DEFORMATION OF LACUSTRINE SHORELINES IN CENTRAL TIBET:
IMPLICATIONS FOR LAKE LEVEL HISTORY, FAULT KINEMATICS, AND
CRUSTAL RHEOLOGY

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Xuhua Shi

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The dissertation of Xuhua Shi was reviewed and approved* by the following:

Eric Kirby  
Associate Professor of Geosciences  
Dissertation Co-Advisor  
Co-Chair of Committee

Kevin P. Furlong  
Professor of Geosciences  
Dissertation Co-Adviser  
Co-Chair of Committee

Donald M. Fisher  
Professor of Geosciences

Robert Crane  
Professor of Geography

Chris J. Marone  
Professor of Geosciences  
Associate Department Head of Graduate Programs

*Signatures are on file in the Graduate School
Quantification of the rheology of Tibetan lithosphere is critical to understanding the
deformational processes that drive the growth of the Tibetan plateau, but remains poorly understood. Here
I constrain effective elastic thickness ($T_e$) of Tibetan crust by measuring surface deformation of the
highstand shorelines (~ 64 m above the lake level in 1976) around Siling Co - the largest lake in central
Tibet - that occurred in response to climatically induced changes in lake levels. Optically-stimulated
luminescence (OSL) dating of numerous shorelines around Siling Co indicate this lake was reached at the
highstand at 6 – 4 ka and subsequently fell to present-day levels. This Middle Holocene lake highstand
was reached or exceeded several times during each period of 100 – 200 ka and 40 – 10 ka.

The highstand shorelines are deflected by 3 – 5 m over wavelengths of 10s of km. Comparison of
measured shoreline deflections with a 3D elastic model provides a $T_e$ of ~ 12 – 14 km in central Tibet.
When combined with geophysical constraints on crustal structure, composition, and heat flow, the strain
rates implied by flexural rebound require a relatively low viscosity middle crust (effective viscosity of ~
$10^{18}$ - $10^{20}$ Pa s) at ~ 20 – 40 km depth, consistent with the view of relatively weak deep crust beneath
Tibet and the notion of channelized crustal flow in this region.

Fault kinematics of the Tibetan plateau provides alternative means to examine the crustal
deformation. Here I document 12 ± 1 m of right-lateral displacement of lacustrine shorelines across the
Gyaring Co fault, one of the primary active strike-slip faults in central Tibet. OSL ages of the shorelines
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CHAPTER 1

INTRODUCTION

The broad region of Himalayan-Tibetan orogenic belts (Figure 1-1) lies at the center of geoscience research for its archetypal India-Asia collision and associated lithospheric deformation, both responsible for the high topography of the Tibetan plateau. This region also provides a great natural laboratory for geoscientists to explore the role of tectonic interaction with climate and erosion in shaping the topography (e.g. Raymo and Ruddiman, 1992; Beaumont et al., 2001; Whipple, 2009; Molnar, 2010 and references therein). Yet, how the Tibetan lithosphere deforms remains an unresolved fundamental question that relates to the growth of high topography of the plateau and its potential influence on paleoclimate change (Molnar et al., 1993; 2010). Three end member models of deformational processes have been proposed to explain how the high topography is generated and maintained. The first model of deformation, known as the rigid block model (Tapponnier et al., 1982; 2001), argues high strain localization along several crustal-scale strike-slip faults (Peltzer and Tapponnier, 1988; Avouac and Tapponnier, 1993) bounding relatively rigid blocks that comprise the plateau (Figure 1-1). Therefore the plateau is uplifted successively toward the east and northeast. This model requires large crustal strength and rapid fault slip (> 20 mm/yr, e.g., Mériaux et al., 2004; Chevalier et al., 2005) along the major strike-slip fault systems to accommodate nearly half of the India-Asian convergence in the interior of the Tibetan plateau (total modern convergence rate: ~ 40 mm/yr, e.g., Paul et al., 2001; Wang et al., 2001; Zhang et al., 2004). Alternatively, the entire lithosphere is approximated as a thin viscous sheet, therefore the deformation is proposed to be diffusively distributed within this region, and low rates of fault slip (a few mm/yr) are required to accommodate the India-Asian convergence (e.g., England and Houseman, 1986; Houseman and England, 1993). The third class of model argues that the deformation mainly occurs in the deep crust via widespread lateral ductile flow (Zhao and Morgan, 1987; Bird, 1991; Royden et al.,
This process causes crustal thickening and outward growth of the plateau and consequent leveling of the high topography (Bird, 1991). In comparison, this model of crustal channel flow requires a strong upper crust, but a weak deep crust. These persisting and vigorous debates over the deformation processes have motivated long-standing research efforts to discriminate the relative contributions of different mechanisms through time to the growth of the Tibetan plateau.

In this thesis, I examine the intracontinental deformation of Tibetan lithosphere from two perspectives (projects), the rheology of Tibetan lithosphere and upper crustal fault kinematics, by studying the deformation of shorelines around two saline lakes in central Tibet (Figure 1-1). The main focus of this thesis is to quantify the rheologic properties of Tibetan crust can test the capability of flow of Tibetan deep crust, as the rheology of the lithosphere fundamentally controls its deformation (e.g., Burgmann and Dresen, 2008). Yet, direct constraints on lithospheric rheology are difficult and most estimates are from studies of postseismic deformation (see Huang et al., 2014 and references therein) after large earthquakes in this region, on decadal timescales. Estimates of crustal rheology of the Tibetan plateau approaching geological timescales are even more difficult. Here I attempt to address this question by studying the flexural deformation of highstand shorelines well-preserved around Siling Co (‘Co’ is a Tibetan word for lake) in central Tibet, in response to millennial change in lake levels (or change in water loads). The idea of this approach is that the magnitude of crustal rebound manifested in shoreline deflection depends on the change in lake levels and the rheologic properties of the lithosphere (e.g., Bills et al., 1994; 2007; England et al., 2013). Therefore, this main project requires two steps to quantify the rheology of Tibetan crust; in other words, I try to answer two questions. First, I establish a relative comprehensive the lake level history (hence the loading history) of Siling Co, from determining elevations of all flights of shorelines by differential global positioning system (GPS) survey and dating all flights of shorelines by multiple dating techniques (\(^{36}\)Cl depth profile dating, optically stimulated luminescence [OSL] dating and U-series dating). The lake level history is required to constrain the crustal rheology but is poorly known. Second, I estimate the effective elastic thickness and viscosity of Tibetan crust from the flexural deformation of the shorelines.
Alternatively, I examine the deformation of Tibetan crust from understanding the fault kinematics that directly relates to crustal deformation. The slip rates of active faults in the Tibetan plateau can directly reflect whether the intracontinental deformation is highly localized along major strike-slip fault systems (Peltzer and Tapponnier, 1988; Avouac and Tapponnier, 1993) or distributed along widespread smaller-scale faults in this region (England and Houseman, 1986; Houseman and England, 1993) to accommodate the India-Asia convergence. Here I constrain the long term slip rate of the right-lateral Gyaring Co fault that traverses the Gyaring Co (Figure 1-1), ~ 70 km SW of Siling Co, from the displacements and ages of the highstand shorelines (of millennial timescales) developed near Gyaring Co. This fault is thought to be a part of so-called “Karakorum-Jiali” fault system (Armijo et al., 1986; 1989) in central Tibet that conjugates to the Altyn Tagh fault at the northern boundary of Tibet (Figure 1-1) to accommodate the India-Asia convergence and facilitate the eastward movement of crustal materials in the rigid block model mentioned above. Moreover, a recent geodetic study suggests that the Gyaring Co fault moves much faster (geodetic rate of > 10 mm /yr, Taylor et al., 2006) than other active faults (with rates of a few mm/yr; Taylor et al., 2006) in central Tibet. However, to date, there is no direct and reliable constraint on the long term geological displacement rate. Therefore the second goal of this thesis is to determine the slip rate of the Gyaring Co fault on millennial timescale and further to differentiate the different modes of deformation within the interior of Tibet.

Structure of the thesis

The overarching structure of the thesis is based on the order of three scientific questions raised above. Chapters 2 and 3 focus on the lake level history of Siling Co, during 100 – 200 ka and 40 – 0 ka, respectively. But the details of the lake history on a highstand (named as the Lingtong highstand) during ~ 6 – 4 ka is presented in Chapter 4. Chapter 4 is an attempt to quantify the crustal rheology in central Tibet using the flexural deformation of the Lingtong highstand shorelines. Chapter 5 represents the study of the Holocene slip rate of the Gyaring Co fault. Because all chapters were written as manuscripts for
publication in professional journals, there are some overlaps between the chapters. Below I summarize the findings of each chapter.

Chapter 2

I determine the Late Pleistocene lake level history of Siling Co from dating and survey of the highest (up to ~ 106 m above modern lake level) group of shorelines above the Holocene Lingtong lake highstand (~ 64 m above modern lake level [Chapter 4]). These shorelines show depositional ages between 100 ka – 200 ka (marine isotope stage [MIS] 5e -6). A tombolo shoreline in the central peninsula of Siling Co, 66 m above modern lake level, yields a $^{36}$Cl depth profile age of ~ 113 ka. U-Th ages of tufa deposits near the Lingtong highstand (~ 65 - 76 m above modern lake level) are ~ 145 – 159 ka and place minimum constraints on lake levels during MIS 6. A high spit shoreline (~ 62 m above modern lake level) in the central peninsula has an age of ~ 178 ka. Finally, the highest beach ridge (~ 106 m above modern lake level) in the northern peninsula shows a $^{36}$Cl depth profile age of ~ 186 ka. These results provide evidence that lake levels at Siling Co reached or exceeded the Holocene Lingtong highstand between 100 ka – 200 ka, and also have two folds of paleoclimatic implications. First, similar lake extensions during interglacial periods of MIS 5e (Eemian) and the Holocene suggests that paleohydrologic conditions were similar during these time periods. Second, even higher lake levels during the MIS 6 glacial probably require enhanced precipitation and/or reduced evaporative loss. Moreover, the new shoreline chronology from this study does not support the hypothesis of a single, integrated lake system (the putative “East Qiangtang Lake”) during the Late Pleistocene on the Tibetan Plateau.

Chapter 2 is in preparation for submission to the journal Quaternary Science Reviews with co-authors Eric Kirby, Kevin Furlong, Kai Meng, Shasta Marrero, John Gosse, Erchie Wang and Fred Phillips.

Chapter 3
Thirty-six new OSL samples from surficial and subsurface shoreline deposits show a general trend of lake declining of Siling Co since 35 ka. In details, this lake has reached or exceeded the Middle Holocene highstand level (~ 4590 – 4595 m) four times during this period, specifically are at ~ 35 ka, ~ 30 – 25 ka, ~ 18 ka and ~ 6 – 4 ka, indicating wet conditions during these periods. Inbetween the periods are relatively or extremely low lake levels at ~ 40 ka, ~ 30 ka, ~ 22 ka and a most recent lake recession since ~ 4 ka, suggesting substantial aridity at these times. Collectively, the results suggest a moderate cyclicity of the lake fluctuation possibly related to paleoclimatic signals. To a first order, the lake level change is predominantly controlled by the glacial evolution. But for higher frequency lake evolution, the lake level change may be controlled partially or predominantly by the monsoon intensity, glacial melting and/or evaporation. In addition, the detailed lake history also provides a history of lake loading and unloading that is necessary for investigate the lithospheric rheology from flexural deformation of the shorelines in response to the climatically induced temporal change in lake loads.

Chapter 3 is in preparation to be submitted to the journal *Geological Society of American Bulletin* with co-authors Eric Kirby, Kevin Furlong, Kai Meng, Ruth Robinson, Haijian Lu and Erchie Wang.

Chapter 4

This study provides some of the first constraints on the rheology of Tibetan middle crust at millennial timescales. Shoreline features associated with the Lingtong highstand complex ~ 60 meters above modern lake level are deflected from horizontal by 3 – 5 meters over wavelengths of tens of kilometers. OSL dating of aggradational shoreline deposits indicate that these Lingtong highstand lake levels were reached at 6 - 4 ka and subsequently fell to modern lake level. Comparison of measured shoreline deflections with a 3D elastic model constrains effective elastic thickness (Te) in central Tibet to 12 – 14 km. When combined with geophysical constraints on crustal structure, composition, and heat flow, the strain rates implied by flexural rebound require a relatively low viscosity middle crust (effective viscosity of ~ $10^{18} - 10^{20}$ Pa s) at ~ 30 – 40 km depth.
Chapter 4 is in preparation to be submitted to the journal *Science* with co-authors Eric Kirby, Kevin Furlong, Kai Meng, Ruth Robinson, Erchie Wang.

**Chapter 5**

Two flights of highstand beach ridge shorelines are found to be displaced by the central segment of the Gyaring Co fault. Both shorelines are displaced by 12 ± 1 m in a right-lateral sense across the singular fault trace. The ages of these shorelines are determined to be between 4.1 and 4.4 ka by OSL dating of the beach sands interbedded with beach gravels. These results provide an average slip rate of 2.1 – 3.2 mm/yr along the central Gyaring Co fault during the latter half of the Holocene. This rate is significantly lower than geodetic estimates (~ 10 – 18 mm/yr; Taylor et al., 2006), implying either that strain accumulation and release along the Gyaring Co fault is temporally variable, or that geodetic data are contaminated by transient deformation. The relatively low slip rate of the Gyaring Co fault is, however, consistent with other conjugate faults in central-western Tibet, and thus supporting the notion that northward convergence of India drives slow, distributed deformation in the plateau interior.

Chapter 5 was submitted to the journal *Geophysical Research Letters* with Eric Kirby, Haijian Lu, Ruth Robinson, Kevin Furlong and Erchie Wang as co-authors.
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Figure 1- 1. The topography and active faults of the Himalaya-Tibetan region, and the study area of this dissertation, Siling Co and Gyaring Co, in central Tibet. Also shown here is the 4500 m elevation contour marked as white polygon. Fault names: HFT – Himalayan Frontal thrust fault; KKF – Karakoram fault; ATF – Altyn Tagh fault; HYF – Haiyuan fault; KLF – Kunlun fault; XSF – Xianshuihe fault; RRF – Red River fault.
CHAPTER 2
HIGH LAKE LEVELS DURING MIS 5E – 6 AROUND SILING CO, CENTRAL TIBET

Abstract

Flights of well-preserved paleoshorelines around lakes in the Tibetan plateau provide important constraints on the history of hydrologic change and reflect paleoclimatic conditions in this region. Previous studies and Chapter 4 have shown that many Tibetan lakes achieved highstand levels during the Late Pleistocene – Middle Holocene. Whether similar extension was reached in the geologic past, however, remains uncertain due to sparse dating of ancient shoreline features. To address that issue, here I focus on exposures of relict, high shorelines above the Holocene Lingtong highstand shorelines (~ 64 m above modern lake level) around Siling Co, in central Tibet. I determined ages of higher shorelines (up to ~ 106 m above modern lake level) using a combination of U-series dating of tufa deposits and 36Cl depth profiles of beach ridges. I obtained a 36Cl depth profile age of ~ 113 ka from a tombolo (~ 66 m above modern level) in central peninsula of Siling Co. U-Th ages of tufa deposits range from 145 – 159 ka and place minimum constraints on lake levels during MIS 6 of 65 - 76 m above modern level. Finally, two even older 36Cl ages were obtained from higher shorelines. A high spit shoreline (~ 62 m above modern lake level) in the central peninsula of Siling Co has an age of ~ 178 ka; and the highest beach ridge (~ 106 m above modern lake level) in the northern peninsula formed at ~ 186 ka. These results provide evidence that lake levels at Siling Co reached or exceeded the Holocene highstand during the MIS 5e (Eemian) interglacial, suggesting that paleohydrologic conditions were similar during these time periods. Moreover, the preservation of higher shorelines developed during MIS 6 suggests the presence of even larger lakes during the penultimate glacial stage. Comparison of my results with available δ¹⁸O records of an ice core in northern Tibet, cave stalagmites in central Tibet and East China and insolation curve at 65°N, suggests
that high lake levels may have developed in response to enhanced precipitation during the Eemian interglacial (as has been argued for the extensive Early Holocene lakes) or in response to deglaciation. High lake levels during the MIS 6 glacial probably require enhanced precipitation and/or reduced evaporative loss. Finally, the new shoreline chronology from this study does not support the hypothesis of a single, integrated lake system (the putative “East Qiangtang Lake”) during the Late Pleistocene on the Tibetan Plateau.

2-1 Introduction

The Tibetan Plateau contains hundreds of natural saline lakes (Ma et al., 2011). Since most of these lakes are internally draining, the occurrence of high lake levels represents a positive water balance among precipitation, river input, surface runoff, thawing permafrost and evaporation. Lake level fluctuations, therefore, can be interpreted to reflect changes in the regional climate.

Constructional paleoshorelines are potential targets for a study to determine lake level histories as these shoreline features are built up at or near still water levels (Gilbert et al., 1890; Adams et al., 1999). The relict shorelines preserved around most saline lakes in Tibet provide an opportunity to study lake level changes and to improve the understanding of Tibetan paleoclimate. Previous studies of paleoshorelines and lake cores in Tibet show that high lake levels mainly existed during the Holocene (Gosse et al., 1991; Avouac et al., 1996; Kong et al., 2007; Lee et al., 2009; Hudson et al., 2013; Chapter 4). Lake histories during the Late Pleistocene time remain poorly understood, although very sparse data from several recent studies report Late Pleistocene ages (e.g., Li et al., 2009; Kong et al., 2011; Hudson et al., 2013), for the highest lacustrine beach ridges around some lakes in Tibet.

Here I use high paleoshorelines around the largest saline lake in central Tibet, Siling Co (‘Co’ is a Tibetan word for lake) (Figure 2-1) to explore its history during the Late Pleistocene. These shorelines lie up to ~ 60 - 100 m above present lake level (also see Chapters 3 and 4), connecting several small
neighboring endorheic lake basins (Figure 2-2). Dating of the shorelines shows high lake levels of Siling Co have been reached multiple times during Marine Isotope Stages (MIS) 5e and 6.

2-2 Background

2-2.1 Study area

The internally-drained Siling Co is presently the largest lake in the central Tibetan plateau, with a lake level at 4530 m in 1976. The lake subsequently rose up by ~ 13 m to ~ 4543 m in elevation in 2010 and now covers an area of ~ 2300 km² (Zhang et al., 2011; Meng et al., 2012a). Because the lake level of 4530 m in 1976 is the first accurate historical record and has been used in Chapter 4 to reconstruct the lake loads, here I refer “modern lake level” to the 1976 record throughout this paper. Four rivers supply water to the lake (Figure 2-1), the Zhajia River that is sourced from glaciers in the Tanggula Range, the Zhagen River that flows through Wuru Co, the Boqu River and the Ali River. The Ali River drains from Co E which is further supplied by the Daergawa River that connects the Mujiu Co, Ren Co Kongma, and Ren Co Ogma (Figure 2-1).

2-2.2 Shoreline classification

Siling Co is ringed by > 30 flights of well-preserved relict shorelines along the lake margin and peninsulas that extend to the lake center (Figure 2-2). The highest of these shorelines are > 60 – 106 m (Meng et al., 2012b; Chapter 4) above modern lake level, indicating more extensive paleolakes that connected several adjacent lake basins (Figure 2-2). Using field surveying by differential GPS and high resolution (0.5 m) satellite imagery, I have mapped numerous paleoshorelines around Siling Co, which were classified into three groups (Figure 2-2) based on their spatial extension, geomorphic characteristics, and sedimentary features (refer to Chapter 4 for detailed descriptions and figures). The first and most
important group is the prominent highstand shoreline at elevation of ~ 4594 m (64 m above modern lake level). Because of the distinct shoreline features on this highstand, for convenience, this highstand is named as the “Lingtong” highstand, from the name of a Tibetan village at the southeastern margin of Siling Co that is immediately close to this highstand shoreline. The more continuous recessional shorelines below the Lingtong highstand constitute the second group (named as “Y” shorelines, for convenience). Those sparsely distributed, degraded shorelines above the Lingtong highstand (Figure 2-2) are classified into the third group (named as “O” shorelines). These high “O” shorelines above Lingtong shorelines appear much older than shorelines below; therefore they have the most potential, and are utilized to elucidate the Late Pleistocene high lake level history of Siling Co in this study.

2-2.3 Previous studies on lake history of Tibetan lakes

The Lingtong lake highstand have been determined to occur during the Middle Holocene age (~ 6 – 4 ka) (Chapter 4), consistent with previous suggestion of a very deep lake during Early-Middle Holocene from lake core geochemistry and chronology (Gu et al., 1993; Kashiwaya et al., 1995), however the Late Pleistocene lake history remains uncertain. Older ages of the shorelines on the Lingtong highstand have been reported recently, including an optical stimulated luminescence (OSL) age of ~ 18 ka (Li et al., 2009) and a $^{10}$Be cosmogenic exposure age, of > 216 ka (Kong et al., 2011) for the same lake highstand. Although such Late Pleistocene high lake levels could possibly exist, Li et al. (2009) show no clear shoreline stratigraphy, making it difficult to demonstrate that the samples are from the lacustrine shoreline features. Kong et al. (2011) sampled bedrock materials immediately above the highstand shorelines to determine the shoreline ages. The bedrock samples likely have experienced more complicated exposure history than the shoreline sediments themselves. The bedrock exposure ages may not be directly related to the lake history.

A recent data compilation of regional lake records in Tibet show high lake levels may occur in four broadly correlative periods during 40 – 28 ka, 19 -15 ka, 13 -11 ka and 9 – 5 ka (Jia et al., 2001).
Studies of highstand paleoshorelines around Longmu Co, Sumxi Co and Ngangla Ringco in western Tibet suggest high lake levels occurred during Late Pleistocene – Early Holocene time (Gasse et al., 1991; Avouac et al., 1996; Kong et al., 2007; Hudson and Quade, 2013), consistent with conclusions of recent paleoshoreline studies in Tangra Yumco (Long et al., 2012; Rades et al., 2013) and Siling Co (Gu et al., 1993; Kashiwaya et al., 1995; Chapter 4). Only a few studies using reliable observations and dating of paleoshorelines reported high lake levels in interior Tibetan Plateau before 40 ka (Madsen et al., 2008; Liu et al., 2010; Fan et al., 2012; Hudson and Quade, 2013). Despite the difficulty of unravelling older lake history, most studies ascribe the high lake levels to an increase in precipitation during time periods of intensified Asian monsoon (Overpeck et al., 1996; Mügler et al., 2010; Li et al., 2011).

2-3 Methods

This study includes an integrated analysis of detailed mapping, surveying, and dating the “O” shorelines above Lingtong highstand to explore the Late Pleistocene lake level history. I follow the approach of Meng et al. (2012b) and Chapter 4 for detailed shoreline mapping, surveying and correlation using high resolution satellite imagery. The elevations of some of the degraded shorelines were previously surveyed using differential GPS as documented in Meng et al. (2012b). Shoreline chronology is determined using $^{36}$Cl depth profile cosmogenic dating of limestone and feldspar (Marrero et al., 2014, in review), U-series dating of tufa (Asmerom and Edwards, 1995) and radiocarbon ($^{14}$C) dating of carbonate shells in the shoreline sediments. Details about the laboratory protocol can be found in Appendix A.

2-4 Geomorphology and stratigraphy of shorelines above the Lingtong lake highstand

Identification of the “O” shorelines above Lingtong highstand is obvious in the field. In places where beach gravels are ubiquitous, the beach ridges are usually dissected by strike-perpendicular, nearly
evenly-spaced gully erosion on both sides of the ridges (Figure 2-3); however, in regions rich in sand deposits, such as along the eastern margin of Siling Co, shoreline degradation is dominated by more diffusive weathering process, therefore a more irregular shape of the shorelines are found in the field.

Very thick (up to ~ 10 m) sediments constitute the “O” shorelines. The sediments are well-consolidated and form a wide, well-paved flat ridge crest (Figure 2-3). The shoreline stratigraphy is typically composed of a relatively thin (~ 15 – 30 cm thick) top soil layer, beach sediments in the middle level and alluviums at the bottom (Figure 2-3). The beach sediments are layered, well-sorted and well-rounded grains. The gravel layers are mostly clast-supported and intercalated with sand layers of variable thickness. Although the sediment composition varies in space, limestone clasts dominate in the sediments. As a result, these shorelines develop very thick (up to ~ 2 cm) carbonate coatings on the beach gravels, extending as deep as 2 m or more.

2-5 High lake levels during MIS 5e – 6

I surveyed and dated the “O” shorelines at seven localities (Figure 2-2). The elevations of shoreline deposits where the samples were collected provide estimates of plaeo-lake levels relative to the sample elevations (Figure 2-5). Constructional shoreline features (CRN-1, 3 and 5) provides good estimates on lake levels (at or below the shoreline features, Chapter 4); and these features show less uncertainties in height than the erosional features (Th-1, -3, and -4) (Figure 2-5). The shoreline spits (CRN-1 and ^14C-11) and tombolo (CRN-5) are very likely built below water level (e.g., Adams and Wesnousky, 1998). The shoreline beach ridge (CRN-3) is expected to form at the lake level. The elevations of the tufa samples provide minimum estimates on the lake level because tufa develops underwater, the lake level at the time of tufa deposition is expected to be higher than the sample elevation. My results show that the shorelines where CRN-5, Th-3 and Th-4 locate in the central peninsula are only up to 2 m above the Lingtong Holocene lake highstand (~ 4594 m, Chapter 4) (Figure 2-5). The shorelines where Th-1 and CRN-1 were collected are ~ 12 m and ~ 8 m above the Lingtong highstand,
respectively. Notice that the tufa sample of Th-1 is collected from a wave-cut bench which is estimated ~12 m above Site 9 (4594 m) in Chapter 4, therefore has the largest uncertainty (~ 2 m) in elevation. The shoreline for CRN-3 is the highest (~ 4636 m) one in northern Siling Co (Meng et al., 2012b).

The seven samples collected from these “O” shorelines yielded six ages (Figure 2-2) (Table 2-1 ~ 3), except the radiocarbon sample ($^{14}$C-11) that gives an infinite age which suggests that the shoreline where this sample was collected is much older than 45 ka. This observation is confirmed by the ages of other six samples. The tombolo shoreline (~ 4596 m in elevation) immediately above Lingtong highstand in the central peninsula has a $^{36}$Cl depth profile age of 113 ± 7 ka (CRN-5) (Figure 2-4). Two tufa samples from similar lake levels (4595 – 4596 m) yield U-Th ages 159 ± 2 ka (Th-3) and 152 ± 5 ka (Th-4). At higher levels, Th-1 (~ 4606 m) has a U-Th age 146 ± 12 ka, similar to two other tufa samples within the age uncertainties. The shoreline spit in the central peninsula (~ 4602 m in elevation and ~ 8 m above the Lingtong highstand) shows an even older age, ~ 178 ± 6 ka (CRN-1). Such old lake history also occurred at a much higher level (~ 4636 m) as recorded in the highest beach ridge in the northern peninsula of Siling Co, which provides a $^{36}$Cl age of 186 ± 21 ka (CRN-3) (Figure 2-4). Notice that the uncertainties of such old $^{36}$Cl ages may actually be as large as 15% of the apparent ages (personal communications with Fred Phillips). Even though, the $^{36}$Cl age ranges still do not overlap much the U-series age ranges. Therefore my results provide somewhat detailed lake history at certain periods during MIS 5e – 6, which certainly suggest that Siling Co has reached high levels similar to, or significantly above the Holocene Lingtong highstand (Chapter 4) during MIS 5e – 6, the time periods much older than the Holocene time.

2-6 Discussions

2-6.1 Comparison of MIS 5e – 6 high lake levels of Siling Co with other records

The ages of shorelines above the highstand are much older than the OSL ages (~ 68 ka) obtained from the highest shoreline (~ 4640 m in elevation, Meng et al., 2012b) in the eastern Siling Co (Li et al.,
2009). The younger OSL ages of Li et al. (2009) may reflect dose saturation, or, a later stage of high lake level. Outside Siling Co, the history of high lake levels during MIS 5e inferred from paleoshorelines have also been reported previously across Tibet (grey bars in Figure 2-6). The Bange Co was ~ 66 m above its lake level in 2010 (~ 4521 m) at ~ 167 ka (Zhao et al., 2011) which is determined from the old lacustrine sediments deposited in northern Bange Co. The old lacustrine sediments deposited at ~ 140 m above present Nam Co was dated as ~ 115 ka (Zhao et al., 2011). In western Tibet, a U-Th age of ~ 211 ka has been obtained from a degraded shoreline above the Holocene highstand around Ngangla Ringco in western Tibet (Hudson and Quade, 2013). In northern Tibet, Lake Qinghai was 50 – 60 m above modern level at around 90 – 100 ka (Madsen et al., 2008; Liu et al., 2010). Lake Gahai in the Qaidam basin in northern Tibet was suggested to be at least 12 m above present during 85 – 72 ka (Fan et al., 2012). These lines of evidence all support existence of high lake levels during MIS 5e – 6 in many parts of the Tibetan plateau.

2-6.2 Implications for hydrologic balance of Siling Co and paleoclimate of Tibet during MIS 5e - 6

My results of high lake levels (Figure 2-5) of Siling Co during MIS 5e – 6 have important implications for paleoclimate during this time period. The high lake levels similar to, or much higher than the Lingtong Holocene highstand indicate a positive water balance, i.e., much larger water input from rainfall precipitation and/or glacial meltwater than evaporation to keep the lake highstand of at least 60 m above modern level. Given both MIS 5e and the Holocene are interglacial periods, the similar lake extension of Siling Co during MIS 5e to the Holocene highstand lake (Chapter 4), suggest similar conditions of hydrologic balance during these two time periods, in relation to precipitation, evaporation and deglaciation. During MIS 6 or start of MIS 7, the lake level was even higher, as is recorded by the highest shoreline (CRN-3) (Figure 2-5) in northern Siling Co, ~ 40 m above the Holocene Lingtong lake highstand. Therefore much more water input or substantial evaporation loss is required to keep such a deep lake.
Previous studies mostly ascribe the high lake levels during Late Pleistocene – Early Holocene to the enhanced precipitation amount due to intensified Indian summer monsoon (Gasse et al., 1991; Overpeck et al., 1996; Kong et al., 2007; Mügler et al., 2010; Li et al., 2011). As the δ\(^{18}\)O of cave stalagmites is thought to reflect the precipitation amount in East Asia (e.g., Wang et al., 2008), here I use the δ\(^{18}\)O of cave stalagmites (Figure 2-6) in the Sanbao cave in eastern China and in the Tianmen cave (see location in Figure 2-1) to evaluate the precipitation associated with Siling Co during MIS 5e – 6.

Figure 2-6 clearly show both high and low δ\(^{18}\)O values during the age range of CRN-5, indicating possible high and low precipitation amounts in the Siling Co region. High lake levels can be caused by high precipitation relative to evaporation. However, during times of low precipitation, the lake water may be mainly supplied by melting of the glaciers in the Tanggula Range through the Zhajia River (Figure 2-1), coincident with high temperature during this Eemian interglacial period. Therefore, high precipitation relative to the evaporation, and/or deglaciation during MIS 5e should be responsible for the Siling Co lake highstand at this time. In comparison, the δ\(^{18}\)O values appear very low through the MIS 6 glacial stage, indicative of a long period of high precipitation amount. Therefore the extremely high lake levels of Siling Co during MIS 6 very likely relate to high precipitation and reduced evaporation.

2-6.3 Implication for the spatial distribution of MIS 5e highstand lakes: did a unified single Ancient Great Qiangtang lake exist?

The great number of lakes that developed in central Tibet, with well-preserved flights of shorelines around these lakes has led some authors to propose that Siling Co, Nam Co, Gyaring Co and other smaller lakes in the eastern part of southern Tibet were unified at the level of ~ 4868 m during MIS 5e (Zhu et al., 2003), forming the ancient great East Qiangtang Lake (E. Qiangtang lake) (Figure 2-7) with a dimension of > 40 000 km\(^2\) (Zhao et al., 2002; 2011; Zhu et al., 2003; Shao et al., 2013). Two lines of evidence are thought to support the concept of E. Qiangtang lake: 1) the finding of so-called lacustrine terraces or deposits developed during ~ 115 ka (\(^{230}\)Th age) at ~ 4868 m in northeastern Nam Co, which
are ~ 150 m above the current lake level (4718 m) of Nam Co and ~ 328 m above modern Siling Co lake level; 2) the water divide in the river valleys that connect Siling Co and Nam Co is lower in elevation than the terraces or deposits.

The high lake level of 4595 m of Siling Co at 113 ka suggests that the lake levels during the period of “E. Qiangtang Lake” in Siling Co and Nam Co do not correlate. From a geomorphological view, even if the lacustrine terrace or deposits at ~ 4868 m exist around Nam Co and the water divide between the Siling Co drainage and Nam Co system is lower than that elevation, this proposed great lake would be out-flowing and the outlet would be unreasonably about tens of kilometers wide (Figure 2-7). In addition, it is not impossible that the Nam Co lake system of high elevation can be connected by the rivers to the Siling Co system at lower elevation, as long as the water input is much larger than the discharge (Figure 2-8). Therefore it is unnecessary to require a unified ancient great lake to explain these observations.

2-7 Conclusions

An integrated study of detailed mapping, correlation, dating of degraded shorelines above the highstand (~ 4594 m) around Siling Co was conducted to explore the lake history of Siling Co during Late Pleistocene. The following conclusions are drawn from this study.

1) Lake levels at Siling Co reached or slightly exceeded the Early Holocene highstand during MIS 5e - 6;

2) Comparison of the timing of high lake levels with other climate indices indicates that high lake levels during MIS 5e may have developed in response to enhanced precipitation during the Eemian interglacial or in response to deglaciation, a similar paleohydrologic condition to Early Holocene;

3) High lake levels during the MIS 6 (penultimate glacial stage) probably require enhanced precipitation and/or reduced evaporative loss;
4) My new shoreline chronology does not support the hypothesis of a single, integrated lake system (the putative “East Qiangtang Lake”) during the Late Pleistocene in Tibet.
2-8 References


Marrero, S. Cosmogenic nuclide systematics and the CRONUScale program. Quaternary Geochronology, in review (2014).


Figure 2-1. The hydrological system of Siling Co. This map shows that Siling Co is mainly supplied by several rivers, including the Zhajia River sourced from the Geladandong glacier and the Daergawa River. Note that Siling Co hydrological system is separated from Nam Co system by a water divide (red dashed line) east of Ringco Ogma. Lake names: B – Bange Co; C – Co E; M – Mugqu Co; K – Ringco Kongma; O – Ringco Ogma; J – Jiuru Co. Also note that Nam Co may connect to Siling Co hydrological system through the pass (4759 m) east of Ringco Ogma if Nam Co lake level rose above such elevation in the past.
Figure 2-2. The map showing three groups of shorelines and sample locations and ages. The thin yellow line and dark blue polygon show the boundary and extension of the paleolake at the Lingtong Holocene (6–4 ka) highstand of ~4594 m (Chapter 4); thick red lines/dots are “O” group shorelines above Lingtong highstand (> 6 ka); the very thin white lines are “Y” group shorelines below Lingtong highstand (< 4 ka). Light blue polygons denote modern lakes. Different symbols denote sample locations for different dating methods: depth profile cosmogenic dating (yellow stars), radiocarbon dating (green squares) and U-series dating (white diamonds).
Figure 2-3. (a) A photo showing three groups of shorelines in the central peninsula of Siling Co (see location in Figure 2-2); (b) An example field photo showing geomorphology of the degraded shorelines above the highstand in northern Siling Co. Well-paved shoreline surface and well-developed gullies by surface erosion are clearly seen in the photo; (c - d) Field photo and sketched stratigraphy of the shoreline deposits above the Lingtong highstand near the center of current Siling Co (the location corresponds to CRN-1 in Figure 2-2).
Figure 2- 4 (previous page). Depth profiles of CRN-1, CRN-3, and CRN-5. The solid curves represent model-predicted depth profile $^{36}\text{Cl}$ concentrations that best fit the measured concentrations. The dashed lines are the uncertainties of the depth profile concentrations with 95% confidence interval. The MAP (maximum a posterior) solutions (based on the Bayesian method) for ages, erosion rates and inheritance of the shoreline deposits are shown as contour plots in each panel.
Figure 2-5. Results of ages of the degraded shorelines above the highstand, suggesting high lake levels are developed during MIS 5e-6. Black and gray symbols represent $^{36}$Cl cosmogenic ages and U-series ages, respectively. The flat box indicates the lake level was at the sample elevation, and the point-up triangles suggest the lake levels were above the sample sites.
Figure 2-6. (a) Stacked mean record $\delta^{18}O$ of foraminifera from 57 global ODP/DSDP cores (Lisiecki and Raymo., 2005); (b) and (c) show stalagmite records of $\delta^{18}O$ from Sanbao cave (Wang et al., 2008) and Tianmen Cave (Cai et al., 2010, see location in Figure 2-1); Black show lake highstands during MIS 5e – 6 inferred from paleoshorelines in this study; grey bars are for previous studies.
Figure 2-7. The extension of modern lakes (cyan polygons), paleo-Siling Co at level of 4602 m (red polygons) and hypothesized ancient great East Qiangtang Lake during MIS 5e (e.g., Zhao et al., 2002; 2011; Zhu et al., 2003; Shao et al., 2013) at level of ~ 4860 m (white polygon). Note that outlet of the proposed great lake at ~ 4860 m (if exists) is of tens of kilometers wide located in NE of the map, which is unreasonable. Outflowing river system: 1. Nu River system; 2. Yarlung River system.
Figure 2- 8. The cartoon showing the lake highstands of Siling Co and Nam Co during MIS 5e – 6, suggesting the highstand lakes in central Tibet may only be connected through rivers during this time.
Table 2-1. $^{36}$Cl ages of shorelines above the Lingtong highstand around Siling Co

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Lat (°)</th>
<th>Long (°)</th>
<th>Elev (m)</th>
<th>Depth (m)</th>
<th>Age (ka)</th>
<th>Uncert* (ka, 2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SL-CRN-1</td>
<td>31.730</td>
<td>88.919</td>
<td>4601.0</td>
<td>2.5</td>
<td>178.4</td>
<td>+4.4 / -6.6</td>
</tr>
<tr>
<td>SL-CRN-3</td>
<td>31.941</td>
<td>88.868</td>
<td>4635.7</td>
<td>2.5</td>
<td>186.3</td>
<td>+21.2 /-18.9</td>
</tr>
<tr>
<td>SL-CRN-5</td>
<td>31.633</td>
<td>88.877</td>
<td>4595.5</td>
<td>2.5</td>
<td>113.4</td>
<td>+7.3 /-6.3</td>
</tr>
</tbody>
</table>

Table 2-2. U-series ages of shorelines above the highstand around Siling Co

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Lat (°N)</th>
<th>Long (°E)</th>
<th>Elev. (m)</th>
<th>$^{238}$U (ppb)</th>
<th>$^{232}$Th (ppb)</th>
<th>$^{238}$Th/$^{232}$Th (activity)</th>
<th>$\delta^{234}$U meas</th>
<th>$\delta^{234}$U initial</th>
<th>$^{230}$Th Age* (ka BP)</th>
<th>Correct. $^{230}$Th Age† (ka BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SL-Th-1</td>
<td>31.520</td>
<td>89.207</td>
<td>4606</td>
<td>1336.5 ± 3.7</td>
<td>1248.3 ± 2.9</td>
<td>18.3 ± 0.1</td>
<td>1.033 ± 0.004</td>
<td>255 ± 1</td>
<td>385 ± 13</td>
<td>170.2 ± 1.6</td>
</tr>
<tr>
<td>SL-Th-3</td>
<td>31.696</td>
<td>88.707</td>
<td>4596</td>
<td>5746.6 ± 14.6</td>
<td>909.7 ± 2.2</td>
<td>105.0 ± 0.4</td>
<td>1.007 ± 0.004</td>
<td>250 ± 1</td>
<td>396 ± 2</td>
<td>162.5 ± 1.4</td>
</tr>
<tr>
<td>SL-Th-4</td>
<td>31.692</td>
<td>88.804</td>
<td>4595</td>
<td>4329.3 ± 12.5</td>
<td>1691.9 ± 4.6</td>
<td>41.2 ± 0.2</td>
<td>0.976 ± 0.004</td>
<td>219 ± 1</td>
<td>345 ± 2</td>
<td>162.0 ± 1.5</td>
</tr>
</tbody>
</table>

* Uncorrected age; † Corrected ages use an initial $^{230}$Th/$^{232}$Th atomic ratio = 5 ± 2.5 ppm. BP: before present, where present is AD 2007. All errors are absolute 2σ. Subsample sizes range from 90 to 120 mg.

