(16 FIGURES AND TABLE FOLLOW TEXT)

# Onset, style and current rate of shortening in the central Tien Shan, Kyrgyz Republic

Kanatbek Abdrakhmatov, Academy of Science, Kyrgyz Institute of Seismology, Bishkek, Kyrgyz Republic Ray Weldon, University of Oregon, Department of Geological Sciences, Eugene, Oregon, U.S.A. Stephen Thompson, University of Washington, Department of Earth and Space Sciences, Seattle, Washington, U.S.A. Douglas Burbank<sup>1</sup>, Pennsylvania State University, Department of Geological Sciences, University Park, Pennsylvania, U.S.A. Charles Rubin, Central Washington University, Department of Geological Sciences, Ellensburg, Washington, U.S.A. Meghan Miller, Central Washington University, Department of Geological Sciences, Ellensburg, Washington, U.S.A. Peter Molnar<sup>2</sup>, Massachusetts Institute of Technology, Department of Earth, Atmosphere, and Planetary Sciences, Cambridge, Massachusetts, U.S.A.

 <sup>1</sup> Now at: University of California Santa Barbara Institute for Crustal Studies and Department of Geological Sciences Santa Barbara, California, U.S.A.
<sup>2</sup> Now at: University of Colorado at Boulder Department of Geological Sciences and CIRES Boulder, Colorado, U.S.A.

#### ABSTRACT

We have determined slip rates on the most active reverse faults, reconstructed an extensive preorogenic erosion surface, constructed local and regional cross sections, and dated syntectonic Tertiary sedimentary rock with magnetostratigraphy along an ~northsouth transect that spans the Kyrgyz portion of the west-central Tien Shan. The cumulative late Quaternary shortening rate along this transect is ~12 mm/yr. The shortening is divided across 5 major fault zones, and the most active faults lie in the interior of the range. Using geometric models developed in other regions of basementinvolved deformation, we estimate 35-80 km of shortening during the late Cenozoic. Apparent simultaneous onset of significant basin formation in at least 3 major basins around 12 Ma is interpreted to mark the onset of the current orogenesis. Given the current shortening rate of about 12 mm/yr, measured across active faults and by GPS, we infer that the rate has increased through time. While we propose an acceleration of shortening, our studies indicate that it has always been of similar style, dominated by north-south shortening across ~east-west trending basement-involved reverse faults. Deformation has been localized to  $\sim 5$  zones that border the largest and deepest Tertiary basins, show the greatest structural relief, and contain the currently most active faults.

## INTRODUCTION

We present a progress report of ongoing investigations of the late Cenozoic tectonic history of the central Tien Shan, the most active example of contractional mountain building in a continental interior. While we perceive our work as an extension of the large body of work of many researchers, we make little attempt to synthesize previous studies in this region, but rather present our preliminary results. The following questions about the geometry and timing of deformation motivate our research: How does the current rate and distribution of shortening, measured by GPS, compare to the rates and distribution of slip over longer geological time scales? When did Cenozoic deformation of the central Tien Shan commence? How was strain distributed during early stages of Cenozoic mountain building? What is the present structural style of deformation, and how has strain distribution evolved over the last several million years? We present techniques and observations that provide a kinematic description of deformation in space and time that will better constrain dynamic interpretations of the initial conditions that localized deformation here, and forces that have resulted in shortening and crustal thickening.

The ongoing collision of India and Eurasia drives deformation in the distant Tien Shan of central Asia. The elongate central and eastern Tien Shan represents a deforming region between two generally stable crustal elements, the Kazakh Platform to the north and the Tarim basin to the south (Figure 1). Currently located ~500-1500 km north of the Indo-Eurasian plate boundary, GPS measurements suggest current shortening across the central Tien Shan absorbs nearly one-half of the total relative plate convergence (DeMets et al., 1994; Abdrakhmatov et al., 1996; Holt et al., 2000).

