Timing and process of river and lake terrace formation in the Kyrgyz Tien Shan

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1. Introduction

The Tien Shan is one of the most active mountain belts on Earth and provides a natural laboratory to study landscape evolution in an intracontinental setting. Active contractional deformation produces variations in rock uplift rate along river systems, and flights of river terraces and alluvial fans are preserved along many river systems, often connecting glacial and lacustrine landforms that provide proxies for climate change (e.g., Aleshinskaya et al., 1971; Thompson et al., 2002; Koppes et al., 2008). This setting provides an excellent place to examine the timing and process of river terrace formation related to climatic and base-level variations.

A river terrace forms when a river cuts through its bed following a period of relative vertical stability or aggradation in response to climatic and tectonic forcings (e.g., Bull, 1991). Rapid, large magnitude variations in base-level caused by sea or lake level changes exert an important control on rivers that drain into oceans or large lakes (Merritts et al., 1994; Pazzaglia and Brandon, 2001). Temporal changes in base-level from tectonic fluctuations also provide a potential means of driving punctuated incision or aggradation along a river. However, most observations of major terrace forming periods align with major Quaternary climate fluctuations, even in areas of tectonic deformation (e.g., Molnar et al., 1994; Bridgland and Westaway, 2008).

The balance between sediment supply and stream power, which
is thought to be the primary control on terrace formation, is dependent in a complex way on climatic factors that have varied temporally and spatially over the late Quaternary (Bull, 1991; Merritts et al., 1994; Pazzaglia and Brandon, 2001; Hancock and Anderson, 2002). For example, shifts to greater precipitation have been inferred to drive a range of outcomes in different landscapes. In some settings, increases in precipitation have driven aggradation by enhancing rates of mass wasting from hillslopes into the fluvial network (e.g., Bull, 1991; Personius et al., 1993). In other landscapes, increased precipitation has caused incision through previously aggraded sediments to create river terraces and incised alluvial fans (e.g., Bull, 1991; Porter et al., 1992; Molnar et al., 1994; Pan et al., 2003; Lu et al., 2010).

In this study we characterize the geometry, stratigraphic relationships, and ages of river and lake terraces in the southern Issyk-Kul basin of the Kyrgyz Tien Shan (Fig. 1). The rivers of the southern Issyk-Kul basin traverse active reverse faults and folds with varying degrees and senses of offset. These rivers drain into Issyk-Kul (Issyk means warm and Kul means lake in Kyrgyz), which is the fifth deepest lake in the world and eleventh largest by volume (Herndendorf, 1990), and has experienced major fluctuations in level and extent throughout the Quaternary. We assess the contributions of base-level and climatic factors in driving terrace formation and incision in the region through synthesizing the timing of southern Issyk-Kul terraces with historic and emerging chronologies of the terrace formation and glaciation from elsewhere in the Tien Shan region. We conclude by highlighting the implications of these remarkably well-preserved river and lacustrine terraces of the Tien Shan for neotectonic and geomorphic studies in the region and elsewhere.

1.1. Study area

The Tien Shan is an east-trending high mountain belt in the interior of Eurasia (Fig. 1). The mountains are forming due to distributed reverse faulting and folding in the Eurasian lithosphere that accommodates convergence between India and Eurasia (e.g., Molnar and Tapponnier, 1975). The Tien Shan is the most active locus of late Cenozoic shortening north of the Himalayan frontal thrust faults, and active deformation is distributed across several major fault systems within and at the edges of the belt (Abdrakhmatov et al., 1996; Thompson et al., 2002).

In a regional sense, the Tien Shan represents a reactivated area of Paleozoic deformation. Following the Paleozoic orogenies, the region was planed off to a low-relief surface (Chediya, 1986). Late Cenozoic deformation has resulted in this surface being warped across a series of mountain ranges cored by crystalline basement and previously deformed Paleozoic sedimentary and metamorphic rocks. The intervening basins have been filled by thick sections of syntectonic sediment (e.g., Abdrakhmatov et al., 2001).

One of the largest basins within the Kyrgyz portion of the Tien Shan is the Issyk-Kul basin, named after the lake. Issyk-Kul, which occupies much of the basin (Fig. 2). Modern Issyk-Kul is internally drained, with a surface area of 6200 km², a maximum depth of 668 m, and a surface elevation of ~1607 m above mean sea level (amsl). The Issyk-Kul basin is bounded to the south by the actively growing Terskey Ala-Tau Range, which includes numerous glaciated peaks that rise above 4000 m amsl (Fig. 2). Rivers in southwestern Issyk-Kul traverse south-vergent structures, which form uphill-facing topography and often show signs of progressive stream capture. The rivers of southeastern Issyk-Kul cut through a large north-vergent fold limb that has been tilting the land surface up to the south over the last few million years (Burgette, 2008).

1.2. Previous Kyrgyz Tien Shan river terrace correlation and dating

Quaternary landforms and associated sedimentary deposits in the Tien Shan have been studied extensively by Soviet and Kyrgyz Quaternary scientists (e.g., Grigorienko, 1970; Grigina and Fortuna, 1981; Trofimov, 1990) and correlated through a relative dating scheme (Fig. 3). This Quaternary framework relies on morphologic similarity, stratigraphic correlation, paleontology, correlation with global climatic change, and a few radiometric ages (e.g., Grigorienko, 1970). The morphostratigraphic approach, assuming landforms can be appropriately correlated from one area to another, offers a first-order method for understanding patterns of paleohydrology and tectonics over extensive areas.

In the Kyrgyz Tien Shan, the Quaternary is divided into four primary divisions: from oldest to youngest these are Q1 to QIV (Grigorienko, 1970). QIV is generally correlative with the Holocene (Melnikova, 1986). QIV, QIII, and QII have been thought to represent climatic cycles in the late to middle Pleistocene (e.g., Trofimov, 1990). In many areas of the Tien Shan, geomorphic surfaces naturally lend themselves to classification in broad divisions, as sets of terraces are often grouped in elevation and are separated by larger elevation differences. Terraces within a group are subdivided with superscript Arabic numbers, again with lower numbers being older (Grigorienko, 1970). For example, the second oldest terrace in the QIII group is called QIII. For flights of terraces that include more terraces than the regional pattern, we use decimals, such as QIII.1, to maintain the correlation for the most prominent terraces in the regional scheme.

Based on our observations and guided by earlier literature, QIV terraces, where present, occur as narrow cut surfaces on older fill deposits or as strath terraces in the hanging walls of the most active faults (Fig. 3). A single QIV terrace is most common, but multiple QIV terraces can be preserved in areas with high uplift rates.

The QIII terraces are the most prominent terraces associated with the modern rivers. The QIII terrace is generally broad and the
best developed. $Q_{III}$ terraces have thick fill deposits except for where rivers traverse areas of rapid incision associated with active structures (Fig. 3). Identification of $Q_{III}$ is generally a key step in making terrace correlations between rivers and basins. There is often a $Q_{III}$ cut terrace developed below the top of the $Q_{II}$. Narrow strath or fill $Q_{III}$ terraces are commonly set into the canyon walls above $Q_{II}$, but well below $Q_{II}$ surfaces. In areas of relatively high incision rate, there are often multiple terraces with $Q_{III}$ geomorphic characteristics between $Q_{II}$ and $Q_{II}$. $Q_{II}$ terraces are the oldest commonly preserved geomorphic surfaces, forming an extensive cap on the Tertiary sedimentary rock (Fig. 3). The highest levels of the $Q_{II}$ terraces do not appear localized with the modern canyons, but form broad bajada or pediment-like surfaces spanning multiple drainages. $Q_{II}$ is the most prominent $Q_{II}$ terrace, and is usually characterized by a fill morphology. Higher discontinuous $Q_{I}$ surfaces are preserved in places, and slightly inset $Q_{II}$ terraces are common. Again, in areas of deep incision, there are commonly more terrace levels than the regionally-common set. In places, workers recognize even higher $Q_{I}$ surfaces that are usually developed on the Paleozoic bedrock, significantly higher than the $Q_{II}$ terraces (e.g. Chediya, 1986), and are too poorly developed to address here.