Table 2-3. Radiocarbon ages of shorelines above the highstand around Siling Co

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Lat (°N)</th>
<th>Long (°E)</th>
<th>Surface Elev. (m)</th>
<th>Depth (m)</th>
<th>Samp. Elev. (m)</th>
<th>Measured Age (ka BP*)</th>
<th>$^{13}$C/$^{12}$C (%)</th>
<th>Conventional Age (ka BP*)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SL-$^{14}$C-11</td>
<td>31.687</td>
<td>88.680</td>
<td>4595</td>
<td>1.3</td>
<td>4593.7</td>
<td>NA</td>
<td>2.1</td>
<td>&gt;43.5</td>
</tr>
</tbody>
</table>

The $^{14}$C activity was extremely low and almost identical to the background signal.

* BP = Before Present (1950 AD).
CHAPTER 3

LATE PLEISTOCENE – HOLOCENE LAKE LEVEL CHANGE OF SILING CO, CENTRAL TIBET, INFERRED FROM PALEOSHORELINES

Abstract

The history of lake expansion and contraction in the interior of the Tibetan plateau is directly related to paleoclimatic change of this broad region. The deepwater lacustrine sediments have previously been utilized to extract paleoclimatic signals in this region. However, the temporal change of lake levels remains difficult to be constrained from the lacustrine sediments. Although well-preserved paleoshorelines throughout Tibet provide good geomorphic markers of lake levels, they have only been well studied until recently in a few lakes. Yet, the lake level change from those previous studies is mostly limited in the Holocene time because of sparse dating and survey of certain shorelines around very few lakes. These limitations highlight the need of systematic geomorphic observations and high quality dating of flights of well-preserved shorelines in single lakes to elucidate a relatively complete lake level change during the Late Pleistocene – Holocene. Here I study the lake level change of Siling Co, the largest lake in the interior of Tibet, using well-preserved flights of shorelines and stream-cut stratigraphy around the lake. The surficial and subsurface shoreline sediments are used to interpret their depositional environments and water depths for inferring the lake levels. The timing of lake levels are determined by optically stimulated luminescence (OSL) dating of samples from the shoreline strata. In total I have collected 36 new OSL samples from around this lake, especially in the central peninsula. In combination with previous OSL ages of the most recent highstand shorelines, the new data from this study provide a complicated lake history since 40 ka.
My results show a general trend of lake declining of Siling Co since 35 ka. But this lake has reached or exceeded the Middle Holocene highstand level (~4590 – 4595 m) four times during this period, specifically are at ~35 ka, ~30 – 25 ka, ~18 ka and ~6 – 4 ka, indicating wet conditions during these periods. Between above periods are relatively or extremely low lake levels at ~40 ka, ~30 ka, ~22 ka and a most recent lake recession since ~4 ka, suggesting substantial aridity at these times. Thus these results suggest a moderate cyclicity of the lake fluctuation possibly related to paleoclimatic signals. To a first order, the lake level change is predominantly controlled by the glacial evolution. But for higher frequency lake evolution, the lake level change may be controlled partially or predominantly by the monsoon intensity, glacial melting and/or evaporation. In addition, the detailed lake history also provides a history of lake loading and unloading that is necessary for investigating the lithospheric rheology from flexural deformation of the shorelines in response to the climatically induced temporal change in lake loads.

3-1 Introduction

The history of saline lakes developed in the Tibetan Plateau can provide important insights into the Quaternary environmental change of this broad region (e.g., Gasse et al., 1991). Previous studies evaluate the lake history and associated paleoclimatic signals (e.g., regional dry/wet conditions, the lake salinity and water depth) by numerous proxies extracted from shallow lacustrine sediment cores, including stable isotopes (e.g., Gasse et al., 1991; 1996; Fontes et al., 1993; Gu et al., 1993; Kashiwaya et al., 1995; Wang et al., 2002; Morrill et al., 2006; Zhu et al., 2008), assemblages of pollen (e.g., Van Campo et al., 1993; Miehe et al., 2014) and ostracod (e.g., Zhu et al., 2010) and mineralogy (e.g., Gasse et al., 1991; Zheng, 1997; Mügler et al., 2010). These studies mostly reveal the lake history of central-southern Tibet since Late-Pleistocene – Holocene transition, and some records can date back to 40 ka.

Although fruitful climatic information can be obtained from these proxies, they are insufficient to reveal the magnitude of lake fluctuation in time, and hence temporal evolution of the lake volume,
another important parameter that can reflect the paleoenvironment. The presence of ubiquitous sequences of spectacular paleoshorelines well-preserved around many lakes in Tibet provides an opportunity to directly constrain the paleolake levels and to infer hydrological balance and paleoclimate. Recent decades have seen progresses on studying the lake level change from shorelines around the lakes (Figure 3-1) in central-southern Tibet, such as Sumxi-Longmu Co (e.g., Gasse et al., 1991; Avouac et al., 1996; Kong et al., 2007), Bangong Co (Gasse et al., 1996), Lagkor Co (Lee et al., 2009), Zabuye Co (Hudson and Quade, 2013), Tangra Yumco (Long et al., 2010; Rades et al., 2013; Miehe et al., 2014), Siling Co (Li et al., 2009; Chapters 2 and 4) and Nam Co (Kong et al., 2011). However, similar to those shallow lake core studies, the lake history extracted from the paleoshorelines is rarely obtained beyond the Holocene.

Two factors may lead to the incapability of extracting a longer period of lake history in this region. First, the highest shorelines around some of the lakes, which possibly record earlier lake history, may lack datable materials to determine the timing of paleolake levels (Chapter 2). Second, although earlier lake level history can potentially be extracted from buried sediments beneath the most recent shoreline beach ridge deposits, as exemplified by studies of the Dead Sea (e.g., Renaut and Owen, 1991; Bookman et al., 2004), Lake Lisan in the Middle East (e.g., Machlus et al., 2000; Bartov et al., 2002) and Lake Lahontan in western United States (e.g., Adams and Wesnousky, 1998), the rarity of such stratigraphic outcrops cross-cutting the paleoshorelines makes it difficult to apply similar studies to Tibet.

Here I utilize well-preserved surficial and subsurface sediments of paleoshorelines and also laterally continuous stratigraphy that cross-cuts a sequence of paleoshorelines around Siling Co (Figure 3-1, 2), to reconstruct the lake history, as complete as possible, from all levels of shorelines. The OSL dating of the sand samples collected in the shoreline deposits is exploited to obtain a framework of shoreline ages.
3-2 Study area and geomorphology of paleoshorelines around Siling Co

The Siling Co (31°30’-32°08’N, 88°31’-89°21’E) is an internally-draining saline lake located in central Tibet (Figure 3-1, 2). It is currently the largest lake in the interior of Tibet with an area of ~ 2300 km² (Meng et al., 2012a; Wan et al., 2014). At present, the lake water is supplied predominantly by several rivers, among which the largest is the Zhajia River sourced from the Geladandong glacier in the Tanggula Mountain (Chapter 2), and partly by the rainfall precipitation (Meng et al., 2012a).

The Siling Co has developed > 30 flights of paleoshorelines since Late Pleistocene that are well preserved around the lake (Meng et al., 2012b; Chapters 2 and 4) (Figure 3-2). The highest shorelines are > 60 – 106 m above its modern lake level of ~ 4530 m in 1976 (Meng et al., 2012a) (I refer to 1976 lake level as modern lake level in the following text), indicating that much more extensive paleolakes have existed and connected several adjacent lake basins, including Wuru Co, Co E, and Bange Co (Figure 3-2). Detailed shoreline mapping with field investigation and high resolution satellite imagery reveal that the sequence of shorelines around Siling Co can be divided into three main groups (Figure 3-2) according to observations of their spatial distribution, geomorphic and sedimentological features (Meng et al., 2012b; Chapters 2 and 4): 1) the Lingtong highstand (Chapters 2 and 4) (Figure 3-2), which is a prominent highstand at ~ 4594 m in elevation, well-distributed around the lake; 2) the recessional, more continuous, shorelines below the highstand, “Y” shorelines (Chapter 2), and 3) highly degraded shorelines above the Lingtong highstand, or “O” shoreline (Chapter 2) which are preserved only at sporadical sites that display a higher degree of soil development in the beach deposits.

3-3 Previous understanding of lake history of Siling Co

The studies of lake cores and paleoshorelines in Tibet in recent decades have shed light on the lake history of Siling Co in certain time periods. Siling Co is previously thought to reach near the Lingtong highstand (~ 4594 m in elevation) (Figure 3-2) at ~ 18 ka (Li et al., 2009). However, a detailed,
most recent surveying and OSL dating of numerous constructional shoreline deposits along the Lingtong highstand and well distributed around the lake shows that the timing of the Lingtong highstand should be at ~ 6 – 4 ka (Chapters 2 and 4), consistent with the implication of an Early – Middle Holocene lake highstand from the lake core geochemistry and chronology (Gu et al., 1993; Kashiwaya et al., 1995). In addition, a much earlier lake history of Siling Co has been found in Chapter 2. By multiple dating ($^{36}$Cl depth profile cosmogenic and U-series dating) and surveying of several constructional and erosional shorelines, Chapter 2 shows, for the first time, that high lake levels, up to ~ 100 above modern lake level, or ~ 40 m above the Lingtong highstand, have been reached multiple times between 100 – 200 ka.

The lake level history between those two periods and after the Middle Holocene, however, is still poorly understood. Although previous data compilation of regional lake records in central-southern Tibet show high lake levels may exist in multiple broadly correlative periods of 40 – 28 ka, 19 -15 ka, 13 -11 ka and 9 – 5 ka (Jia et al., 2001), little information is from Siling Co. A recent study argues a very simple lake recession from the highest shoreline (~ 4640 m, ~ 100 m above modern lake level) since ~ 70 ka, to the Lingtong lake highstand at ~ 4594 m at ~ 18 ka, at the eastern margin of Siling Co (Li et al., 2009). But that study only shows three reliable datapoints along the transect at the eastern lake margin, obviously insufficient to support a simple lake history of Siling Co.

Here I establish a relative complete lake level history of Siling Co by focusing on the surficial and subsurface shoreline deposits of the suite of beach ridges and in the central peninsula (Figure 3-3a) and a spectacular, laterally continuous strata exposed along a seasonal stream (here it is named as Tashi Stream) (Figure 3-3b) that cross-cut the lowest shorelines below the Lingtong highstand.
3-4 Approach

3-4.1 Determining lake levels by observation of shoreline geomorphology and stratigraphy

In this study, I focus on the central peninsula of Siling Co (Figure 3-3a) to study the lake history. This area contains flights of shorelines that include 1) one flight of shoreline tombolo immediately (~ 1 m) above the Lingtong Holocene highstand; 2) the Lingtong highstand shoreline complex (~ 4594 m, Group 3); 3) several flights of shorelines at elevations of ~ 4580 - 4575 m (Group 2) preserved in the middle part of Figure 3-3a and are distinctly separated from those above and below; 4) the lowest suite of shorelines (Group 1) which can be divided into 5 clusters of shorelines (G1d – G1a in Figure 3-3c).

The most important and distinctive feature of this study, as compared with other studies of lakes in Tibet, is the finding of both surficial and laterally continuous shallow shoreline strata cut by the Tashi Stream (Figure 3-3c). The presence of sandy layers within these strata allowed us to determine the timing of the lake levels by OSL dating. In addition, sedimentary facies observations of shoreline strata enabled us to interpret the water depths of the sediments (e.g., Adams and Wesnousky, 1998; Machlus et al., 2000; Bartov et al., 2002), i.e., whether the sample is above, at, or below the lake level. A summary of classification of sedimentary facies and associated interpretation of water depth is given in Table 3-1. The common deposits I found from the shoreline stratigraphy include alluvial fan and debris deposits, foresets and backsets of beach ridges, occasional lagoonal sands, near shore sands, and deepwater lacustrine sands.

For those shorelines without stratigraphic context, I adopt the same method of Chapter 4 and I dug soil pits (with width of 1 m, length of 2 m, and variable depth up to 2.5 m) into the shoreline deposits, trying to show a limited dimension of the strata to collect the sand samples and interpret their water depth from the stratigraphic context. In addition, I also collected samples from shorelines at variable elevations in different portions of the lake (Figure 3-2), to get more information on the lake history of Siling Co.
3-4.2 OSL dating methodology

In the field, the OSL samples were carefully collected from sand layers/lenses within the shoreline deposits. Eolian sands (younger than the age of shoreline deposition) at the top of the shoreline stratigraphy are avoided, as these most recent sediments do not reflect the lake level information. Before collecting the sand samples, the light-affected surface deposits of several centimeters thick were removed to reduce the uncertainty from dose measurement (Aitken, 1998; Rhodes et al., 2011). Sands of medium, fine or silt sizes were drilled using plastic PVC (Polyvinyl chloride) and/or steel tubes, 3 cm or 5 cm in diameter, depending on the consolidation and thickness of the sand layers. These tubes were then wrapped with heavy duty black duct tape to avoid light penetration and keep the water content.

The analysis of luminescence behavior, dose rate estimation and age calculations were conducted at University of St Andrews, following the standard protocol of King et al. (2013) (see Appendix B for details). The age modelling (Arnold and Roberts, 2009) is also described in the Appendix B and most of the sample burial ages (Db) are based on a Minimum Age model (MAM-3) and Central Age model (CAM). The age and dosimetry data are listed in Table 3-1 and B-2, respectively.

3-5 Paleolake levels of Siling Co determined from shoreline deposits

3-5.1 Paleolake levels determined from surficial shoreline deposits

Here I first determine the most recent lake level change from surficial beach and alluvial fan deposits, as these sediments near the top of the shorelines were younger than sediments buried below them. In total I found 9 sites in the central peninsula (Figure 3-3a, c), with surficial deposits to determine the paleolake levels. Six shoreline sites (G2-O18, -17, -44, G1d-O36, G1b-41, -O42) show sand layers intercalating with beach gravels, and 3 sites (G1c-O38, G1b-O10, G1a-O7) show sand layers within the top layers of the youngest alluvial fan deposits (Figure 3-3c). These alluvial fans relate to development of
the shorelines in front of them. The sands in these sediments allow us to get the ages (see details in Table 3-2) of the shorelines, and thus the timing of the paleolake levels. In the following text, I show detailed stratigraphy of two key sites, the higher shoreline G2-O18, on the west and the lower one showing both G1b-O10 and G1b-41, on the east. Detailed descriptions of other sites showing surficial deposits are in Appendix B.

**Group 2 shorelines**

*Site O18*

This site is the highest shoreline among all Group 2 shorelines in the central peninsula (excluding O13 from the Tashi Stream). A ~ 0.8 m deep soil pit has been dug to observe the strata of this beach ridge due to the lack of outcrops. Two main units are observed from the soil depth profile (Figure 3-4). Unit 1 at the bottom is a whitish layer of clean, generally-sorted, and grain-supported rounded gravels. A thin lens of fine sands is intercalated with the upper part of this layer. Unit 2 above consists of two sublayers. The lower one is a medium thick layer of rounded gravels infilled with silt sands and mud. At the top is a brownish layer of mixed silt sands and mud, with plants growing in it.

The top unit with plants is considered as a soil layer. The clean grain-supported whitish gravels at the bottom unit are interpreted as beach deposits, thus reflecting the most recent lake level was at this site, that is, the lake level should be at the elevation this beach ridge.

The sand lens in the bottom unit yields an OSL age of 3.3 ± 0.2 ka, suggesting the lake reached at the height of this shoreline during the Middle-Late Holocene.

*Site O17*

Site O17 is the lowest in Group 2 shorelines, ~ 5 m below Site O18. This site has some similarity in sedimentary facies to Site O18 (see detailed description in Appendix B and Figure B-1). The thin lens of fine sands intercalated with the beach gravels near the surface yields an OSL age of 3.0 ± 0.2 ka,
reflecting the lake level reached at ~ 4580 – 4575 m around 3 ka and was relatively stable for several hundred years.

Site O44

This westernmost site shows two lenses of medium sands intercalated with landward layered gravels that I interpret as backsets of beach ridges (see detailed description of sedimentary facies units in Appendix B and Figure B-2). The OSL age of the medium sands is 3.6 ± 0.3 ka, reflecting the lake level was at this shoreline elevation during the Middle Holocene.

In a short summary, these results of three Group 2 shorelines have clustered ages of 3.6 – 3 ka. While considering that the lake has reached the Lingtong highstand of ~ 4594 m in elevation during ~ 6 – 4 ka (see Chapter 4 for details), my data imply Siling Co dropped by ~ 20 m from the Lingtong highstand (~ 4594 m) to Group 2 shorelines (~ 4580 – 4575 m), during 4 – 3 ka.

Group 1 shorelines

Site O36

Site O36 is a stream-cut outcrop in the lowest of G1d shorelines. Here there is only a limited exposure of strata. A thin layer of coarse sands intercalated with the beach gravel layers near the top (see detailed description of sedimentary facies units in Appendix B and Figure B-3), provides an OSL age of 5.0 ± 0.2 ka, suggesting the lake might be at this site around 5 ka. But because the stratigraphic extension is not clear, I put a low confidence in the lake level at this site.

Site O38

Site O38 is near the top of a possible alluvial fan that traverses the G1c shorelines. The contact between the sedimentary facies units is not very clear (see detailed in Appendix B and Figure B-4). The OSL sample from a thin layer of fine sands in the top alluvial (?) unit yields an age of 3.7 ± 0.2 ka,
suggesting the lake level might be below this site around 4 ka. But I put a low confidence in this interpretation given the unclear stratigraphic context.

*Site O10*

This outcrop shows a strike-perpendicular stratigraphic cross-section. The elevation of this site is ~4565 m. Two OSL samples, O10 and O41, were collected from this cross section. Four sedimentary facies units (Figure 3-5) are observed. Unit 1 at the bottom is a thick pile of layered, whitish-brownish, well-sorted, grain-supported gravels. The gravel layers dip both landward and lakeward. A thin layer of fine-silt sands is intercalated with these gravels but pinches out towards the lake side. Unit 2 is another medium thick layer of whitish, well-sorted, grain-supported gravels, with inclination towards the land. Unit 3 is a topographic high that contains a thin layer of brownish, not well-sorted, mostly grain-supported gravels. Unit 4 is located on the backside (west) of Unit 3, with a base dipping landward. This unit consists of poorly-sorted, matrix-supported gravels mixed with mud and silts. A thin lens of fine sands is embedded into this unit.

Unit 1 of layered, grain-supported gravels is interpreted as a relatively recent beach ridge that shows both backsets and foresets. Similarly, the layered gravels in Unit 2 are also considered as a beach ridge with presence of both backsets and topsets. Unit 3 is probably a topset of the most recently developed beach ridge. The poorly-sorted, matrix-supported gravels in Unit 4 are interpreted as alluvial fan deposits that relate to the development of beach ridge Units 1 – 3.

The OSL age of O41 is 1.7 ± 0.1 ka, and O10 is 1.3 ± 0.2 ka. The stratigraphic observations together with the OSL ages indicate that the lake level was near this site (~4565 m) during 2 – 1 ka.

*Site O42*

Site O42 is the easternmost site among G1b shoreline group, ~3 m below Site O10. The medium sands intercalated with beach gravels (see detailed description in Appendix B and Figure B-5) provide an age of 1.2 ± 0.2 ka, suggesting that the lake was recently at ~4562 m at 1 ka.
Site O7

Site O7 is the highest (~ 4553 m in elevation) among the G1a shoreline group, ~ 10 m below Site 42. The OSL age from the fine/silt sands in the top debris flow unit (see Appendix B and Figure B-6 for details), is 0.7 ± 0.0 ka, suggesting the lake level was below this shoreline around < 1 ka.

From results of the surficial samples in Group 1 shorelines, I can find that, Siling Co generally continued the lake recession from Group 2 shorelines at ~ 4575 – 4580 m to Group 1 shorelines that is ~ 4553 m or lower during 3 – 1 ka. Therefore this lake dropped by ~ 20 m during this 2 ka-period. It continuously retreated by another 23, m from the level of 4553 m to modern lake level of 4530 m (referenced to 1976), during this last thousand years.

Additional observation of the alluvial fans along the Tashi Stream

Along the > 1 km long Tashi Stream, the presence of alluvial fans developed behind beach ridges provides additional inference for the lake level change. The lake level is expected to be either immediately close or below these fresh-looking fans at time of deposition of the fan sediments. Here I use the surface color, freshness and texture to evaluate the degree of fan degradation and hence the relative ages of these fans, an indirect signal for lake level change.

Six phases of alluvial fans (I-VI in Figure 3-3c) can be clearly observed along the Tashi Stream. The G1a shorelines have been traversed by two phases (I and II) of alluvial fans, i.e., Fan I has dissected Fan II. In general, the four alluvial fans (I-IV) developed below (or east of) G1c shorelines (Figure 3-3c) on the east side appear more whitish and fresh, and preserve more fan textures than the two on the west. Within these lowest three fans, the fan on the west appears less fresh, darker, and with less textures than the east, suggesting a gradual increase in the degree of fan degradation, hence a landward increase in the ages of the alluvial fans. Considering that a fan is developed at the same time when a beach ridge is constructed in front of (or east of) the fan, then these observations inform us that this lake has recently retreated continuously from a level where G1c shorelines stand.
Among the upper two alluvial fans (V and VI), the degree of fan degradation appears less distinctive between the two. Apparently Fan VI located between G1d and G1c shorelines (Figure 3-3c) show less textures and blacker color, but the reliability of using fan degradation to infer the relative ages of these fans is reduced. Therefore I tend not to infer the lake level history from these uppermost two fans.

In summary, from geomorphic observations of the alluvial fans and associated beach ridges, I found at least six phases (I–VI in Figure 3-3c) of fan development in this locality, which provides us a qualitative sense of the most recent lake fluctuation at this site, and the lake history of Siling Co since the Middle Holocene has been characterized by a large recession from the highstand to modern lake level.

3-5.2 Paleolake levels determined from shallow shoreline strata

To constrain lake level history beyond the Holocene time, I take the advantage of presence of shallow shoreline stratigraphy in the central peninsula of Siling Co and additional shoreline sites around the lake. In total, I collected 27 new OSL samples (Table 3-2) from these shallow shoreline strata, including 16 samples from the central peninsula (Figure 3-6a, c), and another 11 from the eastern margin and other places around Siling Co.

In the following section, I interpret the lake levels using the shoreline strata at each site. At first I describe the sedimentary facies (see Table 3-1 for a summary), which I use to interpret the depositional environment and the water depths of each sample. These data, combined with the elevations of the samples, give an estimate of the lake level relative to the sample position (see Table B-3 for a summary). Then the OSL ages (Table 3-2) of the samples are presented to infer the timing of the lake levels. Finally I interpret lake level history at each site from observation of the stratigraphy and OSL ages. I show detailed description of the shoreline stratigraphy at four key sites (O14, G2-O43, G1d-O12 and G1a-O15) from the central peninsula to the eastern margin. Details for other sites can be found in the Appendix D.
Central Peninsula

In this specific section, I describe the shoreline sites from the west to east, i.e., from the highest shoreline (O14) to the Group 3 – Lingtong highstand shoreline complex, then to the lowest shorelines cut by the Tashi stream (Figure 3-6a, c).

Site O14

This site has an elevation of ~ 4596 m, immediately (~ 1 m) above the Lingtong highstand shoreline where OSL sample O21 was collected (Figure 3-6a). No strata are exposed along this beach ridge, therefore a 2.5 m deep soil pit was dug into the top deposits (Figure 3-7) to describe the strata, and to collect chronological samples for reconstruction of the lake history.

Four units with distinct sedimentary facies can be observed from this pit (Figure 3-7). Unit 1 is a layer of relatively well-sorted grain-supported rounded beach gravels. It is unknown where the lower boundary of this layer is at depth; Unit 2 contains a thick brownish layer of poorly-sorted rounded pebbles with sands and mud as matrix, this layer retains more moisture than the layers above and below (Figure 3-7). A thin lens of medium-sized sands is intercalated with upper part of this layer; Unit 3 is a very thick pile of whitish, alternating layers of well-sorted grain-supported rounded gravels. A large variation in grain size for each layer can be easily observed from this unit as well; Unit 4 at the top of this shoreline is a ~ 30 cm thick, reddish, wet layer with mixture of mud and (eolian?) silt sands. At the base is a thin sublayer of cobbles with mud and sand as matrix. Also observed in this top unit is the penetration of grass roots into the loose, wet sediments.

Based on the characteristics of the deposits, I interpret Units 1 and 3 as beach ridge deposits, Unit 2 as alluvial fan sediments that sat several meters or immediately above the paleolake level, and Unit 4 as a soil layer. The color and weathering of the bottom of the top unit suggest this shoreline has formed in the Late Pleistocene, about tens of thousands of years.

The medium sands from Unit 3 alluvial fan deposits yield an OSL age of 43.3 ± 3.1 ka. If this age is correct, then it suggests that the lake level was reached above the Middle Holocene highstand (site of
O21, Figure 3-6a) at ~ 43 ka. However, the $^{36}$Cl depth profile age from exactly the same pit shown in Figure 3-5 suggests that these deposits are as old as ~ 113 ka (Chapter 2). The $^{36}$Cl age is believed to be accurate given its consistency with another $^{36}$Cl age (CRN-3, ~ 186 ka) of the highest shoreline (~ 4636 m) in northern peninsula of Siling Co, and also its cross-correlation with the U-series ages of wave-cut erosional shorelines at similar height to this site (Chapter 2). This discrepancy may be explained if the OSL sample of O14 was saturated (Rhodes, 2011). Thus, the shoreline could be much older than the apparent age (43 ka at this site). In this region, any OSL ages older than ~ 40 ka may not reflect the real age of the deposits and are not considered later in interpretation and discussion of the lake history of Siling Co.

**Group 3 shorelines**

Three flights of shorelines at this site (O21, O20 and O19) constitute the Group 3 shoreline complex, with elevation of ~ 4590 m – 4595 m (Figure 3-6b). The highstand shoreline of O21 gives an OSL age of ~ 4 ka (Chapter 4), therefore, in this paper, I only present the soil pit stratigraphy and associated age data of O20 and O19.

**Site O20**

Site O20 is located below O21 (the Lingtong highstand) but above O19 (Figure 3-6a). A lens of fine sands from the bottom alluvial fan sediments (see details in Appendix B and Figure B-7), yields an age of 63.8 ± 4.6 ka, indicating that the lake level was below the sample elevation at a time older than 64 ka (recall that OSL samples older than 40 ka are likely saturated, see discussion above).

**Site O19**

Site O19 is ~ 4 m below Site O20. The OSL sample from a thick pile of fine sands that is interpreted as nearshore environment (see details in Appendix B and Figure B-8), provides an age of 28.5
± 2.0 ka, suggesting that the lake level was probably several meters above the sample position around ~ 29 ka.

In a short summary, the stratigraphic observations at Site O14 and Group 3 shorelines, in combination with their OSL chronology, suggest that Siling Co was once at this high level at least two times before 40 ka, and also reached again this level around 29 ka and during the Middle Holocene (O21, ~ 4 ka) (Chapter 4).

**Group 2 shorelines**

*Site O13*

Site O13 contains stream-cut shoreline strata on the western part of the Tashi Stream (Figure 3-6c). The massive silt sands intercalated with debris flow deposits at the bottom (see details in Appendix B and Figure B-9), provide an OSL age of 13.1 ± 3.2 ka, suggesting that the lake level was below this site the sample position around ~ 13 ka.

*Site O43*

Site O43 is several meters below Site O13 (Figure 3-6c). Four units with distinct sedimentary facies are observed on the ~ 3 m high wall at Site O43. Unit 1 at the bottom shows a thick pile of yellowish, poorly-sorted, wedge-shaped, angular gravels which are intercalated with massive coarse-silt sands and elongated lenses of silt sands (Figure 3-8). A few rounded boulders can be found within this unit. Unit 2 is a whitish layer composed of sub-rounded to subangular pebbles and cobbles coated with pedogenic carbonate. Unit 2 is a relatively thick layer of brownish massive silts mixed with sparse pebbles. Unit 1 is a medium thick, wet layer of poorly-sorted, matrix (mud and silt sands) – supported rounded cobbles and boulders. Plants are less abundant in this top unit than in deposits from shorelines in the central peninsula (e.g., O14, O20, O18 and O17).

I interpret the bottom unit at this site as massive alluvial fan deposits, developed above the lake level. The upper three units are considered as stratified, alluvial fan deposits, possibly related to alluvial
Fan V observed on the surface mentioned above. The absence of a soil layer, similar to O13 of Group 2 shorelines upstream, infers that this alluvial fan is relatively young, or recent erosion has occurred at the surface induced by, for example, wave currents during a lake transgression.

The OSL age of silt sands from the bottom alluvial fan unit, is 27.4 ± 0.9 ka, which, in combination with interpretation of the alluvial fan environment for the sample, suggests the lake level was below the alluvial deposits around ~ 27 ka.

**Group 1d shorelines**

Four flights of shorelines are preserved at Group 1d site (Figure 3-6c). In particular, there is a spectacular, laterally continuous exposure of strata beneath the westernmost, highest shoreline where four OSL samples (O12, O28, O29 and O30) were collected (Figure 3-9). Another subsurface OSL sample, O37 (Figure B-10), was collected from the lowest shoreline in this group.

**Site O12**

This stratigraphic cross-section has ~ 100 m in length and ~ 4 m of maximum height, and contains 11 sedimentary facies units (Figure 3-9), providing rich information about lake level change. Detailed sedimentary observations and the interpretations of lake levels from each unit, from bottom to top, are described below.

**Unit 1** contains poorly-sorted pebbles to boulders with mud and silt sands as matrix. Red layers of probable paleosol (?) are present in this unit. **Unit 2** has a thin – medium thick, wavy layer of brownish, poorly sorted, subrounded-subangular gravel partially grain-supported and partially matrix-supported. A thin lens of brownish-red fine sands was intercalated with this unit. The sample O30 (Figure 3-9) was collected from the sand lens. **Unit 3** is a thin layer of brownish medium sands onlapping onto a very gentle slope atop Unit 1. A bit of laminations can be observed in the sands. This layer pinches out at the bottom of a nearly flat-laying tabular body of Unit 4. The sample O29 (Figure 3-9) was collected from this sand unit. **Unit 4** is a thick pile of brownish, landward-layered, relatively-well sorted, and grain-
supported rounded gravels. These gravels onlap both Unit 2 and 3. **Unit 5** consists of several small chunks of massive, gray, fine sands that filled in the space atop the backsets of beach gravels in Unit 4 and the alluvial channel deposits in Unit 2. The contact between this sand unit and sediments below show significant relief; however, the boundary between these sands and the gravel unit above is flat. The sample O28 was collected from this sand unit. **Unit 6** extends far from the west end to near the east side of this stratigraphic section. This unit has two subunits (see boundary in Figure 3-9d). The western part is a thick whitish layer containing well-sorted, clean, grain-supported rounded gravels. Within this subunit, the western part of these gravels is layered landward, and the eastern part is inclined towards the lake. The east subunit contains two layers. At the bottom is a thin layer of yellow silt sands interbedded with ~1 cm thick layer of gravels. This sand unit has a channel shape and pinches out on both ends by onlapping onto the gravels within the unit. Symmetrical ripple laminations are shown in this bottom sand layer. At the top is a thin layer of relatively well-sorted gravels. The sample O12 was collected from the sands with ripple marks. **Unit 7** is another unit that extends very long along this stratigraphic section (Figure 3-9a, b). It has layering both dipping towards the land on the west and dipping towards the lake on the east. This unit contains whitish, clean, well-sorted, grain-supported gravels. The top of this unit presents a topographic high in the middle and low on the flanks. **Unit 8** is a medium thick layer of yellowish paleosol that onlaps the backsets of the Unit 7 old beach ridge. Flat-laminations are found in this unit. **Unit 9** contains two subunits. The top subunit is composed of landward-dipping layers of clean, well-sorted, grain-supported rounded gravels. The bottom one contains gently-lakeward-dipping layers of rounded pebbles. **Unit 10** is a thick pile of dark paleosol, with sparse pebbles intercalated with it. This layer has covered the Unit 8 paleosol, the old beach ridge of Unit 7, and the beach ridge of Unit 9 that developed most recently. **Unit 11** is a thin layer of poorly-sorted gravels with mud and silt sands as matrix.

The depositional environment of Unit 1 is interpreted to be debris flow deposits based on the mixture of poorly sorted pebbles, boulders and mud matrix. The red weathered sediments indicate these deposits have been exposed above water; therefore, I interpret the lake level was below this debris flow unit at time of its deposition. Unit 2 is interpreted as an alluvial channel deposit according to presence of
the relief of its contact with Unit 1 that may reflect the scouring of channels and also the partially grain supported subrounded gravels. The red weathering of the sediments also indicates they were exposed subaerially; therefore, the lake level is thought to be below this unit at time of its deposition. The sand lenses of Unit 3 are considered as nearshore sands probably below the zone of wave influence, given the geometry of the sand layer and sediments above and below this unit. Therefore, the water depth of this sand layer could be 1 – 15 m (e.g., Machlus, 2000), i.e., the lake level is about 1 – 15 m above this sand layer at time of its deposition. The grain-supported gravel layers of Unit 4 and their geometry suggest they are backsets of a beach ridge. Thus, the lake level should be at or close to the beach ridge at time of its development. For Unit 5, based on the topographic geometry, the contacts of this sand unit with those above and below, the depositional environment of these sands can very likely be lagoon. Therefore, the lake level could be very close or below these lagoonal sands. For Unit 6, the landward-inclining beach gravel layers are interpreted as backsets of the beach ridges; and the lakeward-dipping layers of gravels look like the foresets, the channel-shaped silt sands with ripple laminations in the eastern part of this unit, reflect a nearshore environment with wave influence, which has a water depth of several meters (e.g., Machlus, 2000). The interpretation of nearshore, foresets and backsets of the beach gravels, make us feel confident that Unit 6 is an old beach ridge developed above all units of 1 – 5. The geometry of the gravel layering and the contact relief of Unit 7 suggests that it is an old beach ridge with both backsets and foresets, but buried beneath the surface. The thick layer of paleosol onlapping the backsets of Unit 7 is interpreted as deepwater lacustrine sediments. The bottom subunit of Unit 9 that contains landward-dipping clean grain-supported gravel layers is considered as foresets of a most recent developed beach ridge, and the top subunit of Unit 9 that shows lakeward-dipping gravel layers as the backsets. The relief of the backsets constructed above the foresets may reflect a short-term rising lake level. The thick paleosol layer with sparse pebbles in Unit 10 possibly suggests that these are alluvial fan deposits. Finally, the poorly sorted gravels with matrix of mud and silt sands in Unit 11 are interpreted as alluvial fan deposits.
The OSL age of O30 from Unit 2 alluvial channel sands is 36.7 ± 1.4 ka, O29 from Unit 3 nearshore sands is 22.7 ± 0.9 ka, O28 from Unit 5 lagoonal sands is 17.3 ± 0.6 ka, and O12 from Unit 6 nearshore sands is 6.2 ± 0.2 ka.

This stratigraphic section, as a whole, reflects a detailed lake history. The lake was below Unit 1 debris flow deposits and remained low during deposition of Unit 2 alluvial channel deposits at ~ 37 ka, and then rose above the Unit 3 nearshore sands by about 1 – 15 m (e.g., Machlus, 2000) at ~ 23 ka. Probably at the same time, the Unit 4 beach ridge was built. This process is followed by the development of a lagoon behind Unit 5 beach ridge at ~ 17 ka. After that, there is establishment of two ancient beach ridges in Unit 6 and 7 that are presently buried at depth. At ~ 6 ka, the lake was several meters above Unit 6 nearshore sands. The lake then rose up probably to the middle Holocene (~ 6 – 4 ka, Chapter 4) lake highstand and deposited the Unit 8 deepwater lacustrine mud. Then the lake dropped to this site again and developed the most recent beach ridge in Unit 9. The age of Unit 9, however, is unknown. Finally, alluvial fan sediments of Unit 10 and 11 were deposited at the top.

Site O37

Site O37 is about several meters below, and to the southeast of Site O12 (Figure 3-6c). A large lens of medium-fine sands from the debris flow deposits at the bottom (see details in Appendix B and Figure B-10), gives an OSL age of 11.9 ± 0.6 ka, suggesting that the lake was below this site around 12 ka.

The combination of observations at Site O12 and O37, suggests that after establishing the lagoon behind Unit 4 beach ridge at Site O12 around 17 ka, the lake probably dropped to a level below the debris flow deposits at Site O37 around 13 ka, then it rose up again at these sites during ~ 6 – 4 ka (Chapter 4) and built the ancient beach ridges in Unit 6 at Site O12.
**Group 1c shorelines**

**Site O11**

This site shows a strike-perpendicular stratigraphic exposure that includes four units of sedimentary facies (Figure B-11). The silty sand in the sand wedges that I interpret as possible ice wedge casts near the bottom of the outcrop (see details in Appendix B and Figure B-11), yield an OSL age of 13.2 ± 3.3 ka, implying the lake level was below this unit during development of the ice wedge casts.

