 (Fig. 4).

Clumped isotope values

Isotopic results of paleosol carbonates from the upper Nanxing Formation, including $\delta^{18}\text{O}_{\text{carbonate}}$, $\delta^{13}\text{C}_{\text{carbonate}}$, Δ_{47} , and calculated $\Delta_{47}\text{-T}$ and $\delta^{18}\text{O}_{\text{water}}$ values are listed in Tables 3. Raw data of clumped isotope analysis, including standards measured over the period of sample analysis are listed in Supplementary Table 1. The carbonate samples have Δ_{47} values range from 0.622‰ to 0.669‰ (average = 0.638‰; n=10), which correspond to carbonate precipitation

temperature between 43 °C and 30 °C using the inorganic temperature calibration curve of Tripathi et al. (2015). Δ_{47} -Ts for all upper Nanxing samples average 38 ± 3.5 °C (95% confidence interval; CI) (Fig. 4). The $\delta^{18}\text{O}$ values of soil water ($\delta^{18}\text{O}_{\text{water}}$, VSMOW) calculated using the oxygen fractionation factor of Kim and O'Neil (1997) range from -8.3 to -6.6‰ and average -7.3 ± 0.5 ‰ (95% CI) (Table 3).

Table 3. Stable isotope data including clumped isotope analysis results. Temperatures calculated using the inorganic calibration of Tripathi et al. (2015), generated on the same instrument.

Sample ID	Stratigraphic Section	N	Composite Section Height (m)	$\delta^{13}\text{C}$ (‰, VPDB)	$\delta^{18}\text{O}$ (‰, VPDB)	Δ_{47}^a (‰, ARF)	SE ^b (‰, analytical)	SE ^b (‰, external)	$\Delta_{47}-T^c$ (°C)	SE(1 σ) $\Delta_{47}-T$ (°C)	$\delta^{18}\text{O}_{\text{water}}^d$ (‰, VSMOW)	SE(1 σ) $\delta^{18}\text{O}_{\text{water}}$ (‰, VSMOW)
MK6331	A-A'	1	582	-8.50	-11.97	0.641	0.011	0.011	38	3.2	-7.3	0.6
MK6330	A-A'	1	573	-8.99	-12.31	0.652	0.012	0.012	34	3.5	-8.3	0.7
MK6329	A-A'	1	560	-8.85	-12.31	0.640	0.010	0.010	38	2.9	-7.6	0.6
MK6234	C-C'	1	455	-8.73	-11.12	0.669	0.011	0.011	30	3.1	-8.0	0.6
MK6328 avg	A-A'	2	418	-8.28	-11.99	0.637	0.011	0.011	39	3.3	-7.1	0.6
MK6328				-8.24	-11.88	0.647						
MK6328				-8.31	-12.10	0.626						
MK6233	C-C'	1	410	-9.25	-12.10	0.633	0.008	0.009	40	2.7	-7.0	0.5
MK6232 avg	C-C'	2	395	-9.14	-12.14	0.630	0.009	0.009	41	2.7	-6.9	0.5
MK6232				-9.03	-12.48	0.639						
MK6232				-9.26	-11.80	0.622						
MK6327 avg	A-A'	2	390	-9.03	-12.28	0.623	0.001	0.009	43	2.8	-6.6	0.5
MK6327				-9.16	-12.16	0.625						
MK6327				-8.90	-12.40	0.622						
All sample average									38	$\pm 3.5^f$	-7.3	$\pm 0.5^f$

^a ARF: Absolute Reference Frame.

^b SE: the standard error of Δ_{47} (SE=analytical error for N=1; SE=SD(Δ_{47})/SQRT(N) for N $\geq$ 2). When SE of a sample is less than the observed long term SE of lab standards (0.009), the long-term lab average SE of 0.009 is assigned.

^c Temperature is calculated using the inorganic calibration of Tripathi et al. (2015): $T = \text{SQRT}(50500/(\Delta_{47}-0.1185)) - 273.15$.

^d Oxygen isotopic values of soil water calculated using the oxygen fractionation factor between clacite and water: $103\ln\alpha = 18030/(T+273.15) - 32.42$ from Kim and O'Neil (1997).

^f Errors represent 95% confidence interval of the calculated all sample mean

Figure 4. $\delta^{13}\text{C}_{\text{carbonate}}$, $\delta^{18}\text{O}_{\text{carbonate}}$, $\Delta_{47}\text{-temperature}$, $\delta^{18}\text{O}_{\text{water}}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ values plotted with stratigraphic depth. Filled circles represent samples from the upper Nanxing Formation (Coniacian-Santonian in age) analyzed in this study. Filled squares represent data from the lower Nanxing Formation (Cenomanian-Turonian in age) compiled from Shang et al., in review. Strontium isotope data are all obtained in this study. Purple shading represents 95% confidence interval (CI) of stage-averaged $\Delta_{47}\text{-T}$ and $\delta^{18}\text{O}_{\text{water}}$ means.

Cretaceous climate simulation

The climate model simulations show that summer (i.e., June, July, August, and September) soil temperatures in SE Tibet are remarkably high and stable during the Cenomanian-Santonian time (Fig. 5). Estimated soil temperatures average 45 ± 1.5 °C (95% CI) for Cenomanian; 44 ± 2.6 °C (95% CI) for Turonian; 45 ± 1.5 °C (95% CI) for Coniacian; and 45 ± 1.3 °C (95% CI) for Santonian (Fig. 5). The modeled soil temperatures agree with high Δ_{47} -soil Ts ~ 40 - 45 °C in SE Tibet during late Cenomanian-Turonian time (Shang et al., in review). In addition, the modeled surface sea temperatures (SSTs) show that tropical (~ 5 - 15° N) SSTs in East Asia are relatively stable at ~ 36 °C throughout the simulated geologic stages (Fig. 6). The values are similar to SSTs in equatorial Atlantic indicated by TEX₈₆ proxy (Bornemann et al., 2008; Forster et al., 2007). The simulated summer wind patterns show that atmospheric circulations patterns were relatively unchanged throughout the Cenomanian-Santonian intervals, with winds predominantly blow from northwest to southeast (Fig. 7). The modeled summer precipitation data show that SE Tibet was characterized by arid-semiarid climate conditions with daily precipitation less than 1 mm/day during the simulated geologic interval (Fig. 7).

Figure 5. Model-derived summer (i.e., June, July, August, and September) soil temperature estimates. Simulations represent time-averaged conditions for four geologic time-slices: Cenomanian, Turonian, Coniacian, and Santonian. Soil temperatures are modeled for 25-65 cm depth.