Studies of pollen preserved in terrace gravels and correlative basin deposits show that the plant assemblages of the northern Kyrgyz Tien Shan during times of gravel aggradation were dominated by grasses with minor components of coniferous and small-leaved trees, suggestive of cold, dry conditions (Grigina and Fortuna, 1981). In records from basins with relatively continuous deposition,
the intervening periods have pollen assemblages with greater fractions of tree pollen and more broad-leaved trees, which are interpreted to represent warmer, wetter interglacial climates similar to modern conditions (Grigina and Fortuna, 1981). The palynologic data thus support field relationships that suggest the terrace gravel aggraded contemporaneously with major late Pleistocene glacial advances.

Soviet era age control for the younger Quaternary terrace chronology was largely based on a limited number of numerical ages. The Qu level is dated as Holocene with radiocarbon ages of 7.7 and 1.1 cal ka BP on the lower shoreline near the mouth of the Jergalan River (Fig. 2; Aleshinskaya et al., 1971; Trofimov, 1990); all reported radiocarbon ages are calibrated (Bronk Ramsey, 2009; Reimer et al., 2013), unless otherwise specified. Issyk-Kul lake sediments correlated as QtI near the Tyoup River, yielded a radiocarbon age of 31.3 ± 29.4 cal ka BP (Fig. 2; Aleshinskaya et al., 1971). Thermoluminescence ages on older eastern Issyk-Kul basin sediments correlated with the river terraces suggest ages of 590–440 ka for QtI, 320–280 ka for QtII, and 230–220 ka for QtIII (Trofimov, 1990).

Thompson et al. (2002) collected a suite of charcoal samples for radiocarbon dating from QtIII and younger terraces in five basins west of Issyk-Kul, primarily in fine-grained sediment overlying terrace gravel. The ages of abandonment of QtIII and QtII terrace levels are indistinguishable and consistent between basins in the range of 15.5–13.5 cal ka BP. Two radiocarbon ages within QtII fill along the Kajerty River in the Naryn basin southwest of Issyk-Kul (Fig. 1) are stratigraphically consistent at 47.9 ± 2.6 and 43.2 ± 3.3 cal ka BP, approximately in the middle of a thick QtII gravel fill deposit (Thompson, 2001). Infrared-stimulated luminescence dating of fine sediment capping terraces mapped as QtI in the Chu, Kochkor, and Naryn basins yielded overlapping ages, and assuming synchrony, the best estimate of QtI terrace abandonment is 141 ± 17 ka (Thompson et al., 2002).

Selander et al. (2012) dated terraces along the Toru-Aygir River on the northern shore of Issyk-Kul using in-situ cosmogenic 10Be dating of quartz pebbles collected in depth profiles. The model age for a terrace mapped as QtI is 85.6 ± 7.6 ka. Higher terraces correlated as QtI and QtII yielded ages of 139.9 ± 6.5 ka and 126.4 ± 10.8 ka, respectively.

1.3. Previous work on Issyk-Kul lake history

Subaerial young lakebeds around Issyk-Kul coupled with the distinctive sub-lacustrine channels and terraces revealed in its bathymetry suggest that the lake level has fluctuated dramatically during the late Quaternary (Trofimov, 1978; De Batist et al., 2002). The history of Issyk-Kul is recorded in sediments from the deep basin as well as lake sediments exposed up to ~50 m above the modern lake level (Trofimov, 1990). A radiocarbon age estimate from a late Quaternary highstand at ~50 m higher than the modern lake is 31.3–29.4 cal ka BP and is correlated to the QtI river terraces (Aleshinskaya et al., 1971). Infrared-stimulated luminescence (IRSL) dating of fluvial terrace gravels. Following manual cleaning, charcoal and wood samples were pretreated with acid-alkali-acid, and shell samples were treated with acid-only. Radiocarbon ages were measured by accelerator mass spectrometry (AMS), and results from three laboratories are all consistent with stratigraphic relationships where available (Table 1). All ages are calibrated with IntCal13 (Reimer et al., 2013) implemented in OxCal 4.2 (Bronk Ramsey, 2009), and ages discussed in the text are medians from the calibrated probability distributions in years before 1950 CE, unless otherwise specified. See Table 1 for full details of radiocarbon results.

2. Methods

2.1. Terrace profiling

We have documented the geometry of river terraces using the datum of the top surface of the terrace gravel, as this represents the position of the former floodplain prior to incision. In locations where the surficial sediment of a terrace is in place, rather than terrace gravel, we measured the elevation of the gravel/cover contact. We project the terrace gravel elevation data to smoothly curved valley-parallel lines in cases where the valleys are not linear. We profiled terrace surfaces with three techniques (Burgette, 2008): Quaternary geologic mapping (precision of 5–10 m), total station (precision 2–3 cm), and handheld mapping-grade differential GPS (precision ~0.5 m).

2.2. Radiocarbon dating

We collected mollusk shells, charcoal and plant material from lake deposits and fine grained sediment overlying and underlying fluvial terrace gravels. Following manual cleaning, charcoal and wood samples were pretreated with acid-alkali-acid, and shell samples were pretreated with acid-only. Radiocarbon ages were measured by accelerator mass spectrometry (AMS), and results from three laboratories are all consistent with stratigraphic relationships where available (Table 1). All ages are calibrated with IntCal13 (Reimer et al., 2013) implemented in OxCal 4.2 (Bronk Ramsey, 2009), and ages discussed in the text are medians from the calibrated probability distributions in years before 1950 CE, unless otherwise specified. See Table 1 for full details of radiocarbon results.

2.3. Terrestrial cosmogenic nuclide (TCN) dating

To constrain the ages of older terraces preserved in the southern Issyk-Kul area that are beyond the limit of radiocarbon dating (~45 ka), we collected samples of sediment to measure inventories of cosmogenic 10Be in quartz. We collected most samples from terrace gravels in vertical profiles to assess the inherited proportion of 10Be and verify post-depositional stability of the gravel deposits (Anderson et al., 1996; Phillips et al., 1998). We analyzed the 0.25–0.50 mm size fraction of the matrix of the terrace gravels, and analyzed ~15–30 g samples which contain thousands of individual grains, that well resolve the mean concentration of 10Be for a given depth interval (Phillips et al., 1998). We selected sample sites based on the presence of an undisturbed gravel/cover contact, and minimal cover sediment thickness to ensure the surface TCN inventory had not been altered by erosion of the gravel deposit. We collected samples over ~6 cm depth ranges at intervals of 20–80 cm starting at the uppermost level of fluvial gravel, to a total depth of ~200 cm. The samples were processed in the geochronology laboratory at the University of Cincinnati. Quartz was chemically separated and leached following Kohl and Nishiizumi (1992). Following dissolution and addition of 9Be carrier, Be was extracted using ion exchange chromatography. AMS measurements of 10Be/9Be ratios were made at the PRIME Laboratory, Purdue University.
Concentrations of $^{10}$Be/g quartz are corrected for process blanks, with errors propagated following the procedure outlined by Balco et al. (2008). We determined model ages using the online CRONUScalc program (Marrero et al., 2016), using time-dependent rates from the Lifton et al. (2014b) nuclide-dependent scaling scheme and updated calibration data (Borchers et al., 2016). Production rates are modified for topographic shielding (negligible for all sites here) and the thickness of the sampled interval. The maximum ages are calculated with the CRONUScalc program (Marrero et al., 2016), with the attenuated production rate controlled by modifying the topographic shielding parameter. These cases bracket the most likely scenario of the thickness of the capping sediment progressively accumulating through time.

### 3. Results

#### 3.1. Issyk-Kul lake history

We focus on resolving ages for the two prominent shorelines above the modern lake level. The lower shoreline rings Issyk-Kul at an elevation of ~1622 m asl, consistent with a lake surface at the elevation of the modern sill at the western end of Issyk-Kul. The higher shoreline is more discontinuously preserved as a break in slope at an elevation of 1560–1680 m asl.