**Site O39**

Site O39 sits immediately below the Site O38 at the top of the alluvial (?) sediments (Figure B-4). The OSL age of the nearshore sands at the bottom (see details in Appendix B and Figure B-4), is 24.0 ± 0.9 ka, suggesting that the lake level was above this sample around 24 ka. In comparison with Site O11, the results suggest the lake was below these sites once around 24 ka and another around 13 ka.

**Group 1b shorelines:**

**Site O9**

Site O9 is immediately east of Site O10 (Figure 3-6c). Here only a very small portion of shoreline strata is exposed. The nearshore silty sands at the bottom (see detailed description in Appendix B and Figure B-5) provide an age of 3.8 ± 0.2 ka, suggesting that the lake was possibly above this site around 4 ka. But because of the limited exposure of the stratigraphic context, the confidence of this interpretation is low, and I do not use this site for further discussion of the lake level history.

**Site O8**

Site O8 has a similar simple stratigraphic context to Site O9. The detailed description is shown in Appendix B and Figure B-13. The possible nearshore medium sands at the bottom yield the same age as
O9, 3.8 ± 0.2 ka, suggesting the lake level was probably above this site around 4 ka. Again, the confidence is low because of insufficient stratigraphic context at the bottom of this outcrop.

In a short summary, the lake level was probably high around 4 ka, based on observations at Site O8 and O9, but the confidence for this inference is low, because of insufficiency of stratigraphic exposure at those two sites.

**Group 1a shorelines:**

*Site O15*

This site is the lowest among the three shorelines with OSL samples (Figure 3-6c). Three sedimentary facies units are observed in the 1.5 m deep soil pit (Figure 3-10). Unit 1 at the bottom is a thick layer of yellowish, poorly-sorted, mud-matrix-supported angular gravels that include some cobbles and boulders. A thin layer of fine sands is intercalated with this unit. Unit 2 includes a very thin bottom layer of coarse sands and a medium thick whitish layer of relatively-sorted, grain-supported, and subrounded to sunangular gravels. Unit 3 at the top is a ~15 cm thick layer of mud and silt sands.

The poorly sorted angular gravels with mud matrix in Unit 1 are interpreted as alluvial fan deposits that developed above the lake level. The grain-supported gravels in Unit 2 suggest that this unit is a typical beach ridge. And Unit 3 is a soil layer, which suggests that this shoreline is probably old. The OSL age of O15 from the fine sands is 12.7 ± 2.2 ka, implying the lake was below these deposits around 13 ka.

*Site O16*

Site O16 has a similar stratigraphic context to Site O15 (see Appendix B and Figure B-14 for details). The OSL age of O16 from the sands in the alluvial unit at the bottom is 22.0 ± 1.5 ka, suggesting the lake level was below this unit around 22 ka. The lake reached another lowstand below Site O15 at ~13 ka, which is consistent with observations at Site O13 and O11.
**Other sites around Siling Co**

A number of additional OSL samples were also collected from soil pits or outcrops distributed around Siling Co; the sample locations and ages are shown in Figure 3-11. Obviously only O26 shows a recent lake history, and other samples reflect an earlier history than Holocene. Here I only take O26 from the eastern margin of Siling Co as an example; the detailed descriptions for other samples are shown in Appendix B and Figure B-15~24.

**Site O26**

At this site, two units of sedimentary facies are observed (Figure 3-12). Unit 1 at the bottom is a very thick layer of medium/fine sand layers with very thin laminations. A significant relief exists at the contact between Unit 1 and 2. Symmetrical ripple marks can be found in many parts of this unit; climbing ripple marks are also preserved in the middle part of this unit. Unit 2 is a thick pile of layered, but poorly-sorted mixture of sands and pebbles, with a few lenses of very coarse intercalated sands.

I interpret Unit 1 that developed ripple marks in the sands as nearshore environment within the depth range of fair weather wave influence (about several meters). The thick layers of poorly sorted sands and pebbles in Unit 2 are considered as a sandy/pebbly beach ridge, which is typically observed in the eastern margin of Siling Co. The OSL age of O26 is 4.4 ± 0.2 ka, suggesting the lake level was above this site around 4.4 ka.

**3-6 Late Pleistocene lake level change of Siling Co**

**3-6.1 Holocene lake level change (10 – 0 ka)**

The new data from the central peninsula of Siling Co, in combination with the timing of the lake highstand during Middle Holocene (Chapter 4), provide a solid constraint on the Holocene lake level change of Siling Co, except for the period of ~ 10 – 6 ka. Here I try to reconstruct the lake history based
on the high quality sites to infer the lake levels from the stratigraphic context, those sites of low confidence are not utilized in discussion of the lake level change.

Immediately before the Holocene, during ~ 13 – 11 ka that overlaps the Younger Dryas, the alluvial fan and debris flow deposits at Site O13, O11, O37 and O15, suggest that the lake level was very low (Figure 3-13), only about 10 m above modern lake level (4530 m, Meng et al., 2012a) in 1976. Around 6 ka, the lake reached a relatively high level (~ 4572 m) which is close to the elevation of Group 2 shorelines (e.g., Site O17, O18 and O44). This is followed by the Middle Holocene lake highstand during ~ 6 – 4 ka (Chapter 4). Although Site G1d-O36, G1c-O38, and G1b-O8, O9 show the possibility of a significant drop in the lake level during 6 – 4 ka, I do not utilize these sites in overinterpretation, because of the low confidence of the inference of the lake level due to low quality of the stratigraphic context at these sites and no age clusters around those time periods. Moreover, Site O26, which has a high quality of stratigraphic context, does indicate that the lake was very high above that site, consistent with the Middle Holocene highstand inferred by the age clusters (Figure 3-13). Finally, these seven data points during the last 4 ka strongly suggest that this lake retreated continuously from ~ 4594 m to ~ 4575 m – 4580 m (Group 2 shorelines) at ~ 4 – 3 ka, then to ~ 4560 m (Group 1 shorelines) at ~ 2 – 1 ka, and finally to the modern lake level of ~ 4530 m in 1976 during the last thousand years. The total lake drop from the Middle Holocene highstand is ~ 64 m.

The lake history between ~ 11 – 6 ka, however, is not recorded in the 45 shorelines of my interests. Given that I have documented all flights of recently-developed shorelines and their subsurface strata in the central peninsula, and Chapter 4 has dated a number of highstand shorelines around different parts of the lake, none have been found to develop during this time window. Therefore the probability that this period of lake history recorded in shorelines preserved elsewhere around this lake is low. Three end-member scenarios are proposed here (Figure 3-13): 1) a continuous lake transgression from the very low level (< 4545 m) during ~ 13 – 11 ka near the Younger Dryas, to the level (~ 4570 – 4575 m) close to Group 2 shorelines. This scenario may explain little construction of shoreline deposits during the Early Holocene; 2) an abrupt lake rise at ~ 6 ka after a long time of low lake level during 13 – 6 ka; and 3) a
dramatic lake rise at ~ 13 ka to the lake highstand, followed by another large lake retreat at ~ 6 ka to the level close to Group 2 shoreline elevations. I would expect preservation of either highstand or lowstand shorelines for either of these latter two scenarios, which is no consistent with my observations of an absence of such shorelines in the field. Collectively, the scenario of a continuous lake transgression during ~ 11 – 6 ka seems a more reasonable explanation that is compatible with my field observations, but it is still uncertain.

3-6.2 Lake level change during 40 – 10 ka

Because of the discrepancy between the OSL age (~ 43 ka) and $^{36}$Cl depth profile exposure age (~ 113 ka) of Site O14, which might suggest the OSL samples of ages > 40 ka have been saturated, I categorize the pre-10 ka OSL ages obtained from the shoreline deposits into two groups (Figure 3-14), 40 – 10 ka and pre-40 ka. High confidence is given for ages in the former category in which most samples were collected from the shallow shoreline stratigraphy exposed in the central peninsula, especially the Tashi Stream. Low confidence is assigned for the latter case, which is discussed in Section 6.3.

The first order observation of Figure 3-14 is that the lake level of Siling Co generally declined from ~ 4625 m (O5) to < 4550 m (O15) during 35 – 10 ka, or even to modern lake level of 4530 m. But large lake fluctuations, to a second order, are obvious, with a cyclicity of ~ 5 – 10 ka (Figure 3-14). This finding contradicts previous suggestion of a simple lake recession since ~ 70 ka (Li et al., 2009). In detail, four relative low lake levels are observed during this period, at ~ 40 ka (G1d-O30), ~ 37 ka (G1e-43), ~ 22 ka (G1a-016), and ~ 13 ka (G1a-O15). The last two periods represent the lowest lake levels during 40 – 10 ka.

Inbetween these periods are three high lake levels, at ~ 35 ka (O5), ~ 25 ka (O3), and ~ 18 ka (O2). The last one has the same level (~ 4592 m) as the Middle Holocene lake highstand (~ 4594 m), and the former two sites reflect lake levels much higher than the Lingtong Holocene highstand.
3-6.3 Pre-40 ka lake level change

Several sites (O4, O14, O20, O23, O24 and O27) suggest Siling Co has reached above or close to Lingtong Holocene lake highstand during pre-40 ka. But because of the problem with OSL saturation in such old samples, the exact timing of these high lake levels cannot be determined from OSL dating technique. However, Chapter 2 is able to demonstrate that such high lake levels existed during MIS 5e – 6, using other dating methods, such as $^{36}$Cl cosmogenic inventories or U-series analyses. Therefore, the lake history between 40 ka and 100 ka may be resolved by resorting to these other techniques.

3-7 Discussions

3-7.1 Implications for Holocene paleoclimate of Tibet

For a closed, internally drained lake such as Siling Co, the lake levels (hence the lake volume) reflect the hydrological balance among water supplies (e.g., rainfall precipitation, glacial meltwater, ground water) and evaporation that is related to temperature. The most recent lake highstand of Siling Co during 6 – 4 ka (Figure 3-13) suggests a positive balance between the water inputs over evaporation. This large water supply very likely relate to 1) the increased Asian summer monsoon precipitation during 10 – 4.2 ka, as inferred from lake core geochemistry of Siling Co (Figure 3-15b) (Gu et al., 1993), and 2) substantial input of glacial meltwater from the Tanggula Shan due to the temperature increase in the Middle Holocene (Figure 3-15c) (Thompson et al., 1997). However, most studies argued that the peak monsoon precipitation occurred during the Early Holocene, and there is a general weakening from Early to Middle Holocene. If this is true, then the occurring of lake highstand of Siling Co during the Middle Holocene when the monsoon is weakening (Figure 3-15d), but not the Early Holocene implies significant water supply from the glacial melting. These analyses suggest a warm and wet climate condition of
central Tibet during the lake highstand period of Siling Co in the Middle Holocene, consistent with previous implications (Gu et al., 1993; Morrill et al., 2003; 2006).

The continuous lake retreat since ~ 4 ka (Figure 3-13) reflects large negative water balance, that is, evaporation significantly exceeds the water input from rainfall precipitation and/or glacial melting. This suggests a high probability of cold and dry conditions for the Siling Co region during the Late Holocene that possibly is associated with the weakening of Asian summer monsoon during this period (Gu et al., 1993; 1994; Morrill et al., 2003). This finding is consistent with previous arguments for high aridity in the Late Holocene in many parts of Tibet (e.g., Gu et al., 1993; Avouac et al., 1996; Morrill et al., 2003; 2006; Wu et al., 2006; Zhu et al., 2008; Lee et al., 2009; Mügler et al., 2010; Miehe et al., 2014), which suggests a regional control of the climate in Tibet during the Late Holocene.

The lack of shoreline preservation during 10 – 6 ka makes it difficult to infer paleoclimate during this period. If my reasoning of a lake transgression during this time (see above) (Figure 3-13) is true, then the lake level rise from a very low lake level during ~ 13 – 11 ka (corresponding to the Younger Dryas event) to the most recent lake highstand during the Middle Holocene, suggest a transition from cold-dry to warm-wet climate condition.

This study also shows a spatial variability of the timing of most recent lake highstand in Tibet. In western Tibet, the lake highstand was thought to occur during the Late Pleistocene – Holocene transition for Sumxi-Longmu Co (Gasse et al., 1991; Avouac et al., 1996; Kong et al., 2007), and at ~ 10 – 9 ka and 7 – 6 ka for Bangong Co (Gasse et al., 1996). In central Tibet, the lake highstand occurred before ~ 8 ka in Tangra Yumco (Rades et al., 2013), ~ 6 – 4 ka in Siling Co (Chapter 4), at ~ 5 – 4 ka in Gyaring Co (Chapter 5), ~ 7 – 5 ka at Nam Co, and, ~ 9 – 6 ka in Cuo E. This spatial difference of when the lakes in Tibet reached their highstand during the Holocene may be related to the enduring spatial heterogeneity of monsoon rainfall amount that tied to different monsoon subsystems (e.g., Indian summer monsoon vs. East Asian monsoon, Hudson et al., 2013), or, to the ambiguity in definition of Holocene lake highstand, as the geomorphic expression of such lake highstand has not been well-examined until recently (Meng et al., 2012b; Hudson et al., 2013; Chapters 2 and 4).
3-7.2 Implications for Late Pleistocene (40 ka – 10 ka) paleoclimate of Tibet

The general declining of the lake level from ~ 4625 m (O5) to < 4550 m (O15) (Figure 3-16a), or ~ 75 m of lake drop during 35 – 10 ka, suggest a substantial negative water balance, near the end of last glacial period. To a first order, the period of 35 – 15 ka corresponds to a period with 1) relatively cold climate as recorded in the low δ¹⁸O values of the Guliya ice core in northwestern Tibet (Figure 3-16b) (Thompson et al., 1997), and 2) with a general decrease in monsoon precipitation as reflected by the increase in δ¹⁸O value (Figure 3-16c) (Wang et al., 2001). On the other hand, this period of 35 – 15 ka also corresponds to the timing approaching the last glacial maximum, thus I expect an increase of water storage in mountainous glaciers (e.g., Geladandong glacier in the Tanggula Range to NE to Siling Co, Figure 3-1), and hence less water input in the lakes. Therefore, the glacial process may predominantly control the lake evolution during 35 – 15 ka.

In terms of higher frequency of large lake fluctuations during 40 – 10 ka, the three periods of high lake levels close or above the Holocene highstand at ~ 35 ka (O5), ~ 25 ka (O3), and ~ 18 ka (O2), suggest a large positive water balance and high possibility of wet climate condition during these periods. The mechanisms that drove the wetness and positive water balance during those periods, however, could be different. The high lake level at ~ 35 ka occurred in a period of relatively warm condition in MIS 3, as indicated by the Guliya ice core record (Thompson et al., 1997) (Figure 3-16b) and also relatively high monsoon intensity (Figure 3-16c) (Wang et al., 2001). Therefore, the large water body of Siling Co may be caused by the warm and wet climate, possibly related to strong summer monsoon during this period. In contrast, the lake highstand at ~ 25 ka and ~ 18 ka occurred during MIS 2 glacial period. The monsoon intensity during this period, however, varies considerably, but in general is much lower than around ~ 35 ka (Figure 3-16c), and the temperature is much colder than 35 ka. Therefore, monsoon rainfall, glacial melting and substantial evaporation loss, all may play a role in the high lake levels around 25 ka and 18 ka. The two periods of ~ 35 ka (MIS 3) and ~ 18 ka (MIS 2) of high lake levels constrained from this study are correlative to the proposed lake expansion in Tibet during 40 – 28 ka and 19 – 15 ka,
respectively, based on a regional data compilation (Jia et al., 2001). This comparison gives an additional support for a regional climate control on the lake expansion during these two periods.

Within the period of 40 – 10 ka, two relatively low and another two extremely low lake levels have been observed in this study, which are at ~37 ka (G1d-O30), ~27 ka (G1e-43), ~23 ka (G1a-O16), and ~13 – 11 ka (G1a-O15) (Figure 3-14). The existence of such lake lowstand suggests a very dry climate condition. The mechanisms driving such low lake levels, again, are different. The relative lowstand around 35 ka (05) occurred in a period of high monsoon intensity (Figure 3-16c), but relatively high temperature that may lead to large input of glacial meltwater. Both suggest a relatively large water input around this time. Therefore the only explanation for this low lake level is that the evaporation is significant enough to overcome the large water input during this time. The lowstands at ~27 ka (O3) and ~23 ka (G1a-016) occurred when the monsoon precipitation largely decreases, and the temperature is low. Therefore, although the evaporation may be low in these periods, substantial loss of water input could lead to the lake lowstands. Around 13 – 11 ka (G1a-O15), the monsoon intensity reached its minimum during the Younger Dryas period, the temperature is not as cold as in the periods of ~27 ka and 23 ka, therefore, extremely low lake level during ~13 – 11 ka is probably induced by the low monsoon rainfall at this time.

In summary, the glacially melting process probably dominate the general lake level change of Siling Co during 40 – 10 ka, but in details, the lake evolution may be controlled partially or predominantly by the monsoon intensity, glacial melting and/or evaporation.

3-7.3 Geodynamic implications from the lake level change

The lake level change constrained from the paleoshoreline elevation and chronology also provides a direct observation of the history of lake loading and unloading, which can be utilized to constrain the crustal elastic strength and viscosity by studying the shoreline deflections induced by the crustal flexure in response to the change in lake loads (e.g., Bills et al., 1994; 2007; Chapter 4). Therefore
the duration of high expansion (loading) and contraction (unloading), together with the relaxation
timescale of the flexural isostatic adjustment (Watts, 2001; Turcotte and Schubert, 2002), plays a role in
the magnitude of shoreline deflection within a specific time window. Thus the constraints of lake level
change can provide important information to investigate the crustal rheology of the Tibetan plateau
through viscoelastic response of the lithosphere to such time-varying change in lake loads.

3-8 Conclusions

In this study, I utilized a sequence of paleoshorelines preserved around Siling Co, especially in
the central peninsula and the shoreline chronology to establish a relatively comprehensive history of lake
level change of Siling Co. In total I have collected 36 new OSL samples (32 samples are < 40 ka) from
the surficial and subsurface deposits in the shoreline strata to infer the lake levels relative to the sample
position from interpretation of the sedimentary facies and depositional environment, and also to obtain the
timing of the lake level from the sample ages. In combination with the OSL ages of the most recent
highstand shorelines, these new data provide a complicated lake history since 40 ka.

1) Siling Co generally declined during the past 40 ka. It has reached the lake highstand (~ 4590 – 4595
m) four times from 40 ka, which are ~ 35 ka, ~ 30 – 25 ka, ~ 18 ka and ~ 6 – 4 ka, indicating wet
conditions during these periods. In between these high lake levels are relatively or extremely low
stands at ~ 40 ka, ~ 30 ka, ~ 22 ka and a most recent lake recession since ~ 4 ka, suggesting
substantial aridity at these times. These results collectively suggest a moderate cyclicity of the lake
fluctuation possibly related to paleoclimatic signals.

2) The glacial process generally controls the lake evolution of Siling Co, but in details, the lake level
change may be controlled partially or predominantly by the monsoon intensity, glacial melting
and/or evaporation.
3) Such detailed lake history also provides a history of lake loading and unloading that is necessary for investigate the lithospheric rheology from flexural deformation of the shorelines in response to the climatically induced temporal change in lake loads.
3-9 References


Figure 3- 1. The LandSAT imagery showing geographic locations of saline lakes (black polygons) in Tibet. A few lakes mentioned in the text are labeled here.
Figure 3-2. This map shows the three shoreline groups (at, above or below the lake highstand of ~4594 m in elevation) categorized by Chapter 2 (see text for details). The dark blue polygon outlines the lake extension at its most recent lake (Lingtong) highstand during the Middle Holocene (Chapters 2 and 4), and the light blue polygons show current lake area. Also shown here are locations of the new OSL samples in this study (black circles and labels) and those (white circles and labels) published in Chapter 4.
Figure 3-3 (previous page). (a) GeoEye imagery showing details of the shoreline groups above (red line), at (yellow line) and below (white) the Lingtong highstand, and also a couple of subgroups below the highstand, in the central peninsula of Siling Co. The new OSL sample locations in the surficial shoreline deposits are marked as red circles and labels, whereas one sample (G3-O21) published in (Chapter 4) is labeled in white text and circle. The solid and dashed lines represent the actual and projected paths, respectively, of shoreline topographic survey, shown in the lower panel (b). (c) The map view of beach ridge clusters preserved but cut by the Tashi Stream in the central peninsula of Siling Co. Also shown in the map are several phases of alluvial fans (black polygons with labels of Roman numerals) developed in association with the beach ridges, and locations of surficial (red circles) OSL samples. (d) The topography of beach ridges preserved in the vicinity of the Tashi Stream.
Figure 3-4. The soil pit profile and graphic log and interpretation of depositional environments at Site O18 in the Shoreline Group 2 (G2) (see detailed descriptions of the stratigraphy in the text).
Figure 3-5. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O10 in the Shoreline Group 1b (G1b) (see detailed descriptions of the stratigraphy in the text). Here two OSL samples (O10 and O41) were collected from two different sedimentary facies units.
Figure 3-6 (previous page). (a) GeoEye imagery showing details of the shoreline groups above (red line), at (yellow line) and below (white) the Lingtong highstand, and also a couple of subgroups below the highstand, in the central peninsula of Siling Co. The new OSL sample locations in the shallow shoreline strata are marked as black circles. Notice a 36Cl depth profile age (~113 ka) obtained from the same soil pit of O14 is labeled at the lower left corner of this map. The solid and dashed lines represent the actual and projected paths, respectively, of shoreline topographic survey, shown in the lower panel (b). (c) The map view of beach ridge clusters preserved but cut by the Tashi Stream in the central peninsula of Siling Co. Also shown in the map are several phases of alluvial fans (black polygons with labels of Roman numerals) developed in association with the beach ridges, and locations of subsurface (black circles) OSL samples. (d) The topography of beach ridges preserved in the vicinity of the Tashi Stream.
Figure 3-7. The soil pit profile, graphic log and interpretation of depositional environments at Site O14 above the Lingtong highstand (see detailed descriptions of the stratigraphy in the text).
Figure 3-8. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O43 in the Shoreline Group 1e (G1e) (see detailed descriptions of the stratigraphy in the text).
Figure 3-9. The photos and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O12 in the Shoreline Group 1d (G1d) (see detailed descriptions of the stratigraphy in the text). Here four OSL samples (O12, O28-O30) were collected from different sedimentary facies units.
Figure 3-10. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O15 in the Shoreline Group 1a (G1a) (see detailed descriptions of the stratigraphy in the text).
Figure 3-11. The map shows OSL ages of subsurface samples collected around Siling Co except the central Peninsula. The labels in gray indicate ages older than 40 ka that are not used in discussion of the lake history.
Figure 3-12. The photo and sketch of shoreline stratigraphy, and interpretation of depositional environments at Site O26 in the eastern margin of Siling Co (see location in Figure 3-11 and detailed descriptions of the stratigraphy in the text).
Figure 3-13. Reconstructed lake level change since 15 ka. Numbers 1 – 3 denote three different scenarios of lake history between 11 – 6 ka (see the text for detailed discussions).
Figure 3-14. Reconstructed lake level change during 40 – 10 ka, but only the lake history during 40 – 10 ka (without gray shading) is considered as reliable in this study (see the text for detailed discussions).
Figure 3-15. Comparison between the Holocene lake level change reconstructed in this study (a) and other paleoclimatic proxies. (b) The $\delta^{18}O$ values measured from lacustrine carbonates from a drill core in Siling Co (Gu et al., 1993), which has been used as a proxy of monsoon intensity in this region. (c) The $\delta^{18}O$ values of Guliya ice core (Thompson et al., 1997), which has been used as a proxy of paleotemperature in Tibet. (d) The $\delta^{18}O$ values measured from stalagmites in the Dongge Cave in East China (Dykoski et al., 2005), which has been used as a proxy of monsoon intensity of East China. The blue and red shading represents periods of high and low lake levels, respectively.
Figure 3-16. Comparison between the lake level change during 40 – 10 ka reconstructed in this study (a) and other paleoclimatic proxies. (b) The $\delta^{18}O$ values of Guliya ice core (Thompson et al., 1997), which has been used as a proxy of paleotemperature in Tibet. (c) The $\delta^{18}O$ values measured from stalagmites in the Dongge Cave (blue) (Dykoski et al., 2005) and the Hulu Cave (purple) (Wang et al., 2001) in East China, which has been used as a proxy of monsoon intensity of East China. The blue and red shading represents periods of high and low lake levels, respectively.
<table>
<thead>
<tr>
<th>Facies Symbol</th>
<th>Facies Description</th>
<th>Depositional environment</th>
<th>Water Depth</th>
<th>Examples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Als</td>
<td>Soil developed on top of beach ridges</td>
<td>Soil layers</td>
<td>Above water level</td>
<td>McFadden et al., 1992</td>
</tr>
<tr>
<td>Aldbr</td>
<td>Large poorly sorted cobbles or boulders mixed with massive sands or mud</td>
<td>Debris flow or braided stream</td>
<td>Above water level</td>
<td>Nemec and Steel, 1984; Bartov, 2002;</td>
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<tr>
<td>Alf</td>
<td>Poorly sorted non-stratified to sub-parallel-stratified gravels/pebbles, with matrix of mud, and sands of variable sizes, or mud layer mixed with sparse gravels</td>
<td>Alluvial fan</td>
<td>Above water level</td>
<td>Nemec and Steel, 1984; Bartov, 2002;</td>
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<td>Alc</td>
<td>Clean rounded pebbles with lenses of sands within a scoured channel-like bottom</td>
<td>Alluvial channel</td>
<td>Above water level</td>
<td>Bartov, 2002;</td>
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<tr>
<td>Alw</td>
<td>nearly V-shaped wedges filled with massive sands, vertically cutting beach gravel layers</td>
<td>Alluvial fan</td>
<td>Above water level</td>
<td>Madsen et al., 2008</td>
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<tr>
<td>Alb</td>
<td>Coarse/fine sand layers (possibly cross-bedded) interfingered with poorly sorted gravel lenses</td>
<td>Braided stream</td>
<td>Above water level</td>
<td>Renaut and Owen, 1991; Bartove, 1999; 2002</td>
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<td>Lgn</td>
<td>Fine sand layers on the backsets of beach ridges</td>
<td>Lagoon</td>
<td>0 m (always deposited above or at water level)</td>
<td>Adams and Wesnousky, 1998</td>
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<td>Brb</td>
<td>mostly grain-supported, well sorted, platy pebbly clinoforms dipping landward (backsets), may have intercalated coarse to fine sand lenses/layers</td>
<td>Beach ridge</td>
<td>0 m (always deposited above or at water level)</td>
<td>Renaut and Owen, 1991; Bartove, 1999; 2002</td>
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<td>mostly grain-supported, well sorted, platy pebbly clinoforms dipping lakeward (foresets of beach ridges, spits and tombolos), may have intercalated coarse to fine sand lenses/layers</td>
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<td>0 m</td>
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<td>Brt</td>
<td>mostly grain-supported, well sorted, platy pebbly flat beds, may have intercalated coarse to fine sand lenses/layers</td>
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<td>Ns1</td>
<td>Medium to coarse well-sorted sands, mostly gently-dipping lakeward; symmetric or climbing ripple marks may develop</td>
<td>Near shore wave influenced</td>
<td>1 - 5 m</td>
<td>Sneh, 1979; Machlus, 2000</td>
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<td>Ns2</td>
<td>Fine sand to silt layers, well-sorted</td>
<td>Near shore below wave influence</td>
<td>1 - 15 m</td>
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<td>Dl</td>
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<td>Begin et al., 1974</td>
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<td>Long (°E)</td>
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<td>----------------</td>
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<td><strong>Central Peninsula</strong></td>
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<tr>
<td>XS-SL-OSL-14 O14</td>
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**Table 3- 2. OSL sample data of Siling Co**
### Table 3-2 (cont.). OSL sample data of Siling Co

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<th>Long (°E)</th>
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<th>Depth (m)</th>
<th>Samp Elev (m)</th>
<th>Dose Rate (Gy/ka)</th>
<th>Error (Gy/ka)</th>
<th>Age (Gy)</th>
<th>Age Error (Ka)</th>
<th>Model</th>
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<td><strong>East margin</strong></td>
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<td>4620.6</td>
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CHAPTER 4
CRUSTAL STRENGTH IN CENTRAL TIBET DETERMINED FROM LACUSTRINE SHORELINE DEFLECTION

Abstract

Whether the deep crust beneath the Tibetan Plateau is weak enough to sustain flow on geologic timescales is highly debated. Geophysical observations of the present physical state of Tibetan crust suggest that portions of the middle and lower crust are relatively hot and fluid-rich, conditions that favor low viscosity. Direct determinations of bulk crustal rheology, however, remain relatively few. Using the flexural rebound of lacustrine shorelines developed during a highstand around Siling Co, in central Tibet, I determine the effective elastic thickness (Te) of Tibetan crust and place bounds on allowable viscosity structures compatible with crustal composition and thermal state. Shoreline features associated with a highstand complex ~ 60 meters above present lake level are deflected from horizontal by 3 - 5 meters over wavelengths of tens of kilometers. Optically-stimulated luminescence dating of aggradational shoreline deposits indicate that these maximum lake levels were reached at 6 - 4 ka and subsequently fell to present-day levels. Comparison of measured shoreline deflections with a 3D elastic model constrains Te in central Tibet to 12 - 14 km. When combined with geophysical constraints on crustal structure, composition, and heat flow, the strain rates implied by flexural rebound require a relatively low viscosity middle crust, with an effective viscosity that decreases from \( 10^{20} \) to \( 10^{18} \) Pa s from 20 - 40 km depth. My results confirm notions that the middle crust beneath central Tibet is likely capable of lateral flow over millennial timescales and provide quantitative bounds on the viscosity structure with depth.
4-1 Introduction

The rheology of the deep crust in orogenic systems has long been understood to influence the distribution, rates and style of lithospheric deformation (Argand, 1924; England and Houseman, 1986; Zhao and Morgan, 1987). However, determining the rheologic properties of actively deforming lithosphere is difficult, and much of my understanding of the bulk rheology of continental crust relies on either laboratory estimates of the constitutive behavior of rock materials (Brace and Kohlstedt, 1980) or on measurements of transient surface deformation following large earthquakes (Savage and Prescott, 1978; Pollitz et al., 2001; Bürgmann and Dresen, 2008). To a large degree, debates over the distribution of strength in continental lithosphere (Jackson, 2002; Watts and Burov, 2003) and the related question of whether deformation is continuous between crust and mantle (England and McKenzie, 1982; Wang et al., 2008), owe their persistence to the difficulty of knowing the rheologic properties of the lithosphere in space and time.

Few doubt that the middle and lower portions of the crust of the Tibetan Plateau have undergone deformation during the Indo-Asian collision, but the nature of this deformation is one of the most hotly debated questions in continental dynamics today. On one hand, it is argued that the plateau owes its very existence to widespread lateral flow of the deep crust (Bird, 1991; Royden et al., 1997; Shen et al., 2001; Royden et al., 2008). Abundant geophysical data indicate that the lower crust should be quite weak at high temperatures and in the presence of fluids (Brown et al., 1996; Nelson et al., 1996; Unsworth et al., 2005; Bai et al., 2010; Zhao et al., 2012). Geologic observations from eastern Tibet implicate the large-scale influx of lower crust to explain the growth of plateau topography in that region (Clark and Royden, 2000; Royden et al., 2008), and, numerous aspects of the tectonic evolution of the Himalaya can be explained by channelized flow of crustal rocks from beneath Tibet (Beaumont et al., 2001; Beaumont et al., 2004). On the other hand, however, the correspondence of geodetic velocity fields and the inferred directions of seismic anisotropy in the mantle (Flesch et al., 2005) supports the idea of vertically-coherent deformation throughout Tibetan lithosphere (Bendick and Flesch, 2007; Sol et al., 2007). Geological
observations from xenoliths beneath Tibet suggest that portions of the lower crust may be fluid-poor
(Hacker et al., 2000), and several key aspects of the geology of the Himalaya do not seem to require
channel flow (Harrison, 2006; Robinson and Pearson, 2006). Likewise, the post-seismic response of
lithosphere in northern Tibet does not appear to require unusually low viscosity of the lower crust (Hilley
et al., 2005; Yamasaki and Houseman, 2012). Thus, whether the deep crust beneath Tibet is weak enough
to be capable of widespread flow on geologic timescales is a key unknown in our understanding of this
archtypal orogen.

Here I approach this question of crustal flow by measuring surface deformation around a large
lake, Siling Co (‘co’ is the Tibetan word for lake) (Figure 4-1), in central Tibet, that occurred in response
to climatically induced changes in lake levels. Changes in the volume of lake systems represent a surface
load on the lithosphere, and the consequent flexural response is sensitive to the flexural rigidity of the
elastic portion of the crust, to the viscosity of the deep crust and/or mantle, and to the timescale of loading
or unloading. Moreover, shorelines developed during highstand levels represent a paleo-horizontal datum
from which to measure deflections, as well as provide an estimate of the volume of the load. This
methodology is a proven means of estimating the bulk constitutive behavior of the lithosphere and has
provided key constraints in the Basin and Range (Gilbert, 1890; Crittenden, 1963; Bills et al., 1994a;
Adams et al., 1999), the Andes (Bills et al., 1994b), and most recently, in the Mediterranean (Govers et al.,
2009). It has also been recently use to place bounds on the rheology of the crust in west-central Tibet
(England et al., 2013); as I argue below, my study overcomes some critical challenges faced by this study,
and, consequently, I arrive at a somewhat different conclusion. Central to all of these efforts is the notion
that variations in water loads tend to occur over timescales of several thousand years and across spatial
scales of tens to hundreds of kilometers. The resultant estimates of lithospheric rheology thus can fill a
gap between inferences derived from the geologic evolution of orogens over millions of years (Clark and
Royden, 2000) and those derived from decadal measurements of post-seismic transient surface
deforation (Hilley et al., 2005; Yamasaki and Houseman, 2012; Huang et al., 2014).
In this paper, I evaluate shoreline deformation around Siling Co in the central portion of the Tibetan Plateau. This lake sits at an elevation of ~ 4530 m in 1976 and extends over an area of ~ 1660 km² (Meng et al., 2012a). It is situated just south of the Banggong – Nujiang suture separating the Lhasa and Qiangtang terranes, and lies just to the west of the primary route of the INDEPTH 3 geophysical experiment (Zhao et al., 2001; Ross et al., 2004). Flights of Late Pleistocene – Holocene shorelines are preserved at elevations up to ~ 100 m above present lake level (Li et al., 2009). Here I combine recent mapping and surveying of shoreline elevations (Meng et al., 2012b) with new optically stimulated luminescence (OSL) ages to develop the chronology of shoreline features that allows us to constrain the amplitude, wavelength and timescale of shoreline rebound. I utilize these data in a 3D flexural model to evaluate the flexural rigidity of the Tibetan crust beneath Siling Co. Finally, I combine my results with geophysical constraints on crustal structure, composition, and thermal state to determine the range of viscosities for the middle and lower crust that reconcile all of these data.

4-2 Background

4-2.1 Physical state of Tibetan crust

Numerous geophysical experiments conducted across the Tibetan Plateau over the past two decades provide a wealth of data which place constraints on the structure of physical state of the Tibetan lithosphere (Molnar, 1988; Klemperer, 2006). Of particular relevance are 1) high heat flow in southern Tibet, > 80 mW/m² (Francheteau et al., 1984; Hu et al., 2000; Wang, 2001), consistent with temperatures near or above the wet solidus in the lower crust (Alsdorf and Nelson, 1999; Mechie et al., 2004); 2) marked north-south heterogeneity in the seismic characteristics of the upper mantle – slow (Chen and Molnar, 1983; Brandon and Romanowicz, 1986; Owens and Zandt, 1997) and strongly attenuated (Ni and Barazangi, 1983) shear-waves along paths in northern Tibet. These results suggest that the mantle beneath northern Tibet is relatively hot, whereas high shear-wave velocities beneath southern Tibet indicate that
the upper mantle is perhaps as much as 200-300°C cooler (McNamara et al., 1997); 3) high electrical conductivity across much of Tibet (Wei et al., 2001; Unsworth et al., 2005; Bai et al., 2010; Zhao et al., 2012), indicative of the ubiquitous presence of fluid throughout the crust; 4) high-amplitude seismic reflections (‘bright spots’) associated with P-to-S-wave conversions (Brown et al., 1996; Makovsky and Klemperer, 1999), interpreted to represent the presence of either magmatic (Nelson et al., 1996) or aqueous (Makovsky and Klemperer, 1999) fluids; 5) high seismic attenuation in central (Rodgers and Schwartz, 1998) or even much of Tibetan crust (Ruzakin et al., 1977), a characteristic interpreted to reflect high temperatures and/or partial melt in the crust (Fan and Lay, 2003); 6) reflective, layered lower crust in central Tibet (Ross et al., 2004); 7) radially anisotropic regions of the deep crust (Ozacar and Zandt, 2004; Shapiro et al., 2004; Duret et al., 2010); and 8) low seismic wave speed the middle crust beneath southern (Kind et al., 1996), southeastern (Xu et al., 2007; Yao et al., 2008), or much of peripheral regions of Tibet (Yang et al., 2012).

Although many of these observations are consistent with the possibility that the lower crust of Tibet is quite weak, a number of other observations challenge this perspective. These include: 1) the presence of dry lower crustal granulite xenoliths from northern Tibet (Hacker et al., 2000); 2) disconnected middle crustal low velocity zones (Hetényi et al., 2011); and 3) locally coherent surface displacement velocity vectors and upper mantle anisotropy suggesting similar deformation at different levels within the crust and mantle lithosphere (Flesch et al., 2005; Sol et al., 2007; Copley et al., 2011; León Soto et al., 2012). The conflicting nature of these geophysical observations highlights the need to directly assess the Tibetan deep crustal rheology.

4-2.2 Previous estimates of crustal rheology

Attempts to address the various geophysical observations have motivated studies to directly probe the rheology of Tibetan deep crust. These efforts generally fall into two categories: 1) determinations of the flexural rigidity of the elastic portion of the lithosphere, and 2) models of the viscoelastic response to
loads imposed by large earthquakes. Analyses of topography along rift flanks in southern Tibet suggests a relatively low flexural rigidity ($T_e \sim 2$ to $4 \text{ km}$) (Masek et al., 1994), whereas gravity anomalies and topography across the plateau suggest a wider range of $T_e$ from $\sim 7 \text{ km}$ (Fielding and McKenzie, 2012) to perhaps as high as $\sim 30 \text{ km}$ (Braitenberg et al., 2003; Jordan and Watts, 2005; Chen et al., 2013). Regional differences in these estimates may reflect heterogeneity of crustal strength, potentially associated with strain localization along rift zones (Harrison, 2006; Hetényi et al., 2011).