Figure 6. Model-derived summer (i.e., June, July, August, and September) sea surface temperature estimates. Simulations represent time-averaged conditions for four geologic time-slices: Cenomanian, Turonian, Coniacian, and Santonian.

Figure 7. Model-derived summer (i.e., June, July, August, and September) wind pattern (850 mb) and precipitation estimates. Simulations represent time-averaged conditions for four geologic time-slices: Cenomanian, Turonian, Coniacian, and Santonian.

Interpretation and Discussion

Diagenesis consideration

Isotopic composition of carbonate minerals can be affected by diagenetic processes; therefore, history of diagenesis must be evaluated before meaningful interpretation of any isotopic record. It has been shown that Δ_{47} values of carbonate can be modified by solid-state C-O bond reordering if a burial temperature exceeds 100 °C for an extended period of time (Henkes et al., 2014). In our study site, previous study of paleosol carbonates from the lower Nanxing

Formation (Cenomanian-Turonian in age) reveals that the soil carbonates experienced little diagenetic alterations and preserve their original $\delta^{18}\text{O}$ and Δ_{47} signals based on combined evidence from burial history, carbonate petrography, and isotopic screening of different carbonate phases (Shang et al., in review). Therefore, there is little reason to believe that soil carbonates from the even shallower upper Nanxing Formation (Coniacian-Santonian in age) with estimated maximum burial depth <500 m (Du et al., 1997) are substantially modified by burial diagenesis. However, shallowly buried carbonate minerals could be altered by diagenetic fluids such as ground waters (e.g., Garzione et al., 2004). Observation of petrographic thin-sections shows that the studied carbonate nodules are dominated by red-brown micritic texture with estimated carbonate cement content less than 3% (Fig. 3), which suggests that late-stage carbonate recrystallization or replacement was likely insignificant. The inference is further supported by relatively low carbonate Mg/Ca (<0.03) and Sr/Ca (<0.001) ratios (Table 1) (e.g., Graham et al., 2005).

We noticed that despite carbonate cement only makes up a small percentage of carbonate nodules, the amount of material required for clumped isotope analysis (~10 mg per sample) makes it possible that minor amount of carbonate cement was also included during sampling. Isotopic screening of micrite and co-existing carbonate cement shows that they have very similar $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values, with the differences average 0.14‰ (range =0.01-0.32‰) and 0.21‰ (range =0.04-0.41‰) for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, respectively (Table 2), whereas isotopic values among different paleosol layers can vary up to ~1.2‰ (for $\delta^{18}\text{O}$) and ~0.6‰ (for $\delta^{13}\text{C}$) (Table 2). The similar isotopic patterns have been detected from the underlying Cenomanian-Turonian strata, and are interpreted to reflect early diagenetic origin of carbonate cement formed at near-surface environmental conditions (Shang et al., in review). Here, we suggest that the carbonate cement in

the overlying Coniacian-Santonian strata are likely characterized by a similar early diagenetic history and therefore primarily reflect near-surface environmental conditions.

Interpretation of surface cooling

Soil carbonate forming temperatures measured in this study, combined with soil temperature data from Shang et al. (in review) are used to reconstruct the early Late Cretaceous (Cenomanian-Santonian) surface temperature history for SE Tibet (Fig. 4). Two major temperature trends can be observed from the reconstruction: 1) a warming of soil temperature $\sim 10^{\circ}\text{C}$ at the Cenomanian-Turonian transition; and 2) a subsequent cooling of $\sim 6^{\circ}\text{C}$ at the Turonian-Coniacian boundary. The former trend has been attributed to global warming at the Cenomanian-Turonian interval (Shang et al., in review), whereas the cause of the later temperature trend is not clear and it may be associated with various factors such as global climate, seasonality of carbonate precipitation, paleo-latitude and/or local topography.

Climate consideration

Global climate change may account for the observed cooling of soil temperatures during the Coniacian-Santonian time (Fig. 4). Ideally, the best way to evaluate the climatic effect on soil temperatures for our study site is to look at coeval terrestrial temperature records from other regions; however, such data do not exist. Alternatively, marine temperature records (e.g., Bornemann et al., 2008; Forster et al., 2007; Huber et al., 2002; Linnert et al., 2014; Voigt et al., 2004) could be used as indicator for coeval global climate, however, uncertainty rises as local terrestrial climate may or may not respond consistently with the direction and magnitude of global climate change as indicated by marine records (e.g., Snell et al., 2014). Studies have

shown that land temperatures are often more sensitive to global climate change than marine temperatures (e.g., Lunt et al., 2010; Royer et al., 2012). The amount of temperature amplification on land, however, is not only hard to estimate but varies spatially (location-specific) (e.g., Kump and Pollard, 2008; Lunt et al., 2010; Renssen et al., 2004; Royer et al., 2012; Snell et al., 2013). To constrain this uncertainty for our study site, we estimated the magnitude of soil temperature response to marine temperature change using the Δ_{47} -soil temperature data from the Cenomanian-Turonian strata in the Markam area (Shang et al., in review). Average Markam soil temperatures increased from $\sim 35^{\circ}\text{C}$ in the Cenomanian to $\sim 44^{\circ}\text{C}$ in the Turonian (Fig. 8B). The estimated soil temperature trend closely mirrors coeval SST (Fig. 8C) and sea bottom temperature (SBT) records inferred from proxy data (Table 4), which both show pronounced warming of marine temperatures $\sim 4\text{--}5^{\circ}\text{C}$ during the Cenomanian-Turonian interval (Bornemann et al., 2008; Forster et al., 2007; Huber et al., 2002; Linnert et al., 2014; Voigt et al., 2004). We then compared the amount of temperature change of both soil and marine records in response to the Cenomanian-Turonian global warming (Table 4). The comparison reveals that soil temperatures in our study site are $\sim 2\times$ more sensitive to both SST and SBT changes (Table 4). The estimated ‘soil sensitivity’ value was then utilized to predict the amount of soil temperature change in response to the marine temperature variations during the Coniacian-Santonian time in the Markam area (Forster et al., 2007; Bornemann et al., 2008). Marine temperatures indicated by TEX_{86} proxy (Bornemann et al., 2008; Forster et al., 2007) shows that tropical SSTs stay fairly stable at $\sim 35\text{--}36^{\circ}\text{C}$ during the Turonian-Santonian interval (Fig. 8C). The small SST variation would only cause soil temperature change $\sim 1\text{--}2^{\circ}\text{C}$ when calculated using the estimated soil temperature ‘sensitivity factor’ of 2 (Table 4). Obviously, this small soil temperature variation cannot account the $\sim 6^{\circ}\text{C}$ cooling observed in Δ_{47} -soil T record

(Fig. 8D). It is worth to note that this estimation assumes that the ‘soil sensitivity’ value is relatively unchanged throughout the early Late Cretaceous (Cenomanian-Santonian), albeit more coeval terrestrial temperature data from Tibet and adjacent regions are needed to further test this assumption.