We radiocarbon dated the younger highstand (~1622 m asl) at the bay at the mouth of the Tong River (Fig. 4a) using organic material from two horizons in a package of lacustrine sediment. The elevation of top of the deposit is ~1620 m asl, and the sediment was likely deposited while the lake was at or rising to the 1622 m asl sill level. The lower sample yielded an age of ~1650–1660 CE, and the sample from 3.2 m higher stratigraphically above the modern lake level.

We radiocarbon dates the younger highstand (~1622 m asl) using organic material from two horizons in a package of lacustrine sediment. The elevation of top of the deposit is ~1620 m asl, and the sediment was likely deposited while the lake was at or rising to the 1622 m asl sill level. The lower sample yielded an age of ~1650–1660 CE, and the sample from 3.2 m higher stratigraphically above the modern lake level.
Canyons expose continuous sections of lacustrine sediment between ~1620 and 1660 m amsl. The shoreline is absent where cross-cut by younger river terraces and active alluvial fans. In the eastern end of the Issyk-Kul basin, the shoreline is expressed only as a break in slope cut into the limbs of low relief active folds that have uplifted the Neogene section. We investigated and dated Issyk-Kul lacustrine sections associated with the higher shoreline in three locations along the southern shore and in the northwestern arm of Issyk-Kul: Pristan, Ak-Terek, and Kok-Moynok (Fig. 2).

The Pristan site lies near the mouth of the abandoned course of the Tong River (Fig. 4a). We sampled >15 m of near-shore lacustrine stratigraphy exposed in an irrigation-related gully (Fig. 4c). Laterally continuous beds are traceable over the ~150 m long gully. Roots associated with a surface at a depth of 6.7 m are the only evidence for a significant unconformity breaking the depositional sequence. The section consists of well-sorted, interbedded layers of silt and sand, with some pebble gravel beds at depths >14.5 m and immediately above the horizon with roots (Fig. 4c). Radiocarbon dating of four root and shell samples yield ages consistent with an average rate of deposition of 0.8 m/ka over the period 42–25 cal ka BP (Fig. 4f, Table 1). As the top of dated section is a few meters below the erosional shoreline angle of the highstand, the upper age

Fig. 4. Location, stratigraphy, and radiocarbon dating of late Quaternary shorelines. (a) and (b) Maps of sample locations in the Tong and Ak-Terek areas, respectively (ASTER image, band 3); see Fig. 2 for locations. White and black lines show the Holocene and higher late Pleistocene remnant shorelines, respectively. Numbers of sample locations correspond to Table 1. (c), (d), and (e) Stratigraphic sections showing positions of samples analyzed in this study and relationship to previous IRSL chronology in Ak-Terek area. Sample ages given in ka. (f) Age-elevation history of lake sediment deposition at Pristan site.
of ~25 cal ka BP represents a minimum for the end of the late Pleistocene high lake level. While we lack direct evidence for or against uplift of these lake deposits from the elevation where they were deposited, the average uplift rate must have been lower than the rate of transgression over the >18 ka period of high lake level.

We dated lake sediment between the two shorelines east of the mouth of the Ak-Terek River that were previously dated using a single aliquot IRSL method (Bowman et al., 2004a, Fig. 4b). We sampled mollusk shells from lacustrine sediment exposed in the cliff above the historical (~1622 m asml) shoreline. Thus, in comparison to the Pristan section, these samples come from lower elevation and farther from the local shoreline of the eventual highstand. We radiocarbon dated one sample from the lower part of “station 11” of Bowman et al. (2004a). This sample yielded an age of 42.1 cal ka BP, and lies only 4.6 m below layers that yielded IRSL ages of 15.7 and 14.7 ka (Fig. 4e). Our other samples come from an outcrop ~5.5 km east of the Ak-Terek River mouth (“station 15” of Bowman et al. (2004a); Fig. 4b). Ages of our radiocarbon samples in this section are 42.8 cal ka BP and 38.8 cal ka BP, below and inter-bedded with IRSL samples in the range of 24.0 to 10.5 ka (Fig. 4d). Additionally, we radiocarbon dated oxidized wood preserved in lacustrine fill below the QII and QIII terraces near the mouth of the Ak-Terek River (Fig. 4b). This sample yielded an age of 44.9 cal ka BP, consistent with the oldest ages of the shoreline deposits, although near the limit of the radiocarbon technique.

As the Ak-Terek area dated sections are both low in the total lacustrine section relative to the measured ages, the measured ages are consistent with the lowest exposed lake deposits having been deposited >43 cal ka BP. However, there is a clear mismatch between the ages obtained via the IRSL and radiocarbon methods from the same sediments, with the IRSL ages being systematically younger by ~19 ka. Calibration of the radiocarbon ages cannot explain the discrepancy, as the calibration curve is well constrained for the age range implied by the IRSL dating (Reimer et al., 2013). While all of the radiocarbon samples except for the in situ root of sample RW-PRI-4 at Pristan are from mollusk shells, the excellent age-elevation relationships we obtain from the radiocarbon ages from single outcrops and correlation between sites suggest that significant recycling is unlikely to be responsible for the observed discrepancies. The ~840 yr residence time observed in modern lake carbonate shells (Ricketts et al., 2001), suggests reservoir effects are also unlikely to explain the inconsistency. The IRSL measurements were not corrected for possible fading and there were likely temporal variations in dose rate due to moisture content changes in the lake sediment and erosion of shielding sediment. Given these uncertainties, and the more consistent radiocarbon age-elevation relationships, we prefer the age control from the radiocarbon dating to define the lake history.

The third location where we obtained estimates of lake sediment age is from the Kok-Moynok basin, west of modern Issyk-Kul (Fig. 2). The total thickness of currently subaerial lacustrine material in Kok-Moynok is thicker than observed elsewhere in the Issyk-Kul basin, reaching ~100 m, where the river canyon exposes levels below the modern lake. At the time of the deposition of the Kok-Moynok lakebeds, Issyk-Kul extended farther west, with the Chu River emptying into the lake west of the modern sill (Fig. 2). The similar elevation of the uppermost lake sediments preserved in the Kok-Moynok basin and the upper shoreline in the broader Issyk-Kul basin suggests that these sediments record the same highstand (Burgette, 2008). This inference is corroborated by a radiocarbon age of 27.1 cal ka BP from a sample of shell we collected at a stratigraphic level 70% above the base of the deposit. This age is correlative with the middle to upper portion of the Pristan section (Table 1).

3.2. Fluvial terrace development in southern Issyk-Kul

Flights of well-developed terraces flank most of the rivers that drain out of the Terskey Range into the southern margin of Issyk-Kul. Given the diversity of faults and folds that have actively deformed the southern margin of Issyk-Kul during the Quaternary, different rivers experience varying rates of rock uplift along their lengths, leading to varying patterns of incision and aggradation. We have studied terraces associated with eight southern Issyk-Kul rivers (Burgette, 2008). Here we focus on the three best examples: the Ak-Terek River in the west, and the Barskaun and Jetyn-Oguz Rivers in the eastern basin (Fig. 2).

3.2.1. Ak-Terek River

The Ak-Terek River maintains the only water gap in a 40 km long granitic bedrock ridge, which is uplifted by a south-vergent reverse fault against the regional northward slope down from the Terskey Range (Fig. 2). Downstream of the gorge through the basement block, a set of terraces is well preserved, although deformed by two south-vergent faults and related fold pairs (Fig. 5a). In the northern Ak-Terek area an extensive QII terrace caps the local topography as the terrace gravel is more resistant to erosion than the siltstone and interbedded sandstones of the upper Neogene section (Fig. 5a). Where terrace sediments are well exposed they are composed of cobble gravel that is generally <5 m thick. QII terraces are inset into the broad QI surface, along the modern Ak-Terek River canyon (Fig. 5a). Two prominent terraces, QII and QIII, are preserved in the northern portion of the area, and likely merge into a single aggradational surface in the synclinal Bar-Bulak valley, labeled QIII11. The paired QIII11 terrace appears to bury the more steeply dipping QI surface at the southern edge of the valley (Fig. 5a). The base of this QIII11 fill deposit is not well exposed, but at least 20–30 m of cobble gravel accumulated in the active syncline.