Transient surface deformation following large earthquakes provides constraints on the viscoelastic behavior of the lithosphere. Available data suggest that the lower crust beneath the western (Ryder et al., 2007) and central (Hilley et al., 2005; Hilley et al., 2009; Ryder et al., 2011; Wen et al., 2012) segments of the Kunlun fault (the 1997 $M_w$ 7.6 Manyi and the 2001 $M_w$ 7.9 Kokoxilli earthquakes, respectively) are consistent with steady-state viscosities $>10^{18} - 10^{19} \text{ Pa s}$. However, if low viscosity material is confined to a thin ($< 20 \text{ km}$) channel in the middle crust, somewhat lower viscosities ($\sim 10^{17} - 10^{18} \text{ Pa s}$) may also fit the post-seismic surface deformation (DeVries and Meade, 2013). Although large earthquakes within the central plateau are few, deformation following the 2008 $M_w$ 6.4 Nima-Gaize and the 2008 $M_w$ 6.3 Dangxung normal-faulting earthquakes is compatible with a somewhat lower viscosity of $\sim 10^{17} - 10^{18} \text{ Pa s}$ (Ryder et al., 2010; Bie et al., 2014). In eastern Tibet, post-seismic surface deformation following the 2008 $M_w$ 7.9 Wenchuan event suggests that the lower crust beneath the eastern plateau has a steady-state viscosity of $\sim 10^{18} \text{ Pa s}$ (Huang et al., 2014).

Because the effective viscosity of the lithosphere depends, in part, on the timescale of loading (Watts, 2001; Watts et al., 2013), consideration of the timescales of the surface loads may be important (Bills et al., 1994a; Bills et al., 2007). The apparent absence of shoreline deflection around Zhari Nam Co (Figure 4-1) in south central Tibet (England et al., 2013) places a lower bound on the average viscosity of the ductile crust ($>10^{19} - 10^{20} \text{ Pa s}$). This interpretation may be biased by several factors, including 1) the relatively small size of the load, 2) the simple cylindrical geometry of the basin, such that preserved shorelines are approximately equidistant from the center of the load (England et al., 2013), 3) the lack of high-precision surveys of differences in shoreline elevation, and 4) no age constraint on the shorelines for
4-2.3 Lake level changes in Tibet

Climatically induced changes in lake level act as time-dependent loads imposed on the surface of the lithosphere, and sufficiently large variations in water level can cause observable flexural response of the lithosphere (Bills et al., 1994a; Bills et al., 1994b; Hampel et al., 2010). Many lakes in the central and southern parts of the Tibetan Plateau have flights of shoreline features that extend many tens to hundreds of meters above present lake levels. However, the timing of lake expansion and contraction are only known in a few systems (Kong et al., 2011; Rades et al., 2013). Previous studies from lakes in western Tibet suggest that maximum lake levels occurred during the Pleistocene – Holocene transition (Gasse et al., 1991; Jia et al., 2001; Kong et al., 2007) or the Early Holocene (Rades et al., 2013). This lake expansion is linked to a strong Indian monsoon (Gasse et al., 1991; Jia et al., 2001) and/or an increase in glacial meltwaters (Kong et al., 2007) during this time. At my study site around Siling Co, previous age control was limited; OSL dates from beach deposits range from ~ 70 ka to as young as ~ 6 ka (Li et al., 2009) and generally suggest lake recession since 70 ka. Exposure age dating of bedrock outcrops interpreted to be wave-cut platforms, however, yield ages as old as ~160 ka to ~ 250 ka (Kong et al., 2011) suggesting the possibility of an older and more complicated history.

4-2.4 Crustal and thermal structure beneath central Tibet

Two seismic experiments, INDEPTH 3 (Zhao et al., 2001; Ross et al., 2004), which extended from west of Nam Co, northward to the Tanggula Shan (Figure 4-1) and the 1982 Sino-French expedition (SF-2, Figure 4-1) (Zhang and Klemperer, 2005), provide constraints on crustal and lithospheric properties in the vicinity of Siling Co. Seismic reflection data reveal that the crust south of the Bangong-
Nujiang suture (Figure 4-1) is ~ 65 ± 5 km thick. Sediment thickness along the INDEPTH 3 profile ranges from 1 – 5 km (Ross et al., 2004; Haines and van der Pluijm, 2008; Mechie et al., 2011), and the upper crust beneath the sediment cover is considered granitic in composition (Min and Wu, 1987; Zhang et al., 2011). Seismic velocities in the lower crust suggest a composition consistent with diabase or mafic granulite (Zhang et al., 2011), and the boundary between the felsic upper crust and mafic lower crust in the Siling Co area is interpreted to be at ~ 35 – 40 km at depth (Zhao et al., 2001; Ross et al., 2004; Mechie et al., 2011).

There are few direct constraints on the thermal structure of Tibetan crust. Heat flow measurements in Tibet have been carried out by probe penetration in saline lakes in southern Tibet and in boreholes in several localities in central Tibet. These studies reveal elevated heat flow, ranging from 80 to 110 mW/m² (Francheteau et al., 1984; Shen et al., 1984; Wang and Huang, 1990; Hochstein and Regenauer-Lieb, 1998; Hu et al., 2000; Wang, 2001). Somewhat lower values of 60 – 70 mW/m² were measured in the Lhasa region and a high value of 140 mW/m² was observed in the Lunpola basin (~ 40 km northeast of Siling Co, Figure 4-1). Some indirect observations imply high heat flow and associated elevated crustal temperatures (Klemperer, 2006). These include 1) relatively low intensity rock magnetism from satellite measurements most likely reflecting that the Curie isotherm (~ 550 °C) resides at a shallow depth of ~ 15 km beneath much of interior Tibet (Alsdorf and Nelson, 1999), 2) seismic velocity data suggestive of an α-β quartz transition at ~18 km beneath southern Qiangtang terrane and temperatures in excess of ~700°C (Mechie et al., 2004), and 3) mafic xenoliths from granulite facies rocks in the lower crust of the northern Qiangtang terrane appear to have equilibrated at 800 – 1100 °C and depths of 30 -50 km (Hacker et al., 2000).

4-3 Geomorphology, deflection and age of shorelines around Siling Co

Extensive flights of relict shorelines are preserved around Siling Co and its neighboring lakes. Mapping these shorelines using high resolution (0.5 m nominal resolution) satellite imagery (Figure 4-2)
reveals a prominent group of shoreline features at ~ 4594 m elevation that mark a continuous and distinct boundary between older geomorphic features above the shoreline level and younger features below. The shoreline itself is characterized by both constructional features such as beach ridges, spits, tombolos and cuspsate bars and erosional wave-cut scarps developed in both alluvial fans and bedrock exposures (Figure 4-2). Notably, alluvial fans truncated by wave-cut scarps along this shoreline are dissected by deep gullies and channels, features that suggest erosional degradation of the landscape over a relatively long period of time. Although there are sparse, relict shoreline features present above ~ 4594 m (Li et al., 2009), these are discontinuous around the lake, tend to be poorly preserved, and some even exhibit polygonal “patterned ground” consistent with a protracted period of permafrost activity (Jorgenson et al., 2006; Anderson and Anderson, 2010). In contrast, geomorphic features below the ~ 4594 m shoreline represent groups of continuous beach ridges developed across a strandplain; ridges themselves are fresh, undissected, and only the most active alluvial fans drape these beach ridges. Thus, the shoreline at ~ 4594 m elevation appears to represent a highstand strandline that was extensive around Siling Co, and recession from this level is marked by the deposition of multiple beach ridges, possibly reflecting short-lived stillstands. The wide spatial distribution and continuity of the highstand shoreline on both central peninsulas and along the margins of Siling Co makes this an ideal marker from which to measure variations in shoreline deflection with distance from the center of the former lake.

To precisely determine current relative elevations along the highstand shoreline, I focused on constructional features. These have several advantages over wave-cut cliffs/scarps in determining the position of the ancient lake level (Adams and Wesnousky, 1998): 1) beach barriers take some time to form, and therefore reflect a relatively stable paleolake level, and the elevation of swash surfaces along the shoreface represents a reasonable estimate of mean water level; and 2) wave-cut cliffs retreat during successive undercutting and failure, and it can be difficult to define the shoreline angle that should represent mean water level; 3) well-preserved flat crests of constructional shorelines are relatively easy to survey, and the sediments that comprise constructional features afford the potential for robust estimates of the timing of shoreline development.
I surveyed constructional features at 70 localities along the highstand shoreline complex using differential GPS (Meng et al., 2012a) that allowed us to measure differences in elevation with precision at the decimeter level (see Supplemental Materials). Features that may have been associated with longshore transport (spits and tombolos) were surveyed at the point of attachment to the shoreline. These positions likely represent a minimum estimate on water elevation. Beach ridges and cuspathe bars are considered more reliable, as they likely formed at, or immediately above, the fair weather wave base (Tanner, 1995). I estimate that these features represent mean water level to within ~0.5 – 1 m. In all cases, I occupied numerous survey points (typically 9 observations) on each feature and took the average of all points with >15 minutes occupation time. Of the 70 shoreline survey localities, 56 show lateral continuity of >1 km around the main body of the paleolake (encompassing present day Siling Co and Co E). Fourteen additional sites were surveyed around Wuru Co and Bange Co and represent sites whose correlation to the main highstand shoreline complex is uncertain (Figure 4-3 and Table C-8, and see detailed description in Supplemental Materials).

My results reveal that relative heights of the surveyed shoreline features along the highstand level vary systematically with distance away from the lake center (Figure 4-3d). Deviation from a horizontal datum along a ~30 km long transect from the lake margin to its center ranges up to ~5 m (Figure 4-3d). Measurement uncertainties on any given survey point are small and reflect effects of positioning and baseline processing (~0.1 m, (Meng et al., 2012a)) and the relatively smooth topography along shoreline surfaces (~0.1 - 0.5 m). These uncertainties are smaller, however, than the observed variations in shoreline elevations at a given distance from the lake center (the vertical elevation range of red circles in Figure 4-4). This variation in shoreline elevations reflects what I interpret to be natural variability of shoreline development that reflects site-specific variations in local wave energy, basal topography, sediment supply and duration of activity for shorelines (Gilbert, 1890; Adams and Wesnousky, 1998; Bills et al., 2007). Thus the true, systematic uncertainties, of shoreline elevation at a given position range from 0.5 to perhaps 2 m.
To determine the age of the highstand shoreline complex, I collected 9 samples of medium- and fine-grained sand and silt layers intercalated within beach gravels for OSL chronology (Aitken, 1998), from a variety of locations around the lake (Figure 4-3). Samples were collected from shallow (< 2.5 m) soil pits to ensure that samples represent recent shoreline deposits (see Figure C-1 and methods in Supplemental Materials) and not deeper strata. I took care to avoid aeolian sands that sometimes cover beach deposits. I scraped the wall of the soil pit to remove any material that may have been exposed to light (Aitken, 1998) and collected samples in plastic PVC or steel tubes (3 cm or 5 cm in diameter). Tubes were sealed with black tape to avoid light and preserve moisture content. Analysis of luminescence behavior, dose rate estimation and age calculations were conducted at University of St Andrews (UK), following the protocol of (King et al., 2013).

Samples were analyzed using the single aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000). The equivalent dose (De) was calculated from measurements (n > 50 for each sample) of the luminescence response following stimulation of the natural luminescence and a series of different regenerative doses. Modeling of ages from individual aliquots (Arnold and Roberts, 2009) is described in the Supplemental Materials section, and most of the sample burial ages (Db) are based on a Minimum Age Model (MAM-3). My results reveal that 7 of the 9 samples exhibit ages that cluster between ~ 4 - 6 ka (Table 4-1); two of these samples come from shorelines near the center of the lake (OSL sample numbers 6 and 34 in Figure 4-3) and five samples represent positions along the southeastern and western margin of the lake (OSL-1, 25, 31, 33, 46 in Figure 4-3), confirming my interpretation that the shoreline complex represents features developed during a single highstand. Two samples (OSL-21 and OSL-32) have ages that deviate from this cluster of results. These samples have large overdispersion of individual aliquots and appear to contain multiple components in the distribution of De (see Supplemental Materials). OSL-21 has an age of ~ 3.9 ka based on the younger De component of 3 aliquots using a Finite Mixture Model, which is consistent with the age range of 4 - 6 ka for other samples (Table 4-1; Figure C-2d). However, the older component of the distribution suggests an age of ~7.8 ka. If this component is a reliable burial age, it might suggest that the duration of shoreline development occurred during 8 – 4 ka. The other
sample, OSL-32, has a MAM-3 age of 1.5 ka, significantly younger than the other samples. I observed evidence for bioturbation by rodents at this locality (although not directly in my sample location), and it is possible that this sample may have been influenced by bioturbation. Collectively, the OSL chronology from shoreline complex suggest features were developed during a highstand that occurred from 6 – 4 ka. This is consistent with previous suggestions that high lake levels occurred in central Tibet during the Early Holocene (Rades et al., 2013). Importantly, my results place constraints on the timing of initial recession of the lake (ca. 4 ka), which provides a bound on the timescale of shoreline deflection.

4-4 Rigidity of Tibetan crust from flexural rebound

To determine the effective elastic thickness or flexural rigidity of Tibetan crust, I compare my observed shoreline elevation data with the predicted deflection of an infinite elastic slab overlying an inviscid substrate (Watts, 2001) in response to a spatially varying load that represents the decrease in lake level. I represent the load as an irregular volume consistent with the height and geometry of the highstand shorelines above present lake level, ~ 64 m at Siling Co, ~73 m at Bange Co, ~37 m at Co E, and ~44 m at Wuru Co (see Figure 4-3b-c). I calculate the pattern of shoreline deflection under a range of varying flexural rigidities and compare the results to the observed distributions of elevations (Figure 4-4). Because my data represent minimum bounds on mean water level, I consider the upper envelope of the range of data to be most reliable, as these should be the closest survey points to the paleo-water surface (Bills et al., 1994b).

I assume a uniform elastic thickness (T_e) of the crust beneath Siling Co, and use analytical solutions to the pseudo-3D equation of flexure of a thin elastic spherical shell under a disc-shape load (see details in Supplemental Material) (Brotchie and Silvester, 1969; Watts, 2001). The computation scheme follows the method of (Nakiboglu and Lambeck, 1983). In the scheme, the water load has been discretized to 1-km-diameter cylinders, and the total flexure of the lithosphere is calculated by sum of the flexure in response to loading of each unit cylinder. I account for the geometric differences in the
discretization of the water load by adjusting the original height of the unit cylinder such that the mass of the cylinder is equal to that of a cube (see Supplemental Material).

I calculate the crustal flexural response to lake unloading over a range of flexural rigidities reflecting effective elastic thickness \( (T_e) \) variations from 500 m to 30 km. Because the datum of ‘zero’ deflection is not known \textit{a priori}, I search for a best-fit \( T_e \) that yields the same pattern and magnitude of shoreline deflections as seen in the observed shoreline elevation data. In an ideal case, I expect a one-to-one correlation between the observed and predicted deflections of an initial horizontal datum. As noted previously, my data exhibit variation in shoreline elevation at a given distance from the center of the load up to \( \sim 2 \text{m} \) (Figure 4-4). To account for these, I compare the overall trend of shoreline elevation versus calculated deflection to determine at which elastic thickness the observed and modeled shoreline deflections best show a one-to-one correlation.

Shorelines surveyed around Bange Co and Wuru Co appear to represent separate levels from those around Siling Co (Figure 4-4). Therefore I focus on shorelines developed around the central basin of Siling Co where I are able to robustly correlate the shorelines and I observe a linear correlation of observed elevation and calculated deflection (Figures 4-5 and C-5). When I further consider the depositional settings of each of the sites, I find that shoreline features developed close to mean water elevation such as cuspate bars (filled circles in Figures 4-5 and C-5) exhibit less scatter than do features such as spits, that can develop at somewhat greater water depth (open circles in Figure 4-5).

Comparison of a suite of forward models with varying elastic thickness (Figures 4-5 and C-5) show that deflections are best fit with an effective elastic thickness \( (T_e) \) of \( 13 \pm 1 \text{ km} \). If I consider the upper envelope to the shoreline elevations (Bills et al., 1994b; Bills et al., 2007), the least square regression of data along the upper envelope yield a slope of unity when \( T_e \) is \( 12 – 14 \text{ km} \) (at 95% confidence). It is clear that models with \( T_e \) of less than \( 10 \text{ km} \) systematically overpredict deflections while models with \( T_e \) of \( 15 \text{ km} \) systematically underpredict observed deflections (Figure 4-5). It is important to note that the elastic thickness determined here relies on an implicit assumption that the system has fully
responded to lake withdrawal. If viscoelastic recovery is incomplete, my estimates would overestimate the true flexural rigidity.

I also calculate the bending strain and stress in response to the lake unloading of Siling Co from flexure curves with maximum curvature that corresponds to $T_e$ of 12 – 14 km (Table C-4 in Supplemental Material). Calculation of strain and stress is based on Equations 3.70 and 3.64, respectively, of Turcotte and Schubert (2012). The results show the maximum strain on the order of $10^{-5}$ and the bending stresses of $\sim 3$ MPa. These provide an estimate of the strain rate of $\sim 10^{16}$ s$^{-1}$ over the timescale of lake recession ($\sim 4$ ka).

### 4-5 Strength profile and viscosity of the lower crust

I use above results, especially the effective elastic thickness of $\sim 12 – 14$ km and the strain rate ($\sim 10^{16}$ s$^{-1}$) over the timescale of lake unloading ($\sim 4$ ka) to place constraints on the viscosity of the deep crust beneath central Tibet, from two approaches. First, I use the effective elastic thickness to estimate the depth of compensation within the crust and assume that the observed flexural rebound of the elastic layer represents full relaxation by viscous flow in the middle crust. In this case, I can estimate the maximum viscosity that would allow complete compensation of the removal of the lake load in the available time using a simple approximation (Turcotte and Schubert, 2002).

\[
\eta = \frac{\rho g \lambda}{4\pi \tau_r}, \quad (1)
\]

where $\tau_r$ is the characteristic relaxation time that is approximately 1/3 of the total relaxation time ($\tau$). Here $\tau$ is assumed to be $\sim 5$ ka. $\eta$ is the viscosity of the compensating medium, $\rho$ is the density of the compensating fluid (here assumed to be 2900 kg/m$^3$), $g$ is gravitational acceleration, and $\lambda$ is the flexural wavelength ($\sim 150 – 200$ km) determined from the best estimate of the flexural rigidity. This simple analysis implies that the average viscosity of the middle crust beneath Siling Co is less than $\sim (1 – 3) \times 10^{19}$ Pa s.
Second, I develop a new approach that uses my results and other geophysical data from central Tibet to better constrain the viscosity structure of the middle crust in this region. I determine (internally consistent) strength profiles for the crust that reconcile the effective elastic thickness (12 – 14 km) and strain rate (~ 10^{-16} \text{s}^{-1}) determined in this study with geophysical constraints on crustal structure, composition and heat flow (see details in Supplemental Materials). The relatively low flexural rigidity determined in my analysis places the elastic layer in the upper crust (triangles in Figure 4-6b). I approximate the top bound of the elastic layer at the base of sedimentary basins in central Tibet, ~ 1 – 5 km at depth (Ross et al., 2004; Mechie et al., 2011). Given the allowable range of elastic thickness (12 - 14 km), the base of the elastic layer is estimated to occur between ~13 - 19 km (Figure 4-6b). I estimate the ductile strength of the crust from a range of constitutive flow laws (see details in Supplemental Materials) that account for: 1) the strain rate (~ 10^{-16} \text{s}^{-1}); 2) reasonable assumptions on bulk crustal compositions (Mechie et al., 2011); and 3) a depth dependent temperature profile that honors the crustal composition (Figure 4-6c), the relatively high surface heat flow in the region (80 – 110 mW/m², (Francheteau et al., 1984)), and rare geophysical constraints on crustal temperature at specific depths (Hacker et al., 2000; Mechie et al., 2004).

My analysis of strength profiles suggests that effective viscosities decrease toward the middle-lower crust (Figure 4-6d). The magnitude of the viscosity decrease, of course, depends on the details of the composition and the appropriate constitutive flow law. Given evidence for a granitic composition in the upper crust beneath central Tibet (Haines and van der Pluijm, 2008), I consider the viscosity profile derived from flow laws for wet granite (Hansen and Carter, 1982) or partially molten granite (Rutter et al., 2006) are likely appropriate estimates. In this case, the magnitude of viscosity drop in the middle crust can be as large as two orders of magnitude from 10^{20} \text{Pa s} at 20 km depth to 10^{18} \text{Pa s} at 40 km depth (Figure 4-6d); at a depth of 30 km, the viscosity is estimated to be 10^{19} – 10^{20} \text{Pa s}, which is compatible with my preliminary estimate of viscosity (1 - 3 \times 10^{19} \text{Pa s}) based on the relaxation timescale of lake unloading. Viscosity is expected to decrease even further with depth; by 40 km depth, at the top of the transition to more mafic material, viscosities may be as low as 5 \times 10^{18} \text{Pa s} (Figure 4-6d). This decrease
of viscosity to below $10^{19}$ Pa s between 35 and 40 km, and the sharp transition to a more mafic composition may enable flow within a crustal channel beneath central Tibet (Beaumont et al., 2004).

Viscosities are estimated to be several orders of magnitude greater both above and below this low-viscosity region (Figure 4-6). This effect is dictated by the combination of high heat flow and the relatively sharp transition between a granitic upper crust and a more mafic lower crust (Mechie et al., 2004). The upper limit is dictated by the thermal conditions in the middle crust; my estimates suggest effective viscosities of $10^{20} - 10^{21}$ Pa s at ~ 20 km depth. At depths > 40 km, higher effective viscosities (> $10^{20}$ Pa s) reflect the influence of quartz-poor compositions (Figure 4-6). Although I extrapolate a viscosity profile through the lower crust, it is important to note that my strength profiles are not sensitive to the rheology of the lowest crust and upper mantle. The scale of the surface load and the apparent presence of a weak middle/lower crust conspire such that most of the compensation of flexural deformation would occur within the middle and lower crust.

4-6 Discussion

The fact that the elastic thickness robustly-determined in this study, $T_e = 13 \pm 1$ km, is significantly less than the total crustal thickness ($65 \pm 5$ km, (Zhao et al., 2001; Nábělek et al., 2009)) implies that the crustal strength is substantially reduced. My results are consistent with numerous observations that focal depths of large earthquakes in Tibet are largely confined to the upper ~ 20 km of the crust (Chu et al., 2009). Although rigid Indian lithosphere may contribute to the overall strength profile beneath southern Tibet (Copley et al., 2011), this effect would likely be confined to the plateau south of the Bangong-Nujiang suture (Nábělek et al., 2009). Such thin elastic thickness determined from this study may reflect the decoupling in the middle crust (Burov and Diament, 1995), therefore my study does not provide the strength of the lower crust.

My study provides the first direct constraint on viscosity structure of Tibetan crust over millennial timescale that is compatible with the elastic thickness and other geophysical data of central Tibet.
Previous estimates of viscosity mostly are determined from deformation on seismic timescale (see Tables C-11, 12 and references therein). Thus, my results provide perhaps a closer approximation to viscosities that would characterize crustal flow over millennial timescales (Watts, 2001; Thatcher and Pollitz, 2008; Watts et al., 2013). Although the viscosity profile derived here has a range of allowable values, the viscosity of $10^{20-19}$ Pa s from 20 – 40 km depth and a viscosity drop to < $10^{19}$ Pa s at 35 km depth beneath central Tibet confirm the long-held notion that, to first order, the middle crust beneath the Tibetan Plateau is quite weak (Bird, 1991; Nelson et al., 1996).

My results contrast significantly to the lower bound on the average viscosity (> $10^{19}$ - $10^{20}$ Pa s) of the ductile crust in western-central Tibet, based on apparent absence of shoreline deflection around Zhari Nam Co (Figure 1-1) (England et al., 2013). However, the interpretation of England et al. (2013) may be biased by several factors, including 1) the relatively small size of the load, 2) the simple cylindrical geometry of the basin, such that preserved shorelines are approximately equidistant from the center of the load (England et al., 2013), 3) absence of islands in the lake center to preserve shorelines for examining possible larger magnitude of shoreline deflection that can occur in this region, 4) the lack of high-precision surveys of differences in shoreline elevation which may lead to inconsistent shoreline correlation, and 5) no age constraint on the shorelines for spatial correlation and determination of the deformation timescale. Therefore, our results provide much more reliable constraints on the crustal viscosity of Tibet over millennial timescale.

My results lend support to the proposition that the middle and lower crust beneath portions of the Tibetan Plateau are capable of flow on geological timescales. A number of previous modeling efforts aimed at explaining the present-day topography of the plateau argue for a weak lower/middle crust (Royden et al., 1997; Clark and Royden, 2000; Clark et al., 2005; Copley and McKenzie, 2007; Bendick et al., 2008; Cook and Royden, 2008), with effective viscosities ranging from $10^{16}$ – $10^{22}$ Pa s (also see Table C-13). Thermal-mechanical numerical models parameterized to account for constitutive flow laws (Beaumont et al., 2001) imply that extensive flow in a lower crustal channel may well develop once the effective viscosity reaches values of ~ $10^{19}$ Pa s (Beaumont et al., 2004; Medvedev and Beaumont, 2006).
Although tradeoffs between the viscosity and the thickness of low-viscosity crust make this value only an approximation (Klemperer, 2006; Bendick et al., 2008), a number of studies suggest that this may be an important, quasi-threshold at which widespread flow is likely (Beaumont et al., 2004; Hilley et al., 2005; Huang et al., 2014). In my study, an effective viscosity of $10^{18} - 10^{19}$ Pa s at 35 – 40 km suggests the presence of a layer that is significantly weaker than the upper crust ($\text{viscosity} > 10^{20} - 10^{23}$ Pa s, Figure 4-6d). Thus, my results imply that the middle crust beneath central Tibet is likely capable of channelized flow, but that viscosities at the lower end of suggested values ($\sim 10^{16}$ Pa s (Clark and Royden, 2000) appear unlikely.

The collection of recent studies of Tibetan rheology appears to point to significant spatial heterogeneity in the crustal strength. Although many viscosities values have been obtained from postseismic/interseismic deformation after large earthquakes in portions of Tibet (see Tables C-11, 12 and references therein), here I only compare my estimate of the viscosity over millennial timescale to those determined from relatively longer time records ($> \sim 2$ years) of postseismic/interseismic deformation. The viscosities obtained from shorter records ($< \sim 2$ years) (Table C-12) usually appear much lower than those from the longer time records (Table C-11), likely reflecting the transient, but not steady state crustal rheology (Freed et al., 2012; Wen et al., 2012; Huang et al., 2014). In northern Tibet, viscosities of the lower crust were determined to be $\sim 10^{19} - 10^{20}$ Pa s from analysis of $> 2$-year postseismic or interseismic deformation after the 1997 Mw 7.6 Manyi earthquake (Ryder et al., 2007; Yamasaki and Houseman, 2012) and the 2001 Mw 7.8 Kokoxili earthquake (Hilley et al., 2005; Ryder et al., 2011; Wen et al., 2012). In eastern Tibet, postseismic deformation after the 2008 Mw 7.9 Wenchuan earthquake (4 year record) (Huang et al., 2014) and after the 1973 Ms 7.6 Luhuo earthquake ($\sim 7$-20 year record) (Zhang et al., 2009) provide the lower crustal viscosity of $\sim 10^{18} - 10^{19}$ Pa s. Therefore, in general, northern Tibet show a larger crustal strength than eastern Tibet on seismic timescale. In central Tibet, there are no records of postseismic deformation longer than 2 years. My estimate on middle crustal viscosity of central Tibet, $\sim 10^{20}-10^{18}$ Pa s from 20 – 40 km over millennial timescale, differs from both northern and eastern Tibet (Figure 4-7). Whether these differences reflect the temporal evolution (Watts, 2001; Thatcher and
Pollitz, 2008; Watts et al., 2013) or spatial variations in the crustal strength, remains uncertain. But if the latter case is true, then my results suggest that central Tibet may be weaker than northern Tibet, but a little stronger than eastern Tibet.

4-7 Conclusions

My study of deflection of a Holocene shoreline complex around Siling Co provides some of the first quantitative estimates of the rheology of central Tibetan crust and its depth-dependence over millennial timescales. My results lead us to the following conclusions.

1) Siling Co reached highstand conditions during the early to middle Holocene (4-6 ka). Lake levels fell by ~ 60 m in the past 4 ka.

2) Shorelines developed during the highstand have been deflected systematically by up to ~5 m.

3) Forward modeling of rebound in response to a spatially-distributed load constrains effective elastic thickness to 13 ± 1 km in central Tibet.

4) Rationalization of existing constraints on crustal structure, composition, and heat flow with the strain rates implied by shoreline deflection implies an abrupt decrease in viscosity from > $10^{20}$ Pa s in the upper crust to relatively low viscosity ($10^{18}$ Pa s) by 40 km depth.

Thus, my results imply that the middle/lower crust beneath central Tibet appears capable of widespread, and likely channelized, flow on geologic timescales.
4-8 References


Figure 4-1. Geological setting of the study area. Shown are two seismic profiles around Siling Co. SF-2: Sino-French 2; INDEPTH-3 (International Deep Profiling of Tibet and the Himalaya. BNS: Bangong-Nujiang suture; TB: Tarim basin; QB: Qaidam basin; SB: Sichuan basin.
Figure 4-2. Geomorphic features of shorelines around Siling Co. a) high resolution GeoEye satellite imagery showing the highstand depositional shorelines which are characterized by their continuation with highstand wave-cut scarps that dissect alluvial gullies above the highstand and lower continuous shorelines; b) the photo showing highstand cuspate bar and the back-bar depression behind it; c) the surface of the cuspate bar.
Figure 4-3 (previous page). (a) Localities of surveyed depositional highstand shorelines and OSL samples. The blue area represents the extent of the paleo-highstand of Siling Co and the cyan area is the present lake area. The red circles show localities of highstand shorelines that are well-correlated around the main body of paleo-Siling Co, with filled red circles for shoreline features formed at or immediately above the still water level and open red circle for those deposited likely below still water level; filled white and orange circles display locations of highstand shorelines around Wuru Co and Bange Co, respectively, where the correlation is uncertain. Sample numbers are labeled off the sample locations pointed by black arrows. (b) Reconstructed water load between the paleo-highstand and the lake level in 1976 along profile ABC. (c) Map view of the 3D geometry of the reconstructed water load; (d) Observed elevation difference of the highstand shorelines in southern peninsula of Siling Co and Co along profile D-E. This figure shows up to ~5 m of deflection over distance of 30 km. Also shown are the uncertainty of the elevations on this highstand that is mainly induced by natural processes forming the shoreline deposits (see text for details).
Figure 4-4. Comparison of observed shoreline elevations and model-predicted shoreline deflections for all surveyed constructional shorelines around Siling Co, Co E, Wuru Co and Bange Co. It is clear that the shorelines around the main body of paleo-Siling Co are within a narrow trend of slope equal to 1. However, the shorelines below the highstand (dark green) and around other lakes (e.g., white for Wuru Co and orange for Bange Co) that are currently isolated from Siling Co lose the pattern, which may be ascribed to discontinuous geomorphic correlation or the shorelines formed at different levels.
Figure 4-5 (previous page). (a-e) Predicted crustal deflection pattern with elastic thickness of 7, 10, 13, and 15 km (from top to bottom) by forward elastic modeling. (f-j) Comparison of deflections ($T_e = 7, 10, 13$ and 15 km in sequence) at the highstand shoreline localities with observed elevations at those locations. Filled red circles display the highstand shorelines that most likely form at or immediately close to the water level (e.g., cuspatate bars and beach ridges) and open red circles represent those shorelines very likely form below the water level. Shaded rectangles show the slope trend of the data points.
Figure 4-6. (a) Strength envelopes of central Tibetan lithosphere constructed based on elastic core thickness ($T_e = 12-14$ km) and strain rate (this study), and assumed crustal compositions and thermal parameters (previous studies); (b) part of the strength envelope showing the elastic core of central Tibetan crust. The upper and lower gray region show the top (at depth of $~1-5$ km) and bottom (at depth of $~13-19$ km) of the elastic core; (c) temperature profile compatible with $T_e$ and strain rate (this study) and crustal structure and compositions (previous studies); (d) viscosity profiles of central Tibetan lithosphere. The red, blue and magenta lines represent strength envelopes in the ductile part of the upper crust, lower crust and upper mantle, respectively. Different line styles denote different rock compositions (see details in Supplemental Materials); TZEC: top zone of elastic core; BZEC: bottom zone of elastic core. Experimental rock rheology: 1. wet Heavitree quartzite (Jaoul et al., 1984); 2. wet granite (Hansen and Carter, 1982); 3. partially molten granite (Rutter et al., 2006); 4. wet Simpson quartzite (Koch et al., 1989); 5. wet Black hills quartzite (Gleason and Tullis, 1995); 6. dry Maryland diabase (Mackwell et al., 1998); 7. dry Columbia diabase (Mackwell et al., 1998); 8. dry anorthite (Rybacki and Dresen, 2000); 9. undried Pikwitonei granulite (Wilks and Carter, 1990); and 10. dry dunite (Hirth and Kohlstedt, 1996). Thermal parameters for this example temperature profile: depth scale of heat production ($b$): 35 km; thermal conductivity ($k$): 2.5 W/m/K; surface heat production ($A_0$): 2.5 uW/m$^3$; surface heat flow ($q_0$): 90 mW/m$^2$; strain rate ($\dot{\varepsilon}$): $1.18 \times 10^{-16}$ s$^{-1}$. 
Figure 4-7. Comparison of viscosities for different regions and variable timescales determined from postseismic deformation after large earthquakes (labeled at bottom) and lacustrine shoreline deformation. Open and grey bars show viscosities determined from ≤ 2-year, and > 2-year record, respectively, of postseismic deformation. Orange bars denote millennial timescale viscosity range. Depth-dependent viscosities are shown as hatched patterns, with the depth range labeled off the bars. Heavy lines indicate the viscosity at 40 km depth. Point-up or -down black arrows indicate viscosities with a lower or upper bound, respectively. Green-shaded boxes show possible viscosity ranges that may represent steady state rheology for different regions in the Tibetan Plateau. References: B-14: Bie et al., 2014; D-13: DeVries and Meade, 2013; E-13: England et al., 2013; H-05: Hilley et al., 2015; H-09: Hilley et al., 2009; H-14: Huang et al., 2014; R-07: Ryder et al., 2007; R-10: Ryder et al., 2010; R-11: Ryder et al., 2011; S-11: Shao et al., 2011; W-12: Wen et al., 2012; Y-12: Yamasaki and Houseman et al., 2012; Z-09: Zhang et al., 2009.
Table 4-1. Field data and ages of OSL samples from the highstand shorelines around Siling Co

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Lat (°N)</th>
<th>Long (°E)</th>
<th>Elev (m)</th>
<th>Depth (m)</th>
<th>N ( aliquots)</th>
<th>De (Gy)</th>
<th>Error (Gy)</th>
<th>Dose Rate (Gy/ka)</th>
<th>Error (Gy/ka)</th>
<th>Age (ka)</th>
<th>Error (ka)</th>
<th>Age Model*</th>
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<tr>
<td>XS-SL-OSL-O1A</td>
<td>31.524</td>
<td>89.216</td>
<td>4594</td>
<td>2.2</td>
<td>52</td>
<td>14.98</td>
<td>0.90</td>
<td>2.37</td>
<td>0.09</td>
<td>6.3</td>
<td>0.5</td>
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<td>89.216</td>
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<td>52</td>
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<tr>
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<td>MAM-3</td>
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* FMM – Finite mixture model; MAM – Minimum age model; numbers denote the component of each age model
CHAPTER 5

HOLOCENE SLIP RATE ALONG THE GYARING CO FAULT, CENTRAL TIBET

Abstract

The degree to which active deformation within the Tibetan Plateau is localized along major strike-slip faults versus distributed throughout the crust has been debated for decades. Within the interior of the high plateau, a network of strike-slip faults act to transfer displacement among north-south trending graben systems. Although geodetic measurements of interseismic deformation along most of these fault systems suggest relatively slow strain accumulation, active slip along the right-lateral Gyaring Co fault has been proposed to range from 10 – 18 mm/yr (Taylor and Peltzer, 2006), but reliable geologic constraints on rate of slip along this fault are sparse. Here I determine the slip rates of the Gyaring Co fault in central Tibet using flights of displaced lacustrine shorelines along the eastern shore of Zigui Co, a small lake developed along the fault zone. Two arcuate beach ridge complexes are displaced by 12 ± 1 m in a right-lateral sense across the singular fault trace. Optically-stimulated luminescence dating of well-sorted beach sands intercalated with beach gravels yields tightly clustered ages between 4.1 and 4.4 ka. These data imply an average slip rate of 2.1 – 3.2 mm/yr along the central Gyaring Co fault during the latter half of the Holocene. Although I cannot rule out the possibility that strain accumulation and release along the Gyaring Co fault are temporally variable, the relatively low slip rate determined here supports the proposition that active deformation in the plateau interior is distributed among numerous, slowly moving faults.
5-1 Introduction

Intracontinental deformation within the Himalaya-Tibetan orogen presently accommodates ~ 40 mm/yr of north-south convergence between India and Eurasia (Paul et al., 2001; Wang et al., 2001). Approximately half of this convergence is accommodated by shortening across the Himalaya (Lavé and Avouac, 2000; Bettinelli et al., 2006; Ader et al., 2012), and the remainder is distributed throughout the adjacent Eurasian lithosphere. Whether active deformation is primarily localized along major strike-slip fault systems (Figure 5-1) (Peltzer and Tapponnier, 1988; Avouac and Tapponnier, 1993) or occurs as distributed strain within the interior of the plateau (England and Houseman, 1986) remains a point of debate. The former hypothesis implies that rapid slip along the major strike-slip faults absorbs a large fraction of the total convergence (e.g., Mériaux et al., 2004; Chevalier et al., 2005), whereas the latter hypothesis implies relatively low displacement rates (e.g., Brown et al. 2002; Cowgill et al., 2009) along the numerous active faults throughout Tibet.