Geodesy and historical seismicity provide clues about the present-day deformation field across the central Tien Shan. The ~12 mm/yr GPS-measured north-south shortening rate across the northern 2/3 of the central Tien Shan is evenly distributed (Abdrakhmatov et al., 1996). Focal mechanisms from moderate and large earthquakes show primarily thrust and reverse faulting with P-axes oriented ~north-south, consistent with the geodetically-measured maximum shortening direction and the overall direction of Indo-Eurasian plate convergence (Shirokova, 1974; Ni, 1978; Tapponnier and Molnar, 1979; Nelson et al., 1987; Ghose et al., 1998). Molnar and Ghose (2000) find a rate of seismic moment release across the central Tien Shan consistent with the rate of shortening measured by GPS; although uncertainties are large, the seismicity that they used is more focused near the edges of the belt.

The central Tien Shan display a basin-and-range topography that reflects late Cenozoic relief caused by distributed reverse faulting and folding (Schulz, 1948; Sadybakasov, 1972; Makarov, 1977; Tapponnier and Molnar, 1979; Chediya, 1986; Sadybakasov, 1990; Trifonov et al., 1997) (Figures 1 and 2). The generally east-west trending ranges of the central Tien Shan define folded and fault-bounded blocks that comprise previously deformed Paleozoic and Mesozoic rocks (e.g., Burtman, 1975; Knauf, 1976) separated by basins of syntectonic Cenozoic sediment. The unconformity that separates bedrock from Cenozoic strata has been preserved or exhumed throughout much of the Tien Shan, providing critical constraints on young deformation within the ranges.

Stratigraphic studies and cooling ages have been interpreted to suggest a wide variety of ages for the initiation of late Cenozoic mountain building in the Tien Shan. Terrestrial sedimentation near the base of the Tertiary section has been interpreted as Oligocene (e.g., Makarov, 1977) to middle Miocene (Bullen et al., in press; this study) in age. Fission tracks in detrital apatite grains indicate cooling of source rock ~14 Ma near the southern margin of the central Tien Shan (Sobel and Dumitru, 1997), and combined fission track and U-Th-He analyses from the northern margin are consistent with rapid exhumation beginning ~11 Ma (Bullen et al., in press). In the eastern Tien Shan, exhumation and likely associated relief production occurred in early Miocene time (~20-24 Ma), as indicated by apatite and Ar/Ar cooling ages (Hendrix et al., 1994; Sobel and Dumitru, 1997; Yin et al., 1998) and stratigraphic and structural studies (Yin et al., 1998; Burchfiel et al., 1999). Burchfiel et al. (1999) argued, however, that most of the deformation where they worked occurred in late Pliocene or Quaternary time.

The preliminary results presented in this paper constitute a geological summary of an extensive geological and geophysical investigation. This investigation, funded in part by the Continental Dynamics program of the U.S. National Science Foundation, aims to understand deformation in continental interior settings. We do not review data from the geology group that have been published (Burbank et al., 1999; Bullen et al., in press), are too preliminary (e.g., our work on the southern margin in China (Rubin et al., 2000; Scharer et al., 2000)), or that do not pertain to the questions that motivate this paper.

The results are summarized in three sections; each section presents only a few examples that convey our techniques, our assumptions, and general interpretations. Most study locations occur along an approximately north-south transect between the Chu basin east of Bishkek, and the Aksay basin (Figure 2). Late Quaternary slip rates on active faults demonstrate that deformation is concentrated on a few major faults that are distributed across the belt. Geologic cross-sections illustrate geometric and kinematic examples of fault-related folding. These cross-sections, which rely on a reconstruction of the erosion surface contoured from a digital elevation model, constrain total crustal shortening. The final section presents the paleomagnetic stratigraphy of syntectonic sediment in the Chu, Kochkor, Naryn, and At-Bashi basins, which together with facies information constrain the timing and distribution of basin formation. The discussion presents a simplified hypothesis for the geologic evolution of the central Tien Shan that, although clearly inadequate in detail, we hope is revealing as a general framework.