The youngest prominent terraces are subtly distinct paired QII1 and QII2 surfaces that diverge downstream from a common surface in the Bar-Bulak synclinal valley (Fig. 5a). Interbedded fluvial and lacustrine sediment below the QII2 surface extends below the level of the modern river and tributary gullies along most of the river. Where visible, the basal unconformity of the QII2 fill is irregular, burying underlying topography. The QII1 terrace is cut into the QII2 fill on the east side of the river, and has a strath morphology on the west side of the river where ~2 m of gravel overlie Neogene rock. At the mouth of the river, the uppermost QII1 terrace gravel is sandwiched between two distinct packages of lacustrine sediment, which we interpret to have been deposited during the late Pleistocene high stand of Issyk-Kul. The inset QII1 terrace cuts through the lacustrine sediment and extends north to where it is truncated by the ~1620 m late Holocene shoreline (Fig. 5a).

3.2.1.1. Ak-Terek profiles. Profiles of the Ak-Terek river terraces constructed from total station surveys and topographic and bathymetric maps show the results of terrace formation during progressive tectonic deformation and lake level fluctuations (Fig. 5c and d). The linear portion of the modern river profile is 1.2 km long, and becomes concave over the northern 4 km of the profile, near Lake Issyk-Kul (Fig. 5c and d). Below the modern lake surface, the bathymetry shows an alluvial fan deposited at the mouth of the Ak-Terek River, biected by a canyon. Extrapolation of the younger QI1 terrace profiles to the top of the alluvial fan suggests that the QII1 and/or QIII2 Ak-Terek River were the source of the fan sediment. The northern edge of the alluvial fan is convex, steepening toward the central lake floor. The canyon incised into the fan extends to ~100 m below the modern lake level, and it the offshore portion of this canyon projects to the profile of the modern river upstream of the
Fig. 5. (a) Quaternary geology and geomorphology of the Ak-Terek River area. (b) Map showing broader area and profile line, with distances indicated in km. (c) Long profile of river terraces along the Ak-Terek River and positions of Quaternary structures. Filled symbols show points surveyed with a total station, and open symbols were digitized from 10 m contour topographic maps. (d) More detailed view of central profile, with slopes of terrace remnants, dated sample locations, and ages indicated.
flat reach that is graded to the modern lake surface (Fig. 5c). The submerged channel suggests that the local concavity of the northernmost modern Ak-Terek River is due to an ongoing adjustment to the latest Holocene highstand via deposition in the lowest reach. Terrace remnants are quite planar except for offsets across localized structures, which suggests the nonlinear modern river profile is not representative of the conditions when the terraces were abandoned (Fig. 5d). The slope of the QIII terrace is 0.7° on both sides of the Kyzyl-Tau anticline, lower angle than the modern incised river, but steeper than 0.6° QII fill surface.

All of the QII and QIII terraces are progressively deformed across the Kyzyl-Tau anticline-syncline pair, and the pre-QIII terraces are also offset by the faulted fold to the north. Although it is difficult to match the higher surfaces across the fault to the north, there has been considerable offset across this structure. However, the QIII surface, which is clearly warped across the southern Kyzyl-Tau fold pair, exhibits little to no deformation at the northern fault. The QII terraces are offset by an additional fold pair that dies out west of the QIII terrace remnants. Terraces slope northward at progressively steeper angles with age across the entire profile length, and correlative terraces have similar angles away from the active structures. This long-wavelength progressive northward tilting reflects the uplift of the Terskey Range (Burgette, 2008).

3.2.1.2. Ak-Terek chronology. The geologic and geomorphic relationships between the dated lake sections and the younger terraces strongly constrain the ages of QII and QIII. The gravel of the QIII terrace is clearly older than the youngest late Pleistocene lakebeds. However, the upper part of the QIII terrace gravel also overlies older lakebeds at the mouth of the river, as well as upriver. We have two direct radiocarbon ages for the lacustrine and alluvial fill below QII, showing they are ~43 cal ka BP or older (Fig. 5d; Table 1). Ages for charcoal from a burn horizon in cover sediment on the QIII terrace yield limiting young ages of 34.9 and 37.2 cal ka BP, indicating the top of the QII fill must have been at least locally abandoned by that time. Abandonment of the inset QII level is broadly bracketed by the youngest sediment of the late Pleistocene high stand (<25 cal ka BP), and the late Holocene high stand at the modern sill elevation.

To define the ages of the older QII and QIII terraces we collected 10Be profile samples from pits 0.5 km north of the Kyzyl-Tau anticlinal hinge (Fig. 5). The sampled QII terrace forms the local hilltop, and the QIII sample site is at the distal fringe of a wedge of slopewash and colluvial sediment shed off of the riser up to QIII 1 0Be depth profiles show well-constrained exponential decreases in the concentration of 10Be with depth, consistent with the relative ages of the terraces (Fig. 6a and Table 2). Since the samples came from the interiors of terrace remnants, there is little possibility that the coarse gravel has been eroded since the abandonment of either surface. The low 10Be concentrations at the maximum sampled depths suggest a minimal inherited contribution to the observed inventory, implying relatively rapid erosion and transport rates in the upstream portion of the catchment (Fig. 6a).

The greatest uncertainty associated with converting the measured 10Be concentrations to model ages relates to the history of the fine-grained cover material overlying the terrace gravel. The lack of gullies developed on the smooth surface of the cover wedges suggests that there has not been significant degradation of the capping wedges. By assuming a time-averaged zero thickness for the shielding material, we calculate robust minimum 10Be depth profile ages of 216 and 127 ka for QII and QIII, respectively (Table 2). If the current thicknesses of the covering sediments formed instantly upon terrace abandonment, the ages would be 279 and 189 ka, respectively, using a cover sediment density of 1.5 g/cm³ (Table 2). A scenario where the fine-grained cap grew monotonically through time would yield an age bracketed by these two estimates.

A 10Be age of 196–140 ka was obtained for granitic cobbles on the surface of the QIII terrace near the mouth of the Ak-Terek River (Fig. 5, Table 2). This age range reflects clast degradation rates from 0 to 0.001 mm/yr, and 1σ uncertainties from the CRONUS calculator (Marrero et al., 2016) propagated from analytical and calibration uncertainties. As this age is at least 20 ka younger than the robust minimum of the profile age for this terrace upstream, it appears that the surface cobbles were unlikely to have been exposed at the surface over the entire terrace history. A greater clast surface denudation rate is unlikely to explain the discrepancy, as the bracketing maximum erosion rate we use implies the sampled cobbles lost nearly 18 cm of material. Such erosion seems unlikely given the rounded, equant appearance of the clasts and their
the Terskey Range (Fig. 2). The Barskaun River has deposited an
central Issyk-Kul, emerging north out of a deep glacial trough in
3.2.2. Barskaun River
river and dated late Quaternary terraces representing different
of the slope-age relationship (Fig. 7), consistent with the modern
Errors are 1
2 and QIII
3 terraces dated in the basins to west (Thompson
rates of incision of Ak-Terek River from the dated terrace sequence and rates of
rate of 0.001 mm/yr for surface samples and present cover sediment thickness over history of terrace. The terrace ages are consistent with relatively constant rates of
Similarity to deeper, pristine portions of the terrace gravel.
The terrace ages are consistent with relatively constant rates of river incision and deformation over the late Pleistocene when averaged over multiple cycles of terrace formation (Fig. 7). Aban-
donment of the QII terraces is well defined to occur after the 25 cal ka BP age of the youngest late Pleistocene lake sediment that we dated, and we use a 12 cal ka BP minimum age based on incision of QII and QIII terraces dated in the basins to the west (Thompson et al., 2002). The good
Fig. 7. Rates of incision of Ak-Terek River from the dated terrace sequence and rates of
tilting for the two age scenarios. The linearity of the age evolution suggests that the
shielding history of the two upper terraces dated with 10Be did not differ substantially.
merge at the steeper rangefront. For both the QIII and QII surfaces, there is little clast-supported gravel overlying the matrix-supported boulder gravel across which the surfaces are cut.