This fundamental difference in deformational behavior in Tibet has received extensive attention in recent years, particularly along structures developed at the margin of the plateau. A preponderance of evidence now suggests that many of these faults have slip rates on the order of ~1 cm/yr, when averaged over the Late Pleistocene – Holocene timescales. Along the Altyn Tagh fault, slip rates determined at nearly a dozen different sites indicate average rates of ~ 10 mm/yr (Figure 5-1) (Washburn et al., 2001; Cowgill et al., 2007; 2009; Zhang et al., 2007; Gold et al., 2009; Chen et al., 2012; 2013). Studies along the central Kunlun fault (Van der Woerd et al., 1998; 2000; 2002; Li et al., 2005) indicate slip rates of ~ 8 – 10 mm/yr, and the Karakorum fault appears to have slip rates between ~ 5 – 8 mm/yr (Brown et al., 2002; 2005; Chevalier et al., 2005a; 2005b; 2011; 2012). These modest slip rates determined from offset geomorphic markers are similar to those inferred from space geodesy (e.g., Wright et al., 2004; Bendick et al., 2000; Meade et al., 2007; Thatcher, 2007; Elliott et al., 2008; Loveless and Meade, 2011; He et al., 2013). Although it has been suggested that some faults may exhibit temporal variations in slip rate (e.g., Chevalier et al., 2004), most of the data appear to be consistent with a relatively constant slip rate through
time (e.g., Gold and Cowgill, 2009; He et al., 2013). Spatial variations of slip rates along some of these major structures (Kirby et al., 2007; Zhang et al., 2007; Harkins et al., 2010) are associated with distributed deformation throughout the plateau (Kirby and Harkins, 2013). Collectively, these studies make a strong case that the degree of strain localization along major strike-slip faults largely reflects heterogeneity in the strength of Tibetan crust/lithosphere (e.g., England and Houseman, 1985; Molnar and Dayem, 2010).

Within the interior of the Tibetan Plateau, geodetic and geologic observations suggest that active deformation is accomplished along a network of conjugate strike-slip fault systems (Taylor et al., 2003) and associated N-S-trending rift systems (Figure 5-1). Geodetic data suggest that rates of E-W extension are quite low, ranging from < 1 to 4 mm/yr (Chen et al., 2004a; Gan et al., 2007; Elliott et al., 2010; Liang et al., 2013). This is consistent with relatively slow slip rates determined along graben-bounding normal faults (Armijo et al., 1986; Molnar and Lyon-Caen, 1989; Yin et al, 1999; Blisniuk and Sharp, 2003). Likewise, rates of slip along NW-SE and NE-SW striking faults inferred from geodetic data also appear to be modest: 1) InSAR results indicate left-lateral slip rates < ~ 6 mm/yr of the Riganpei Co fault (Taylor and Peltzer, 2006) and, 2) similar observations across the Beng Co and Dongqiao faults suggest slip rates of 1-3 mm/yr (Garthwaite et al., 2013). Although early workers suggested rapid slip along a through-going fault system (the “Karakorum-Jiali fault zone” – Armijo et al., 1986; 1989), the present day strain field appears to be characterized by slow N-S shortening and E-W extension accommodated along a distributed array of conjugate strike-slip fault systems (Taylor et al., 2003; Chen et al., 2004b).

An exception may occur along the dextral Gyaring Co fault (GCF, Figure 5-1), where slip rates are thought to be high (Taylor and Peltzer, 2006; Armijo et al., 1989). The GCF represents one of the primary conjugate faults in the region (Taylor et al., 2003) and Armijo et al. (1989) conclude it shows Holocene slip rates > ~ 20 mm/yr on the basis of fresh scarps and displaced geomorphic features. Although these geomorphic features were not directly dated, analysis of stacked InSAR data appears to confirm rapid strain accumulation across the fault system, with slip rates ranging from ~10 – 18 mm/yr (Taylor and Peltzer, 2006). Because there are no direct constraints on the average geologic slip rate of the
GCF, the significance of these high apparent rates remains uncertain. Here I place quantitative bounds on the Holocene slip rate of the GCF by reconstructing displaced lacustrine shorelines around Zigui Co, a lake immediately northwest of Gyaring Co (‘co’ is the Tibetan word for lake, Figure 5-1).

5-2 Background

The GCF (Figure 5-1a) is a WNW-striking fault that extends along the northern shore of Gyaring Co (Figure 5-1b) and forms the northern boundary of several major N-S striking grabens (Figure 5-1). The fault extends ~ 250 km and consists of several segments exposed between major lakes that cover the fault trace (Figure 5-1b). The fault trace is prominent and marked by fresh scarps that displace alluvial fans and fan terraces, channels, and lacustrine shorelines in a right-lateral sense (Armijo et al., 1989; Taylor and Peltzer, 2006).

The GCF has long been considered a part of a proposed regional shear zone (the ‘Karakorum-Jiali fault zone’, Armijo et al., 1989) that is suggested to act in concert with the left-lateral Altyn Tagh and Kunlun faults to accommodate eastward extrusion of northern Tibet from in front of the Indian indenter (Armijo et al., 1989). However, other studies consider the GCF one of a network of conjugate fault systems in central Tibet (Taylor et al., 2003; Taylor and Peltzer, 2006). The fault appears to merge toward the northwest with the NE-striking, left-lateral Riganpei Co fault and, toward the southeast, the GCF appears to be kinematically linked with the Shenza rift system (Figure 5-1).

Recent analysis of stacked InSAR interferograms across the northern segment of this fault suggests allowable slip rates that range from ~ 10 – 18 mm/yr (Taylor and Peltzer, 2006), similar to the inference of > 20 mm/yr from displaced geomorphic features (Armijo et al., 1989). Moreover, these authors argued that rapid slip along the fault is consistent with the occurrence of an M ~ 7 earthquake that occurred in 1934 in the general vicinity of the Shenzha rift (Armijo et al., 1989). It is not clear, however, whether velocities observed in the geodetic data (Taylor and Peltzer, 2006) could be associated with long-wavelength artifacts in the satellite data, post-seismic transients (e.g., Ryder et al., 2010), or perhaps even...
surface deformation associated with changes in lake and glacier loads. Likewise, the location of the 1934 event is still debated; whether this event occurred along the southeastern branch of the GCF (Armijo et al., 1989) or was perhaps related to normal faulting in the Shenzha graben (Wu et al., 1990) is uncertain.

5-3 Holocene Slip Rate along the Gyaring Co Fault

5-3.1 Zigui Co Fault Site

My study site is located along the central segment of the fault system, between Gyaring Co and Zigui Co (Figure 5-1b). Approximately 5 km east of the present-day shoreline of Zigui Co, the active fault trace is marked by two primary strands that comprise a right step in the fault system (Figure 5-2). The nature of the scarps, and the rhomboid topographic depression formed between the fault strands, confirm this as an extensional step, compatible with right-lateral displacement along the fault (Figure 5-2). The northern margin of the pull-apart is marked by a prominent south-facing scarp that separates dissected alluvial fans from the basin (Figure 5-2). Near the western end of the pull-apart, this fault strand displaces several younger generations of alluvial fans and inset fan-terrace surfaces (Figure 5-2). Toward the west, the northern fault strand appears to carry most of the displacement; near the margin of Zigui Co, scarps along the southern fault strand are small and discontinuous (Figure 5-2).

I observed two groups of relict shoreline features that I interpret to mark former, higher levels of Zigui Co (Figure 5-2). The highest of these features are wave-cut scarps developed discontinuously along the alluvial fans south of the GCF. These wave-cut scarps trend WNW, along the slope of the alluvial apron (G4 in Figure 5-2) and are substantially degraded. The second, lower set of shorelines are marked by prominent beach ridges and spits, indicative of deposition of beach and shoreface deposits during a high lake stage. Two primary ridges form an arcuate trace that trends N-S along the eastern margin of the lake (G3 in Figure 5-2). In contrast to the higher shorelines, these features are fresh, undissected, and act as barriers to modern drainage; fine-grained sand and silt is ponded to the east of the beach ridges (Figure
To the west of these shoreline groups are lower shorelines and beach ridges associated with the modern level of Zigui Co (G2 and G1 in Figure 5-2).

I focus my study on the G3 shoreline group for two reasons. First, this group forms an extensive linear marker that intersects the GCF at a high angle (Figures 5-2 and 5-3). South of the fault, the beach ridges are remarkably straight, extending south from the scarp for several hundred meters. North of the fault, the beach ridges again extend for several hundred meters and then merge into a wave-cut scarp developed in older alluvium (Figure 5-3). The linear character of the ridges, absent embayments and cusps, and their near orthogonal orientation with respect to the fault trace make these ideal markers to reconstruct displacement along the GCF. Second, because the G3 shorelines represent depositional features, the age of these shorelines is not subject any ambiguity sometimes associated with fluvial terrace risers (e.g., Cowgill et al., 2007); the age of deposition of the shoreline places a direct constraint on the average rate of slip since then.

The beach ridges exhibit cross-sectional topography typical of constructional shorelines: a main ridge with a flat top (~1 – 2 m wide), a lake-ward front scarp and a subtle depression (or swale) on the land-ward side (light-colored region in Figure 5-3a). Pits excavated in the beach ridges reveal distinct stratigraphy: layers of rounded, well-sorted gravels, intercalated with thin sandy layers (Figure D-1). In contrast, the back-ridge depressions (Figure 5-3a) are filled with eolian sands interbedded with alluvial sand and gravel. In both depositional settings, I determine the ages of the shorelines using optically-stimulated luminescence (OSL) dating.

5-3.2 Reconstructing Fault Slip

Using GeoEye satellite images (nominal resolution ~ 0.5 m), I project the center lines (Figure 5-3d) of the ridge flats (~1 – 2 m wide) (Figure 5-3e) into the fault trace. I also surveyed the position of these intersections along the fault with a measuring tape in the field, and I consider the uncertainty on this projection to be ~1 m. Both beach ridges exhibit a ~ 30° clockwise bend in their trace as they approach
the fault (Figure 5-3c, 3d). Because the shorelines are nearly linear for > 500 m north and south of the fault trace, and because those to the north of the fault are consistently displaced to the east relative to the southern shorelines, I argue that this bend must reflect near-fault deformation. I do not believe that it was an embayment in the shoreline. Moreover, I note that the sense of bending is consistent with “drag” along a right-lateral fault.

Projecting the linear crest of the beach ridges yields estimates of displacement along the GCF of 11.5 m along shoreline 1 (S1) and 11.9 m along shoreline 2 (S2). Given that the strike separation was measured manually by tape and that width of the ridge crest line is ~ 1 m, I assign an uncertainty of 1 m for all measurements. Thus, my best estimate of lateral slip along the GCF is ~ 12 ± 1 m. In addition to the strike-slip component, the GCF also exhibits a vertical separation at this site, with the southern side of the fault 0.5 – 1 m higher than the northern side (Figure 5-3e).

Importantly, I do not find evidence for displacement of the youngest alluvial and lacustrine features along this segment of the GCF. Although the geometry of G2 shorelines is suggestive that they have been cut by the fault (Figure 5-2), the fault trace is buried by recent alluvium, and the curvilinear traces of the G2 shoreline features (Figure 5-2) make reconstruction uncertain. It is clear, however, that the G1 shoreline, immediately adjacent to the modern lake margin is not displaced across the fault (Figure 5-2). Thus, these features provide a lower bound for the most recent surface-rupturing event along the fault. Similarly, I observe scarps developed in relatively young alluvial terraces developed across the fault trace bounding the northwestern margin of the pull-apart basin (Figures 5-2 and D-2). The youngest of these terraces is not cut by the fault (Figure D-2). Unfortunately, I were unable to obtain appropriate material for dating of either the G1 shorelines or this youngest alluvial surface. I can say, however, that the absence of scarps within these deposits implies that the 1934 M ~ 7 Shenzha earthquake did not contribute to displacement of the G3 shorelines. Because the epicenter of the 1934 event is located ~ 70 km southeast of my site, this may reflect termination of the rupture father east along the GCF.

Alternatively, it may be consistent with the suggestion that the 1934 event occurred within the Shenzha
graben (Wu et al., 1990). Regardless, these data suggest that my slip estimate is not likely influenced by a relatively recent, historic or pre-historic, earthquake.

5-3.3 Shoreline Chronology and Fault Slip Rate

Three samples were collected: two from the sandy layers intercalated with the beach gravel layers within pits (Figure D-1) in shorelines S1 and S2 (GR1 and GR2 in Figure 5-3b) and another from the eolian sand deposits in the depression east of shoreline S1 (GR3 in Figure 5-3a). Samples were prepared using standard OSL procedures and analyzed using the single aliquot regenerative dose (SAR) protocol and small aliquots (see methods in the Appendix D) (Murray & Wintle, 2000). The environmental dose rates (Dr) were calculated using the concentrations of U, Th, K and Rb were measured directly using solution ICP-MS (Thermo X-Series), a cosmic-dose component after Prescott and Hutton (1994) and an internal alpha dose rate of 5% from the decay of U and Th after Sutton and Zimmerman (1978).

All three samples are characterized by large overdispersion values (broad distributions of equivalent dose, see Figures 5-4 and D-3) (Galbraith and Roberts, 2012) of 21 – 23%; samples with overdispersion greater than 20%, are assumed to reflect heterogeneous bleaching before deposition. An age model was selected for each sample following the criteria of Arnold and Roberts (2009). Two of the samples (GR-2 and GR-3) are modelled using the three component minimum age model (MAM-3) and one sample (GR-1) is based on a central age model (CAM) (Galbraith et al., 1999) using the RStudio Luminescence package (Kreutzer et al., 2012).

Collectively, these results suggest that the shorelines S1 and S2 were deposited at 4.1 ± 0.1 ka and 4.1 ± 0.3 ka, respectively (Table D-2). Deposition of the eolian sands appears to have been synchronous with the development of the beach ridges at 4.4 ± 0.3 ka (Table D-2). These tightly clustered age constraints yield estimates of the average slip rate during the latter part of the Holocene that range from 2.1 – 3.2 mm/yr. Thus, my results require relatively modest slip rates along the GCF during the Holocene time.
5–4 Discussions and Implications

My findings show a large discrepancy between the Holocene slip rate and geodetically determined slip rate along the GCF (e.g., Taylor and Peltzer, 2006). It is increasingly recognized that secular variation in fault slip (e.g., Wallace, 1987), potentially associated with earthquake clusters (e.g., Rockwell et al., 2000), may influence the time-averaged rate of slip along intracontinental faults (Chevalier et al., 2005; Rittase et al., 2014). To evaluate the possible contributions of earthquake histories to my estimate of slip rate, I consider two alternative scenarios. First, I consider it unlikely that the entire 11–12 meters of displacement accrued in a single earthquake. Maximum coseismic slip during large strike-slip events - the Mw 7.6 Manyi earthquake in 1997 (~ 7 m, Peltzer et al., 1999), the Mw 7.9 Kokoxili earthquake in 2001 (~ 7–8 m, Xu et al., 2006), the Mx 7.9 Fuyun earthquake in 1931 (~ 5–7 m; Klinger et al., 2011), and even the Mw 7.9 Fort Tejon earthquake of 1857 along the San Andreas fault (~ 5–6 m; Zielke et al., 2010) - were all on the order of ~ 6 meters. Thus, it seems reasonable that the total displacement of ~ 11–13 m at the Zigui Co site reflects at least two events and perhaps more (if characteristic earthquakes had magnitudes Mw of 6–7). Thus, my slip rates probably represent a reasonable average over at least two, and perhaps more, seismic cycles.

Although I do not have direct constraints on the age of the youngest shoreline (G1), the absence of scarps cutting these features forces us to consider the maximum slip rate allowable by my data. If I presume that displacement along this segment of the fault accrued during two events with slip of ~ 6 m, then it is possible that a future event of similar size would lead to a total displacement of ~ 18 m. Thus, even if the fault is late in the earthquake cycle, my data would imply that the average slip rate during the latter half of the Holocene could be no greater than 3.6–4.5 mm/yr. I argue that the large differences between my determination of fault slip rate and the geodetic data is not an artifact of a limited rupture history at my site. If the geodetic data are a complete reflection of strain accumulation along the GCF, it would appear that this fault is being loaded at a far greater rate than strain has been released. My results
highlight the need for additional constraints on slip rate farther back in time to evaluate the potential for secular variations in displacement.

From a regional perspective, my slip rate of 2.1 – 3.2 mm/yr along the GCF during the Holocene is consistent with slip rates determined along several other faults in central-western Tibet. Geodetic measurements of interseismic strain determined from InSAR suggest that many of the conjugate strike slip faults are moving at modest rates (Figure 5-1): (1) the Lamu Co fault in western Tibet (~ 2 – 4 mm/yr; Taylor and Peltzer, 2006); (2) the Riganpei Co fault (~ 3 – 11 mm/yr; Taylor and Peltzer, 2006); and (3) the Beng Co fault (~ 1 – 4 mm/yr; Garthwaite et al., 2013). Although I do not directly address the question of whether the Karakorum – Jiali fault zone is a continuous structural feature during the Cenozoic (Armijo et al., 1989), my data require that displacement rates along this system are modest, at least at present (e.g., Chung et al., 2008). Collectively, my results add to the growing perspective that active deformation within the interior of the Tibetan Plateau experiences broadly distributed deformation in response to the ongoing convergence of India with Asia (e.g., England and Houseman, 1986), but at relatively slow rates.

5-5 Conclusions

Displaced lacustrine shorelines developed around Zigui Co in central Tibet place bounds on the Holocene slip rate along the dextral GCF. My results show that the GCF has accrued 12 ± 1 m of displacement since the deposition of the shorelines at ~ 4.1 – 4.4 ka. These results represent the first direct estimate on Holocene slip rate of the GCF of ~ 2.1 – 3.2 mm/yr. Modest slip rates along this fault zone are consistent with rates of displacement along other strike-slip faults in central Tibet, but are significantly slower than geodetic measures of strain accumulation (Taylor and Peltzer, 2006). Thus, my results both support the notion that active deformation within the interior of the Tibetan Plateau is characterized by slow, distributed deformation and highlight the need for slip rates determined over longer timescales along this and other fault systems in central Tibet.
5-6 References


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Figure 5-1. The tectonic setting of the interior Tibetan plateau. Also shown here are the estimates of slip rates of fault systems in this region. The open circle indicate the possible location for the 1934 Shenzha earthquake rupture proposed by Armijo et al. (1989). Fault names: LCF – Lamu fault; RPCF – Riganpei Co fault; GCF – Gyaring Co fault; BCF – Beng Co fault. Lake names: N – Nam Co; S – Siling Co; T – Tangra Yumco; Z – Zhari Namco; Black numbers in the shaded boxes denote geodetically-derived slip rates and white numbers for geologically-derived long term slip rates. References: (a) Zhang et al., 2007; (b) Cowgill et al., 2009; (c) Bendick et al., 2000; (d) Shen et al., 2001; (e) Brown et al., 2002; (f) Chevalier et al., 2012; (g) Wright et al., 2004; (h) Jade et al., 2004; (i) van der Woerd et al., 2000; (j) Zhang et al., 2004; (k) Taylor et al., 2006.
Figure 5-2. GeoEye imagery (top) and sketch map (bottom) showing 1) fault displacements and scarps of the central segment of the Gyaring Co fault and 2) three groups of shorelines developed at the southeastern margin of Zigui Co.
Figure 5-3. (a) GeoEye imagery showing three groups of shorelines at the southeastern margin of Zigui Co where the highstand shorelines were displaced; (b) Closer view of the fault displacement on the highstand shorelines; (c) Even close view of the fault displacement in the imagery; (d) Sketch of reconstruction of fault displacement on top of the imagery; (e) Field photo showing shoreline offset.
Figure 5-4. Radial plots showing the distribution of equivalent dose (De) for each sample. Gray shading shows the aliquots used for modeling the OSL ages. Dr: dose rate; n is the number of aliquots analyzed for each sample.
APPENDIX A

Appendix A provides details about $^{36}$Cl depth profile, U-series, and radiocarbon ($^{14}$C) dating of the degraded shorelines above the Holocene highstand.

A-1 $^{36}$Cl depth profile dating

For shorelines that are mostly composed of limestone clasts and absent of sand layers, the OSL dating (of quartz) technique cannot be used. In such cases, I exploited depth profile (Anderson et al., 1996) dating of $in situ$ $^{36}$Cl in the shoreline deposits. This method utilizes the pattern of exponential decrease of $^{36}$Cl production down depth as a result of combination of different $^{36}$Cl production pathways: spallation of target elements (mainly Ca and K, and Ti and Fe to a lesser extent), negative muon capture (e.g., by $^{39}$K and $^{40}$Ca) and thermal neutron capture (e.g., by $^{35}$Cl) (Gosse and Phillips, 2001; Dunai, 2010). Because of such pattern of production, the post-deposition production of $^{36}$Cl in a rock at great depth (e.g., > 2 m) is negligible (Anderson et al., 1996); therefore the $^{36}$Cl concentration at such depth should reflect the inherited $^{36}$Cl prior to deposition – very important information to determine the age of the deposits. Therefore the depth profile dating could reduce uncertainties from inheritance of cosmogenic nuclides and yield more accurate age for the deposits where the inheritance is independently unknown (Anderson et al., 1996). In addition, using limestone for $^{36}$Cl dating has the advantage that $^{36}$Cl production pathways in limestone is relatively simple than other high chlorine-contained rocks – significant contributions from spallation and muon capture, but only relatively minor contribution from thermal neutron capture by $^{35}$Cl (especially for limestone with low total chlorine concentration [< 50 ppm], which is exactly the case for my samples) (Dep et al., 1994; Stone et al., 1996).

For dating I targeted undisturbed shoreline deposits after their deposition, according to the following observations: 1) The top surfaces of the shorelines are very flat such that I do not need to worry
about the age uncertainties from the geometry factor of the shielding surface; 2) The flat surfaces have been well paved, which indicates nearly no erosion on those surfaces and helps avoid large age uncertainties of unknown erosion rates, if any; 3) No bioturbation (e.g., beach clasts with soil carbonate coatings at depth have been transported to the surface by burrowing of rodent animals) is observed on the surface; and 4) The vertical cross-sections show sedimentary structures (e.g., imbricate beach clasts) indicative of original deposition.

A-1.1 Field sampling

To collect samples for the depth profile dating, pits were dug with dimensions of 2 m in length, 1.5 m in width and 2.5 m in depth (Table A-1). Such profiles are deep enough to catch the lowest \(^{36}\text{Cl}\) concentration to determine the inheritance of \(^{36}\text{Cl}\) in the deposits (Gosse and Phillips, 2001). Seven to eight horizontal sample bands of 5-10 cm thick were selected across the walls of vertical profiles. The vertical intervals are unevenly distributed, with more dense layers (three within the uppermost half meter) in the uppermost parts of the pits; this is to catch the pattern of exponential decrease of the \(^{36}\text{Cl}\) concentration down depth.

Samples were collected from the pit walls from bottom to top to avoid mixing clasts of high \(^{36}\text{Cl}\) concentration at top with those of much lower \(^{36}\text{Cl}\) concentration near the bottom of the pits. In the field, limestone clasts were excavated from selected horizontal sample layers, and were then sieved through 5 mm and 2 mm – size sieves to get clasts of 2 – 5 mm in size for all samples except CRN1-L2, L3 and L5. These three samples are clasts of 5 – 10 mm in size due to lack of 2 – 5 mm size in the sample layers. To avoid the influence of different clast size (hence clast surface area and volume) on the concentration of \(^{36}\text{Cl}\) in these three samples, I collected more than 2000 clasts for each sample so that the sample can represent the average \(^{36}\text{Cl}\) concentration in a specific sample layers. Because of this careful field sampling and chemical processing, the final result of depth profile concentration of CRN1 (See Figure 2-3, 4 in the main paper) indicate that clasts of different size may not cause incompatible cosmogenic nuclide
concentrations (that deviate from the exponential pattern) through a depth profile, as long as the total number of clasts is large enough to average the concentration.

In order to measure the bulk chemistry of the shoreline deposits, bulk samples were collected inbetween each horizontal sample layers. Bulk densities of deposits across the profiles were determined in the field by a elod method (Blake and Hartge, 1986).

A-1.2 Mass shielding correction

Partial shielding of the cosmic-ray flux by surrounding topographic and snow covers has been estimated to correct the production rates of $^{36}$Cl at the sample sites. The azimuths (45° interval) and dips of horizons were measured by compass and inclinometer to calculate the topographic shielding using the method of Dunne et al. (1999). Because the snow cover history of Tibet is not known, therefore I were forced to estimate the snow cover shielding based on monthly-average snow depths according to records from 72 stations in central-eastern Tibet during 1953 – 1999 (Wei et al., 2002; Yang et al., 2008). These records suggest a maximum snow depth of < 50 cm from October to May and zero snow cover from June to September in a year, which gives a snow shielding factor of 0.97. The final shielding factor (Table A-1) is then calculated by integrating the topographic and snow cover shielding (Gosse and Phillips, 2001).

A-1.3 Chemical processing

My samples were chemically prepared in the laboratory at New Mexico Tech and I follow the protocol of the lab which can be found in Appendix B of Marrero (2009). The chemical preparation of $^{36}$Cl samples in general includes 1) chemical etching, 2) hand-picking of limestone clasts, 3) crushing samples, 4) in-house determination of total chlorine in samples, 5) dissolving samples, 6) precipitating silver chloride (AgCl) in rock samples and blank solutions, 7) removing isobar ($^{35}$S) of $^{36}$Cl and, 8) AMS (Acceleration Mass Spectrometry) measurements of isotope ratios.
1) Chemical etching

The goal to chemically etch my samples is mainly to remove other isotopes and the thick (from < 1 mm – 1 cm) carbonate coatings on the surface of clasts. These post-depositional coatings can contaminate $^{36}$Cl samples to be prepared at the end. Before etching, the clasts were homogenized and split to make sure that the samples are fully mixed and representative of the average $^{36}$Cl concentration for the specific sample. During etching, pure HNO$_3$ acid was used to remove the thick carbonated coatings. After etching, the clasts were rinsed using 18 MΩ deionized water which help remove the meteoric chloride and dissolved carbonate coatings. This process of etching and rinsing was repeated for many times depending on the thickness of the carbonate coatings. The etched samples were then dried up in an oven.

2) Hand-picking limestone clasts

Usually little pieces of carbonate coatings remain after multiple times of etching and rinsing of the samples. To avoid contamination of these coatings in determination of $^{36}$Cl concentration, I removed the last little pieces of coatings by hand using tweezers and hand-picked ~ 900 – 1600 clean pure limestone clasts (~ 70 – 80 grams) from each sample. This total number of beach clasts is much larger than the number (~ 150) (Phillips et al., 2003) necessary to represent the average $^{36}$Cl concentration and inheritance for a specific sample. Although a careful attention has been paid, a very small fraction of non-carbonate clasts that well resemble limestone were picked and mixed with the limestone samples to be crushed and dissolved for Cl extraction. This non-carbonate fraction, however, is left over after dissolution, and the chemical composition of those leftovers was subtracted during later calculation (see below).

3) Crushing samples

The clean beach clasts after chemical etching were pulverized, and grains of 100 – 1000 microns were sieved out for further dissolution to release the chlorine in the rock. This size of
samples also helps remove meteoric chloride on the surface of the samples. At the same time, a very small fraction of these crushed samples were then powdered using a pair of mortar and pestle to determine: 1) approximate total chlorine concentration in the rock, and 2) composition of target elements (Ca, K, Ti, Fe) (Table A-4) of these carbonate samples by XRF (X-Ray Fluorescence) analysis. In addition, the bulk samples are also powdered to determine the bulk rock composition of both major and trace elements (B, Gd, Sm, U, Th, Li, Cr) (Table A-2) by XRF method. The XRF analysis has been done in a SGS lab in Toronto, Canada and the analytical results are shown in Table A-3.

4) In-housing determination of total chlorine in samples

The total chlorine concentration was measured by an ion-specific electrode method (Aruscavage and Campbell, 1983; Elsheimer, 1987). This process is to estimate the masses of rock samples and spike to be used in the dissolution to extract enough AgCl samples to determine the $^{36}\text{Cl}$ concentration, using the principle of isotope dilution mass spectrometry (IDMS) (Desilets et al., 2006). The results show that most of my samples are low in chlorine concentration (< 50 ppm).

5) Dissolving samples

According to the total chlorine concentration in the limestone and associate information of my sample sites, and utilizing a spreadsheet code, CHLOE (Phillips and Plummer, 1996), I determined to use ~ 20 - 25 grams of rock samples and ~ 3 grams of spike (0.9991±0.01 or 0.9999±0.01 mg $^{35}\text{Cl}$/g solution) (Table A-4) to extract enough $^{36}\text{Cl}$. These samples were separated from the homogenized samples and dissolved with ~ 30 – 35 mL of pure HNO₃ and ~ 210 – 265 mL 18 MΩ deionized water. The brown containers for dissolution were carefully sealed to prevent escape of chlorine from the dissolution. This dissolution process lasted for 3 days for complete extraction of chlorine from the rock samples. After dissolution, the brown solution was transferred to clean Teflon bottles and centrifuged for several times for purification. Some non-carbonate leftovers after
dissolution were then rinsed, dried up and powdered again for XRF analysis to determine their composition for further correction of target element composition in the dissolved carbonate samples (Schimmelpfennig et al., 2009).

6) Precipitating AgCl in rock samples and blank solutions

The chlorine in the purified solution was precipitated in Teflon beaker by adding 10 mL silver nitrite (AgNO₃). This process was done by placing the beakers onto a warm hotplate for > 8 hours. To subtract background chlorine in the lab that will be incorporated into the final AgCl samples for AMS measurement, blank solutions were made every other batch of ³⁶Cl samples using ~ 3 grams of spike and ~ 30 – 35 mL pure HNO₃. These blank solutions were then precipitated as AgCl at the same time as, and following the same method of, precipitation of AgCl from the sample solutions.

7) Removing isobar (³⁶S) of ³⁶Cl

The precipitated AgCl was centrifuged, rinsed, dissolved and reprecipitated for multiple times. Then 1.2 mL barium nitrite [Ba(NO₃)₂] was added for each solution to remove ³⁶S, an isobar of ³⁶Cl by precipitation of barium sulfate Ba(SO₄) for 7 days. After removal of ³⁶S the solution of AgCl was centrifuged, rinsed, dissolved and reprecipitated for multiple times. The final precipitates of AgCl were then dried in the oven.

8) AMS measurements

The dry samples of AgCl were packed up for AMS measurements of ratios of ³⁶Cl/Cl and ³⁵Cl/³⁷Cl at PRIMB lab (Elmore and Phillips, 1987; Muzikar et al., 2003). The analytical results are shown in Table A-4.
In addition, water content of the bulk samples was measured by weighing the masses before and after the samples was dried.

**A-1.4 Depth profile age computation**

The age of shoreline deposits was determined by a best fit between my measured $^{36}$Cl concentrations across a depth profile and the model-predicted concentrations at corresponding sample layers. This computation is based on the inheritance of $^{36}$Cl, post-deposition age and erosion rate of the deposits, $^{36}$Cl production rates at the site of interest that are calculated based on a given assumed scaling scheme (Gosse and Phillips, 2001). Here I exploited a newly-developed $^{36}$Cl depth profile calculator – CRONUScalc (Marrero et al., 2014, in review) to model depth profile concentrations and determine the ages of the shoreline deposits around the Siling Co. This depth profile calculator has incorporated the most recent improvements on calibration of $^{36}$Cl production rates (Table A-5) (Marrero, 2012) and scaling schemes (Lifton et al., 2014) that involves modeling of temporal variations of geomagnetic field and solar modulation on cosmic ray flux (Marrero, 2012) through the CRONUS-Earth (Cosmic-Ray prOduced NUclide Systematics on Earth). In addition, because this new calculator adopted the Bayesian method and maximum a posterior (MAP) solution (Marrero et al., 2014, in review), it has the advantage to simultaneously solve three key parameters that are best compatible with each other in cosmogenic depth profile dating: erosion rate, age and inheritance. Before running this calculator, several calculations need to be done.

1. Because interactions of cosmic ray fluxes with rock (hence the production rate of $^{36}$Cl) exponentially decreases, in fact, with depth times bulk density of the rock, or cumulative mass per unit area ($\text{g/cm}^2$, Gosse and Phillips (2001)), I calculated such cumulative overburden mass based on my measurements of bulk sample densities and thicknesses. In this way, the $^{36}$Cl concentrations in samples from different depths of variant lithology and bulk densities can be compared with each other and be used in depth profile calculations.
2. To correct the chemical compositions of the target elements (Ca, K, Ti and Fe) in the dissolved samples, the chemical composition of these elements in undissolved residues need to be subtracted from those measured before carbonate dissolution (Schimmelpfennig et al., 2009). This can be simply done through a mass balance before and after the dissolution, using the masses and chemical compositions of two sets of samples: the total samples before dissolution and the undissovled residues after dissolution (Schimmelpfennig et al., 2009).

3. Because background $^{36}$Cl and Cl were incorporated during the chemical processing (especially through addition of nitric acid for dissolving carbonates), therefore they were subtracted from the total samples. The final $^{36}$Cl and total Cl concentrations in the samples were calculated based on the isotope dilution method (Desilets et al., 2006), using isotope ratios of $^{36}$Cl/Cl, $^{35}$Cl/$^{37}$Cl (Table A-4), measurement of masses of nitric acid and spikes in the blank solutions (Table A-4) (Schimmelpfennig et al., 2009; Marrero, 2012).

Finally, I used the following initial numeric bounds of age, erosion and inheritance (Table A-6), to run the calculator. These parameter ranges are large enough given previous estimate on ages of the shorelines. Using the Bayesian method and maximum a posterior (MAP) solution, this calculator treats all three parameters as unknown random variables and finds a best solution that the modeled $^{36}$Cl concentrations using each set of the three parameters show a minimum $\chi^2$ goodness of fit (see details in Chapter 3 of Aumer (2010) and Section 3.6.2 of Marrero (2012a)) with my measured $^{36}$Cl concentrations. The uncertainties of each parameter are provided as 68% (1σ) and 95% (2σ) of confidence interval contours (See Figure 2-4 in the main paper).

A-1.5 Depth profile ages and uncertainties

The uncertainties of depth profile ages come from many sources (Gosse and Phillips, 2001), including natural uncertainties from estimation of snow cover shielding, calibration of production rates, scaling factors due to variable latitudes, altitudes, uncertain temporal variations of geomagnetic field and
solar modulation, and human-sourced uncertainties from physical measurements (e.g., of topographic shielding, bulk density, sample depth) and lab chemical processing (e.g., measurements of sample and spike masses, volumes of acids and water, contamination of carbonate coatings) and AMS measurements. As a result of recent advancements of cosmogenic nuclide researches, especially within the CRONUS framework, such as calibration of production rates of $^{36}\text{Cl}$ (Marrero, 2012), $^{10}\text{Be}$ and $^{26}\text{Al}$ (cf. Balco et al., 2008) and improvement of scaling scheme that involves time-dependent variation of geomagnetic field and solar modulation (Lifton and Sato, in prep.), the uncertainties of cosmogenic nuclide dating has been improved accordingly. As shown in Table A-7 and See Figure 2-4 in the main paper, the age uncertainties for my samples are less than 2% for confidence interval of 68% (one standard deviation - $1\sigma$). Even for 95% of confidence interval ($2\sigma$), the uncertainties are less than 4%.

A-2 U-series dating

For some of the wave-cut cliffs immediately above the highstand where tufa deposits are present (see locations in Figure 2-2 in the main text), I use U-series dating of $^{230}\text{Th}$ to estimate the age of the lake levels above the highstand. The tufa samples are analyzed at the Radiogenic Isotope Laboratory at University of New Mexico, following the analytical protocol of Asmerom and Edwards, (1995).

A-3 Radiocarbon ($^{14}\text{C}$) dating

Organic materials can rarely be found in shoreline deposits around Siling Co. But occasionally I found from a depth profile pit one broken piece of snail shell. This shell was cemented on limestone clasts in the beach deposits of a degraded constructional shoreline developed in the central peninsula of Siling Co (Figure 2-2). This shell sample was analyzed at Beta Analytic Inc., following the standard pretreatment protocol (http://www.radiocarbon.com/pretreatment-carbon-dating.htm). More details about
the dating method can be found at the link (http://www.radiocarbon.com/PDF/Beta-AMS-methodology.pdf).
A-4 References


### Table A-1. Locations, physical measurements for three depth profile samples

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<th>Site ID</th>
<th>Lat (˚)</th>
<th>Long (˚)</th>
<th>Elev (m)</th>
<th>Shielding</th>
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<table>
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<th>Cumu. Depth to Top (g/cm²)</th>
<th>Uncert (g/cm²)</th>
<th>Water Content (wt. %)</th>
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</table>

* Cumu. Depth to Top: the cumulative sample depth times the bulk density at the specific depth.
Table A-2. Trace element and target oxide composition† of bulk rock samples from three depth profiles

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Trace Element Composition in Bulk Rock</th>
<th>Target Oxide/Element in the Sample</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>B (ppm)</td>
<td>Sm (ppm)</td>
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</tr>
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</tr>
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<td>3.00</td>
</tr>
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† the composition of target elements has been corrected by subtraction of those in silicate leftovers after dissolving crushed limestone samples of 0.25 – 1 mm in size.
Table A-3. Oxide composition of bulk rock samples from three depth profiles

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>SiO$_2$ (wt. %)</th>
<th>TiO$_2$ (wt. %)</th>
<th>Al$_2$O$_3$ (wt. %)</th>
<th>Fe$_2$O$_3$ (wt. %)</th>
<th>MnO (wt. %)</th>
<th>MgO (wt. %)</th>
<th>CaO (wt. %)</th>
<th>Na$_2$O (wt. %)</th>
<th>K$_2$O (wt. %)</th>
<th>P$_2$O$_5$ (wt. %)</th>
<th>LOI$^*$ (wt. %)</th>
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<td>CRN1-L1</td>
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<td>0.71</td>
<td>44.70</td>
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<td>0.47</td>
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<td>0.71</td>
<td>44.70</td>
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* LOI – loss on ignition, here considered being the composition of CO$_2$. 
### Table A-4. Sample, spike masses and AMS measurements for three depth profile samples

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<tr>
<th>Layer Name</th>
<th>Masses of rock and spike analyzed</th>
<th>AMS measurements</th>
<th></th>
<th></th>
</tr>
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<tr>
<td></td>
<td>Rock (g)</td>
<td>Spike (g)</td>
<td>$^{36}\text{Cl}/^{35}\text{Cl}$ (10$^{-15}$)</td>
<td>$^{36}\text{Cl}$ uncert</td>
</tr>
<tr>
<td>CRN1-L1</td>
<td>25.0640</td>
<td>2.9923</td>
<td>19000</td>
<td>400</td>
</tr>
<tr>
<td>CRN1-L2</td>
<td>25.2969</td>
<td>3.0033</td>
<td>18000</td>
<td>400</td>
</tr>
<tr>
<td>CRN1-L3</td>
<td>24.9982</td>
<td>3.0087</td>
<td>16000</td>
<td>400</td>
</tr>
<tr>
<td>CRN1-L4</td>
<td>25.5895</td>
<td>2.9951</td>
<td>14700</td>
<td>700</td>
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<tr>
<td>CRN1-L5</td>
<td>24.8352</td>
<td>3.0085</td>
<td>10950</td>
<td>210</td>
</tr>
<tr>
<td>CRN1-L6</td>
<td>25.2068</td>
<td>3.0140</td>
<td>8110</td>
<td>130</td>
</tr>
<tr>
<td>CRN1-L7</td>
<td>24.4976</td>
<td>3.0297</td>
<td>3780</td>
<td>60</td>
</tr>
<tr>
<td>CRN1-L8</td>
<td>24.7634</td>
<td>2.9934</td>
<td>2490</td>
<td>40</td>
</tr>
</tbody>
</table>

| CRN3-L4    | 60.4869                           | 1.8270           | 5030  | 270   | 5.92  | 0.09  |
| CRN3-L5    | 59.9961                           | 1.9401           | 3600  | 140   | 6.53  | 0.09  |
| CRN3-L6    | 60.4638                           | 1.7887           | 2150  | 50    | 5.20  | 0.10  |
| CRN3-L7    | 60.6132                           | 1.8414           | 1530  | 40    | 6.31  | 0.23  |
| CRN3-L8    | 74.6719                           | 1.8487           | 1210  | 50    | 5.61  | 0.13  |

| CRN5-L1    | 21.7475                           | 2.9960           | 21200 | 280   | 31.00 | 0.50  |
| CRN5-L2    | 20.6923                           | 3.0236           | 18100 | 400   | 40.26 | 1.50  |
| CRN5-L3    | 21.2690                           | 2.9962           | 19900 | 400   | 36.22 | 0.78  |
| CRN5-L4    | 22.2113                           | 3.0082           | 14200 | 400   | 39.28 | 1.08  |
| CRN5-L5    | 25.5481                           | 3.0076           | 14180 | 160   | 31.36 | 0.63  |
| CRN5-L6    | 22.1862                           | 2.9950           | 8510  | 230   | 41.00 | 1.66  |
| CRN5-L7    | 21.3079                           | 3.0298           | 8800  | 500   | 34.41 | 0.60  |
Table A- 5. Calibration parameters for $^{36}$Cl*

<table>
<thead>
<tr>
<th>Pathway</th>
<th>Production rate of $^{36}$Cl (atom g$^{-1}$ yr$^{-1}$)</th>
<th>Uncertainty (atom g$^{-1}$ yr$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>spallation of Ca (PsCa0)</td>
<td>56</td>
<td>2.2</td>
</tr>
<tr>
<td>spallation of K (PsK0)</td>
<td>157</td>
<td>6</td>
</tr>
<tr>
<td>Thermal/epithermal neutron capture (Pf0)</td>
<td>704</td>
<td>141</td>
</tr>
</tbody>
</table>

* Results from Marrero (2012)

Table A- 6. Input parameter bounds for depth profile computation

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Age (ka)</th>
<th>Erosion rate (g/cm$^2$/ka)*</th>
<th>Inheritance (ka)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bounds</td>
<td>0 ~ 300</td>
<td>-10 ~ 10</td>
<td>0 ~ 300</td>
</tr>
</tbody>
</table>

* the erosion rate is normalized to rock density, for a density of 2 g/cm$^3$, the erosion rate of 10 g/cm$^2$/ka equals to 0.05 mm/yr or 50 m/Ma; a negative erosion rate means aggradation.