# RESULTS

### Late Quaternary fault slip rates

A systematic study of late Quaternary slip rates on faults provides a kinematic description of deformation over a relatively short time span, thus providing a 'snapshot' in geologic time for how crustal deformation is partitioned across a region. Quantitative determination of slip rate across several faults allows an analysis of the partitioning and total rate of deformation. Comparison of rates determined by GPS (Abdrakhmatov et al., 1996) with those determined by basin analysis and balanced cross-sections allow us to address questions such as whether rates have changed over time, and to what degree deformation is distributed.

We have evaluated late Quaternary slip rates of eight faults along a ~north-south transect through the Kyrgyz central Tien Shan. The faults occupy four different intermontane basins and the northern foreland: Chu, Kochkor, Naryn, At-Bashi, and Aksay (Figure 2). All of these faults show primary evidence for Holocene or late Quaternary slip, indicating a relatively high slip rate. We use 'late Quaternary' to mean the most recent ~140,000 years, or the time since the later stages of global marine oxygen isotope stage 6 (Imbrie et al., 1984). Thermoluminescence and optically-stimulated luminescence dating shows that the QII(2) river terrace within the Tien Shan formed at this time. We assign a late Quaternary age to fault activity associated with surface deformation of a river terrace, alluvial fan, or other landform of late Pleistocene or Holocene age. These criteria may be biased to faults that lie in or at the margins of intermontane basins, where late Quaternary landforms are more apparent. Evidence for recent fault activity in mountainous topography is difficult to detect, although our reconnaissance efforts and final result indicate that faults in such areas do not contribute much to the total amount of shortening over the late Quaternary.

Calculations of slip rate on three faults that lie within the interior of the central Tien Shan demonstrate the variety of techniques we use. For the Akchop Hills fault in Kochkor basin, we calculate slip rates from the differences in height of river terraces across bends in the fault that are mapped and quantified by dips of Tertiary strata. The measured vertical offset of a radiocarbon-dated QIII(2) terrace and direct observation of an exposed fault constrain the slip rate of the Central Naryn fault at the Kadjerty River. A trench excavated across the Oinak-Djar fault ("Quaternary fault" of Makarov (1977)) in the western At-Bashi basin reveals evidence for multiple earthquake ruptures during the past ~10,000 years, and a radiocarbon-dated layer traced across the fault zone allows estimation of the slip rate. Uncertainties in the slip-rate calculations are determined by a Monte Carlo method. Each variable in the calculation is expressed with its own probability distribution, based on field measurements or laboratory uncertainties. The results of the Monte Carlo simulation are expressed as a histogram that yields a most probable rate and 95% confidence limits that are not necessarily symmetric about the most probable value.

We surveyed river terrace surfaces and fault scarps with a laser-distance theodolite (total station) and with pairs of Global Positioning System receivers set up to allow differences in position to be measured. For river terraces covered with thick or variable deposits of loess or colluvium, we surveyed the contact between fluvial gravel and overlying fine-grained deposits. We explicitly assume that, when rivers incised their floodplains to form terraces, incision occurred quickly, and that the terraces originally formed parts of planar surfaces over distances of several hundred meters to a few kilometers.

### Akchop Hills fault, Kochkor Basin

The South Kochkor fault separates the Terskey Ala-Too range to the south from the Kochkor basin (Fedorovich, 1935; Schulz, 1948; Tarasov, 1970; Sadybakasov, 1972)

(Figures 2 and 3). Although this fault has displaced late Quaternary river terraces along the Djuanarik River, a greater amount of late Quaternary fault displacement has occurred within the basin several kilometers north, producing the Akchop Hills. Tertiary strata exposed along the Djuanarik River in the hangingwall of the Akchop Hills fault are folded into sections of relatively constant dip, separated by narrow axial surfaces (Figure 3). These hangingwall folds result from slip over a footwall with sharp bends in the fault. An extensive literature provides a framework for quantitative evaluation of fault-related folding of layered strata (e.g., Suppe, 1983; Mount et al., 1990; Suppe and Medwedeff, 1990; Suppe et al., 1992; Shaw et al., 1994; Medwedeff and Suppe, 1997), which is applicable to deformation in the southern Kochkor basin and in other Tertiary basins of the central Tien Shan.