An extensive QIII terrace is inset well into the coarse sediment capped by the QII and QIII alluvial surfaces, which forms the prominent fan surface at the mouth of the canyon (Fig. 8a). The Barskaun QIII terrace coalesces with an alluvial fan terrace at the mouth of the Tamga River to the west. Exposures below the QIII terrace show >100 m of gravel, with no exposure of underlying Neogene sedimentary rock. In contrast to the deposit underlying the QII terrace, the QIII gravel is clast-supported, with rounded clasts that range from pebble to boulder size (up to 4 m in maximum dimension). The gravel deposit is generally massive with the exception of a mud-cemented layer that stands out in relief from the surrounding walls ~2–3 km upstream from the river mouth.

The fan shaped QIII surface is pristine with no significant sublevels. The northern edge of the terrace surface is truncated by a 2–20 m cliff down to the late Holocene shoreline. The only suggestion of a higher lake level preserved on the surface is a small remnant of sediment with a maximum elevation of ~1648 m amsl that forms a wedge thickening toward the lake near the eastern edge of the QIII alluvial fan (Fig. 8a).

3.2.2.1. Barskaun profiles. Barskaun River terrace profiles collected with GPS and from topographic maps are projected to a smoothly curved line that generalizes the modern Barskaun inner canyon and alluvial fan axis (Fig. 8b and c). The profiles collected from the west and east sides of the QIII surface show that this terrace is precisely paired across the modern canyon. The overall shape of the QIII terrace in profile is well approximated by a slope of 1.8°, with a possible convexity of ~0.1° over the 10 km surveyed reach (Fig. 8c). Bathymetric data from the top of the sub-lacustrine fan surface project to the subaerial portion of the QIII surface (Fig. 8c). The alluvial fan steepens below a depth of 50 m to a mean slope >3°. Surveyed points collected on the top of the resistant mud-cemented bed in the QIII gravel yielded a slope of 2.8°. This steeper internal stratigraphy suggests that constant-age horizons in the QIII gravel may not parallel the top of the terrace surface. This could result from the alluvial fan prograding northward rather than through parallel vertical growth over its entire modern extent, or that the progressive northward tilting extends north across the lake shoreline.

In contrast to the linear profile of the QII terrace, the modern Barskaun River shows a more complex geometry that we infer is transient (Fig. 8c). In the upper reach of the profile, the modern lies only ~20 m below the top of the QIII gravel. The river steepens from a slope of ~2.0° in this southern reach to a maximum value of ~2.6° at a profile distance of 6.2 km. North of the steepest reach, the profile is abruptly concave, decreasing to a slope of 0.3° over a profile distance of 3 km (Fig. 8c). This reach of the river coincides with the transition from a straight to braided channel habit. The lake bathymetry reveals a canyon cut through the axis of the offshore fan at the mouth of the Barskaun River. This submerged canyon is incised to >100 m below the modern lake surface, and its profile projects to the steeper subaerial reach.

We interpret this complex profile of the modern river and offshore channel to have formed initially by incision of a canyon through the QIII alluvial fan following lake regression. The subsequent late Holocene highstand raised base level and caused deposition of a wedge of sediment in the downstream portion of the canyon. The sub-lacustrine steep face of this wedge represents foreset beds of a growing delta, potentially analogous to the growth of the lower part of the QIII fan, as shown by its basal bathymetry and the geometry of the indurated bed.

The older terraces that cap the topography along the rangefront dip much more steeply northward than the QIII terrace (Fig. 8c). The highest QII surface dips 5.6° along the profile and ends to the south at a much steeper hillslope developed across the older Quaternary sediment. The profiled QIII terraces show a localized steepening near the rangefront, but do not match one another in detail across the canyon, likely in part due to the change in range front orientation across the Barskaun River (Fig. 8a and c). In spite of the complications of the terrace geometries, it is clear that terraces are progressively tilted with age along the rangefront.

3.2.2.2. Barskaun chronology. Over most of the broad QIII terrace, there is < 0.5 m of cover sediment overlying the terrace gravel. However, on the far western edge of the QIII alluvial fan, west of the Tamga River, a QIII terrace remnant sheds fine-grained colluvial and slipwash deposits onto QIII gravel (Fig. 8a). We collected three radiocarbon samples through this 4.95 m thick wedge near the riser to the QIII terrace where exposed in a young gully. The lowest radiocarbon age of 37.8 cal ka BP is on terrestrial gastropod shells from a 30 cm thick package of well-bedded sand and silt immediately overlying the QIII gravel (Table 1). Shells from an extremely shell-rich layer midway through the section (2.5 m above the gravel) yielded a radiocarbon age of 8.0 cal ka BP. The uppermost sample we dated is charcoal and comes from 1.4 m higher, ~1 m below the surface of the colluvial section, yielding an age of 3.1 cal ka BP.

The ages from the section overlying the QIII gravel at Tamga suggest that deposition on the terrace following abandonment spanned >30 ka, with half of the thickness accumulating in the past 8 ka, before the recent onset of gullying. The setting and stratigraphy of the dated section suggests that aggradation of the QIII surface had reached the upper surface, at least at this location on the Barskaun fan, prior to ~38 cal ka BP, during the middle to early part of the late Pleistocene lake highstand. The fact that evidence of this highstand is not preserved across the QIII alluvial fan (except perhaps the local outcrop at the northeastern Barskaun alluvial fan, Fig. 8a) suggests that the Barskaun and Tamga Rivers remained near the top of the QIII fill until well into the subsequent deep lake regression. This occupation of the surface must have occurred over a long enough period to allow migration of flow over nearly the entire surface, leaving only isolated remnants of the original aggradational surface like the sampled location, which lies in the protected lee of the older QIII terrace outcrop. The offshore continuation of the QIII alluvial fan in the bathymetry also argues for the river maintaining a relatively static profile until well into the regression.

3.2.3. Jety Oszug River

The greatest number of terraces in the Issyk-Kul basin are preserved above the Jety Oszug River in the 11 km between the base ment unconformity and the rangefront (Fig. 9a). The rangefront at Jety-Oszug is defined by the intersection of an exceptionally broad and well-preserved north-dipping QIII surface and the low-gradient floor of the modern Issyk-Kul basin. North of the rangefront, the river runs on the top of basin sediments.

The QIII terrace is the most prominent terrace in the vicinity of the Jety Oszug River (Fig. 5a). This terrace is highly planar and extends for 20 km along-strike to the east (Fig. 10a). The topographic surface of this QIII terrace does not exhibit a fan shape and cannot be clearly associated with deposition from any individual modern streams. West of the Jety Oszug River, the QIII surface is locally visible, but it has been cut by numerous minor fault scarps (the most clear examples are mapped on Fig. 9a) and more thoroughly dissected by gullies. Several QII sublevels with southward-increasing separation are inset into the main surface in the Jety Oszug River canyon near the rangefront. Generally, the thicknesses of the Jety Oszug terrace
Gravels are poorly defined, as the lower contacts of the terrace gravel with the coarse-grained Neogene sediment are poorly exposed on the vegetated risers. However, on the steep canyon walls near the southern limits of the upper QII surfaces, there are...
Fig. 9. (a) Late Quaternary river terraces along the Jety Oguz River. A well preserved, broad QII surface caps the rangefront along this portion of the eastern Issyk-Kul basin, and multiple levels of QIII terraces are preserved along the modern river canyon. See Fig. 5 for legend. (b) Location of profile line, showing distance (km) and extent of Fig. 9b. (c) Long elevation profile of terraces along the Jety Oguz River and locations of active structures. (d) More detailed portion of profile with terrace names and angles indicated. North-dipping terrace remnants intersect at the rangefront and show a consistent pattern of steepening with age.
maximum thicknesses of 10–20 m of gravel overlying Neogene sediment. As the vertical separations between the different QIII levels are greater than the gravel thicknesses for the upper QII terraces, the gravel of the inset terraces was most likely deposited sequentially between incision events, (as opposed to multiple cut surfaces developed on a single fill). Exposure of colluvial/aeolian deposits at the outer edge of the QII terrace remnants show there is generally <5 m of sediment overlying the gravel surface except against north-side-up fault scarp and infilled gullies cut into the gravel surface.