Table A- 7. MAP solutions for ages, erosion rates and inheritance of the depth profiles

<table>
<thead>
<tr>
<th>Profile ID</th>
<th>Site ID</th>
<th>Age (ka)</th>
<th>1σ (ka)</th>
<th>2σ (ka)</th>
<th>Erosion (g/cm$^2$/ka)</th>
<th>1σ</th>
<th>Inheritance (ka)</th>
<th>1σ</th>
<th>Shoreline Level</th>
</tr>
</thead>
<tbody>
<tr>
<td>CRN1</td>
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<td>178.4</td>
<td>+2.1</td>
<td>+4.4</td>
<td>0</td>
<td>0.1</td>
<td>16.0</td>
<td>+0.7</td>
<td>above highstand</td>
</tr>
<tr>
<td>CRN3</td>
<td>66</td>
<td>186.3</td>
<td>+11.8</td>
<td>+21.2</td>
<td>1.4</td>
<td>0.4</td>
<td>5.6</td>
<td>+3.6</td>
<td>above highstand</td>
</tr>
<tr>
<td>CRN5</td>
<td>23</td>
<td>113.4</td>
<td>+4.3</td>
<td>+7.3</td>
<td>0</td>
<td>0.1</td>
<td>98.3</td>
<td>+1.9</td>
<td>above highstand</td>
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</tbody>
</table>
APPENDIX B

Appendix B includes two parts. Part 1 describes detailed OSL methodology. Part 2 provides descriptions of sedimentary facies and interpretation of depositional environment and water depths of additional sample sites that have not been (fully) discussed in the main body of the paper. Together with the OSL sample age, paleo-lake level history is reconstructed at each site. The OSL dosimetry tables are also attached in the back.

B-1 OSL Laboratory Processing

B-1.1 Sample preparation

Samples were prepared using standard OSL procedures. Material was desiccated at 50 °C to enable calculation of water content, and then sieved to extract the 180-212 µm grain size fraction. Approximately 10 g of the 180-212 µm grain size fraction was treated with 30% HCl for 30 minutes to remove CaCO3. Samples were agitated throughout the treatment, and once complete, HCl was replaced with 30% H2O2 to remove organic material. The duration of H2O2 treatment varied between samples, dependent upon the amount of organic material present, and two H2O2 treatments were necessary for some samples with a high organic content. Once effervescence ceased, the H2O2 was decanted, the sample was washed four times with deionised water and desiccated at 50 °C. Quartz is extracted from polymineral sediment residues through density separations using LST fastfloat (sodium heteropolytungstate dissolved in deionized water). Heavy minerals (> 2.68 g cm-3) were separated from the lighter fraction, and the target 2.58-2.68 g cm-3 fraction was further separated from the < 2.58 g cm-3 material. The target fraction was washed five times with deionised water to ensure removal of all LST.
Final separates were dried and etched with 40% HF for 40 minutes to remove any contaminating feldspar; all samples were agitated at 5 minute intervals throughout treatment. The etched quartz was treated with 30% HCl for 30 minutes to remove any carbonates produced during HF etching.

### B-1.2 Luminescence measurements

All analyses were carried out using either a TL-DA-15 or TL-DA-20 Risø reader, equipped with an EMI 9235QA photomultiplier and 7.5 mm Hoya U-340 filter. Blue (470±20 nm) and infrared (~870 nm) diodes operated at 90% and 40% power respectively, were used for stimulation and irradiation was achieved using a 90Sr/90Y beta source. Readers were calibrated using quartz prepared at the Risø National Laboratory in Denmark. Quartz was applied to stainless steel discs (10 mm Ø, 1 mm thick) using silicon grease and the aliquot size was regulated using a small (2 mm Ø, ~35 grain) mask.

Samples were analysed using the single aliquot regenerative dose (SAR) protocol (Murray & Wintle, 2000). The equivalent dose (De) was calculated from measurements of the luminescence response following stimulation of the natural luminescence (Ln) and a series of different regenerative doses (Lx). The Ln and Lx measurements are normalised by measurement of the luminescence response (Tx) to a constant test dose (TD). The ratio Lx/Tx is used to compensate for sensitivity changes of the quartz throughout analysis and Lx/Tx measurements are used to obtain a range of values which bracket Ln/Tx, allowing De interpolation with minimal associated errors (Banerjee et al., 2000).

The sample is heated prior to making the luminescence measurements in order to reduce the contribution of luminescence from unstable trap, which cause erroneous dose determinations. The temperature of the pre-Lx pre-heat (PH1) and the pre-Tx pre-heat (PH2) must be empirically determined for each sample under analysis. This was achieved through analysis of 8 disks with a dose-recovery pre-heat plateau experiment (Murray and Wintle, 2003), using a range of PH1 and PH2 temperatures ranging from 150 – 220 °C. The influence of a hot-bleach at 280 °C was also investigated (Murray and Wintle, 2003) but was not required to counter any thermal transfer or recuperation.
The SAR protocol with a PH1 of 200 °C used on most of the samples is provided in Table B-1. Aliquot acceptance criteria used are 1) recycling ratios within 10% of unity; 2) signal intensities ≥ 3 σ above background; 3) infra-red (IR) depletion ratio within 20% of unity (Duller, 2003); 4) De uncertainty ≤ 20 % and 5) recuperation within 10% of the normalised maximum dose. The acceptance thresholds are generally very high for the samples (85 – 100%) reflecting the sensitivity of the quartz analysed. All samples have at least 50 accepted aliquots.

B-1.3 Environmental Dose rate determination

The environmental dose rates (Dr) were calculated for each sample from the unsieved portions of the original sample; concentrations of U, Th, K and Rb were measured directly using solution ICP-MS (Thermo X-Series), a cosmic-dose component after Prescott and Hutton (1994) and an internal alpha dose rate of 5% from the decay of U and Th after Sutton and Zimmerman (1978). External α-dose rates were ignored as the alpha irradiated portion of quartz grains was removed by etching. The conversion factors of Adamiec and Aitken (1998) and beta-particle attenuation factors after Mejdahl (1979) and Readhead (2002a, b) have been used. Sample water content was calculated following desiccation at 50 °C, and an uncertainty of 5 % assumed. Table B-2 contains the dosimetry data of OSL samples from the highstand shorelines.

B-1.4 Analysis of results

Most samples are characterized by large overdispersion values (broad De distributions) which describe the spread in the data not accounted for by analytical uncertainties (Galbraith and Roberts, 2012). Samples with overdispersion, greater than 20%, are assumed to reflect heterogeneous bleaching before deposition and the De data were analysed within Excel to calculate overdispersion and statistical parameters that are used to model the appropriate burial age (Db). I adopted the age model selection
criteria of Arnold (2009). Most samples are modelled using the three component minimum age model (MAM-3; Galbraith et al., 1999) and the RStudio Luminescence package (Kreustzer et al., 2013). The data are distributed and in multi-grain analyses the clear differentiation of separate populations is not straightforward (Arnold, 2009), however the finite mixture model (FMM) has been adopted in two cases where the MAM3 model fails to capture the lowest population of De values (Galbraith and Green, 1990). Arnold and Roberts (2009) proposed that the FMM model should only be used on single grain age data because there is the possibility of selecting false populations when multi-grain aliquots are used. However, Rodnight et al. (2006) used the same model to calculate ages for fluvial sediments in South Africa based on agreement of FMM ages and independent radiocarbon ages. In the absence of secondary age dating, I use the FMM model for two ages when MAM3 fails to capture the lowest population and to explore the age of older populations which may have been reworked into younger shoreline deposits.

**B-2 Stratigraphic descriptions for additional sites**

**B-2.1 Lake levels determined from surficial shoreline deposits**

**Group 2 shorelines**

*Site O17*

Site O17 is located below O18, constituting one of the Group 2 shorelines in the central peninsula of Siling Co. Two units of sedimentary facies are observed from the 1.7 m deep soil pit (Figure B-1). Unit 1 is a very thick layer that contains many interbedded thin sublayers of well-sorted, grain-supported rounded gravels and sands. Unit 2 at the top is a thin layer mostly composed of mud and silt sands.
I interpret Unit 1 of well-sorted, grain-supported gravels and sands as typical beach ridges, such that the lake level was at or very close at the elevation of this unit, and Unit 2 of mud and silt sands as soil layer at the top.

The OSL sample, O17, from a thin layer of fine sands yields an OSL age of 3.0 ± 0.2 ka, suggesting that the lake level was at the sample position around ~3 ka.

**Site O44**

Site O44 is the lowest of the G2 shorelines cut by the Tashi Stream (Figure 3-3c). Five units of sedimentary facies are observed at this site (Figure B-2). Unit 1 at the bottom of this outcrop contains landward-dipping layers of clean grain-supported, rounded gravels. Unit 2 is a medium-to-coarse-sized sand body that wedged into Unit 1 gravels. Unit 3 is a thick layer of landward-dipping layers of clean, grain-supported, rounded pebble-sized gravels. Two thin lenses of medium sands are intercalated within the gravel layers. An OSL sample (O44) was collected from the sand lenses. Likewise, no soil layer was found at the top surface at Site O44. Unit 4 is a layer of relatively poorly sorted gravels with variable sizes from pebbles to cobbles, with a small portion of mud and silt sands as matrix. Unit 5 is a thin layer of mixture of mud and silt sands.

I interpret the landward-dipping grain-supported rounded gravels in Unit 1 as backsets of a beach ridge, and the thick medium-to-coarse-sized sand wedge of Unit 2 as possible ice wedge cast deposits, or alluvial channel deposits. The thick landward-dipping gravel layers in Unit 3 is considered as backsets of a typical beach ridge. The relatively poorly sorted gravels in Unit 4 suggest an alluvial fan deposition. The thin layer of muddy silt at the top is interpreted as alluvial soils.

The medium sands in Unit 3 beach ridge give an OSL age of 4.5 ± 0.3 ka. This reflects the lake level was at this shoreline elevation during the Middle Holocene.
**Group 1d shorelines**

*Site O36*

Site O36 is a stream-cut outcrop in the lowest of G1d shorelines (Figure B-3). Here there is only a limited exposure of strata. Three units of sedimentary facies are shown in this outcrop. Unit 1 at the bottom is a thick layer of poorly-sorted, mud-matrix-supported gravels. Unit 2 is a thick pile of alternating layers of grain-supported gravels. A thin layer of coarse sands is intercalated with this unit. Unit 3 at the top includes two sublayers of poorly-sorted subangular gravels with mud and silt sands as matrix at the bottom and a top sublayer of mud with sparse pebbles within it.

Unit 1 at the bottom and Unit 3 at the top that contain poorly sorted sediments with muddy matrix are interpreted as alluvial fan deposits. The top sublayer of Unit 3 is a soil layer. Unit 2 gravels are typical of beach ridge deposits, suggesting the lake level was at the elevation of Unit 2 during deposition of the coarse sands. Given the presence of a thin soil layer at the top of this site and no preservation of soil layer at other most-recently developed beach ridges (e.g., O12, O13 and O44), the Unit 2 at this site could be an ancient beach ridge.

The OSL age of Unit 2 beach sands is determined to be 5.0 ± 0.2 ka, suggesting the lake level was near the elevation of these shorelines during 5 ka. But because the stratigraphic extension is not clear, I put a low confidence in the lake level at this site.

**Group 1c shorelines**

*Site O38/O39*

Site O38 is near the top of a possible alluvial fan that traverses the G1c shorelines. The contact between the sedimentary facies units is not very clear. Two OSL samples, O38 and O39, are obtained from this outcrop (Figure B-4). Here the cross-section shows four units of sedimentary facies. Unit 1 at the bottom presents landward-dipping layers of relatively well sorted, mostly grain-supported rounded
gravels. Unit 2 is a thin layer of fine sands onlapping Unit 1 gravels with a variable slope. Unit 3 is a thick pile of lakeward layering of relatively-well sorted, mostly grain-supported rounded gravels, with a little bit of silt sands as matrix. Unit 4 is a thick pile of not-well sorted mud-matrix-supported gravels mixed with lenses of fine sands. But the contact between Unit 3 and 4 is not very clear.

I consider the landward-dipping gravel layers in Unit 1 as backsets of a beach ridge, and the find sands in Unit 2 that gently onlapping to Unit 1 as nearshore sands below the wave influence. The lakeward-dipping grain-supported gravels in Unit 3 are interpreted as foresets of a beach ridge. The mud-matrix supported, not well sorted gravels in Unit 4 are considered as alluvial fan deposits. But because the contact between Unit 3 and 4 is not clear, I lower my confidence in interpretation of the depositional environment of Unit 4.

The OSL age of O39 from the nearshore sands in Unit 2 is $24.0 \pm 0.9$ ka, and O38 is $3.7 \pm 0.2$ ka. Together with the observations of the depositional environment, these ages suggest that the lake level was above the nearshore sands in Unit 2 around 24 ka, but below the sands in Unit 4 (low confidence) at $\sim 4$ ka. But I put a low confidence in this interpretation given the unclear stratigraphic context.

**Group 1b shorelines**

*Site O42*

Only one unit of sedimentary facies is observed from this stream-cut outcrop (Figure B-5). It shows several sublayers of interbedded well-sorted, grain-supported gravels and medium, coarse sands, which is interpreted as beach ridge deposits. Thus, the lake level is expected to be at or close to this site. The OSL sample, O42, from the medium sands this unit yields an age of $1.2 \pm 0.2$ ka, suggesting that the lake level was at the sample elevation around 1 ka.
**Group 1a shorelines**

*Site O7*

Two units of sedimentary facies are observed from this < 1 m high outcrop (Figure B-6). Unit 1 is a thick layer composed of poorly-sorted, matrix-supported rounded gravels, with many cobbles and boulders in this unit. A body of fine to silt sands was wedged into this unit. Unit 2 is a medium thick layer consists of a mixture of mud, silt sands and sparsely distributed gravels.

I interpret Unit 1 that consists of poorly sorted, matrix-supported gravels as debris flow deposits, and Unit 2 with mixture of mud and sands as alluvial fan deposits. The OSL sample, O7, from Unit 1 yields an age of 0.8 ± 0.0 ka, suggesting that the lake level was below the debris flow unit around 1 ka.

**B-2.2 Lake levels determined from shallow shoreline strata**

**Group 3 shorelines**

*Site O20*

Site O20 is located below O21 (the Lingtong highstand) but above O19 (Figure 3-3a in the text). Three units of sedimentary facies are observed from the 1.7 m deep soil pit (Figure B-7). Unit 1 at the bottom, like the deposits of Unit 2 at Site O14, contains a layer of yellowish, wet, poorly-sorted rounded pebbles/cobbles mixing with sands and mud. The contact of this unit with the unit below is not seen in this pit. A thin lens of fine sands was found in the upper part of this unit. Unit 2 in general is a very thick pile of whitish, relatively well-layered gravels. Two sublayers are present in this unit. The lower part is characterized by a thick bed of grain-supported rounded to subangular pebbles and cobbles, and in the upper part are interbedded thinner layers with gravels of smaller grain size. At the top is a thin layer of yellow-brownish mixture of silts, mud and sparsely distributed pebbles. Plants are clearly seen growing in this unit.
I interpret the poorly sorted gravels mixed with mud and sands in Unit 1 at the bottom of this soil pit as alluvial fan that was developed above the lake level with some distance. The grain-supported gravel layers in Unit 2 are considered as beach ridges. Unit 3 with mixture of mud and silt sands is considered as a soil layer.

The OSL age of fine sands from the bottom alluvial deposits is $63.8 \pm 4.6$ ka, indicating the lake level was below the sample elevation at a time older than 64 ka (recall that OSL samples older than 40 ka are likely saturated, see discussion above).

**Site O19**

Site O19 is below Site O20 in elevation. Four units of sedimentary facies are observed from this 1.2 m soil pit (Figure B-8). Unit 1 at the bottom is a $>30$ cm thick layer (the contact with the unit below not seen in this pit) of yellowish fine sands. Unit 2 is a thick whitish layer of clean, grain-supported rounded gravels. It consists a few alternating sublayers characterized by grain size. One sublayer in the middle is not-well sorted and contains a few cobbles. Unit 3 is a medium thick layer of rounded pebbles, with mud and silt sands as matrix. Unit 4 is a thin layer mostly composed of mud and silt sands.

I interpret the thick layer of fine sands in Unit 1 at the bottom as possible nearshore environment below the wave influence, such that the water depth of this unit could be several meters or even deeper than 10 m (e.g., Machlus et al., 2000). Unit 2 of grain-supported rounded gravels is interpreted as beach ridge deposits. Unit 3 is considered as alluvial fan deposits based on the mixing of mud and silt sands within the gravels. Unit 4 of mud and silt sands suggests it is a soil layer at the top.

The OSL sample, O19, was collected from the bottom unit and provides an OSL age of $28.5 \pm 2.0$ ka, suggesting that the lake level was probably several meters above the sample position at time of deposition of the near shore fine sands at $\sim 29$ ka.
Group 2 shorelines

*Site O13*

Site O13 is a shoreline spit cross-cut by the seasonally-flowing Tashi Stream (Figure 3-3b in the main body of text). It has some distance away from Site O18 and O17. But because this shoreline has a similar elevation (~ 4582 m) to Site O18, I categorize it into Group 2 shorelines as well. Basically two units of sedimentary facies are observed from this cross-section of ~ 3.5 m in height (Figure B-9). Unit 1 shows poorly-sorted matrix-supported rounded to angular cobbles and boulders at the bottom. Immediately above is a sublayer of poorly-sorted angular gravel wedges intercalated with massive coarse-to silt-sized sands. Unit 2 at the top is a thick pile of whitish, alternatingly-layered, clean, grain-supported, rounded gravels of different grain sizes, inclining towards the lake.

The top gravel unit with lakeward clinoforms is typical of foresets of beach ridges which should sit at or immediately above the lake level. The bottom unit with matrix-supported cobbles and boulders is interpreted as debris flow deposits that developed above the lake level. Notice that, this site, unlike all previous ones, has no presence of a soil layer, indicating the age of the top beach ridge is possibly younger than all recently-constructed beach ridges mentioned above. But the presence of a sharp boundary between the top beach gravels and bottom debris flow deposits suggests the debris unit may be much older than the top beach ridge.

I collected an OSL sample of silt sands from the upper part of Unit 1 debris deposits. The age of the sample is 13.1 ± 3.2 ka, suggesting the lake level is below this site. It is unknown, however, how low the lake was from this site around 13 ka. This question is answered later from other shoreline sites.

Group 1d shorelines

*Site O37*
Three units of sedimentary facies are observed from the ~ 2.5 - 3 m high outcrop cut by the Tashi Stream at Site O37 (Figure B-4). Unit 1 contains a mixture of poorly-sorted rounded gravels, with mud and silt sands as matrix. Several irregular-shaped bodies of medium sands are wedged into this unit. Unit 2 is a thin layer of whitish pebbles that are mostly grain-supported, but not well-sorted. Unit 3 contains three sublayers. Both sublayers at the top and bottom of this unit are characteristic of brownish mixture of massive mud, silt sands and sparsely distributed angular-shaped gravels. The middle sublayer contains poorly-sorted matrix-supported gravels, with a few cobbles in it.

I interpret Unit 1 of poorly-sorted, matrix-supported gravels as debris flow deposits; therefore the lake level was below this unit in elevation. The thin layer of grain-supported pebbles in Unit 2 looks like beach deposits, but I do not have high confidence in this interpretation here. Unit 3 with poorly sorted gravels and mud/silt sediment portions is considered as stratified alluvial fan deposits.

The OSL sample, O37, from the medium sands in Unit 1 yields an age of 11.9 ± 0.6 ka, suggesting that the lake level was below the sample position around ~ 12 ka.

**Group 1c shorelines**

**Site O11**

This site shows a strike-perpendicular stratigraphic exposure that includes four units of sedimentary facies (Figure B-11). Unit 1 at the bottom is a thick pile of landward-dipping layers of relatively-well-sorted, mostly grain-supported rounded gravels. Unit 2 has several point-down triangular-shaped wedges of silt sands. Unit 3 is a thick pile of brownish-looking well-sorted, grain-supported, rounded gravel layers with variable inclinations. Unit 4 is a thick pile of mud and silts contains sparse pebbles and sand lenses.

I interpret Unit 1 with landward-dipping gravel layers as backsets of the beach ridge; Unit 2 of sand wedges as probable ice wedge casts; Unit 3 of well sorted and grain-supported gravels as a typical beach ridge with presence of both backsets and foresets; and Unit 4 with mud and silt mixture as alluvial...
fan deposits. The OSL age of O11 from the silt sands in the ice wedge cast is $13.2 \pm 3.3$ ka, implying the lake level was below this unit during development of the ice wedge casts.

**Group 1b shorelines**

*Site O9*

Site O9 is immediately east of Site O10. Here only a small portion of shoreline strata is exposed. Two units are observed at this site (*Figure B-12*). Unit 1 at the bottom is a layer of grayish silt sands. Unit 2 at the top is a thick layer of well-sorted, mostly grain-supported, rounded pebbles, with a little bit of sands within this layer.

Unit 1 of grayish silt sands is considered as possible nearshore below the wave influence. But because of the limited exposure of the stratigraphic context, the confidence of this interpretation is low. Unit 2 of well sorted, grain-supported pebbles is interpreted as beach ridge deposits. The OSL age of O9 from Unit 1 is $3.8 \pm 0.2$ ka, suggesting that the lake level is possibly above this site around 4 ka. But because of the limited exposure of the stratigraphic context, the confidence of this interpretation is low, and I do not use this site for further discussion of the lake level.

*Site O8*

Two units of sedimentary facies are observed from the ~ 1.7 m high outcrop cut by the Tashi Stream at Site O8 (*Figure B-13*). Unit 1 is a thin layer of medium sands, but its contact with the unit below is shown here. Unit 2 contains four sublayers of gravels. The bottom sublayer is medium thick, mostly grain-supported but with a few sand matrix. The upper three sublayers are clean, well-sorted, grain-supported rounded gravels with layering both towards the lake and land. No soil layer is preserved at the top.
Unit 1 with thin layer of medium sands could be nearshore environment below the wave influence. But because the contact is not clear here, I put a low confidence of this interpretation. Well sorted and grain-supported gravels in Unit 2 are typical of backsets or foresets of beach ridges.

The OSL sample, O8, from the medium sands in Unit 1 yields an age of 3.8 ± 0.2 ka, suggesting that the lake level was possibly above this unit around 4 ka. Again, the confidence is low because of insufficient stratigraphic context at the bottom. Therefore I do not take site into account in discussion of the lake history of Siling Co.

Group 1a shorelines

*Site O16*

Three sedimentary facies units are observed in this ~ 1.7 m deep soil pit (Figure B-14). Unit 1 at the bottom is a thick layer of yellowish, poorly-sorted, mud-matrix-supported angular gravels that include some cobbles and boulders. A thin layer of fine sands is intercalated with this unit. Unit 2 includes three thin-medium thick sublayers. The top and bottom sublayers contain relatively well-sorted, grain-supported rounded gravels. The middle sublayer is not well-sorted, with a few cobbles in it, but the gravels are clean and grain-supported. Unit 3 at the top is a ~ 20 cm thick layer of mud and silt sands.

I interpret Unit 1 characterized by poorly sorted, matrix-supported angular gravels at the bottom as alluvial fan deposits that developed above the lake level; and Unit 2 with well sorted, grain-supported gravels as a beach ridge; Unit 3 with a mixture of mud and silt sands is a soil layer, which suggests that this shoreline is probably old.

The OSL age of O16 from the fine sands is 22.0 ± 1.5 ka, implying the lake was below these deposits around 22 ka.
**East margin**

*Site O27*

This site is the easternmost one among all OSL sites. Four sedimentary facies units are observed in this ~ 2 m high road-cut outcrop (Figure B-15). Unit 1 at the bottom is a very thick layer of yellowish, laminated fine sands. Unit 2 is a medium thick layer containing landward-dipping, well-sorted, mostly grain-supported and rounded gravels, with layering towards the land. Unit 3 has two sublayers, with a very thin layer of very coarse sands at the bottom and well-sorted, mostly grain-supported, rounded gravels at the top. Unit 4 at the top is a thin layer of mud and silt sands.

I interpret Unit 1 with laminated fine sands at the bottom as deepwater lacustrine sands; Unit 2 with landward-dipping well sorted, grain-supported gravels as backsets of a beach ridge; Unit 3 with flat-lying grain-supported gravels as the topset of a beach ridge; and Unit 4 with a mixture of mud and silt sands as a soil layer.

The OSL age of O27 from the fine sands in Unit 1 is 52.8 ± 1.7 ka, implying the lake was away above the sample elevation around 53 ka.

*Site O23*

The shoreline at this site is the highest one (~ 4640 m in elevation) along the eastern margin of Siling Co. Two sedimentary facies units are observed in this ~ 1.7 m deep soil pit at Site O23 (Figure B-16). In general, the entire pit shows a reddish color, and the sediment textures are not as clear as sites previously observed. Unit 1 at the bottom is a very thick pile of interbedded sublayers of well-sorted rounded gravels, which are mostly grain-supported, but with a proportion of sand matrix. Several thin layers of fine and medium sands are intercalated with this unit. Unit 2 is a ~ 20 cm thick layer of mud and silt sands.
I interpret Unit 1 of interbedded rounded gravel layers as beach ridge deposits, and Unit 2 with a mixture of mud and silt sands as a soil layer. In general, the reddish color and textures of the sediments suggest a large degree of shoreline degradation at this site.

The OSL age of O23 from the fine-medium sands near the bottom of Unit 1 is 43.0 ± 4.5 ka, implying the lake was at the sample elevation around 43 ka.

Site O24

The shoreline at this site is also very high, but below Site O23 along the eastern margin of Siling Co. Similar to Site O23, two sedimentary facies units are observed in this ~ 2.5 m deep soil pit (Figure B-17). In general, the entire pit shows a reddish color at the top and brownish color at the bottom, and the sediment textures are a little better to observe than Site O23, but still not as clear as sites in the central peninsula. Unit 1 at the bottom is a very thick pile relatively well-sorted and rounded gravels, which are mostly grain-supported, but with a proportion of sand matrix. A thin layer of fine sands is intercalated with this unit. Unit 2 is a ~ 40 cm thick layer of mud and silt sands.

I interpret Unit 1 of thick, relatively well-sorted, rounded gravels as beach ridge deposits; and Unit 2 with a mixture of mud and silt sands as a soil layer. In general, the reddish color at the top of this pit and the sediment textures also suggest a large degree, but less than Site O23, of shoreline degradation at this site.

The OSL age of O24 from the fine sands in the upper part of Unit 1 is 39.3 ± 12.2 ka, implying the lake was at the sample elevation around 39 ka. Notice that the age uncertainty is too large, > 25% of the age. Therefore I do not use this age when interpreting the lake history of Siling Co.

Others

Site O45
This site is a shoreline spit close to the Holocene lake highstand and is located in the central islands of Siling Co. Three sedimentary facies units are observed in this ~ 2.5 m deep soil pit at Site O45 (Figure B-18). Unit 1 at the bottom is a very thick layer of relatively poor-sorted gravels with mud and sands as matrix. A thin lens of coarse sands is intercalated with this unit. Unit 2 is a very thick layer that consists of multiple alternating gravel sublayers of different grain size. The gravels in some sublayers are not well-sorted, but they are clean and rounded, mostly grain-supported. Some layering of the gravels can be observed from this unit. Unit 3 is a ~ 20 cm thick layer mixed with mud and silt sands.

I interpret Unit 1 of relatively poor-sorted gravels with mud and sands as matrix as alluvial deposits atop the paleo-lake level at time of deposition of these sediments. Unit 2 that contains clean grain-supported, rounded gravels is considered as beach ridge deposits. And Unit 3 with a mixture of mud and silt sands is interpreted a soil layer.

The OSL age of O45 from the sand lens in Unit 1 is 26.4 ± 1.3 ka, suggesting the lake level was below the sample elevation around 26 ka.

Site O35

This is another site located in the central islands of Siling Co, but above the Holocene lake highstand. Three sedimentary facies units are observed in this ~ 2.5 m deep soil pit at Site O35 (Figure B-19). Unit 1 at the bottom is a very thick layer of relatively poor-sorted gravels with mud and sands as matrix. A thin lens of coarse sands is intercalated with this unit. Unit 2 is a very thick layer that consists of multiple alternating gravel sublayers of different grain size. The gravels in some sublayers are not well-sorted, but they are clean and rounded, mostly grain-supported. Unit 3 is a ~ 30 cm thick layer mixed with mud and silt sands.

I interpret Unit 1 of poorly sorted gravels with matrix of mud and sands as alluvial deposits. Therefore the lake level was below this unit at time of deposition of the alluvial sediments. Unit 2 of clean and grain-supported gravels are considered as beach ridge deposits. And Unit 3 with a mixture of mud and silt sands is considered as a soil layer.
The OSL age of O35 from the sand lens in Unit 1 is $29.4 \pm 1.7$ ka, suggesting the lake level was below the sample elevation around 29 ka.

**Site O22**

This shoreline site is below the Holocene lake highstand. Four sedimentary facies units are observed in this ~ 2.5 m high outcrop (Figure B-20). Unit 1 at the bottom is a very thick pile of laminated fine sands. Unit 2 consists of two sublayers of red paleosol at the top and bottom of this unit and poorly-sorted, matrix-supported angular gravels. Unit 3 contains relatively well-sorted grain-supported, rounded gravels. Unit 4 is a layer mixed with mud and silt sands.

I interpret Unit 1 of thick pile of laminated fine sands as deepwater lacustrine sands. Therefore the lake level was much higher than this site at time of deposition of these sands. Unit 2 with poorly sorted angular gravels with mud matrix is considered as alluvial fan deposits or soils. Unit 3 of well sorted and grain-supported gravels is typical of beach ridge deposits. And Unit 4 with a mixture of mud and silt sands is considered as a soil layer.

The OSL age of O22 from the sands in Unit 1 is $19.9 \pm 1.1$ ka, suggesting the lake level was way above the sample elevation around 20 ka.

**Site O4**

This site is on the land side slope of the highest shoreline (~ 4636 m) above the Holocene lake highstand in the northern peninsula of Siling Co. Here I basically observed one sedimentary facies unit dipping to the land (Figure B-21). This unit is characterized by the interfingerring of medium sands and poorly-sorted sub-angular gravels. Thin cross-bedding is also found within the sand body.

Based on these observations, this unit probably reflects the deposition on the steep wall of a braided stream. The OSL age of sample O4 from the cross-bedded sands is $55.8 \pm 1.9$ ka, suggesting the lake level was below this unit around 56 ka.
Site O5

This site is close to but below Site O4 in the northern peninsula of Siling Co. The stream-cut exposure of the shoreline strata is very high, but basically I observed one sedimentary facies unit (Figure B-22). This unit is characterized by interbedded layers of relatively well-sorted, mostly grain-supported, rounded gravels. There are some thin layers of medium-fine sands intercalated with this unit.

Based on these observations, this unit is interpreted as beach ridge deposits. The OSL age of sample O5 from the sand layers is 34.5 ± 3.1 ka, suggesting the lake level was at the sample elevation around ~ 35 ka.

Site O3

Two sedimentary facies units are observed at Site O3 (Figure B-23). Unit 1 is a thick layer of medium sands. Its contact with the unit below is not shown here. Unit 2 at the top is a layer relatively well-sorted, mostly grain-supported rounded gravels, with some sands as matrix.

The thick layer of medium sands in Unit 1 possibly reflect a nearshore environment below the wave influence. And Unit 2 with matrix-supported gravels are considered as possible alluvial sediments. The OSL sample, O3, from Unit 1 yields an age of 40.5 ± 8.2 ka, suggesting the lake level was above this unit around 41 ka.

Site O2

This site is located several meters below the highstand shoreline surface in northern peninsula of Siling Co. Three sedimentary facies units are observed at Site O2 (Figure B-24). Unit 1 is a very thick pile of medium sands. Unit 2 contains a thick layer of relatively well-sorted, mostly grain-supported, and rounded gravels with layering dipping to the land. Unit 3 is the same as Unit 2, except the gravel layering dips to the lake.

The thick pile of medium sands in Unit 1 possibly reflect the nearshore environment below the wave influence. And the landward-dipping gravel layers in Unit 2 and lakeward-dipping gravel layers in
Unit 3 are considered as the backset and foreset of beach ridges, respectively. The OSL sample, O2, from Unit 1 yields an age of $17.7 \pm 1.3$ ka, suggesting the lake level was above the sample elevation around 18 ka.
B-3 References


Figure B-1. The soil pit profile and graphic log and interpretation of depositional environments at Site O17 in the Shoreline Group 2 (G2) (see detailed descriptions of the stratigraphy in the text).
Figure B-2. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O44 in the Shoreline Group 1e (G1e) (see detailed descriptions of the stratigraphy in the text).
Figure B-3. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O36 in the Shoreline Group 1d (G1d) (see detailed descriptions of the stratigraphy in the text).
Figure B- 4. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O38 in the Shoreline Group 1c (G1c) (see detailed descriptions of the stratigraphy in the text). Here two OSL samples (O38-O39) were collected from two different sedimentary facies units.
Figure B-5. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O42 in the Shoreline Group 1b (G1b) (see detailed descriptions of the stratigraphy in the text).
Figure B-6. The photo and sketch of shoreline stratigraphy and interpretation of depositional environments at Site O7 in the Shoreline Group 1a (G1a) (see detailed descriptions of the stratigraphy in the text).
Figure B-7. The soil pit profile and graphic log and interpretation of depositional environments at Site O20 in the G3 highstand shoreline complex (see detailed descriptions of the stratigraphy in the text).
Figure B-8. The soil pit profile and graphic log and interpretation of depositional environments at Site O19 in the G3 highstand shoreline complex (see detailed descriptions of the stratigraphy in the text).
Figure B-9. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O13 in the Shoreline Group 2 (G2) (see detailed descriptions of the stratigraphy in the text).
Figure B-10. The photos and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O37 in the Shoreline Group 1d (G1d) (see detailed descriptions of the stratigraphy in the text).
Figure B- 11. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O11 in the Shoreline Group 1c (G1c) (see detailed descriptions of the stratigraphy in the text).
Figure B-12. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O9 in the Shoreline Group 1b (G1b) (see detailed descriptions of the stratigraphy in the text).
Figure B-13. The photo and sketch of shoreline stratigraphy cut by the Tashi Stream, and interpretation of depositional environments at Site O8 in the Shoreline Group 1b (G1b) (see detailed descriptions of the stratigraphy in the text).
Figure B-14. The soil pit profile and graphic log and interpretation of depositional environments at Site O16 in the Shoreline Group 1a (G1a) (see detailed descriptions of the stratigraphy in the text).
Figure B-15. The photo and sketch of a road-cut shoreline stratigraphy, and interpretation of depositional environments at Site O27 above the highstand in the easternmost margin of Siling Co.
Figure B-16. The soil pit profile and graphic log and interpretation of depositional environments at Site O23 above the highstand along the eastern margin of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-17. The soil pit profile and graphic log and interpretation of depositional environments at Site O24 above the highstand along the eastern margin of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-18. The soil pit profile and graphic log and interpretation of depositional environments at Site O45 at the highstand in the central islands of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-19. The soil pit profile and graphic log and interpretation of depositional environments at Site O35 above the highstand in the central islands of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B- 20. The photo and graphic log of the shoreline stratigraphy, and interpretation of depositional environments at a quarrying site, O22, below the highstand in the central island of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-21. The photo and sketch of a stream-cut shoreline stratigraphy, and interpretation of depositional environments at Site O4 above the highstand in the northern peninsula of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-22. The photo and sketch of a stream-cut shoreline stratigraphy, and interpretation of depositional environments at Site O5 above the highstand in the northern peninsula of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-23. The soil pit profile and graphic log and interpretation of depositional environments at Site O3 above the highstand in the northern peninsula of Siling Co (see detailed descriptions of the stratigraphy in the text).
Figure B-24. The soil pit profile and graphic log and interpretation of depositional environments at Site O2 at the highstand in the northern peninsula of Siling Co (see detailed descriptions of the stratigraphy in the text).
### Table B-1. Quartz SAR protocol for all Tibet samples

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Table B-2. Dosimetry data of the OSL samples collected from around Siling Co.