Table 4. Comparison of variations of soil and marine temperatures in respond to Cenomanian-Turonian global warming event

	Average soil temperature (°C)	Average tropical sea surface temperature (°C)	Average sub-tropical sea surface temperature (°C)	Average sub-tropical sea bottom temperature (°C)
Location (paleo-latitude)	Markam, Southeastern Tibet (~20-25° N)	Demerara rise, Atlantic Ocean (~5° N)	European shelf-sea basins (26-36° N)	Black Nose, North Atlantic (~30° N)
Proxy type	Δ_{47} (soil carbonate)	TEX ₈₆ (marine organics)	$\delta^{18}\text{O}$ (plankton)	$\delta^{18}\text{O}$ (benthic foram)
Time Interval				
Cenomanian	~ 35	~ 31	~ 18	~ 14
Turonian	~ 44	~ 35	~ 23	~ 18
Reference	Shang et al., in review	Forster et al., 2007; Bornemann et al., 2008	Voigt et al., 2004	Huber et al., 2002
$\Delta T_{\text{Cenomanian-Turonian}}$	~ +9	~ +4	~ +5	~ +4

The inference that global climate change had limited impact on soil temperatures of SE Tibet is also supported by climate simulations showing relatively constant soil temperatures in SE Tibet area (Fig. 5), as well as stable SSTs of east Tethys Ocean throughout the Cenomanian-Santonian time interval (Fig. 6). Note that there are two data points in the proxy-derived SST record (Forster et al., 2007) showing that tropical SSTs cooled to ~33 °C in the latest Santonian (Fig. 8C). However, this latest Santonian cooling event cannot account for the observed Coniacian-Santonian cooling revealed by Markam paleosol record because cooling of soil temperatures preceded cooling of SSTs by ~5 Ma (Fig. 8B, C).

Figure 8. (A) Climate model (HadCM3L) output of average soil temperature (at soil depth between 25 cm and 65 cm) for SE Tibet during the Cenomanian, Turonian, Coniacian and Santonian time slices. Red shading marks 95% confidence interval (CI) of the modeled mean soil temperature for each modeled geologic stage. (B) Plot of Δ_{47} -soil temperature with stratigraphic height. Filled circles represent samples analyzed in this study, while filled squares represent data compiled from Shang et al., in review. Purple shading represents 95% CI of the stage-averaged Δ_{47} -T value. (C) The Cenomanian-Santonian tropical sea surface temperatures (SSTs) from organic geochemical analysis (i.e., TEX_{86}) of ocean drilling cores (Bornemann et al., 2008; Forster et al., 2007). Blue shading marks the place where >95% data points lie. (D) Relative temperature changes of Δ_{47} -soil T and SST during the Cenomanian-Santonian interval. Note the decreasing soil temperature trend compared with relatively stable SST trend in the Turonian-

Santonian interval suggests that the observed soil cooling was unlikely resulted from contemporaneous global climate change.

Global climate change could alter seasonality of rainfall and therefore change soil carbonate growth seasons (e.g., Quade et al., 2013). Studies have shown that seasonality of soil carbonate precipitation has a vital effect on interpretation of Δ_{47} -soil Ts as temperatures recorded by soil carbonates mainly reflect environmental condition from which they precipitated (e.g., Hough et al., 2014; Peters et al., 2013; Quade et al., 2013). Model-generated precipitation data show that SE Tibet was located in an arid zone with summer precipitation less than 1 mm/day throughout the entire Cenomanian-Santonian interval (Fig. 7). The relatively low summer rainfall amount combined with high summer soil temperatures shown by both Δ_{47} proxy and climate model (Figs. 4, 5) suggest that soil dewatering and subsequent precipitation of soil carbonates mostly happened in summer months (e.g., Breecker et al., 2009; Quade et al., 2013). Therefore, the Δ_{47} -Ts of Markam soil carbonates consistently reflect warm summer soil temperatures during the Cenomanian-Santonian time. Together, multiple lines of evidence discussed above suggest that global climate change cannot account for the observed soil temperature cooling in SE Tibet during the Turonian-Coniacian time (Fig. 8).

Finally, global climate change could also affect $\delta^{18}\text{O}$ values of precipitation and therefore affects $\delta^{18}\text{O}$ values of soil water ($\delta^{18}\text{O}_{\text{water}}$). Climatic variables that can influence meteoric $\delta^{18}\text{O}$ compositions include atmospheric circulation pattern, air temperature and vapor source (Rowley, 2007). Simulation of Cretaceous wind patterns over the area of SE Tibet show that circulation patterns were unchanged during the Cenomanian-Santonian interval, with wind blowing constantly from northwest to southeast carrying moistures derived from the east Tethys Ocean

(Fig. 7). The uniform circulation patterns suggest vapor sources were unlikely changed during the studied time interval. Therefore, the observed decrease of $\delta^{18}\text{O}_{\text{water}}$ values at the Turonian-Santonian interval (Fig. 4) is unlikely caused by global climate change, especially when considering the very stable coeval global temperatures inferred from both marine records and climate simulations (Fig. 8A, C).

Tectonic consideration

Besides global climate change, land surface temperatures can be influenced by tectonic forces expressed as lateral or vertical movement of continental plates. Horizontal plate movement (or continental drift) can change surface temperatures of a region by moving the place from one climatic zone to another. The effect of continental drift on soil temperatures in SE Tibet can be assessed by paleogeographic reconstructions. Paleo-latitude estimate suggests position of SE Tibet was relatively stable at $\sim 20\text{--}25^\circ \text{N}$ during Late Cretaceous (Lippert et al., 2014); therefore paleo-latitudinal variations should have minimum effects on soil temperatures in our study area. On the other hand, vertical plate movement expressed as surface elevation changes can affect surface temperature and $\delta^{18}\text{O}_{\text{water}}$ values as well (e.g., Rowley and Garzzone, 2007). In our study site, average Δ_{47} -soil temperature and $\delta^{18}\text{O}_{\text{water}}$ values decreased $\sim 6^\circ \text{C}$ and $\sim 1.0\text{‰}$, respectively during the Turonian-Santonian time interval (Fig. 4), which cannot explained by coeval global climate change or paleogeographic variations as previously discussed. Therefore, it is logical to speculate that the observed systematic decrease of both soil temperature and $\delta^{18}\text{O}_{\text{water}}$ values was due to local elevation change, especially when similar implications that part of Tibet may had been an uplifted region as early as Late Cretaceous have

been drawn from previous studies (e.g., Kapp et al., 2007).