Five distinct QIII terraces are inset below the extensive QII surface. The QIII terraces are generally unpaired except for the QII terraces, which is only a few meters above the modern river level at the rangefront, and connects to a large fan-shaped surface north of the rangefront. In the canyon, QIII terrace gravel deposits are covered by thick (10–50 m) accumulations of cover material shed from outcrops of weakly indurated Neogene sedimentary rock exposed in the risers above. The gravel/colluvium contacts of the higher QIII terraces are generally distinct as a marked, continuous exposure of colluvial/aeolian deposits at the outer edge of the QII terrace remnants show there is generally <5 m of sediment overlying the gravel surface except against north-side-up fault scarp and infilled gullies cut into the gravel surface.

3.2.3.2. Jety Oguz chronology. We sampled the QII terrace east of the Jety Oguz River for 10Be profile dating ~10 m east of the top of the terrace, suggests a negligible inherited contribution.

We determine an absolute minimum 10Be depth profile age for the QII terrace of 129 ± 14 ka by using the unshielded intact gravel as a limiting case (Fig. 6b). Using the scenario of the present covering sediment geometry having been present since the time of the terrace abandonment yields an age estimate of 219 ka, assuming a cover sediment density of 1.8 g/cm³. We use a greater density than at Ak-Terek due to the greater concentration of large clasts incorporated in the cover sediment. This provides a maximum limiting age for the likely scenario of the cover sediment monotonically increasing in thickness through time. The 10Be concentration at the lowest sampled depth of 230 cm is 7% of that at the top of the profile, similar to the estimated 10Be production rate at this depth (Table 2). This coupled with the low 10Be concentration of the modern Jety Oguz River sand (~1% of the upper profile concentration) suggests a negligible inherited contribution.
to the age (Table 2). Progressive growth of the cover sediment is supported by the relatively low concentration of $^{10}$Be from the sample within the cover. Using end-member shielding scenarios to bracket the history of the 22 cm of sediment presently overlying this sample returns an age range of 184 to 142 ka. With only a single sample we cannot define the inherited $^{10}$Be concentration for the cover, but we favor the interpretation that the present cap was deposited ~180 ka, which supports the older end of the age range we calculate for the gravel.

4. Discussion

4.1. Synthesis of Issyk-Kul terrace history

We integrate the data from the locations presented here, our other observations (Burgette, 2008), and previous work (Section 1.2) to synthesize a history of terraces in the Issyk-Kul basin. Our data for terraces older than QIII place broad constraints on the timing of formation. The best data are from the Ak-Terek River, where $^{10}$Be profile ages give estimates for the timing of the formation of two pre- QIII terraces with QIV morphologies. Roughly equal amounts of incision and tectonic deformation in the intervals between the abandonment of QIII and earlier QIV and later QIV are consistent with the approximately equal intervals derived from the $^{10}$Be dating (Fig. 7). The numerical ages suggest the major terraces formed on a cycle with a period of ~100 ka over the past 300 ka (Fig. 11).

The late Pleistocene highstand of Issyk-Kul was an important event in the most recent river terrace-forming cycle (the past ~120 ka). Based on radiocarbon age and field relationships, Lake Issyk-Kul rose above its modern sill at about 43 cal ka BP or slightly earlier, and reached a highstand ~33 m above the modern sill by ~25 cal ka BP (Fig. 12). Based on the geometry and chronology of the lake deposits, the sill for this highstand must have been downstream of the western Issyk-Kul basin, in Boam Canyon. One possible source of the dam is a landslide in the gorge. Alternatively, the sill may have grown through colluvial sedimentation in the gorge while the Chu River was routed into a closed, low standing Lake Issyk-Kul. Once this high sill was breached post-25 cal ka BP, the Chu River likely re-occupied its modern route west of Lake Issyk-Kul, and the closed lake experienced a regression of 50–100 m below present (Fig. 12). The lake subsequently filled back to overflow levels in the latest Holocene, based on our dates and historical accounts (Ricketts et al., 2001).

The development of QIII terraces is much better defined than the older terraces, and is generally consistent for all of the river valleys along the southern margin of Issyk-Kul. The earliest event in the relative chronology for the construction of QIII is a preceding interval of incision and valley widening greater than the Holocene. Even along reaches where preserved higher terraces show a clear trend of fluvial incision over the late Quaternary, the base of the QIII fill is not exposed in the modern inner gorges. The base of the QIII fill is only exposed in the areas of the highest rock uplift rates, such as near anticlinal hinges along the Ak-Terek River and the southernmost terrace remnants along the northward-tilting Jety Oguz area. One possible cause of such deep incision is a more extreme lake regression than the 50–100 m below modern level regression documented for the early Holocene (e.g., Zabirov et al., 1973; Gebhardt et al., 2016). The bathymetry of Issyk-Kul has low gradient areas >150 m below the modern sill that have been interpreted as terraces (De Batist et al., 2002). Seismic reflection data have also been interpreted as showing deposition of deltaic sediment over 400 m below the modern lake level (Gebhardt et al., 2016).

Radiocarbon ages near the top of the QIII fill and interfingering older lake beds are consistently >40 cal ka BP, near or beyond the limits of radiocarbon. Given the very low $^{14}$C concentrations and stratigraphic positions of the samples in the upper part of the fill, our radiocarbon ages place little constraint on the initiation or duration of the QIII fill, (e.g., >100 m thickness at the Barskaun River). The top of the QIII fill was at least locally abandoned near the edges of the river valleys by 38–37 cal ka BP, based on the oldest ages from cover sediment on QIII at the Ak-Terek and Barskaun Rivers. The QIII cut-strath terrace at Ak-Terek is inset only ~5 m below the top of the QIII gravel where we dated the cover wedge on QIII (Fig. 5d); however, this terrace post-dates the termination of the late Pleistocene lake highstand (~25 ka) 5 km downstream at the Ak-Terek mouth. Likewise, the QIII alluvial fan at the mouths of the Barskaun and Tamga rivers appears to have been maintained near its upper surface until regression of the lake below the modern sill. Based on Thompson et al. (2002), deeper incision likely occurred ~15–13 cal ka BP.

Field relationships at Ak-Terek show that deposition of the QIII gravel overlapped in time with deposition of the higher lake deposits above the modern sill elevation. However, the continuous

![Fig. 11. Correlation of Issyk-Kul river terrace formation with terrace chronologies from the central Kyrgyz Tien Shan (Thompson et al., 2002) and Toru Aygir River, northern Issyk-Kul (Selander et al., 2012). Terrace names are as mapped locally. Timing of Issyk-Kul lake high stands is based on our observations. Timing of glacial advances from dated moraines in the Kyrgyz Tien Shan (Koppes et al., 2008; Narama et al., 2007; Zech, 2012; Lifton et al., 2014a; Blomdin et al., 2016), and global climate from a marine oxygen isotope compilation (Imbrie et al., 1984). Gray bands show times when global curve shows greater than average $^{18}$O values (times of greater global ice volume). MIS column indicates marine oxygen isotope stages. Generally, the current set of dates from the top of the QIII fill from Kyrgyz Tien Shan terraces is consistent with synchrony of aggradation. Major aggradational terrace forming events for older terraces appear to have occurred once per global glaciation, although the morphologic correlations do not match the numerical dating correlation.](http://example.com/figure11.png)
were abandoned ~15.7−13.5 cal ka BP, and incised by narrow modern gorges (Thompson, 2001; Thompson et al., 2002). In the Naryn basin, charcoal leaves from the middle of the alluvial fill of the Kajerty River QII terrace yielded ages of ≥43 ka BP (Thompson, 2001). On the north side of Issyk-Kul, a surface mapped as QII yielded age estimates of 85.6 ± 7.6 ka and 105−57 ka using TCN and luminescence dating, respectively (Bowman et al., 2004b; Selander et al., 2012). Similar to our data from southern Issyk-Kul, this suggests the QII strath and much or the entire fill were developed early in the last glacial cycle. The age from the uppermost portion of the gravel shows that the Kajerty River was at the top of its fill at ~14 cal ka BP, in contrast to the Issyk-Kul rivers that subtly incised below the top of the QII fill prior to 35 cal ka BP. However, the dated samples at the Kajerty River were collected near the modern river and distant from the riser to the terrace above, so while incision is well defined locally, most of the fill may have been deposited earlier and the river had remained near the top of the fill until the deep incision occurred. Ages from QII terraces below the QII surfaces are indistinguishable from the main QII surface, suggesting that incision proceeded relatively rapidly (Thompson, 2001). Another difference between the Issyk-Kul river terraces and those from the central Kyrgyz Tien Shan is presence of an early to middle Holocene QII terrace associated with most of the rivers studied by Thompson (2001). The general lack of this terrace in the Issyk-Kul basin likely reflects the incision of the Issyk-Kul streams below the modern base level in response to the regression of the lake during the early Holocene.