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<th>Error (ppm)</th>
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Table B-2 (cont.). Dosimetry data of the OSL samples collected from around Siling Co.

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<th>Error (ppm)</th>
<th>Th (ppm)</th>
<th>Error (ppm)</th>
<th>K (%)</th>
<th>Error (%)</th>
<th>Rb (ppm)</th>
<th>Error (ppm)</th>
<th>H$_2$O (wt. %)</th>
<th>Error (wt. %)</th>
<th>Cosmic (mGya$^{-1}$)</th>
<th>Error (10%)</th>
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<tr>
<td>XS-SL-OSL-29</td>
<td>2.20</td>
<td>0.07</td>
<td>5.20</td>
<td>0.16</td>
<td>0.66</td>
<td>0.02</td>
<td>49.80</td>
<td>4.98</td>
<td>7.7</td>
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<td>0.04</td>
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<td>XS-SL-OSL-30</td>
<td>2.20</td>
<td>0.07</td>
<td>2.60</td>
<td>0.08</td>
<td>0.49</td>
<td>0.01</td>
<td>33.80</td>
<td>3.38</td>
<td>1.9</td>
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<td>0.07</td>
<td>9.54</td>
<td>0.29</td>
<td>1.58</td>
<td>0.05</td>
<td>105.14</td>
<td>10.51</td>
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<tr>
<td>XS-SL-OSL-36</td>
<td>2.70</td>
<td>0.08</td>
<td>3.40</td>
<td>0.10</td>
<td>0.28</td>
<td>0.01</td>
<td>18.20</td>
<td>1.82</td>
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<td>0.08</td>
<td>9.60</td>
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<td>1.96</td>
<td>0.06</td>
<td>114.70</td>
<td>11.47</td>
<td>2.6</td>
<td>0.1</td>
<td>0.44</td>
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<td>10.30</td>
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<td>0.08</td>
<td>9.00</td>
<td>0.27</td>
<td>1.58</td>
<td>0.05</td>
<td>104.90</td>
<td>10.49</td>
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<td>0.5</td>
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<td>11.90</td>
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<td>131.30</td>
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<td>119.70</td>
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<td>XS-SL-OSL-42</td>
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<td>0.06</td>
<td>0.25</td>
<td>0.01</td>
<td>15.40</td>
<td>1.54</td>
<td>11.5</td>
<td>0.6</td>
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<tr>
<td>XS-SL-OSL-43</td>
<td>3.00</td>
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<td>13.10</td>
<td>0.39</td>
<td>2.00</td>
<td>0.06</td>
<td>123.20</td>
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<td>0.3</td>
<td>0.12</td>
<td>0.01</td>
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<td>2.30</td>
<td>0.07</td>
<td>4.50</td>
<td>0.14</td>
<td>0.44</td>
<td>0.01</td>
<td>30.10</td>
<td>3.01</td>
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<td>0.12</td>
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<td>XS-SL-OSL-45</td>
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<td>Sample ID</td>
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<td>Depositional Environment</td>
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<tr>
<td><strong>Central Peninsula</strong></td>
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<tr>
<td>XS-SL-OSL-14</td>
<td>Thin medium sand lenses intercalated with a thick pile of poorly-sorted alluvial sediments; above is the thick pile of well-sorted clean beach gravel layers, and soil layers at the surface.</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
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<tr>
<td>XS-SL-OSL-20</td>
<td>Thin fine sand lenses intercalated with a thick pile of poorly-sorted alluvial sediments; above is the thick pile of well-sorted beach gravel layers, and soil layers near the surface.</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
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<tr>
<td>XS-SL-OSL-19</td>
<td>A thick pile of well-sorted fine sands below beach gravel layers</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
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<tr>
<td>XS-SL-OSL-18</td>
<td>A thin sand lenses intercalated with the thick pile of well-sorted clean beach gravels, above is a soil layer at the surface</td>
<td>Beach ridge</td>
<td>At lake level</td>
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</tr>
<tr>
<td>XS-SL-OSL-17</td>
<td>A medium thick layer of intercalated with a very thick pile of well-sorted layers of beach gravels</td>
<td>Beach ridge</td>
<td>At lake level</td>
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<tr>
<td><strong>Tashi Stream</strong></td>
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<tr>
<td>XS-SL-OSL-13</td>
<td>thick massive coarse to silt sand body wedged into poorly-sorted boulders and angular gravels; above is a very thick bed of well-sorted clean beach gravels that are inclined towards the lake</td>
<td>Debris flow</td>
<td>Above lake level</td>
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</tr>
<tr>
<td>XS-SL-OSL-43</td>
<td>A thin to medium thick layer of fine to silt sand layer intercalated with poorly-sorted angular gravels, with a few boulders present in the deposits; above are three layers of alluvial soils and fan gravels</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
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<tr>
<td>XS-SL-OSL-44</td>
<td>A thin layer of medium to coarse sands intercalated with well-sorted, clean beach gravels near the top of the site; above is poorly-sorted fan gravels</td>
<td>Beach ridge</td>
<td>At lake level</td>
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<tr>
<td>XS-SL-OSL-30</td>
<td>A thin layer of fine sand intercalated with reddish, weathered, and poorly-sorted gravels</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
<td></td>
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</tr>
<tr>
<td>XS-SL-OSL-28</td>
<td>Massive medium sand body between the backsets of a beach ridge below and foresets of a beach ridge above; a scoured erosional surface is developed at the bottom of the sand body, a few coarser gravels are present in the sand body</td>
<td>Lagoon</td>
<td>Above lake level</td>
<td></td>
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</tr>
<tr>
<td>XS-SL-OSL-29</td>
<td>A thin layer of well-sorted medium lower sands deposited on a very gentle slope, above is a medium thick layer of well-sorted clean beach gravels</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
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</table>
Table B-3 (cont.). Description of sediment facies, depositional environments and water depths of the OSL samples

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Description</th>
<th>Depositional Environment</th>
<th>Water Depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tashi Stream</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>XS-SL-OSL-12</td>
<td>A thin to medium thick layer of silt sands deposited within a wide U-shaped channel; symmetrical ripple marks are developed in the sand layer, above which is another thin medium layer of relatively well-sorted beach gravels in the channel</td>
<td>Near shore (wave influenced)</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-36</td>
<td>A thin layer of coarse sands intercalated with well-sorted clean beach gravels; developed on the shoreface of the beach ridge</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-37</td>
<td>Thick massive medium to coarse sand body wedged into poorly-sorted cobbles and boulders</td>
<td>Debris flow</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-11</td>
<td>V-shaped silt sand bodies wedged into relatively well-sorted beach gravels; above is a very thick pile of well-sorted clean beach gravel layers</td>
<td>Alluvial fan (ice wedge casts)</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-38</td>
<td>Thin lenses of fine sands intercalated with relatively stratified gravels with mud and sands as matrix</td>
<td>Alluvial fan (possible)</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-39</td>
<td>A thin layer of fine sands developed on the slope of backsets of the beach gravels, and below foresets of another beach gravels</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-40</td>
<td>A thick lenses of fine-silt sand body wedged into very poorly-sorted sediments of gravels to boulders</td>
<td>Debris flow</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-41</td>
<td>A thin layer of fine lower sands intercalated with thick pile of beach gravels above and below</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-10</td>
<td>A thin layer of fine-silt sands intercalated with relatively stratified gravels with mud and sands as matrix</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-42</td>
<td>A thin layer of coarse sands intercalated with well-sorted clean beach gravels; developed on the shoreface of the beach ridge</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-09</td>
<td>A medium thick layer of silt sands below a thick pile of well-sorted beach gravels</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-08</td>
<td>A medium thick layer of medium sands below a very thick pile of well-sorted beach gravels</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-07</td>
<td>A medium thick lenses of fine-silt sands wedged into poorly-sorted layers of gravels to boulders, with sands and mud as matrix</td>
<td>Debris flow</td>
<td>Above lake level</td>
</tr>
</tbody>
</table>
### Table B-3 (cont.). Description of sediment facies, depositional environments and water depths of the OSL samples

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Description</th>
<th>Depositional Environment</th>
<th>Water Depth</th>
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</thead>
<tbody>
<tr>
<td><strong>Tashi Stream</strong></td>
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<tr>
<td>XS-SL-OSL-16</td>
<td>A thin layer of fine sands intercalated with poorly-sorted, angular gravel layers, with a few boulders present in the sediments</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-15</td>
<td>A thin layer of fine sands intercalated with poorly-sorted, angular gravel layers, with a few boulders present in the sediments</td>
<td>Alluvial fan</td>
<td>Above lake level</td>
</tr>
<tr>
<td><strong>East margin</strong></td>
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</tr>
<tr>
<td>XS-SL-OSL-27</td>
<td>A thick pile of thickly laminated fine sands below a thick pile of well-sorted beach gravels</td>
<td>Lacustrine</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-23</td>
<td>A thin layer of medium to fine sands intercalated with a thick pile of weathered beach gravels</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-24</td>
<td>A thin layer of medium to fine sands intercalated with a thick pile of weathered beach gravels</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-26</td>
<td>A very thick pile of laminated medium-fine sands, with crossed-bedding and climbing ripple marks developed near the top of this thick sand body; above is a very thick pile of lakeward inclined sandy gravels</td>
<td>Near shore</td>
<td>Below lake level</td>
</tr>
<tr>
<td><strong>Others</strong></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>XS-SL-OSL-35</td>
<td>A thin lenses of coarse sands intercalated with a thick pile of poorly-sorted gravels and cobbles, with sands and mud as matrix</td>
<td>Alluvial fan</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-22</td>
<td>A thick pile of fine sands below thick layers of beach gravels and paleosol</td>
<td>Lacustrine</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-33</td>
<td>A medium thick layer of fine sands intercalated with a thick pile of poorly-sorted beach gravels, supported by sand matrix</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-04</td>
<td>A thick pile of medium sands with cross-bedding, steeply (&gt; 20°) interfingered with poorly-sorted sub-angular gravels</td>
<td>Braided stream?</td>
<td>Above lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-05</td>
<td>A thin layer of fine sands intercalated with a thick pile of relatively well sorted beach sands and gravels</td>
<td>Beach ridge</td>
<td>At lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-03</td>
<td>A thick pile of medium sands below a layer of poorly-sorted matrix-supported gravels</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
</tr>
<tr>
<td>XS-SL-OSL-02</td>
<td>A very thick pile of medium sands, below a thick pile of relatively well-sorted beach sands and gravels</td>
<td>Near shore (below wave influence)</td>
<td>Below lake level</td>
</tr>
</tbody>
</table>
APPENDIX C

Appendix C includes detailed description of methods of all analyses in this study (OSL; forward elastic modeling of crustal flexure; strength envelope analysis and shoreline survey) and associated figures and tables.

C-1 OSL dating

C-1.1 Field sampling

Given that no natural exposures of lacustrine stratigraphy were found along the highstand shoreline complex, depth profile pits (> 0.5 m – 2.5 m deep) were dug by hand to characterize the nature of the deposit and locate samples. All samples were carefully collected from sand layers/lenses that are intercalated within the upper beach gravel deposits (Figure C-1), such that the age of deposition is a reasonable measure of the timing of shoreline occupation. Eolian sands (younger than the age of shoreline deposition) at the top of the shoreline stratigraphy were avoided for sampling. Before collecting the sand samples, several centimeters of the exposure surface were removed to eliminate possible uncertainty from light exposure (Aitken, 1998). I sampled fine to medium sand, and in a few instances, coarse silt using plastic PVC (Polyvinyl chloride) and or steel tubes, 3 cm or 5 cm in diameter, depending on the consolidation and thickness of the sand layers. These tubes were then wrapped with heavy duty black duct tape to avoid light penetration and maintain sample water content.
C-1.2 Laboratory processing

Sample preparation

Samples were prepared using standard OSL procedures. Material was desiccated at 50 °C to enable calculation of water content, and then sieved to extract the 180-212 µm grain size fraction. Approximately 10 g of the 180-212 µm grain size fraction was treated with 30% HCl for 30 minutes to remove CaCO3. Samples were agitated throughout the treatment, and once complete, HCl was replaced with 30% H2O2 to remove organic material. The duration of H2O2 treatment varied between samples, dependent upon the amount of organic material present, and two H2O2 treatments were necessary for some samples with a high organic content. Once effervescence ceased, the H2O2 was decanted, the sample was washed four times with deionised water and desiccated at 50 °C. Quartz is extracted from polymineral sediment residues through density separations using LST fastfloat (sodium heteropolytungstate dissolved in deionized water). Heavy minerals (> 2.68 g cm-3) were separated from the lighter fraction, and the target 2.58-2.68 g cm-3 fraction was further separated from the < 2.58 g cm-3 material. The target fraction was washed five times with deionised water to ensure removal of all LST. Final separates were dried and etched with 40% HF for 40 minutes to remove any contaminating feldspar; all samples were agitated at 5 minute intervals throughout treatment. The etched quartz was treated with 30% HCl for 30 minutes to remove any carbonates produced during HF etching.

Luminescence measurements

All analyses were carried out using either a TL-DA-15 or TL-DA-20 Risø reader, equipped with an EMI 9235QA photomultiplier and 7.5 mm Hoya U-340 filter. Blue (470±20 nm) and infrared (~870 nm) diodes operated at 90% and 40% power respectively, were used for stimulation and irradiation was achieved using a 90Sr/90Y beta source. Readers were calibrated using quartz prepared at the Risø National Laboratory in Denmark. Quartz was applied to stainless steel discs (10 mm Ø, 1 mm thick) using silicon grease and the aliquot size was regulated using a small (2 mm Ø, ~35 grain) mask.
Samples were analysed using the single aliquot regenerative dose (SAR) protocol (Murray and Wintle, 2000). The equivalent dose (De) was calculated from measurements of the luminescence response following stimulation of the natural luminescence (Ln) and a series of different regenerative doses (Lx). The Ln and Lx measurements are normalized by measurement of the luminescence response (Tx) to a constant test dose (TD). The ratio Lx/Tx is used to compensate for sensitivity changes of the quartz throughout analysis and Lx/Tx measurements are used to obtain a range of values which bracket Ln/Tx, allowing De interpolation with minimal associated errors (Banerjee et al., 2000).

The sample is heated prior to making the luminescence measurements in order to reduce the contribution of luminescence from unstable trap, which cause erroneous dose determinations. The temperature of the pre-Lx pre-heat (PH1) and the pre-Tx pre-heat (PH2) must be empirically determined for each sample under analysis. This was achieved through analysis of 8 disks with a dose-recovery pre-heat plateau experiment (Murray and Wintle, 2003), using a range of PH1 and PH2 temperatures ranging from 150 – 220 °C. The influence of a hot-bleach at 280 °C was also investigated (Murray and Wintle, 2003) but was not required to counter any thermal transfer or recuperation.

The SAR protocol with a PH1 of 200 °C used on most of the samples is provided in Table C-1. Aliquot acceptance criteria used are 1) recycling ratios within 10% of unity; 2) signal intensities ≥ 3 σ above background; 3) infra-red (IR) depletion ratio within 20% of unity (Duller, 2003); 4) De uncertainty ≤ 20 % and 5) recuperation within 10% of the normalised maximum dose. The acceptance thresholds are generally very high for the samples (85 – 100%) reflecting the sensitivity of the quartz analysed. All samples have at least 50 accepted aliquots.

**Environmental Dose rate determination**

The environmental dose rates (Dr) were calculated for each sample from the unsieved portions of the original sample; concentrations of U, Th, K and Rb were measured directly using solution ICP-MS (Thermo X-Series), a cosmic-dose component after Prescott and Hutton (1994) and an internal alpha dose rate of 5% from the decay of U and Th after Sutton and Zimmerman (1978). External α-dose rates were
ignored as the alpha irradiated portion of quartz grains was removed by etching. The conversion factors of Adamiec and Aitken (1998) and beta-particle attenuation factors after Mejdahl (1979) and Readhead (2002a; 2002b) have been used. Sample water content was calculated following desiccation at 50 °C, and an uncertainty of 5 % assumed. Table C-2 contains the dosimetry data of OSL samples from the highstand shorelines.

Analysis of results

Most samples are characterized by large overdispersion values (broad De distributions) which describe the spread in the data not accounted for by analytical uncertainties (Galbraith and Roberts, 2012). Samples with overdispersion, greater than 20%, are assumed to reflect heterogeneous bleaching before deposition and the De data were analyzed within Excel to calculate overdispersion and statistical parameters that are used to model the appropriate burial age (Db). I adopted the age model selection criteria of Arnold and Roberts (Arnold and Roberts, 2009). Most samples are modeled using the three component minimum age model (MAM-3, Galbraith et al., 1999)) and the RStudio Luminescence package (Kreutzer et al., 2012). The data are distributed and in multi-grain analyses the clear differentiation of separate populations is not straightforward (Arnold and Roberts, 2009), however, the finite mixture model (FMM) has been adopted in two cases where the MAM3 model fails to capture the lowest population of De values (Galbraith and Green, 1990). Arnold and Roberts (Arnold and Roberts, 2009) proposed that the FMM model should only be used on single grain age data because there is the possibility of selecting false populations when multi-grain aliquots are used. However, Rodnight et al (Rodnight et al., 2006) used the same model to calculate ages for fluvial sediments in South Africa based on agreement of FMM ages and independent radiocarbon ages. In the absence of secondary age dating, I use the FMM model for two ages when MAM3 fails to capture the lowest population and to explore the age of older populations which may have been reworked into younger shoreline deposits (OSL-21). The radial and kernel density plots for each sample are shown in Figure C-2.
C-2 Forward modeling to constrain the effective elastic thickness of Tibetan crust

I approach the determination of the effective elastic thickness ($T_e$) of Tibetan lithosphere by finding the best fit between the measured shoreline elevations and predicted deflections of the highstand shorelines in response to the lake unloading since the Early - Middle Holocene (Table 4-1 in the main document). I run a series of forward elastic models with a range of $T_e$.

As an initial approximation, I assume a uniform $T_e$ (with no spatial variation) of the crust beneath Siling Co. The general equation (refer to Eq. 3.52 in (Watts, 2001)) for the deflection of a thin elastic spherical shell can be written as

$$\nabla^4 w + \left( \frac{I}{\beta^4} \right) w = \frac{q}{D}, \quad (S-1)$$

Where $w$ is the radial displacement, $q$ is the load applied on the elastic plate, $D$ is the flexural rigidity, $\beta$ is the three-dimensional flexure parameter (refer to Eq. 2.6 in (Watts, 2001)) and $\nabla^4$ is the biharmonic operator in the surface coordinates of the shell.

I adopt the analytical solutions (refer to Eq. 3.54 and Eq. 3.55 in (Watts, 2001)) to the pseudo-3D equation of flexure of a thin elastic spherical shell under a disc-shaped load (Brotchie and Silvester, 1969; Watts, 2001), which are also written as below.

For a disc-shaped load with height of $h$, radius of $R_d$, and density of load $\rho_{\text{load}}$ (here is the water density), the deflection beneath the disc load is

$$w = \frac{h \rho_{\text{load}}}{(\rho_m - \rho_{\text{inf ill}})} \left[ \left( \frac{R_d}{\beta} \right) \text{Ber} \left( \frac{r}{\beta} \right) - \left( \frac{R_d}{\beta} \right) \text{Kei} \left( \frac{r}{\beta} \right) \text{Bei} \left( \frac{r}{\beta} \right) + 1 \right] \quad (\text{for } r \leq R_d), \quad (S-2)$$

Outside the disc load, the deflection is

$$w = \frac{h \rho_{\text{load}}}{(\rho_m - \rho_{\text{inf ill}})} \left[ \left( \frac{R_d}{\beta} \right) \text{Ber} \left( \frac{r}{\beta} \right) \text{Ker} \left( \frac{r}{\beta} \right) - \left( \frac{R_d}{\beta} \right) \text{Kei} \left( \frac{r}{\beta} \right) \text{Bei} \left( \frac{r}{\beta} \right) + 1 \right] \quad (\text{for } r > R_d), \quad (S-3)$$

In both equations S-2 and S-3, $\rho_{\text{inf ill}}$ is the density of the material filled in the displaced space (here is the air density). $\text{Kei}$ and $\text{Ker}$ represent the Kelvin function of zero order and $\text{Kei}'$ and $\text{Ker}'$
represent the Kelvin function of first order. \( Bei \) and \( Ber \) represent the Bessel function of zero order and \( Bei' \) and \( Ber' \) represent the Bessel function of first order.

The computation scheme follows the method of Nakiboglue and Lambeck (1983). In the scheme, the water load has been discretized to 1-km-diameter cylinders with varying heights, and the total flexure of the lithosphere is calculated by sum of the flexure in response to loading of each unit cylinder. However, the discretization of the water load as unit cylinders leads to a geometric deficit of mass (with a factor of \( 4/\pi \)) from unit cuboid load (area = 1 km\(^2\)) of the same height as the cylinders. The calculation is described as:

\[
M_{cylinder} = \pi \left( \frac{L}{2} \right)^2 h = \frac{\pi}{4} (L^2 h) = \frac{\pi}{4} M_{square}, \quad (S-4)
\]

where \( M_{cylinder} \) is the mass of a unit cylinder, \( M_{square} \) the mass of a square cell and \( L \) the diameter of the cylinder or unit length of the cell, 1 km. Because of the beauty of symmetric flexure response under a cylinder load, such mass deficit is compensated by adjusting the original height \( h \) of the unit cylinder to \( (4/\pi) h \), such that the mass of the cylinder is equal to that of a cube.

I validated this discretization method for modeling the flexure using a large cylinder load with diameter of 33 km. The results of flexure curves from the analytical solution before and after discretizing the load into unit cylinders of 1 km \( \times \) 1 km in area are indistinguishable from one another. Therefore this flexural modeling by discretization of large loads is considered to be sufficiently accurate to model the flexural response under a load of much more complicated 3D geometry, such as the water load of Siling Co.

The water load that has been removed from the Tibetan crust during lake recession is defined as the volume between the paleo-highstand at the Early Holocene and modern topography and/or present lake level (Figure C-3). The paleo-highstand lake level is assumed to be close to elevation of the highstand shorelines at the margin of Siling Co (least deflected by the lithospheric flexure as a result of the lake unloading). I chose the lake levels in 1976 to represent modern levels because these are earliest records currently available. I reconstructed this load by: 1) using ArcGIS to extract the load from the 90
m digital elevation model (DEM) obtained from Shuttle Radar Topographic Mission (SRTM); 2) resampling the extracted load of 90 m into resolution of 1 km, and 3) correcting the load geometry in 1976 by comparison of the lake extent through time using the LandSAT imagery of Siling Co area obtained in 1976, 1999 and 2008 (see details in Meng et al (Meng et al., 2012a)). The resulted load has 303907 unit cells of 1 km × 1 km, ~ 304 km³ in water volume. The maximum load height for Siling Co is 64 m, Bange Co 73 m, Co E 37 m and Wuru Co 44 m (Figure C-3).

I modeled the crustal flexure in response to lake unloading with a wide range of effective elastic thickness (Tₑ) from 500 m to 30 km, trying to find a Tₑ at which the magnitude of shoreline deflections exhibits the same range of elevation variation as my observations. Other input model parameters are shown in the Table C-3. The deflection pattern, as seen in Figure C-4, mimics the shape of the load geometry when Tₑ is very small (e.g., Tₑ = 2 km); when Tₑ increases (e.g., Tₑ = 13 km), the magnitude of deflection decreases and the pattern becomes less sensitive to the details of the load geometry.

The model results of predicted flexure of lithosphere are compared with observed elevations of the highstand shoreline at a specific site (Bills et al., 2007), instead of along a transect that involves several surveyed highstand shorelines. The reason to do this is that the highstand shorelines have been only preserved at limited locations and are not distributed exactly along a specific transect. However, deflections of shorelines at different locations, even within short distances, can be significantly different due to the complex geometry of the lake load (Figure C-3). By comparing observed and model-predicted shoreline deflections at specific sites, ideally I expect to see that the data points in the plots of shoreline elevation vs. predicted deflection exhibit a 1:1 correlation (Bills et al., 2007). This line may not intercept with the origin of the coordinate in that the datum of zero deflection is not known a priori. The results of predicted shoreline deflections under a wide range of Tₑ vs. observed shoreline elevations are shown in Figure C-5.
C-3 Strength envelope analysis and viscosity structure of Tibetan crust

Strength envelopes (Goetze and Evans, 1979; Brace and Kohlstedt, 1980; Watts, 2001) of central Tibetan lithosphere can be constructed based on current understanding of the crustal composition and temperature structures, the elastic core thickness which corresponds to the effective elastic thickness of a specific crustal layer and the strain rate. Once all these geophysical observations are compatible with all other observations, the temperature structure, together with the strain rate and understanding of crustal compositions, can be used to construct the depth-dependent viscosity profile of the lithosphere on the timescale of the lake unloading of Siling Co.

In this study, I use the strain rate and effective elastic thickness (corresponding to thickness of the elastic core of the upper crust beneath central Tibet, see details below), as fixed constraints to construct strength envelopes of central Tibetan lithosphere with inputs of thermal parameters. The strength limit of the brittle part of the lithosphere is defined by the Byerlee’s law (Byerlee, 1978; Brace and Kohlstedt, 1980). Here I use a crustal density of $\rho = 2700$ kg/m$^3$ in calculation the frictional strength in the brittle uppermost crust. However, as shown in Table C-4, the maximum stress due to the lake unloading is less than 4 MPa, a number that may not exceed the strength limit in the brittle zone, therefore, my analysis does not depend on Byerlee’s frictional strength nor on an assumed seismogenic thickness.

Before constructing the strength envelopes of the ductile part of the lithosphere, several parameters and their ranges need to be defined, including 1) the top thickness of the elastic core, 2) the strain rate on the timescale of the lake loading of Siling Co, 3) the crustal compositional structure, and 4) value ranges of the thermal parameters.

Here I assume the top of this elastic layer is at the base of the sedimentary cover, $\sim 1 – 5$ km in the coverage of the INDEPTH III profile (Haines et al., 2003; Mechie et al., 2004; Ross et al., 2004), for two reasons. First, the sediments should be relatively anelastic at shallow depths. Second, my study area is located relatively far away from large active strike-slip fault systems which might act to reduce the
elasticity of upper crust. Considering the elastic core thickness of \( T_e = 12 – 14 \, \text{km} \), the base of the elastic core is \( \sim 13 – 19 \, \text{km} \).

The strain rate is estimated from the strain (on the order of \( 10^{-5} \)) that causes the maximum deflection (the maximum curvature of the flexure curves, on the order of \( 10^{-9} \, \text{m}^{-1} \)) and the timescale of lake unloading of Siling Co, \( \sim 6 – 4 \, \text{ka} \) (Table 4-1 in the manuscript). Calculation of strain and stress is based on Equations 3.70 and 3.64, respectively, of Turcotte and Schubert (Turcotte and Schubert, 2002). These parameters are listed in the Table C-4.

Following estimates of crustal compositions and layering in central Tibet (Min and Wu, 1987; Haines et al., 2003; Mechie et al., 2004; Zhang et al., 2011), here I assume the upper crust is composed primarily of felsic rocks and mafic rocks reside in the lower crust. Because of the heterogeneity of crustal composition, choosing creep parameters obtained from one type of rock/mineral in one location will lead to bias choice of the lithology. Instead, I selected several types of the quartz-rich or mafic rocks for upper and lower crust, respectively. These rocks/minerals were obtained from variable locations and show different flow law parameters obtained from several laboratories (Table C-5) therefore may better reflect the variability of the crustal strength and viscosity as a function of rock composition.

Among the selected quartz-rich felsic rocks, the wet Black Hills quartzite (Gleason and Tullis, 1995) and wet Simpson quartzite (Koch et al., 1989) are somewhat stronger than the wet Heavitree quartzite (Jaoul et al., 1984). Experiments using wet granite (Hansen and Carter, 1982) and partially molten granite (Rutter et al., 2006) lie somewhere in the middle of the range of experimental data. For mafic compositions, strength ranges from dry clinopyroxene (Bystricky and Mackwell, 2001), through dry Maryland and Columbia diabase (Mackwell et al., 1998), to Pikwitonei granulite (Wilks and Carter, 1990) and dry anorthite (Rybacki and Dresen, 2000). All creep parameters from these experiments are listed in Table C-5.

Finally, the thermal structure of the Tibetan lithosphere is estimated according to Equation 4.31 of Turcotte and Schubert (Turcotte and Schubert, 2002), with adjusted parameter symbols shown below (Table C-5).
\[ T(z) = T_0 + \frac{q_0 z}{k} - \frac{A_0 b^2}{k} \left( \frac{z}{b} + e^{-\frac{z}{b}} - 1 \right), \quad (S-5) \]

where the parameters are

- \( T(z) \) – temperature (K) as a function of depth \( z \) (km)
- \( T_0 \) – surface temperature of 0 °C in this study
- \( q_0 \) – surface heat flow (mW/m²)
- \( k \) – thermal conductivity (W/m/K)
- \( A_0 \) – heat production (µW/m³)
- \( b \) – depth scale of heat production (km)

I use a range of values for the thermal parameters (Table C-6) that are compatible with available observations. The value ranges are 2 – 3 µW/m³ for heat production \((A_0)\), 2 – 3 W/m/K for thermal conductivity \((k)\) and 80 – 110 mW/m² for surface heat flow \((q_0)\). For the depth scale \((b)\) for heat production which exponentially decreases with depth, \( b \) is 10 – 15 km for a normal continental crust of ~ 35 km thick, a \( b \) of ~ 30 km is reasonable for a total crustal thickness of 65 ± 5 km. In fact, Jiménez-Munt and Platt (Jiménez-Munt and Platt, 2006) used \( b \) of 55 km for crustal thickness larger than 40 km.

Similarly in this study, I use a range of 20 – 50 km for the total heat production depth, although all model fittings (see details below) to the elastic core thickness require only 20-35 km for \( b \), with relatively high thermal conductivity (2.5 – 3 W/m/K) and heat production (2.5 – 3 µW/m³).

The strength envelopes of central Tibetan lithosphere are then constructed to fit my estimate of the elastic core thickness of 12 – 14 km, with all physical parameters defined above. The criteria for such model fitting include two aspects. First, with adjusting inputs of variable thermal parameters, the resulted strength profiles for the upper crust need to mostly intersect the base of elastic layer defined by the maximum bending stresses (for specific \( T_e \)) induced by lake unloading of Siling Co (Fig. 4-6 in the main document). As mentioned above, the base of the elastic layer has a range of depth from 13 to 19 km, due to the unknown top of the elastic layer and the uncertainties of the elastic core thickness (effective elastic
thickness). Second, the resulted depth-dependent temperature profile need to be compatible with indirect estimates of temperatures at different levels of depth in the crust (Alsdorf and Nelson, 1999; Hacker et al., 2000; Mechie et al., 2004).

C-4 Viscosity structure of central Tibetan lithosphere

The depth-dependent profile of effective viscosity of the crust beneath central Tibet, on the time scale of the lake loading and unloading of Siling Co, is then constructed based on the strain rate and the thermal structure determined from above, using the equation below.

\[
\eta_{\text{eff}}(z) = \frac{\sigma}{2\dot{\varepsilon}} = \frac{1}{2} A \frac{Q}{n R T(z)} e^{\frac{Q}{n R T(z)}} \left( \frac{L}{n} \right),
\]

where \( \eta_{\text{eff}} \) and \( T \) are, respectively, effective viscosity and temperature as a function of depth \( z \); \( \sigma \) and \( \dot{\varepsilon} \) are the differential stress and strain rate, respectively; \( R \) is the gas constant and \( A, n, Q \) are rock material properties that vary with deformation mechanism.

The layering of the resulted effective viscosity is based on the previous geophysical observations of crustal structure (Zhao et al., 2001; Ross et al., 2004; Mechie et al., 2011) as listed in Table C-7.

C-5 Data of shorelines around Siling Co area

As investigation of the magnitude of deflection of the highstand shorelines around Siling Co region constitutes a major part of this study, correlation and DGPS (differential global positioning system) survey by of these shorelines have been carefully conducted. The methods of DGPS survey and results of shoreline survey data have been documented in detail in Meng et al. (2012b) and are translated in Tables C-8 ~ C-10.
C-6 References


Figure C-1. An example of stratigraphic context of highstand shoreline and OSL sampling site. (a) The photo showing the layering of beach gravels in a depth profile pit for OSL sample XS-SL-O25. Notice that the beach deposits are grain-supported, clean, and well-rounded and sorted sands/pebbles/gravels. The layers with variable grain size are interbedded with each other. (b) Stratigraphic log of the pit. Relative locations of six layers (A-F) of gravels are shown in (a). Note that the OSL sample is collected from the sand layer within the top beach gravel deposits. (c) A closer look at the OSL sample locations and the sediments.
Figure C-2. Radial plots showing the equivalent dose of OSL sample O1A (left) and O1B (right). The age calculation is also shown in the plots.
Figure C-2 (cont.). Radial plots showing the equivalent dose of OSL sample O6 (left) and O21 (right). The age calculation is also shown in the plots.
Figure C-2 (cont.). Radial plots showing the equivalent dose of OSL sample O25 (left) and O31 (right). The age calculation is also shown in the plots.
Figure C-2 (cont.). Radial plots showing the equivalent dose of OSL sample O32 (left) and O33 (right). The age calculation is also shown in the plots.
Figure C-2 (cont.). Radial plots showing the equivalent dose of OSL sample O4 (left) and O46 (right). The age calculation is also shown in the plots.
Figure C-3. The geometry of the water load of Siling Co used in the forward elastic modeling of the crust deflection. This water load is defined between the paleo-highstand shoreline and lake level in 1976 (lower panel), which has been removed from the lake basin since the lake recession from the highstand during early Holocene. The map view of the water load is shown in the upper panel.
Figure C-4. Model-predicted crustal deflection in response to the lake unloading of Siling Co for the same load in Figure C-3 in the manuscript. As shown in the figures, if the crust is very thin (e.g., $T_e = 2$ km), the deflection pattern mimics the load geometry and the maximum deflection reaches up to $> 22$ m; in contrast, if the crust is a little thicker (e.g., $T_e = 13$ km), the deflection pattern soothes out and is independent of the load geometry, with the expense of reduced maximum crust deflection ($< 10$ m). Filled circles denote four categories of highstand shorelines: red for those continuous correlated around the main body of paleo-Siling Co, orange and white for those discontinuously correlated around Bange Co and Wuru Co, respectively. And dark green color shows the shorelines below the highstand. The black polygons show the lake levels in 1976 and the white polygons reflect the paleo-Siling Co defined by the highstand shorelines.
Figure C-5. Comparison of observed shoreline elevations and model-predicted shoreline deflections for highstand constructional shorelines around the main body of paleo-Siling Co. Different panels represent the results from a series of effective elastic thickness of the crust (Te = 7 and 10 km). The filled circles represent depositional shorelines that are at or immediately above the still water level (e.g., beach ridges, cuspate bars); while the open circles reflect those shorelines more likely below the water level during the highstand (e.g., offshore parts of spits and tombolos). Clearly, most beach ridges and cuspate bars (filled circles) are distributed within a narrower elevation range (~2 m).
Figure C-5 (cont.). Comparison of observed shoreline elevations and model-predicted shoreline deflections for highstand constructional shorelines around the main body of paleo-Siling Co. Different panels represent the results from a series of effective elastic thickness of the crust (Te = 11 and 12 km). The filled circles represent depositional shorelines that are at or immediately above the still water level (e.g., beach ridges, cuspate bars); while the open circles reflect those shorelines more likely below the water level during the highstand (e.g., offshore parts of spits and tombolos). Clearly, most beach ridges and cuspate bars (filled circles) are distributed within a narrower elevation range (~ 2 m).
Figure C-5 (cont.). Comparison of observed shoreline elevations and model-predicted shoreline deflections for highstand constructional shorelines around the main body of paleo-Siling Co. Different panels represent the results from a series of effective elastic thickness of the crust (Te = 13 km). The filled circles represent depositional shorelines that are at or immediately above the still water level (e.g., beach ridges, cuspate bars); while the open circles reflect those shorelines more likely below the water level during the highstand (e.g., offshore parts of spits and tombolos). Clearly, most beach ridges and cuspate bars (filled circles) are distributed within a narrower elevation range (~ 2 m).
Figure C-5 (cont.). Comparison of observed shoreline elevations and model-predicted shoreline deflections for highstand constructional shorelines around the main body of paleo-Siling Co. Different panels represent the results from a series of effective elastic thickness of the crust (Te = 14 and 15 km). The filled circles represent depositional shorelines that are at or immediately above the still water level (e.g., beach ridges, cuspate bars); while the open circles reflect those shorelines more likely below the water level during the highstand (e.g., offshore parts of spits and tombolos). Clearly, most beach ridges and cuspate bars (filled circles) are distributed within a narrower elevation range (~ 2 m).
Figure C-5 (cont.). Comparison of observed shoreline elevations and model-predicted shoreline deflections for highstand constructional shorelines around the main body of paleo-Siling Co. Different panels represent the results from a series of effective elastic thickness of the crust (Te = 17 and 20 km). The filled circles represent depositional shorelines that are at or immediately above the still water level (e.g., beach ridges, cuspate bars); while the open circles reflect those shorelines more likely below the water level during the highstand (e.g., offshore parts of spits and tombolos). Clearly, most beach ridges and cuspate bars (filled circles) are distributed within a narrower elevation range (~ 2 m).
Table C-1. Quartz SAR protocol for all Tibet samples

<table>
<thead>
<tr>
<th>Natural/Regenerative Dose</th>
<th>TL (PH1)</th>
<th>IRSL (final cycle only)</th>
<th>OSL</th>
<th>Test Dose</th>
<th>TH (PH2)</th>
<th>OSL</th>
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</thead>
<tbody>
<tr>
<td>5, 10, 20, 30, 0, 5, 5 Gy</td>
<td>200˚C, 10 s, 5˚C/s</td>
<td>20˚C, 40 s, 5˚C/s</td>
<td>125˚C, 40 s, 5˚C/s, 90% power (Lx)</td>
<td>5 Gy</td>
<td>180˚C, 10 s, 5˚C/s</td>
<td>125˚C, 40 s, 5˚C/s, 90% (Tx)</td>
</tr>
</tbody>
</table>

Table C-2. Dosimetry data of OSL samples from highstand shorelines

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>U (ppm)</th>
<th>Error (ppm)</th>
<th>Th (ppm)</th>
<th>Error (ppm)</th>
<th>K (%)</th>
<th>Error (%)</th>
<th>Rb (ppm)</th>
<th>Error (ppm)</th>
<th>H₂O (wt. %)</th>
<th>Error (wt. %)</th>
<th>Cosmic (mGya⁻¹) (10%)</th>
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</thead>
<tbody>
<tr>
<td>XS-SL-O1A</td>
<td>1.74</td>
<td>0.052</td>
<td>7.55</td>
<td>0.227</td>
<td>1.31</td>
<td>0.039</td>
<td>79.35</td>
<td>2.38</td>
<td>10.84</td>
<td>5.00</td>
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<td>0.05</td>
<td>7.36</td>
<td>0.221</td>
<td>1.33</td>
<td>0.04</td>
<td>79.51</td>
<td>2.39</td>
<td>10.39</td>
<td>5.00</td>
<td>0.40</td>
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<td>0.14</td>
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<td>4.02</td>
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<td>0.31</td>
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<td>0.06</td>
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<td>5.00</td>
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<td>0.31</td>
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<td>0.05</td>
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<td>1.13</td>
<td>0.03</td>
<td>87.18</td>
<td>8.72</td>
<td>0.84</td>
<td>0.77</td>
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<td>Parameter</td>
<td>Symbol</td>
<td>Value</td>
<td>Unit</td>
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</tr>
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<td>Gravity acceleration rate</td>
<td>g</td>
<td>9.8</td>
<td>m/s²</td>
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<td>Young’s modulus</td>
<td>E</td>
<td>$8 \times 10^{10}$</td>
<td>Pa</td>
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<td>Poisson’s ratio</td>
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<tr>
<td>Density of water</td>
<td>ρ&lt;sub&gt;w&lt;/sub&gt;</td>
<td>1000</td>
<td>kg/m³</td>
<td></td>
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<td>Density of lower crust</td>
<td>ρ&lt;sub&gt;lc&lt;/sub&gt;</td>
<td>2900</td>
<td>kg/m³</td>
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<tr>
<td>Density of upper mantle</td>
<td>ρ&lt;sub&gt;m&lt;/sub&gt;</td>
<td>3400</td>
<td>kg/m³</td>
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<tr>
<td>Range of effective elastic thickness</td>
<td>T&lt;sub&gt;e&lt;/sub&gt;</td>
<td>0.5 ~ 30</td>
<td>km</td>
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<td>Total load cells</td>
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<td></td>
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</tr>
<tr>
<td>Area of cell for modeling</td>
<td>A&lt;sub&gt;cell&lt;/sub&gt;</td>
<td>1×1</td>
<td>km²</td>
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<td></td>
</tr>
<tr>
<td>Volume of water load</td>
<td>V&lt;sub&gt;w&lt;/sub&gt;</td>
<td>303.907</td>
<td>km³</td>
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<tr>
<td>Max load height for the Siling Co</td>
<td>h&lt;sub&gt;sl&lt;/sub&gt;</td>
<td>64</td>
<td>m</td>
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<td></td>
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<tr>
<td>Max load height for the Bange Co</td>
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<td>73</td>
<td>m</td>
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<td>Max load height for the Co E</td>
<td>h&lt;sub&gt;ce&lt;/sub&gt;</td>
<td>37</td>
<td>m</td>
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<tr>
<td>Max load height for the Wuru Co</td>
<td>h&lt;sub&gt;wr&lt;/sub&gt;</td>
<td>44</td>
<td>m</td>
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Table C-4. Bending stress, strain and strain rate on the timescale of the highstand lake loading

<table>
<thead>
<tr>
<th>Te (km)</th>
<th>max κ (10^9 m^-1)</th>
<th>max strain (10^-5)</th>
<th>max stress (MPa)</th>
<th>max strain rate (10^-16 s^-1)</th>
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</thead>
<tbody>
<tr>
<td>12</td>
<td>6.6336</td>
<td>3.9802</td>
<td>3.4</td>
<td>1.26</td>
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<tr>
<td>13</td>
<td>5.7395</td>
<td>3.7307</td>
<td>3.2</td>
<td>1.18</td>
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<tr>
<td>14</td>
<td>5.0115</td>
<td>3.5081</td>
<td>3.0</td>
<td>1.11</td>
</tr>
</tbody>
</table>

κ: curvature of flexure curves

Table C-5. Assumed crustal composition and their rheological parameters

<table>
<thead>
<tr>
<th>Rock/mineral</th>
<th>A (MPa^n s^-1)</th>
<th>n</th>
<th>Q (kJ/mol)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Crust</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>wet Simpson quartzite</td>
<td>5.1×10^-6</td>
<td>2.6</td>
<td>145</td>
<td>Koch et al. (1989)</td>
</tr>
<tr>
<td>wet Heavitree quartzite</td>
<td>2.9×10^-3</td>
<td>1.8</td>
<td>151</td>
<td>Jaoul et al. (1984)</td>
</tr>
<tr>
<td>wet granite</td>
<td>2×10^-4</td>
<td>1.9</td>
<td>141</td>
<td>Hansen and Carter (1982)</td>
</tr>
<tr>
<td>partially molten granite</td>
<td>4.1×10^-2</td>
<td>1.8</td>
<td>220</td>
<td>Rutter et al. (2006)</td>
</tr>
<tr>
<td>Lower Crust</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>dry Maryland diabase</td>
<td>8</td>
<td>4.7</td>
<td>485</td>
<td>Mackwell et al. (1998)</td>
</tr>
<tr>
<td>dry Columbia diabase</td>
<td>190</td>
<td>4.7</td>
<td>485</td>
<td>Mackwell et al. (1998)</td>
</tr>
<tr>
<td>dry anorthite</td>
<td>5×10^{12}</td>
<td>3</td>
<td>648</td>
<td>Rybacki and Dresen (2000)</td>
</tr>
<tr>
<td>Pikwitonei granulite</td>
<td>1.4×10^4</td>
<td>4.2</td>
<td>445</td>
<td>Wilks and Carter (1990)</td>
</tr>
<tr>
<td>UM dry dunite</td>
<td>4.85×10^4</td>
<td>3.5</td>
<td>535</td>
<td>Hirth and Kohlstedt (1996)</td>
</tr>
</tbody>
</table>

UM: upper mantle; Abbr.: abbreviations for different rocks/minerals.