The inference that our study area may experienced surface uplift during Late Cretaceous is supported by evidence of strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) ratios and structural features. The strontium isotope results show that soil carbonate $^{87}\text{Sr}/^{86}\text{Sr}$ values gradually increase up-section from ~ 0.7104 in the Cenomanian-Turonian interval to ~ 0.7112 in the Coniacian-Santonian interval, coincident with decreases of both soil temperature and $\delta^{18}\text{O}_{\text{water}}$ values (Fig. 4). In continental setting, $^{87}\text{Sr}/^{86}\text{Sr}$ values of sedimentary rocks are typically related to weathering of rubidium-rich minerals (e.g., K-feldspar, hornblende, micas) in source areas and therefore can be used as a proxy for weathering intensity (Mulch et al., 2010). Meanwhile, geologic bodies usually bear distinct $^{87}\text{Sr}/^{86}\text{Sr}$ signatures and thus the $^{87}\text{Sr}/^{86}\text{Sr}$ signals preserved in sedimentary rocks can be used as provenance tracers. Detrital zircon age and paleo-current data show that sediments of the Nanxing Formation were sourced primarily from the recycled Upper Triassic strata in Qamdo basin, as well as volcanic rocks of the Jiangda-Weixi arc (JWA) (Shang, 2014). The JWA plutonic rocks have $^{87}\text{Sr}/^{86}\text{Sr}$ ratios range from 7.099 to 0.7113 (Deng et al., 2014; Zi et al., 2012), whereas Late Triassic crystalline rocks of the Yidun terrane (i.e., a major sediment contributor for Upper Triassic Qamdo strata) (Shang et al., 2014) are characterized by lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios ranging from 7.051 to 0.7059 (Deng et al., 2014). We suggest that the increase of $^{87}\text{Sr}/^{86}\text{Sr}$ values in Markam paleosols from ~ 0.7104 to ~ 0.7112 reflects a change of source areas during the Coniacian-Santonian interval. The $^{87}\text{Sr}/^{86}\text{Sr}$ increase was probably associated with enhanced exhumation and weathering of the JWA, implying active tectonics in our study area during the Coniacian-Santonian time. The interpretation is also consistent with increasing sandstone/mudstone ratios up-section (Fig. 4), which suggests enhanced siliciclastic influx during Coniacian-Santonian time. Another piece of evidence for active tectonics in our study site

during Late Cretaceous is the syntectonic depositions revealed by lateral variations of stratal thickness, or ‘growth strata’. The structure is expressed by a gradual decrease of measured bed dip angles, with average dip angles changing from ~65 degree in the Cenomanian-Turonian interval to ~40 degree in the Coniacian-Santonian (Fig. 9). Together, the increase of $^{87}\text{Sr}/^{86}\text{Sr}$ values and syn-tectonic depositions both indicate active tectonisms, and by inference, early Late Cretaceous (Coniacian-Santonian) topographic growth in SE Tibet.

Figure 9. Syntectonic deposition of the Late Cretaceous Naxing Formation indicated by variation of dip angles. Cen.-Tur. —Cenomanian-Turonian strata; Con.-San. —Coniacian-Santonian strata.

Estimation of Late Cretaceous paleo-elevation change for SE Tibet

During the period of Turonian-Santonian, an average soil temperature drop of $\sim 6^\circ\text{C}$ and a $\sim 1.0\text{‰}$ decrease of average $\delta^{18}\text{O}_{\text{water}}$ values (Fig. 10) were likely caused by local surface uplift. If we assume a present-day global average temperature-elevation gradient of $\sim 6^\circ\text{C}/\text{km}$ (Wolfe, 1992), the observed cooling would likely corresponds to a surface elevation gain of $\sim 1\text{ km}$. However, taking the extreme equable mid to Late Cretaceous climate condition in to consideration (e.g., Huber et al., 1995), a temperature-elevation gradient of $\sim 6^\circ\text{C}/\text{km}$ is probably an overestimation, causing underestimation of paleo-elevation changes. Similarly, the present-day global average oxygen isotopic lapse rate is $\sim 2.8\text{‰}/\text{km}$ (Poage and Chamberlain, 2002), however, model simulations show that warmer climates tend to produce much shallower oxygen isotopic lapse rates (e.g., $\sim 1.2\text{‰}/\text{km}$ when surface temperature $> 30^\circ\text{C}$) (Rowley and Garzzone, 2007). During early Late Cretaceous, summer surface temperatures exceeded 30°C in our study site (Fig. 4; Table. 3) and the $\sim 1.0\text{‰}$ decrease of $\delta^{18}\text{O}_{\text{water}}$ likely corresponds to a surface elevation gain $\sim 1\text{ km}$.

Figure 10. Cross-plot of $\Delta_{47}\text{-soil}$ temperature and $\delta^{18}\text{O}_{\text{water}}$ values of Markam paleosol carbonates. Red dot represents samples from the Turonian interval (compiled from Shang et al.,

in review) and blue dot marks the data analyzed in this study.

It is worth to note that, however, we are unable to estimate the pre-uplift topography of the study site due to lack of control points from low altitude regions (i.e., coeval terrestrial temperatures close to sea-level). However, abundant conglomerates and localized provenance of detrital zircons (Shang et al., 2014), combined with relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of paleosol carbonates (>0.70900) (Fig. 4) suggest that the study area must have been uplifted above the sea level before Late Cretaceous. We suggest that a moderate topography >500 m was probably established prior to the Cenomanian time. The inference is also consistent with Late Cretaceous paleogeographic reconstructions provided by Geotec Inc. showing SE Tibet was characterized by a moderate topography $\sim 1\text{km}$ during the Cenomanian-Santonian time interval (Fig. 2). Together, we suggest that paleo-elevation in our study site was unlikely lower than 1.5 km by the Santonian stage if we add this ‘pre-uplift’ topography into our reconstruction.

Tectonic implication

Our study reveals an episode early Late Cretaceous (Coniacian-Santonian) topographic growth in SE Tibet which broadly supports the Cretaceous ‘proto-plateau’ hypothesis that part of Tibet may have been uplifted prior to the Indo-Asian collision (e.g., Kapp et al., 2005; 2007; Wang et al., 2008; Sun et al., 2015). However, unlike most studies which infer topographic growth in the Lhasa block (e.g., Wang et al., 2008; Sun et al., 2015). The source to sink relationship indicated by paleocurrent and detrital zircon data suggests that at least the southern Yidun terrane was an elevated topography during early Late Cretaceous, whereas a more extensive regional uplift including the Yidun and eastern Songpan-Ganzi terranes can be inferred

from rapid rock cooling histories (e.g., Tian et al., 2014), implying high topography may have been spatially more extensive.