The ages of Issyk-Kul river terraces are consistent with the timing of regional glaciation and suggest the end of aggradation is temporally decoupled from the timing of significant incision. Although the age control leaves uncertainty regarding the timing of accumulation of the QII gravel, there is strong evidence for aggradation being largely complete prior to the global last glacial maximum (as defined by Clark et al. (2009)) during marine oxygen isotope stage [MIS] 2 at ~26−19 ka). The uppermost fill was also at least locally abandoned ~20 ka prior to the likely timing of deep incision (16−13 ka) during the most recent global glacial-interglacial transition at ~19−10 ka (Thompson et al., 2002, Fig. 11). The 16−13 ka river incision dated by Thompson et al. (2002) slightly precedes the ~12.5 ka age of the most extensive late Quaternary deglaciation in the Tien Shan (Takeuchi et al., 2014). Hence, the slight start of aggradation of the QII gravel is likely linked to the most prominent latest Pleistocene glaciation in the Tien Shan. The emerging ages of glacial features throughout the Tien Shan suggest that the largest latest Pleistocene glacial advance occurred during MIS 4 (60−80 ka) rather than during the most recent global ice volume maximum of MIS 2 (Abramowski et al., 2006; Narama et al., 2007, 2009; Koppes et al., 2008; Zhao et al., 2010; Zech, 2012). More rapid accumulation of loess in the northern Tien Shan during MIS 2 relative to the rest of the late Quaternary has been interpreted to reflect arid conditions during MIS 2, supporting previous inferences from studies of glacial chronology (Youn et al., 2014).

Studies from the Terskey Range are conflicting with regard to the extent of glaciation during MIS 2, with the 10Be exposure dating study of Koppes et al. (2008) finding no evidence of advance far beyond the limits of modern glaciers, and the OSL dating studies of Narama et al. (2007, 2009) yielding multiple dates in the 29−21 ka range on moraines only ~350 m above the elevations of the MIS 4 terminal moraines. TCN studies of moraines associated with the highest peaks southeast of the Terskey Range found prominent MIS 2 age moraines with elevations consistent with more moderate advance of glaciers and evidence of more extensive pre-MIS 2 glaciers (Lifton et al., 2014a; Blondin et al., 2016). Although there are no published ages of the lowest moraines in the Barskana valley, the QII terrace is graded to the lowest preserved moraines.
This field relationship suggests that aggradation of the QII gravel was related to the 80–70 ka local maximum glacial advance and subsequent glacial interval. The association of late Pleistocene terminal moraines and outwash terraces is widely observed in other mountainous landscapes (e.g., Reheis et al., 1991; Pinter et al., 1994; Chadwick et al., 1997). Increased sediment production related to larger glacial extent is observed in modern landscapes (Hallet et al., 1996), and increased sediment flux into river systems provides a likely explanation for the aggradation.

Guided by the inferences from QIII linking the start of aggradation to the local last glacial maximum (~80–60 ka) and deep incision to the global deglacial transition, we can better assess our numerical ages for the older river terraces. Although uncertainties regarding the cover sediment history make the error bars broad, the robust minimum age limit for the QII terrace suggests its aggradation occurred prior to the MIS 5e interglacial (Fig. 11). The minimum age for the higher QIII surface indicates deposition prior to the MIS 7 interglacial. These broad chronometric constraints, coupled with the evidence that such terrace forms develop during periods of glaciations, suggest that the three recent major Ak-Terek terrace levels formed during each of the last three global glaciations. Apparent asynchrony of the latest Pleistocene glacial advances in the Tien Shan with respect to the global ice volume history (Koppes et al., 2008; Narama et al., 2009; Lifton et al., 2014a) preclude us from confidently making greater refinements of the terrace ages based on global climate records. The global climate regulation of terrace forming events is also supported by the approximately equal amounts of incision and deformation that occurred between QII and the terraces above and below (Fig. 7). The MIS 6 age of formation for QIII terrace at Ak-Terek may correlate with terraces that have QII morphologies along rivers in the central Kyrgyz Tien Shan dated by Thompson et al. (2002; Fig. 11). Their luminescence ages for sediments capping a terrace mapped as QII give a best estimate of terrace abandonment at 140.7 ± 8.5 ka, assuming synchronous incision occurred in three basins.

The situation at Jety Oguz, where we dated the QII terrace, is more complicated to resolve. The uncertainty age range of this terrace (219–129 ka) results from the relatively thick cover sediment at the sampled site. Based on the geochronology, the Jety Oguz QII would correlate with the QIII terrace at Ak-Terek, and many of the other Jety Oguz terraces would represent more minor terraces that were not formed or preserved in other areas. Alternatively, the QII terrace could be older than the sequence we dated at Ak-Terek. In the context of the total ~0.5 km incision of the Jety Oguz River, the dated QII terrace is inset only subtly into the expansive QI surface. The broad extent of the QII terraces in both the eastern and western Issyk-Kul areas suggests a fundamental difference in the geomorphic system during their development compared to the more recent terraces, perhaps driven by a common climatic regime. At Ak-Terek, the slope of the QII surface is approximately three times that of the QIII surface based on the initial slope from the regression shown in Fig. 7. If the deformation rate and style has remained constant, the age of the QII surface would be in the ~750 ka range. The expansive QIII terrace at Jety Oguz (Figs. 9 and 10) is the seventh to ninth major terrace above the river, and if there is one major terrace per ~100 ka glacial cycle, the QII surfaces could correlate along the entire basin. In this case, the much younger age from the Jety Oguz QII surface could result from the sampled location having had a thicker cover over much of its history. A time-averaged thickness of ~60–100 cm greater than modern cover thickness would explain the discrepancy (depending on the overlying sediment density and age difference between QII and QIII). In this case the switch in the nature of terrace formation would coincide with the ~800 ka mid-Pleistocene transition from ~40 ka to ~100 ka glacial cycles (e.g., Pisias and Moore, 1981; Ruddiman et al., 1986). The more rapid climatic oscillations of the earlier Pleistocene may have promoted broad planation rather than deep valley entrenchment (Bridgland and Westaway, 2008). However, if accumulation of sediment at the sampled Jety Oguz QII site has been monotonic and the age is correct, the conditions required for very broad planation must have been caused by more local factors that caused the generation of similar terrace morphologies in different parts of the Issyk-Kul basin at different times. Future dating of additional terrace levels at Jety Oguz and elsewhere would resolve which possibility is more likely.

4.3. Inferred controls on the terrace forming process

The general temporal link we find between alluvial aggradation during glaciation in headwaters areas and incision during deglaciation matches inferences made in previous studies of terraces in the Kyrgyz and Chinese Tien Shan (Molnar et al., 1994; Thompson et al., 2002; Poisson and Avouac, 2004; Lu et al., 2010) as well as many other mountainous areas elsewhere. However, the relatively well-dated history of southern Issyk-Kul QII terraces shows a different pattern of terrace development than previously inferred. In the conceptual model of Bull (1991), the switch from gravel aggradation to incision represents a threshold process, in which rivers switch relatively quickly from aggradation to deep incision back to the tectonically-controlled base level. Guided by this idea, one would expect the age for the uppermost surface of an aggradational deposit to be relatively close to the age of incision below it. Owing to the difficulty in getting multiple ages through a sequence of aggradational and degradational events to truly define the rates of these processes, there are few studies that offer good constraints for fill terrace deposits. Weldon (1986) used radiocarbon ages from within and above fill terrace deposits along Cajon Creek in southern California to show that a pulse of aggradation migrated upstream through time, with the peak fill closely followed by a switch to down-cutting. Pederson et al. (2006) obtained multiple ages from late Quaternary aggradational events along the Colorado River in the Grand Canyon, from which they interpreted fairly rapid switches from aggradation to incision.