Table C-6. Thermal parameter ranges for strength envelope and viscosity analyses

<table>
<thead>
<tr>
<th>A₀ (μW/m³)</th>
<th>k (W/m/K)</th>
<th>q₀ (mW/m³)</th>
<th>b (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2 – 3</td>
<td>2 – 3</td>
<td>80 – 110</td>
<td>20 – 50  (mostly 25 – 35)</td>
</tr>
</tbody>
</table>

Table C-7. Assumptions of lithospheric compositions and layering

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth (km)</th>
<th>Rock*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper crust</td>
<td>0 – (35 ~ 40)</td>
<td>wet quartzite, wet granite or partially molten granite</td>
</tr>
<tr>
<td>Lower crust</td>
<td>(35 ~ 40) – (65 ± 5)</td>
<td>dry diabase, anorthite, clinopyroxene or undried granulite</td>
</tr>
<tr>
<td>Upper mantle</td>
<td>&gt; 65 ± 5</td>
<td>dry olivine</td>
</tr>
</tbody>
</table>

* All rheological parameters for these rocks/minerals and related references are in Table C-4.
**Table C-8. Field survey data of the highstand shorelines around Siling Co**

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Lat  (°)</th>
<th>Long (°)</th>
<th>Elev (m)</th>
<th>Error (m)</th>
<th>Description</th>
<th>Lake Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>31.880</td>
<td>89.413</td>
<td>4594.2</td>
<td>0.1</td>
<td>SW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>5</td>
<td>31.635</td>
<td>89.511</td>
<td>4593.7</td>
<td>0.1</td>
<td>NNE-facing beach ridge</td>
<td>Bange Co</td>
</tr>
<tr>
<td>8</td>
<td>31.524</td>
<td>89.216</td>
<td>4594.2</td>
<td>0.1</td>
<td>NW-facing curved beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>9</td>
<td>31.520</td>
<td>89.204</td>
<td>4594.0</td>
<td>0.1</td>
<td>NNW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>10</td>
<td>31.502</td>
<td>89.124</td>
<td>4593.9</td>
<td>0.0</td>
<td>NW-facing curved beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>17</td>
<td>31.477</td>
<td>88.812</td>
<td>4591.6</td>
<td>0.1</td>
<td>NE-pointing and NW-facing spit</td>
<td>Co E</td>
</tr>
<tr>
<td>21</td>
<td>31.558</td>
<td>88.878</td>
<td>4594.4</td>
<td>0.1</td>
<td>NNE-trending eroded tombolo</td>
<td>Siling Co</td>
</tr>
<tr>
<td>31</td>
<td>31.712</td>
<td>88.733</td>
<td>4595.8</td>
<td>0.0</td>
<td>NW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>36</td>
<td>31.682</td>
<td>88.733</td>
<td>4595.8</td>
<td>0.1</td>
<td>SW-pointing spit</td>
<td>Siling Co</td>
</tr>
<tr>
<td>40</td>
<td>31.685</td>
<td>88.682</td>
<td>4594.3</td>
<td>0.1</td>
<td>NEE-facing and NNW-pointing spit</td>
<td>Co E</td>
</tr>
<tr>
<td>41</td>
<td>31.687</td>
<td>88.683</td>
<td>4593.7</td>
<td>0.0</td>
<td>SW-pointing, SE-facing curved spit</td>
<td>Co E</td>
</tr>
<tr>
<td>42</td>
<td>31.673</td>
<td>88.679</td>
<td>4595.1</td>
<td>0.1</td>
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<td>Co E</td>
</tr>
<tr>
<td>43</td>
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<td>88.673</td>
<td>4595.5</td>
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<td>Siling Co</td>
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<tr>
<td>47</td>
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<td>Siling Co</td>
</tr>
<tr>
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<td>Co E</td>
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Table C-8 (cont.): Field survey data of the highstand shorelines around Siling Co

<table>
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<tr>
<th>Site ID</th>
<th>Lat (°)</th>
<th>Long (°)</th>
<th>Elev (m)</th>
<th>Error (m)</th>
<th>Description</th>
<th>Lake Name</th>
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<td>Siling Co</td>
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<td>Wuru Co</td>
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<td>4594.9</td>
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<td>125</td>
<td>32.007</td>
<td>88.807</td>
<td>4593.9</td>
<td>0.0</td>
<td>NW-trending tombolo</td>
<td>Guojialun</td>
</tr>
<tr>
<td>127</td>
<td>31.996</td>
<td>88.823</td>
<td>4594.6</td>
<td>0.1</td>
<td>NE-trending tombolo, ERODED</td>
<td>Guojialun</td>
</tr>
<tr>
<td>128</td>
<td>31.991</td>
<td>88.823</td>
<td>4593.6</td>
<td>0.0</td>
<td>NNW-trending tombolo</td>
<td>Siling Co</td>
</tr>
<tr>
<td>129</td>
<td>31.967</td>
<td>88.848</td>
<td>4594.4</td>
<td>0.0</td>
<td>NE-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>130</td>
<td>31.942</td>
<td>88.879</td>
<td>4594.2</td>
<td>0.1</td>
<td>NE-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>134</td>
<td>31.858</td>
<td>88.774</td>
<td>4594.2</td>
<td>0.0</td>
<td>SW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>135</td>
<td>31.858</td>
<td>88.770</td>
<td>4594.0</td>
<td>0.2</td>
<td>S-facing curved cuspate bar</td>
<td>Siling Co</td>
</tr>
<tr>
<td>136</td>
<td>31.832</td>
<td>88.783</td>
<td>4594.4</td>
<td>0.0</td>
<td>NE-facing curved cuspate bar</td>
<td>Siling Co</td>
</tr>
<tr>
<td>137</td>
<td>32.093</td>
<td>88.968</td>
<td>4593.1</td>
<td>0.1</td>
<td>SE-facing curved beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>138</td>
<td>32.155</td>
<td>89.091</td>
<td>4593.3</td>
<td>0.0</td>
<td>S-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>139</td>
<td>32.157</td>
<td>89.122</td>
<td>4593.1</td>
<td>0.0</td>
<td>S-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>146</td>
<td>32.234</td>
<td>89.532</td>
<td>4592.8</td>
<td>0.1</td>
<td>SWW-facing beach ridge</td>
<td>Zanzong</td>
</tr>
<tr>
<td>147</td>
<td>32.089</td>
<td>89.179</td>
<td>4593.7</td>
<td>0.0</td>
<td>NNW-pointing and SWW-facing spit</td>
<td>Siling Co</td>
</tr>
<tr>
<td>152</td>
<td>32.021</td>
<td>89.203</td>
<td>4593.7</td>
<td>0.1</td>
<td>NNW-pointing spit</td>
<td>Siling Co</td>
</tr>
<tr>
<td>155</td>
<td>31.891</td>
<td>89.298</td>
<td>4593.6</td>
<td>0.2</td>
<td>SW-pointing and SE-facing spit</td>
<td>Siling Co</td>
</tr>
<tr>
<td>161</td>
<td>31.850</td>
<td>89.448</td>
<td>4593.9</td>
<td>0.0</td>
<td>SW-facing beach ridge</td>
<td>Siling Co</td>
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<tr>
<td>162</td>
<td>31.820</td>
<td>89.490</td>
<td>4593.3</td>
<td>0.1</td>
<td>SW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>168</td>
<td>31.914</td>
<td>89.831</td>
<td>4592.8</td>
<td>0.0</td>
<td>SW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>172</td>
<td>31.785</td>
<td>89.704</td>
<td>4593.5</td>
<td>0.0</td>
<td>NW-facing curved beach ridge</td>
<td>Bange Co</td>
</tr>
<tr>
<td>173</td>
<td>31.728</td>
<td>89.688</td>
<td>4593.7</td>
<td>0.1</td>
<td>SWW-facing beach ridge</td>
<td>Bange Co</td>
</tr>
<tr>
<td>174</td>
<td>31.582</td>
<td>89.733</td>
<td>4593.6</td>
<td>0.1</td>
<td>W-facing beach ridge</td>
<td>Bange Co</td>
</tr>
<tr>
<td>175</td>
<td>31.568</td>
<td>89.700</td>
<td>4593.3</td>
<td>0.0</td>
<td>N-facing and E-pointing spit</td>
<td>Bange Co</td>
</tr>
</tbody>
</table>
### Table C-9. Field survey data of the shorelines above the highstand shorelines around Siling Co

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Lat (°)</th>
<th>Long (°)</th>
<th>Elev (m)</th>
<th>Error (m)</th>
<th>Description</th>
<th>Lake Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>31.600</td>
<td>89.376</td>
<td>4633.0</td>
<td>0.0</td>
<td>NWW-facing bar</td>
<td>Siling Co</td>
</tr>
<tr>
<td>7</td>
<td>31.579</td>
<td>89.310</td>
<td>4596.9</td>
<td>0.1</td>
<td>NWW-facing degraded beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>18</td>
<td>31.498</td>
<td>88.816</td>
<td>4608.6</td>
<td>0.1</td>
<td>NW-facing beach ridge</td>
<td>Co E</td>
</tr>
<tr>
<td>22</td>
<td>31.578</td>
<td>88.953</td>
<td>4617.4</td>
<td>0.1</td>
<td>SSE-pointing and NEE-facing spit</td>
<td>Yagedong</td>
</tr>
<tr>
<td>23</td>
<td>31.633</td>
<td>88.877</td>
<td>4595.5</td>
<td>0.0</td>
<td>NNW-trending tombolo</td>
<td>Siling Co</td>
</tr>
<tr>
<td>25</td>
<td>31.668</td>
<td>88.913</td>
<td>4619.4</td>
<td>0.1</td>
<td>S-pointing and E-facing spit</td>
<td>Siling Co</td>
</tr>
<tr>
<td>28</td>
<td>31.731</td>
<td>88.919</td>
<td>4601.6</td>
<td>0.2</td>
<td>SW-pointing and NW-facing spit</td>
<td>Siling Co</td>
</tr>
<tr>
<td>45</td>
<td>31.694</td>
<td>88.578</td>
<td>4597.4</td>
<td>0.2</td>
<td>NWW-trending tombolo</td>
<td>Siling Co</td>
</tr>
<tr>
<td>46</td>
<td>31.716</td>
<td>88.512</td>
<td>4601.4</td>
<td>0.2</td>
<td>SEE-facing curved and degraded</td>
<td>Siling Co</td>
</tr>
<tr>
<td>52</td>
<td>31.861</td>
<td>88.774</td>
<td>4617.9</td>
<td>0.2</td>
<td>E-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>65</td>
<td>31.919</td>
<td>88.846</td>
<td>4636.2</td>
<td>0.1</td>
<td>NEE-facing highly degraded beach</td>
<td>Siling Co</td>
</tr>
<tr>
<td>66</td>
<td>31.942</td>
<td>88.871</td>
<td>4628.3</td>
<td>0.0</td>
<td>NEE-facing highly degraded beach</td>
<td>Siling Co</td>
</tr>
<tr>
<td>68</td>
<td>32.144</td>
<td>89.074</td>
<td>4598.7</td>
<td>0.0</td>
<td>SW-facing bar/tombolo</td>
<td>Siling Co</td>
</tr>
<tr>
<td>71</td>
<td>32.088</td>
<td>89.182</td>
<td>4596.1</td>
<td>0.1</td>
<td>SW-facing degraded beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>72</td>
<td>32.047</td>
<td>89.231</td>
<td>4607.7</td>
<td>0.4</td>
<td>SSE-facing beach ridge</td>
<td>Siling Co</td>
</tr>
<tr>
<td>73</td>
<td>32.008</td>
<td>89.206</td>
<td>4618.4</td>
<td>0.0</td>
<td>NEE-facing highly degraded beach</td>
<td>Siling Co</td>
</tr>
<tr>
<td>180</td>
<td>31.675</td>
<td>89.416</td>
<td>4640.1</td>
<td>0.1</td>
<td>W-facing beach ridge</td>
<td>Siling Co</td>
</tr>
</tbody>
</table>

### Table C-10. Field survey data of the shorelines below the highstand shorelines around Siling Co

<table>
<thead>
<tr>
<th>Site ID</th>
<th>Lat (°)</th>
<th>Long (°)</th>
<th>Elev (m)</th>
<th>Error (m)</th>
<th>Description</th>
<th>Lake Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>32</td>
<td>31.666</td>
<td>88.803</td>
<td>4592.1</td>
<td>0.0</td>
<td>E-pointing spit</td>
<td>Co E</td>
</tr>
<tr>
<td>33</td>
<td>31.685</td>
<td>88.800</td>
<td>4593.5</td>
<td>0.0</td>
<td>NEE-facing ridge or scarp top?</td>
<td>Co E</td>
</tr>
<tr>
<td>35</td>
<td>31.658</td>
<td>88.785</td>
<td>4592.3</td>
<td>0.3</td>
<td>NNW-trending tombolo</td>
<td>Co E</td>
</tr>
<tr>
<td>37</td>
<td>31.658</td>
<td>88.789</td>
<td>4593.1</td>
<td>0.3</td>
<td>NE-pointing and SE-facing spit</td>
<td>Co E</td>
</tr>
<tr>
<td>45</td>
<td>31.694</td>
<td>88.573</td>
<td>4590.4</td>
<td>0.1</td>
<td>NWW-trending tombolo</td>
<td>Siling Co</td>
</tr>
<tr>
<td>86</td>
<td>31.658</td>
<td>88.846</td>
<td>4591.3</td>
<td>0.0</td>
<td>SW-facing beach ridge</td>
<td>Co E</td>
</tr>
<tr>
<td>150</td>
<td>32.049</td>
<td>89.203</td>
<td>4591.3</td>
<td>0.1</td>
<td>SW-facing beach ridge</td>
<td>Siling Co</td>
</tr>
</tbody>
</table>
Table C-11. Crustal viscosity data of Tibetan crust determined from different methods, for data records of > 2 years.

<table>
<thead>
<tr>
<th>Timescale (yr)</th>
<th>Technique</th>
<th>Event</th>
<th>Record Time</th>
<th>Depth</th>
<th>Region</th>
<th>Viscosity (Pa s)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$10^{-1} - 10^2$ (Postseismic)</td>
<td>InSAR</td>
<td>1997 $M_w$ 7.6 Manyi</td>
<td>&gt; ~ 3 yrs</td>
<td>&gt; 15 km</td>
<td>N Tibet</td>
<td>$\sim 10^{19}$</td>
<td>Ryder et al., 2007</td>
</tr>
<tr>
<td></td>
<td>GPS and InSAR</td>
<td>2001 $M_w$ 7.8 Kokoxili</td>
<td>&gt; 2 yrs</td>
<td>Lower crust</td>
<td></td>
<td>$\sim 10^{19}$</td>
<td>Ryder et al., 2011</td>
</tr>
<tr>
<td></td>
<td>InSAR</td>
<td>2001 $M_w$ 7.8 Kokoxili</td>
<td>&gt; 2-6 yrs</td>
<td>&gt; 15 km</td>
<td></td>
<td>(2-5) $\times 10^{19}$</td>
<td>Wen et al., 2012</td>
</tr>
<tr>
<td></td>
<td>GPS and InSAR</td>
<td>2008 $M_w$ 7.9 Wenchuan</td>
<td>&gt; ~ 2 yrs</td>
<td>45-60 km</td>
<td>E Tibet</td>
<td>1 $\times 10^{18}$</td>
<td>Huang et al., 2014</td>
</tr>
<tr>
<td></td>
<td>Leveling</td>
<td>1973 $M_s$ 7.6 Luhuo</td>
<td>&gt; 7 - 20 yrs</td>
<td>&gt; 31 km</td>
<td></td>
<td>$\sim 10^{19}$</td>
<td>Zhang et al. 2009a</td>
</tr>
<tr>
<td>$10^1 - 10^3$ (Interseismic)</td>
<td>GPS and geological data</td>
<td>2001 $M_w$ 7.9 Kokoxili</td>
<td>Long term</td>
<td>middle to lower crust and mantle</td>
<td>N Tibet</td>
<td>1 $\times 10^{19} - 2\times 10^{21}$</td>
<td>Hilley et al., 2005</td>
</tr>
<tr>
<td></td>
<td>GPS</td>
<td>2001 $M_w$ 7.9 Kokoxili</td>
<td>Short term</td>
<td>middle to lower crust</td>
<td></td>
<td>$\geq 10^{18}$</td>
<td>Hilley et al., 2009</td>
</tr>
<tr>
<td>$10^3 - 10^5$</td>
<td>Lacustrine shoreline rebound</td>
<td>5 - 10 ka (?)</td>
<td>middle/lower crust</td>
<td>Central Tibet</td>
<td>&gt; 10^{19} - 10^{20} (?)</td>
<td>England et al., 2013</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>4 ka</td>
<td>20-40 km</td>
<td>Central Tibet</td>
<td>$10^{18} - 10^{20}$ (*)</td>
<td>This study</td>
<td></td>
</tr>
</tbody>
</table>

* This represents depth-dependent effective viscosity.

Table C-12. Crustal viscosity data of Tibetan crust determined from postseismic deformation with records of < 2 years.

<table>
<thead>
<tr>
<th>Timescale (yr)</th>
<th>Technique</th>
<th>Event</th>
<th>Record Time</th>
<th>Depth</th>
<th>Region</th>
<th>Viscosity (Pa s)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$10^{-1} - 10^2$ (Postseismic)</td>
<td>InSAR</td>
<td>1997 $M_w$ 7.6 Manyi</td>
<td>&gt; 25 - 305 days</td>
<td>&gt; 15 km</td>
<td>N Tibet</td>
<td>4 $\times 10^{18}$</td>
<td>Ryder et al., 2007</td>
</tr>
<tr>
<td></td>
<td>InSAR</td>
<td>1997 $M_w$ 7.6 Manyi</td>
<td>&gt; 165 days</td>
<td>20-60 km</td>
<td></td>
<td>$2 \times 10^{20} - 10^{18}$ (*)</td>
<td>Yamasaki and Houseman, 2012</td>
</tr>
<tr>
<td></td>
<td>GPS</td>
<td>1997 $M_w$ 7.6 Manyi</td>
<td>730 days</td>
<td>20-40 km</td>
<td></td>
<td>$\leq 3 \times 10^{18}$</td>
<td>DeVries and Meade, 2013</td>
</tr>
<tr>
<td></td>
<td>GPS</td>
<td>2001 $M_w$ 7.8 Kokoxili</td>
<td>&gt; 1 yr</td>
<td>Lower crust</td>
<td></td>
<td>10^{17}</td>
<td>Zhang et al. 2009a</td>
</tr>
<tr>
<td></td>
<td>InSAR</td>
<td>2001 $M_w$ 7.8 Kokoxili</td>
<td>333 days</td>
<td>20-40 km</td>
<td></td>
<td>$\leq 3 \times 10^{18}$</td>
<td>DeVries and Meade, 2013</td>
</tr>
<tr>
<td></td>
<td>InSAR</td>
<td>2008 M 6.4 Nima-Gaize</td>
<td>&gt; 9 months</td>
<td>middle/lower crust</td>
<td>Central Tibet</td>
<td>&gt; 3 $\times 10^{17}$</td>
<td>Ryder et al., 2010</td>
</tr>
<tr>
<td></td>
<td>InSAR</td>
<td>2008 M 6.3 Damxung</td>
<td>&gt; 1.6 yrs</td>
<td>middle/lower crust</td>
<td></td>
<td>&gt; 1 $\times 10^{18}$</td>
<td>Bie et al., 2014</td>
</tr>
<tr>
<td></td>
<td>GPS</td>
<td>2008 $M_w$ 7.9 Wenchuan</td>
<td>&gt; 14 days</td>
<td>middle/lower crust</td>
<td>E Tibet</td>
<td>4 - 7 $\times 10^{17}$</td>
<td>Shao et al., 2011</td>
</tr>
</tbody>
</table>

* This represents depth-dependent effective viscosity.

Table C-13. Model-determined crustal viscosity data of Tibetan crust.
<table>
<thead>
<tr>
<th>Timescale (yr)</th>
<th>Technique</th>
<th>Record Time</th>
<th>Depth</th>
<th>Region</th>
<th>Viscosity (Pa s)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$10^5 - 10^6$ (geodynamic and topography)</td>
<td>Topography</td>
<td>Ma</td>
<td>middle crust</td>
<td>S Tibet</td>
<td>$10^{18} - 10^{19}$</td>
<td>Beaumont et al., 2001</td>
</tr>
<tr>
<td></td>
<td>Topography</td>
<td>Ma</td>
<td>lower crust</td>
<td>central Tibet</td>
<td>$10^{16}$</td>
<td>Clark and Royden, 2000</td>
</tr>
<tr>
<td></td>
<td>Topography</td>
<td>Ma</td>
<td>lower crust</td>
<td>Plateau</td>
<td>$2 \times 10^{18}$</td>
<td>Clark et al., 2005</td>
</tr>
<tr>
<td></td>
<td>Topography+GPS</td>
<td>Ma</td>
<td>middle crust</td>
<td>Plateau</td>
<td>$10^{18} - 10^{19}$</td>
<td>Beaumont et al., 2004</td>
</tr>
<tr>
<td></td>
<td>Topography</td>
<td>Ma</td>
<td>lower crust</td>
<td>Plateau</td>
<td>$10^{19}$</td>
<td>Cook and Royden, 2008</td>
</tr>
<tr>
<td></td>
<td>magnetotelluric</td>
<td>Short term</td>
<td>middle crust</td>
<td>S Lhasa</td>
<td>$2.5 \times 10^{18} - 3 \times 10^{20}$</td>
<td>Rippe and Unsworth, 2010</td>
</tr>
<tr>
<td></td>
<td>magnetotelluric</td>
<td>Short term</td>
<td>middle crust</td>
<td>N Lhasa</td>
<td>$1 \times 10^{20} - 3.5 \times 10^{21}$</td>
<td>Rippe and Unsworth, 2010</td>
</tr>
<tr>
<td></td>
<td>magnetotelluric</td>
<td>Short term</td>
<td>middle crust</td>
<td>Qiangtang</td>
<td>$\sim 10^{18} - 10^{20}$</td>
<td>Rippe and Unsworth, 2010</td>
</tr>
<tr>
<td></td>
<td>magnetotelluric</td>
<td>Short term</td>
<td>middle crust</td>
<td>E Tibet</td>
<td>$\sim 2 \times 10^{17} - 1 \times 10^{20}$</td>
<td>Rippe and Unsworth, 2010</td>
</tr>
</tbody>
</table>
APPENDIX D

D-1 OSL Methodology

D-1.1 Field sampling

Given that no stream-cut stratigraphy was found in these highstand shorelines, depth profile pits (> 0.5 m – 2.5 m deep) have been dug to observe a limited dimension of vertical and lateral stratigraphy. All samples were carefully collected from sand layers/lenses that are intercalated with the upper beach gravel deposits (Figure D-1) that can reflect ages of the most recent highstand shorelines hence lake levels. Eolian sands (younger than the age of shoreline deposition) at the top of the shoreline stratigraphy are definitely avoided. Before collecting the sand samples, the light-affected surface deposits of several centimeters thick were removed to reduce the uncertainty from dose measurement (Aitken, 1998). Sands of medium, fine or silt sizes were drilled using plastic PVC (Polyvinyl chloride) and or steel tubes, 3 cm or 5 cm in diameter, depending on the consolidation and thickness of the sand layers. These tubes were then wrapped with heavy duty black duct tape to avoid light penetration and keep the water content.

D-1.2 OSL Laboratory Processing

D-1.2.1 Sample preparation

Samples were prepared using standard OSL procedures. Material was desiccated at 50 °C to enable calculation of water content, and then sieved to extract the 180-212 µm grain size fraction. Approximately 10 g of the 180-212 µm grain size fraction was treated with 30% HCl for 30 minutes to remove CaCO3. Samples were agitated throughout the treatment, and once complete, HCl was replaced with 30% H2O2 to remove organic material. The duration of H2O2 treatment varied between samples,
dependent upon the amount of organic material present. Once effervescence ceased, the H₂O₂ was decanted, the sample was washed four times with deionised water and desiccated at 50 °C. Quartz is extracted from polymineral sediment residues through density separations using LST fastfloat (sodium heteropolytungstate dissolved in deionized water). Heavy minerals (> 2.68 g cm⁻³) were separated from the lighter fraction, and the target 2.58-2.68 g cm⁻³ fraction was further separated from the < 2.58 g cm⁻³ material. The target fraction was washed five times with deionised water to ensure removal of all LST. Final separates were dried and etched with 40% HF for 40 minutes to remove any contaminating feldspar; all samples were agitated at 5 minute intervals throughout treatment. The etched quartz was treated with 30% HCl for 30 minutes to remove any carbonates produced during HF etching.

D-1.2.2 Luminescence measurements

All analyses were carried out using either a TL-DA-15 or TL-DA-20 Risø reader, equipped with an EMI 9235QA photomultiplier and 7.5 mm Hoya U-340 filter. Blue (470±20 nm) and infrared (~870 nm) diodes operated at 90% and 40% power respectively, were used for stimulation and irradiation was achieved using a 90Sr/90Y beta source. Readers were calibrated using quartz prepared at the Risø National Laboratory in Denmark. Quartz was applied to stainless steel discs (10 mm Ø, 1 mm thick) using silicon grease and the aliquot size was regulated using a small (2 mm Ø, ~35 grain) mask.

Samples were analysed using the single aliquot regenerative dose (SAR) protocol (Murray & Wintle, 2000). The equivalent dose (De) was calculated from measurements of the luminescence response following stimulation of the natural luminescence (Ln) and a series of different regenerative doses (Lx). The Ln and Lx measurements are normalised by measurement of the luminescence response (Tx) to a constant test dose (TD). The ratio Lx/Tx is used to compensate for sensitivity changes of the quartz throughout analysis and Lx/Tx measurements are used to obtain a range of values which bracket Ln/Tx, allowing De interpolation with minimal associated errors (Banerjee et al., 2000).

The sample is heated prior to making the luminescence measurements in order to reduce the contribution of luminescence from unstable trap, which cause erroneous dose determinations. The
temperature of the pre-Lx pre-heat (PH1) and the pre-Tx pre-heat (PH2) must be empirically determined for each sample under analysis. This was achieved through analysis of 8 disks with a dose-recovery pre-heat plateau experiment (Murray and Wintle, 2003), using a range of PH1 and PH2 temperatures ranging from 150 – 220 °C. The influence of a hot-bleach at 280 °C was also investigated (Murray and Wintle, 2003) but was not required to counter any thermal transfer or recuperation.

The SAR protocol with a PH1 of 200 °C used on most of the samples is provided in Table D-1. Aliquot acceptance criteria used are 1) recycling ratios within 10% of unity; 2) signal intensities ≥ 3 σ above background; 3) infra-red (IR) depletion ratio within 20% of unity (Duller, 2003); 4) De uncertainty ≤ 20 % and 5) recuperation within 10% of the normalised maximum dose. The acceptance thresholds are generally very high for the samples (85 – 100%) reflecting the sensitivity of the quartz analysed. All three samples have at least 50 accepted aliquots.

**D-1.2.3 Environmental Dose rate determination**

The environmental dose rates (Dr) were calculated for each sample from the unsieved portions of the original sample; concentrations of U, Th, K and Rb were measured directly using solution ICP-MS (Thermo X-Series), a cosmic-dose component after Prescott and Hutton (1994) and an internal alpha dose rate of 5% from the decay of U and Th after Sutton and Zimmerman (1978). External α-dose rates were ignored as the alpha irradiated portion of quartz grains was removed by etching. The conversion factors of Adamiec and Aitken (1998) and beta-particle attenuation factors after Mejdahl (1979) and Readhead (2002a, b) have been used. Sample water content was calculated following desiccation at 50 °C, and an uncertainty of 5 % assumed. Tables D-2 and D-3 contain the dosimetry and age data of OSL samples from the highstand shorelines.
D-2 References


Figure D-1. Example photos showing the beach gravel deposits in the highstand shorelines along the SE margin of Zigui Co.
Figure D-2. GeoEye imagery (left) and sketch (right) showing the displaced old alluvial terraces (T2-T4) and also the undisplaced youngest terrace (T1). Red lines show the fault traces and hatched polygons represent the terrace risers.
Figure D-3. (a-c) The density function and cumulative frequency of equivalent dose (De) for OSL samples GR-1, 2, and 3, respectively.
Table D-1. Quartz SAR protocol for all Tibet samples

<table>
<thead>
<tr>
<th>Natural/Regenerative Dose</th>
<th>TL (PH1)</th>
<th>IRLS (final cycle only)</th>
<th>OSL</th>
</tr>
</thead>
<tbody>
<tr>
<td>5, 10, 20, 30, 0, 5, 5 Gy</td>
<td>200°C, 10 s, 5°C/s</td>
<td>20°C, 40 s, 5°C/s</td>
<td>125°C, 40 s, 5°C/s, 90% power (Lx)</td>
</tr>
<tr>
<td>Test Dose</td>
<td>5 Gy</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TH (PH2)</td>
<td>180°C, 10 s, 5°C/s</td>
<td></td>
<td></td>
</tr>
<tr>
<td>OSL</td>
<td>125°C, 40 s, 5°C/s, 90% (Tx)</td>
<td></td>
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</tbody>
</table>

Table D-2. OSL sample locations and age calculation data of the displaced shorelines along the Gyaring Co fault.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Lat (°N)</th>
<th>Long (°E)</th>
<th>Elev. (m)</th>
<th>Samp_Depth (m)</th>
<th>N (aliquots)</th>
<th>De (Gy)</th>
<th>Error (Gy)</th>
<th>Dose Rate (Gy/kyr)</th>
<th>Error (Gy/kyr)</th>
<th>Age (ka BP)</th>
<th>Uncert_2σ (ka)</th>
<th>Age Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>GR1</td>
<td>31.347</td>
<td>87.985</td>
<td>~4665</td>
<td>1.8</td>
<td>53</td>
<td>16.6</td>
<td>0.5</td>
<td>4.1</td>
<td>0.1</td>
<td>4.1</td>
<td>0.1</td>
<td>CAM</td>
</tr>
<tr>
<td>GR2</td>
<td>31.347</td>
<td>87.985</td>
<td>~4665</td>
<td>1.1</td>
<td>50</td>
<td>11.6</td>
<td>0.7</td>
<td>2.8</td>
<td>0.1</td>
<td>4.1</td>
<td>0.3</td>
<td>MAM-3</td>
</tr>
<tr>
<td>GR3</td>
<td>31.350</td>
<td>87.985</td>
<td>~4665</td>
<td>15.0</td>
<td>53</td>
<td>15.9</td>
<td>0.9</td>
<td>3.6</td>
<td>0.1</td>
<td>4.4</td>
<td>0.3</td>
<td>MAM-3</td>
</tr>
</tbody>
</table>

* MAM – Minimum age model; CAM – central age model; numbers denote the component of each age model

Table D-3. Chemical data of OSL samples of the displaced shorelines along the Gyaring Co fault.

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>U (ppm)</th>
<th>Error (ppm)</th>
<th>Th (ppm)</th>
<th>Error (ppm)</th>
<th>K (%)</th>
<th>Error (ppm)</th>
<th>Rb (ppm)</th>
<th>Error (ppm)</th>
<th>H₂O (wt. %)</th>
<th>Error (wt. %)</th>
<th>Cosmic (mGya⁻¹)</th>
<th>Error -0.10</th>
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</thead>
<tbody>
<tr>
<td>GR1</td>
<td>2.90</td>
<td>0.09</td>
<td>12.90</td>
<td>0.39</td>
<td>2.37</td>
<td>0.07</td>
<td>139.20</td>
<td>13.92</td>
<td>6.73</td>
<td>0.34</td>
<td>0.43</td>
<td>0.04</td>
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<tr>
<td>GR2</td>
<td>2.30</td>
<td>0.07</td>
<td>9.30</td>
<td>0.28</td>
<td>1.35</td>
<td>0.04</td>
<td>91.30</td>
<td>9.13</td>
<td>5.50</td>
<td>0.28</td>
<td>0.47</td>
<td>0.05</td>
</tr>
<tr>
<td>GR3</td>
<td>2.90</td>
<td>0.09</td>
<td>11.10</td>
<td>0.33</td>
<td>2.01</td>
<td>0.06</td>
<td>125.00</td>
<td>12.50</td>
<td>9.43</td>
<td>0.47</td>
<td>0.54</td>
<td>0.05</td>
</tr>
</tbody>
</table>
XUHUA SHI

EDUCATION

- Ph.D. Geosciences, Pennsylvania State University. 2014
- M.S. Geosciences, Pennsylvania State University. 2011
- M.S. Geology, Graduate University of Chinese Academy of Sciences, 2007
- B.S. Geology, China University of Geosciences, Beijing. 2004

PROFESSIONAL EXPERIENCES


TEACHING EXPERIENCES

- Graduate Teaching Assistantships – Nature Disasters and Geology of National Parks

SELECTED HONORS AND AWARDS

- Scholten-Williams-Wright Scholarship in Field Geology (2013)
- ExxonMobil Geosciences Student Research Grant (2012)
- First Prize (poster) in Penn State Geosciences Graduate Student Colloquium (2011)
- Geological Society of America Student Research Grant (2010)
- ConocoPhillips Student Scholarship (2002)

SELECTED PUBLICATIONS