Topographic growth of the SE Tibetan Plateau has been attributed to ductile flow of the lower crust from the thickened central Tibetan lithosphere to the margin of SE Tibet during middle to late Miocene time (Clark and Royden, 2000). However, recent study suggests the eastward migration of lower crustal flow was probably not a major cause of uplift of SE Tibet as paleoaltimetry study shows the margin of SE Tibet was at its current elevation ~3 km by late Eocene (Hoke et al., 2014). Our study provides further evidence that a Late Cretaceous ‘proto-plateau’ with elevation >1.5 km was probably established in SE Tibet by Santonian (~83 Ma) and the modern day Tibetan Plateau may inherit some elevation from this paleo-topography.

Conclusion

By reconstructing soil temperature history, our study reveals an episode of Late Cretaceous (Coniacian-Santonian) topographic growth in SE Tibet. Paleo-elevation in our study site was likely >1.5 km by Santonian (~83 Ma), whereas a more extensive high topography may have existed in the Yidun and eastern Songpan-Ganzi terrane as inferred from sediment provenance and thermochronology evidence. Late Cretaceous topographic growth in SE Tibet was probably associated with compressional strain field created by the Lhasa-Qiangtang collision. Our study provides evidence that the modern-day high elevation of SE Tibet is partially inherited from topography developed prior to the India-Asian collision.

Reference

- Blakey, R.C., 2008. Gondwana paleogeography from assembly to breakup—A 500 my odyssey. Geological Society of America Special Papers, 441: 1-28.
- Blisniuk, P.M. and Stern, L.A., 2005. Stable isotope paleoaltimetry: a critical review. American Journal of Science, 305(10): 1033-1074.
- Bornemann, A., Norris, R.D., Friedrich, O., Beckmann, B., Schouten, S., Damsté, J.S.S., Vogel, J., Hofmann, P. and Wagner, T., 2008. Isotopic evidence for glaciation during the Cretaceous supergreenhouse. Science, 319(5860): 189-192.
- Breecker, D., Sharp, Z. and McFadden, L.D., 2009. Seasonal bias in the formation and stable isotopic composition of pedogenic carbonate in modern soils from central New Mexico, USA. Geological Society of America Bulletin, 121(3-4): 630-640.
- Carrapa, B., Huntington, K.W., Clementz, M., Quade, J., Bywater-Reyes, S., Schoenbohm, L.M. and Canavan, R.R., 2014. Uplift of the Central Andes of NW Argentina associated with upper crustal shortening, revealed by multiproxy isotopic analyses. Tectonics, 33(6): 1039-1054.
- Clark, M.K. and Royden, L.H., 2000. Topographic ooze: Building the eastern margin of Tibet by lower crustal flow. Geology, 28(8): 703-706.
- Clarke, L.J. and Jenkyns, H.C., 1999. New oxygen isotope evidence for long-term Cretaceous climatic change in the Southern Hemisphere. Geology, 27(8): 699-702.
- Craig, H. and Gordon, L.I., 1965. Deuterium and oxygen 18 variations in the ocean and the marine atmosphere.
- Currie, B.S., Rowley, D.B. and Tabor, N.J., 2005. Middle Miocene paleoaltimetry of southern Tibet: implications for the role of mantle thickening and delamination in the Himalayan orogen. Geology, 33(3): 181-184.
- Dansgaard, W., 1964. Stable isotopes in precipitation. Tellus A, 16(4).
- Decelles, P.G., Kapp, P., Ding, L. and Gehrels, G., 2007. Late Cretaceous to middle Tertiary basin evolution in the central Tibetan Plateau: Changing environments in response to tectonic partitioning, aridification, and regional elevation gain. Geological Society of America Bulletin, 119(5-6): 654-680.
- Deng, B., Liu, S., Li, Z., Cao, J. and Sun, W., 2012. Late cretaceous tectonic change of the eastern margin of the Tibetan Plateau—results from multisystem thermochronology. Journal of the Geological Society of India, 80(2): 241-254.
- Deng, J., Wang, Q., Li, G., Li, C. and Wang, C., 2014. Tethys tectonic evolution and its bearing on the distribution of important mineral deposits in the Sanjiang region, SW China. Gondwana Research, 26(2): 419-437.
- Dennis, K.J., Affek, H.P., Passey, B.H., Schrag, D.P. and Eiler, J.M., 2011. Defining an absolute reference frame for ‘clumped’ isotope studies of CO₂. Geochimica et Cosmochimica Acta, 75(22): 7117-7131.
- Dewey, J.F. and Burke, K.C., 1973. Tibetan, Variscan, and Precambrian basement reactivation: products of continental collision. The Journal of Geology: 683-692.