Our data from the southern Issyk-Kul rivers show a different situation, where the river aggraded to the top of QIII and then remained near that level for a period of at least ~10 ka, and likely >20 ka (Fig. 12). This suggests that the rivers maintained a quasi-equilibrium long profile in spite of long-term tectonically-induced base level fall. Perhaps the tectonic base level fall was in part balanced by base level rise to the late Pleistocene high stand of the lake, but it motivates careful dating of incision and aggradation events in this and other regions.

4.4. Implications for tectonic geomorphology and neotectonics

4.4.1. Terrace correlation

Our numerical ages for terraces coupled with previous work, allow us to test the effectiveness of intra- and inter-basin terrace correlations made in the Kyrgyz Tien Shan on the basis of distinctive geomorphology. The set of numerical ages suggests that the terrace associated with the youngest significant aggradational event is correctly correlated as QIII around our study areas in the Tien Shan. The correlation is robust in spite of the QIII surface having a strath terrace morphology in places of higher uplift rates and being a fill terrace in other locations. The generally complete preservation of this latest Pleistocene terrace allows us to trace the top of the terrace gravel from locations with thin fluviial deposits into areas of thicker fill. Merritts et al. (1994) show that longitudinal variations in terrace character result from proximity to a
fluctuating base level control as well. These observations motivate cautious use of strath versus fill morphology as a correlation tool and careful longitudinal tracing of terraces in areas with discontinuous outcrops.

For terraces from the penultimate aggradational cycle, terrace correlations are more dependent on the preservation of terraces associated with particular rivers. The inter-basin terrace mapping in the central Kyrgyz Tien Shan by Thompson et al. (2002) is supported by consistent numerical ages of the abandonment of what was mapped as the QI terrace. However, along the rivers they studied, there were only two major levels of terraces with similar extensive morphologies. The mapped outcrop widths of the QI terraces are similar to those of the QII terraces, and the QII terraces do not appear to cap the topography to the extent of those from the Issyk-Kul basin. The Kyrgyz terrace correlation scheme, at least as applied here, involves identifying the two most prominent terraces and calling them QI and QII, with the QII surface near the local topographic culminations. The remaining terrace names are then interpolated between the major levels. As a result, the potential for miscorrelation is high if there are major variations in incision rates and the associated removal of uplifted terraces by erosion. If only the QII and QIII terraces west of the Ak-Terek River were currently preserved, we would have correlated the terrace flight in a different fashion, and comparison to incision rates from the rivers of the central Kyrgyz Tien Shan would be more straightforward.

The number and ages of terraces that survive along a given river is controlled by the pattern of river migration and lateral erosion following terrace abandonment. Terrace preservation and pairing is analogous to the problem of moraine preservation (Gibbons et al., 1984), and a consistent migration direction of a river or progressive valley narrowing promotes preservation of long terrace sequences. Preservation of the three major levels of late Pleistocene terraces along the Ak-Terek River from the last 300 ka was promoted by a net eastward migration of the river and the greater erosional resistance of the terrace gravel in comparison to the underlying late Cenozoic mudstones. The QI and QII terraces are unpaired, so we cannot assess how much wider the paleo-floodplains were at the time of terrace abandonment. The morphological distinction of terrace pairing has been used to identify climatically-controlled terraces (Bull, 1991; Merritts et al., 1994). However, the >10 m of gravel overlying the bedrock straths, and the lack of discernable cross-valley slope argue for the QIII Ak-Terek terraces representing major climatically-controlled events. The migration of the Ak-Terek River up-plunge relative to active structures was likely promoted by the inverted relationship between gravel and “bedrock” erodibility (Burgette, 2008).

4.4.2. Initial terrace slopes

Understanding the initial slope of terraces at the time of abandonment is key to using deformed terraces as quantitative markers of angular deformation. Most studies assume that the down-valley slope of the modern river provides the best estimate of the initial terrace geometry as well (e.g., Thompson et al., 2002; Amos et al., 2007). This assumption is likely valid in many cases when comparing a strath terrace to a modern river that is in a period of lateral planation across bedrock. Observations of differences in slope between aggrading and incising rivers suggest caution in making such comparisons between terraces and rivers in different stages of the terrace forming cycle (Leopold and Bull, 1979; Merritts et al., 1994). Another example of non-parallelism between a modern river and an initial terrace slope occurs when the modern river profile is in a transient state due to localized incision via knickpoint migration or local aggradation (e.g., Weldon, 1986). In addition to biasing estimates of tilting, non-parallelism between terraces and the modern stream profile will make calculations of incision rate vary longitudinally along a river. This effect adds to the temporal biases others have noted in calculating incision rates from terraces and rivers that are in different states of development (e.g., Gardner et al., 1987; Hancock and Anderson, 2002; Gallen et al., 2015).

Terrace and river profiles from the southern Issyk-Kul rivers illustrate all of these issues related to estimating incision and deformation. There are local concavities near almost every river mouth due to relatively recent lake level rise and the resultant aggraded wedges of sediment. Local convexities in the modern river profiles exist along both the Ak-Terek and Barskaun Rivers, related to either lithologic constrasts and/or transient incision patterns. Converting terrace profiles into height above the modern river would produce spurious curvature in terraces along these reaches that could be falsely attributed to tectonic deformation, especially in locations with more subdued signals. Although there is question whether an aggradational alluvial surface can be truly considered graded and in equilibrium (Phillips, 1991; Muto and Swenson, 2005), the consistent linearity of the terrace profiles suggests that the terraces represent a more ordered, stable geometry than the transient condition of the lower parts of the modern rivers.

The best evidence we have that the upper surfaces of the Issyk-Kul river terraces were formed with an essentially consistent gradient, especially in comparison to the curved modern river, comes from the Ak-Terek River (Fig. 5d). Long profiles of the three major late Pleistocene terraces are highly linear, and the consistent northward tilting of the Neogene sediment, the QI terraces, and the QII terraces gives a well-resolved tectonic forcing. The consistent slope-age relationship shown in Fig. 7 suggests the terraces formed at an initial slope of ~0.6°. An ambiguity for the older terraces remains in that the uppermost surface of the QIII deposit has a <0.1° lower slope than the younger cut QIII surface. The QI and QII gravel surfaces could represent either the top of the fill, analogous to QIII, or a cut surface planed across the fill in a time of relative stability such as the QII surface.

5. Conclusions

Our observations of the geology and chronology of terraces in the southern Issyk-Kul basin contribute to the understanding of process and timing of terrace formation in the Kyrgyz Tien Shan. Our numerical dating tests the Soviet-era correlation scheme, and we find that identification of the youngest major aggradational terrace (QIV) is likely correct in most cases. This terrace aggraded during and after the peak in the last major glaciation of the Tien Shan (~80–60 ka), which preceded the last global ice-volume maximum at 26–19 ka (Clark et al., 2009). This asynchrony between regional and global climate records illustrates the potential for major uncertainty by simply applying the global Quaternary climate curves to specific fluvial terraces. Our 10Be ages on older terraces at Ak-Terek and Jeti Oguz support the finding of Thompson et al. (2002) from the central Kyrgyz Tien Shan that there was one major aggradational terrace-forming event per ~100 ka glacial cycle, and that deep incision likely occurs during the major glacial-interglacial transitions. Although the few ages on terraces older than 50 ka support relatively synchronous aggradation and abandonment in the area, the terrace correlations we make based on radiometric dating are different in some cases than had been previously inferred based on morphologic criteria. Differential preservation of terraces between rivers and basins complicates accurately making relative correlations.

The presence of the large lake Issyk-Kul both aids and complicates deciphering the controls on river terrace development and abandonment. Our radiocarbon ages on lacustrine sediment at the
References