- Dewey, J.F., Shackleton, R.M., Chengfa, C. and Yiyin, S., 1988. The tectonic evolution of the Tibetan Plateau. *Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, 327(1594): 379-413.
- Digital Global, 2015. Source <https://www.google.com/maps/@29.6886303,98.5912691,2205m/data=!3m1!1e3>
- Du, D., Luo, J. and Li, X., 1997. Sedimentary evolution and palaeogeography of the Qamdo Block in Xizang. *Lithofacies Palaeogeogr*, 17(4): 1-17.
- Eiler, J.M., 2007. “Clumped-isotope” geochemistry—The study of naturally-occurring, multiply-substituted isotopologues. *Earth and Planetary Science Letters*, 262(3): 309-327.
- Eiler, J.M. and Schauble, E., 2004. ^{18}O ^{13}C ^{16}O in Earth’s atmosphere. *Geochimica et Cosmochimica Acta*, 68(23): 4767-4777.
- England, P. and Searle, M., 1986. The Cretaceous-Tertiary deformation of the Lhasa block and its implications for crustal thickening in Tibet. *Tectonics*, 5(1): 1-14.
- Fan, M., Hough, B.G. and Passey, B.H., 2014. Middle to late Cenozoic cooling and high topography in the central Rocky Mountains: Constraints from clumped isotope geochemistry. *Earth and Planetary Science Letters*, 408: 35-47.
- Forster, A., Schouten, S., Baas, M. and Damsté, J.S.S., 2007. Mid-Cretaceous (Albian–Santonian) sea surface temperature record of the tropical Atlantic Ocean. *Geology*, 35(10): 919-922.
- Garzanti, E., 2008. Comment on “When and where did India and Asia collide?” by Jonathan C. Aitchison, Jason R. Ali, and Aileen M. Davis. *Journal of Geophysical Research: Solid Earth* (1978–2012), 113(B4).
- Garzione, C.N., Auerbach, D.J., Smith, J.J.-S., Rosario, J.J., Passey, B.H., Jordan, T.E. and Eiler, J.M., 2014. Clumped isotope evidence for diachronous surface cooling of the Altiplano and pulsed surface uplift of the Central Andes. *Earth and Planetary Science Letters*, 393: 173-181.
- Garzione, C.N., Dettman, D.L. and Horton, B.K., 2004. Carbonate oxygen isotope paleoaltimetry: evaluating the effect of diagenesis on paleoelevation estimates for the Tibetan plateau. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 212(1): 119-140.
- Garzione, C.N., Quade, J., DeCelles, P.G. and English, N.B., 2000. Predicting paleoelevation of Tibet and the Himalaya from $\delta^{18}\text{O}$ vs. altitude gradients in meteoric water across the Nepal Himalaya. *Earth and Planetary Science Letters*, 183(1): 215-229.
- Ghosh, P., Adkins, J., Affek, H., Balta, B., Guo, W., Schauble, E.A., Schrag, D. and Eiler, J.M., 2006. ^{13}C – ^{18}O bonds in carbonate minerals: A new kind of paleothermometer. *Geochimica et Cosmochimica Acta*, 70(6): 1439-1456.
- Gordon, C., Cooper, C., Senior, C.A., Banks, H., Gregory, J.M., Johns, T.C., Mitchell, J.F. and Wood, R.A., 2000. The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments. *Climate dynamics*, 16(2-3): 147-168.
- Gradstein, F.M., Ogg, J.G. and Smith, A.G., 2004. A geologic time scale 2004, 86. Cambridge University Press.
- Graham, S.A., Chamberlain, C.P., Yue, Y., Ritts, B.D., Hanson, A.D., Horton, T.W., Waldbauer, J.R., Poage, M.A. and Feng, X., 2005. Stable isotope records of Cenozoic climate and topography, Tibetan plateau and Tarim basin. *American Journal of Science*, 305(2): 101-118.

- Henkes, G.A., Passey, B.H., Grossman, E.L., Shenton, B.J., Pérez-Huerta, A. and Yancey, T.E., 2014. Temperature limits for preservation of primary calcite clumped isotope paleotemperatures. *Geochimica et cosmochimica acta*, 139: 362-382.
- Henkes, G.A., Passey, B.H., Wanamaker, A.D., Grossman, E.L., Ambrose, W.G. and Carroll, M.L., 2013. Carbonate clumped isotope compositions of modern marine mollusk and brachiopod shells. *Geochimica et Cosmochimica Acta*, 106: 307-325.
- Hetzel, R., Dunkl, I., Haider, V., Strobl, M., von Eynatten, H., Ding, L. and Frei, D., 2011. Peneplain formation in southern Tibet predates the India-Asia collision and plateau uplift. *Geology*, 39(10): 983-986.
- Hoke, G.D., Liu-Zeng, J., Hren, M.T., Wissink, G.K. and Garzione, C.N., 2014. Stable isotopes reveal high southeast Tibetan Plateau margin since the Paleogene. *Earth and Planetary Science Letters*, 394: 270-278.
- Hough, B.G., Fan, M. and Passey, B.H., 2014. Calibration of the clumped isotope geothermometer in soil carbonate in Wyoming and Nebraska, USA: Implications for paleoelevation and paleoclimate reconstruction. *Earth and Planetary Science Letters*, 391: 110-120.
- Hren, M.T., Bookhagen, B., Blisniuk, P.M., Booth, A.L. and Chamberlain, C.P., 2009. $\delta^{18}\text{O}$ and δD of streamwaters across the Himalaya and Tibetan Plateau: Implications for moisture sources and paleoelevation reconstructions. *Earth and Planetary Science Letters*, 288(1): 20-32.
- Hren, M.T. and Sheldon, N.D., 2012. Temporal variations in lake water temperature: Paleoenvironmental implications of lake carbonate $\delta^{18}\text{O}$ and temperature records. *Earth and Planetary Science Letters*, 337: 77-84.
- Huang, K., Opdyke, N.D., Li, J. and Peng, X., 1992. Paleomagnetism of Cretaceous rocks from eastern Qiangtang terrane of Tibet. *Journal of Geophysical Research: Solid Earth* (1978–2012), 97(B2): 1789-1799.
- Huber, B.T., Hodell, D.A. and Hamilton, C.P., 1995. Middle–Late Cretaceous climate of the southern high latitudes: stable isotopic evidence for minimal equator-to-pole thermal gradients. *Geological Society of America Bulletin*, 107(10): 1164-1191.
- Huber, B.T., Norris, R.D. and MacLeod, K.G., 2002. Deep-sea paleotemperature record of extreme warmth during the Cretaceous. *Geology*, 30(2): 123-126.
- Huntington, K.W., Saylor, J., Quade, J. and Hudson, A.M., 2015. High late Miocene–Pliocene elevation of the Zhada Basin, southwestern Tibetan Plateau, from carbonate clumped isotope thermometry. *Geological Society of America Bulletin*, 127(1-2): 181-199.
- Jarvis, I., Lignum, J.S., Gröcke, D.R., Jenkyns, H.C. and Pearce, M.A., 2011. Black shale deposition, atmospheric CO_2 drawdown, and cooling during the Cenomanian-Turonian Oceanic Anoxic Event. *Paleoceanography*, 26(3).
- Kapp, P., DeCelles, P.G., Gehrels, G.E., Heizler, M. and Ding, L., 2007. Geological records of the Lhasa-Qiangtang and Indo-Asian collisions in the Nima area of central Tibet. *Geological Society of America Bulletin*, 119(7-8): 917-933.
- Kapp, P., Murphy, M.A., Yin, A., Harrison, T.M., Ding, L. and Guo, J., 2003. Mesozoic and Cenozoic tectonic evolution of the Shiquanhe area of western Tibet. *Tectonics*, 22(4).
- Kapp, P., Yin, A., Harrison, T.M. and Ding, L., 2005. Cretaceous-Tertiary shortening, basin development, and volcanism in central Tibet. *Geological Society of America Bulletin*, 117(7-8): 865-878.

- Kent-Corson, M.L., Ritts, B.D., Zhuang, G., Bovet, P.M., Graham, S.A. and Chamberlain, C.P., 2009. Stable isotopic constraints on the tectonic, topographic, and climatic evolution of the northern margin of the Tibetan Plateau. *Earth and Planetary Science Letters*, 282(1): 158-166.
- Kim, S.-T. and O'Neil, J.R., 1997. Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochimica et Cosmochimica Acta*, 61(16): 3461-3475.
- Kump, L.R. and Pollard, D., 2008. Amplification of Cretaceous warmth by biological cloud feedbacks. *Science*, 320(5873): 195-195.
- Leier, A., McQuarrie, N., Garzione, C. and Eiler, J., 2013. Stable isotope evidence for multiple pulses of rapid surface uplift in the Central Andes, Bolivia. *Earth and Planetary Science Letters*, 371: 49-58.
- Leier, A.L., Kapp, P., Gehrels, G.E. and DeCelles, P.G., 2007. Detrital zircon geochronology of Carboniferous–Cretaceous strata in the Lhasa terrane, Southern Tibet. *Basin Research*, 19(3): 361-378.
- Linnert, C., Robinson, S.A., Lees, J.A., Bown, P.R., Pérez-Rodríguez, I., Petrizzo, M.R., Falzoni, F., Littler, K., Arz, J.A. and Russell, E.E., 2014. Evidence for global cooling in the Late Cretaceous. *Nature communications*, 5.
- Lippert, P.C., Van Hinsbergen, D.J. and Dupont-Nivet, G., 2014. Early Cretaceous to present latitude of the central proto-Tibetan Plateau: A paleomagnetic synthesis with implications for Cenozoic tectonics, paleogeography, and climate of Asia. *Geological Society of America Special Papers*, 507: SPE507-01.
- Lunt, D.J., Haywood, A.M., Schmidt, G.A., Salzmann, U., Valdes, P.J. and Dowsett, H.J., 2010. Earth system sensitivity inferred from Pliocene modelling and data. *Nature Geoscience*, 3(1): 60-64.
- Lunt, D. J., Farnsworth, A., Loptson, C., Foster, G. L., Markwick, P., O'Brien, C. L., Pancost, R. D., Robinson, S. A., and Wrobel, N., 2015, Palaeogeographic controls on climate and proxy interpretation, *Clim. Past Discuss.*, 11, 5683-5725, doi:10.5194/cpd-11-5683-2015, 2015.
- Mack, G.H., James, W.C. and Monger, H.C., 1993. Classification of paleosols. *Geological Society of America Bulletin*, 105(2): 129-136.
- Markwick, P., 2007. The palaeogeographic and palaeoclimatic significance of climate proxies for data-model comparisons. *Deep-time perspectives on climate change: marrying the signal from computer models and biological proxies*: 251-312.
- Meyer, H.W., 1992. Lapse rates and other variables applied to estimating paleoaltitudes from fossil floras. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 99(1): 71-99.
- Meyer, H.W., 2007. A review of paleotemperature–lapse rate methods for estimating paleoelevation from fossil floras. *Reviews in Mineralogy and Geochemistry*, 66(1): 155-171.
- Mulch, A., Uba, C.E., Strecker, M.R., Schoenberg, R. and Chamberlain, C.P., 2010. Late Miocene climate variability and surface elevation in the central Andes. *Earth and Planetary Science Letters*, 290(1): 173-182.
- Murphy, M., Yin, A., Harrison, T., Dürr, S., Chen, Z., Ryerson, F., Kidd, W., Wang, X. and Zhou, X., 1997. Did the Indo-Asian collision alone create the Tibetan plateau? *Geology*, 25(8): 719-722.

- NGDR, 1999 National Geophysical Data Center, 1999. Global Land One-kilometer Base Elevation (GLOBE) v.1. Hastings, D. and P.K. Dunbar. National Geophysical Data Center, NOAA. doi:10.7289/V52R3PMS [access date: 2015-03-16].
- Norris, R.D., Bice, K.L., Magno, E.A. and Wilson, P.A., 2002. Jiggling the tropical thermostat in the Cretaceous hothouse. *Geology*, 30(4): 299-302.
- Otofujii, Y.-i., Inoue, Y., Funahara, S., Murata, F. and Zheng, X., 1990. Palaeomagnetic study of eastern Tibet-deformation of the Three Rivers region. *Geophysical Journal International*, 103(1): 85-94.
- Pan, G.-t., Ding, J., Yao, D.-s. and Wang, L.-q., 2004. Geological map of the Qinghai-Xizang (Tibet) Plateau and adjacent areas. Chengdu Cartographic Publishing House.
- Passey, B.H., Levin, N.E., Cerling, T.E., Brown, F.H. and Eiler, J.M., 2010. High-temperature environments of human evolution in East Africa based on bond ordering in paleosol carbonates. *Proceedings of the National Academy of Sciences*, 107(25): 11245-11249.
- Peters, N.A., Huntington, K.W. and Hoke, G.D., 2013. Hot or not? Impact of seasonally variable soil carbonate formation on paleotemperature and O-isotope records from clumped isotope thermometry. *Earth and Planetary Science Letters*, 361: 208-218.
- Poage, M. and Chamberlain, C., 2002. Stable isotopic evidence for a pre-Middle Miocene rain shadow in the western Basin and Range: Implications for the paleotopography of the Sierra Nevada. *Tectonics*, 21(4): 16-1-16-10.
- Poage, M.A. and Chamberlain, C.P., 2001. Empirical relationships between elevation and the stable isotope composition of precipitation and surface waters: considerations for studies of paleoelevation change. *American Journal of Science*, 301(1): 1-15.
- Polissar, P.J., Freeman, K.H., Rowley, D.B., McInerney, F.A. and Currie, B.S., 2009. Paleoaltimetry of the Tibetan Plateau from D/H ratios of lipid biomarkers. *Earth and Planetary Science Letters*, 287(1): 64-76.
- Powell, C.M. and Conaghan, P., 1973. Plate tectonics and the Himalayas. *Earth and Planetary Science Letters*, 20(1): 1-12.
- Quade, J., Breecker, D.O., Daëron, M. and Eiler, J., 2011. The paleoaltimetry of Tibet: An isotopic perspective. *American Journal of Science*, 311(2): 77-115.
- Quade, J., Eiler, J., Daëron, M. and Achyuthan, H., 2013. The clumped isotope geothermometer in soil and paleosol carbonate. *Geochimica et Cosmochimica Acta*, 105: 92-107.
- Quade, J., Garzione, C. and Eiler, J., 2007. Paleoelevation reconstruction using pedogenic carbonates. *Reviews in Mineralogy and Geochemistry*, 66(1): 53-87.
- Ray, S., Gault, H. and Dodd, C.G., 1957. The separation of clay minerals from carbonate rocks. *Am. Mineral*, 42: 681-686.
- Reid, A.J., Wilson, C.J. and Liu, S., 2005. Structural evidence for the Permo-Triassic tectonic evolution of the Yidun Arc, eastern Tibetan Plateau. *Journal of Structural Geology*, 27(1): 119-137.
- Renssen, H., Beets, C., Fichet, T., Goosse, H. and Kroon, D., 2004. Modeling the climate response to a massive methane release from gas hydrates. *Paleoceanography*, 19(2).
- Rohrmann, A., Kapp, P., Carrapa, B., Reinert, P.W., Guynn, J., Ding, L. and Heizler, M., 2012. Thermochronologic evidence for plateau formation in central Tibet by 45 Ma. *Geology*, 40(2): 187-190.
- Rohrmann, A., Strecker, M.R., Bookhagen, B., Mulch, A., Sachse, D., Pingel, H., Alonso, R.N., Schildgen, T.F. and Montero, C., 2014. Can stable isotopes ride out the storms? The role

- of convection for water isotopes in models, records, and paleoaltimetry studies in the central Andes. *Earth and Planetary Science Letters*, 407: 187-195.
- Rowley, D.B., 1996. Age of initiation of collision between India and Asia: A review of stratigraphic data. *Earth and Planetary Science Letters*, 145(1): 1-13.
- Rowley, D.B., 2007. Stable isotope-based paleoaltimetry: Theory and validation. *Reviews in Mineralogy and Geochemistry*, 66(1): 23-52.
- Rowley, D.B. and Garzzone, C.N., 2007. Stable isotope-based paleoaltimetry. *Annu. Rev. Earth Planet. Sci.*, 35: 463-508.
- Royer, D., Pagani, M. and Beerling, D., 2012. Geobiological constraints on Earth system sensitivity to CO₂ during the Cretaceous and Cenozoic. *Geobiology*, 10(4): 298-310.
- Saylor, J., Quade, J., Dettman, D., DeCelles, P., Kapp, P. and Ding, L., 2009. The late Miocene through present paleoelevation history of southwestern Tibet. *American Journal of Science*, 309(1): 1-42.
- SHANG, F., 2014. DETRITAL ZIRCON GEOCHRONOLOGY FROM THE MESOZOIC QAMDO (CHANGDU) BASIN, SOUTHEASTERN TIBET: IMPLICATIONS FOR THE PALEO-AND MESO-TETHYS EVOLUTION, 2014 GSA Annual Meeting in Vancouver, British Columbia.
- Snell, K.E., Koch, P.L., Druschke, P., Foreman, B.Z. and Eiler, J.M., 2014. High elevation of the 'Nevadaplano' during the Late Cretaceous. *Earth and Planetary Science Letters*, 386: 52-63.
- Snell, K.E., Thrasher, B.L., Eiler, J.M., Koch, P.L., Sloan, L.C. and Tabor, N.J., 2013. Hot summers in the Bighorn Basin during the early Paleogene. *Geology*, 41(1): 55-58.
- Studnicki-Gizbert, C., Burchfiel, B., Li, Z. and Chen, Z., 2008. Early Tertiary Gonjo basin, eastern Tibet: Sedimentary and structural record of the early history of India-Asia collision. *Geosphere*, 4(4): 713-735.
- Sun, G., Hu, X., Sinclair, H.D., BouDagher-Fadel, M.K. and Wang, J., 2015. Late Cretaceous evolution of the Coqen Basin (Lhasa terrane) and implications for early topographic growth on the Tibetan Plateau. *Geological Society of America Bulletin*: B31137. 1.
- Tian, Y., Kohn, B.P., Hu, S. and Gleadow, A.J., 2014. Postorogenic rigid behavior of the eastern Songpan-Ganze terrane: Insights from low-temperature thermochronology and implications for intracontinental deformation in central Asia. *Geochemistry, Geophysics, Geosystems*, 15(2): 453-474.
- Tripathi, A.K., Eagle, R.A., Thiagarajan, N., Gagnon, A.C., Bauch, H., Halloran, P.R. and Eiler, J.M., 2010. 13 C–18 O isotope signatures and 'clumped isotope' thermometry in foraminifera and coccoliths. *Geochimica et Cosmochimica Acta*, 74(20): 5697-5717.
- Tripathi, A.K., Hill, P.S., Eagle, R.A., Mosenfelder, J.L., Tang, J., Schauble, E.A., Eiler, J.M., Zeebe, R.E., Uchikawa, J. and Coplen, T.B., 2015. Beyond temperature: Clumped isotope signatures in dissolved inorganic carbon species and the influence of solution chemistry on carbonate mineral composition. *Geochimica et Cosmochimica Acta*, 166: 344-371.
- Voigt, S., Gale, A.S. and Flögel, S., 2004. Midlatitude shelf seas in the Cenomanian-Turonian greenhouse world: Temperature evolution and North Atlantic circulation. *Paleoceanography*, 19(4).
- Volkmer, J.E., Kapp, P., Guynn, J.H. and Lai, Q., 2007. Cretaceous-Tertiary structural evolution of the north central Lhasa terrane, Tibet. *Tectonics*, 26(6).

- Wang, C., Zhao, X., Liu, Z., Lippert, P.C., Graham, S.A., Coe, R.S., Yi, H., Zhu, L., Liu, S. and Li, Y., 2008. Constraints on the early uplift history of the Tibetan Plateau. *Proceedings of the National Academy of Sciences*, 105(13): 4987-4992.
- Wang, X.-S., Hu, R.-Z., Bi, X.-W., Leng, C.-B., Pan, L.-C., Zhu, J.-J. and Chen, Y.-W., 2014. Petrogenesis of Late Cretaceous I-type granites in the southern Yidun Terrane: New constraints on the Late Mesozoic tectonic evolution of the eastern Tibetan Plateau. *Lithos*, 208: 202-219.
- Wilson, C.J. and Fowler, A.P., 2011. Denudational response to surface uplift in east Tibet: Evidence from apatite fission-track thermochronology. *Geological Society of America Bulletin*, 123(9-10): 1966-1987.
- Wolfe, J.A., 1992. An analysis of present-day terrestrial lapse rates in the western conterminous United States and their significance to paleoaltitudinal estimates, US Geological Survey, US, GPO; Books and Open-File Reports Section [distributor].
- Zhang, K.-J., Zhang, Y.-X., Tang, X.-C. and Xia, B., 2012. Late Mesozoic tectonic evolution and growth of the Tibetan plateau prior to the Indo-Asian collision. *Earth-Science Reviews*, 114(3): 236-249.
- Zi, J.W., Cawood, P.A., Fan, W.M., Wang, Y.J. and Tohver, E., 2012. Contrasting rift and subduction-related plagiogranites in the Jinshajiang ophiolitic mélangé, southwest China, and implications for the Paleo-Tethys. *Tectonics*, 31(2).