Profiles of the preserved river terraces along the west bank of the Djuanarik River and a geologic cross-section provide evidence that the decreasing bedding dips at the southern end of the Akchop Hills coincides with a decrease in dip of the underlying Akchop Hills fault (Figure 4). The terrace profiles across the south end of the Akchop Hills reveal abrupt kinks that separate approximately straight sections (Figure 4A). Axial surfaces in the underlying Neogene strata that coincide with kinks in the terraces project to bends in the fault below (Figure 4B). The amount of displacement on the fault since the terrace formed and the fault dip determine the height of the terrace above the river (Figure 5). Flat sections of the river terraces are progressively higher above the Djuanarik River to the north, coinciding with a greater fault dip that is indicated by steeply dipping Neogene beds.

The southern synclinal bend (labeled (i), Figure 4) reflects fault-bend folding above a thrust fault that steepens from  $9\pm2^{\circ}$  to  $18\pm2^{\circ}$  (Figures 4 and 5). Displacement on the underlying fault is constrained by the ~64 m difference in height of the QII(2) terrace across axial surface (i). Figure 5C shows that the 9° dip of the QII(2) terrace backlimb measured in the field is also the most probable angle  $\psi$  predicted by the equations that are derived from this thrust fault geometry. The backlimb in terrace QIII(1) is longer and less angular than predicted by strict fault-bend folding, and may be due to a finite width of the axial surface. We did not make enough measurements of the QIII(2) terrace to evaluate the width or angularity of the backlimb.

Thermoluminescence and infrared-stimulated luminescence analyses of massive silt from the base of an ~4 m thick deposit overlying the fluvial gravel of the QII(2) terrace yield a combined age of 128±26 ky B.P. (thousand years before present) since last exposure to light and/or heat. Calibrated radiocarbon ages of charcoal collected from the QIII(2) terrace in several intermontane basins suggest that the terraces formed between 13.5-15.5 cal ky B.P., during the transition between global glacial and interglacial conditions (Imbrie et al., 1984; Fairbanks, 1989). The age of the QII(2) terrace is consistent with formation between 128-140 ka, which coincides with the penultimate transition between global glacial and interglacial conditions (Imbrie et al., 1984; Henderson and Slowey, 2000; Shackleton, 2000). The 64±5 m change in height of the QII(2) terrace across axial surface (i), the 9±2° and 18±2° dips of the underlying fault, and the inferred 128-140 ka age of the terrace yield a slip rate of 3.1 + 1.6/-0.7 mm/yr (Figure 5). The 95% confidence limits represent the propagation of uncertainties from each measurement, and do not reflect uncertainties in the proposed structural geometry. The agreement between the observed and predicted dip of the backlimb of the QII(2) terrace, however, supports our proposed structural geometry.

#### Central Naryn fault, Naryn basin

The Naryn basin contains abundant evidence for Quaternary shortening (Schulz, 1948; Makarov, 1977; Sadybakasov, 1990; Burbank et al., 1999). Much late Quaternary deformation is associated with two south-vergent thrust faults that reach the surface in the northeastern and north-central portions of the basin (Figure 2) and cut the QIII(2) and other terraces of the Kadjerty River. The southern fault is the Central Naryn fault of Schultz (1948), and we call the northern fault the Kadjerty fault.

The Central Naryn fault is visible in outcrop along the west bank of the Kadjerty River, where it cuts a ~20 m-high scarp across the QIII(2) terrace (Figure 6). A north-south terrace profile in the hangingwall of the fault shows an abrupt bend ~750 m north of the scarp (labeled "hinge" in Figure 6B) and an anticline near the fault scarp. North of the bend, the terrace surface dips  $0.9^{\circ}$  S, similar to the  $0.9^{\circ}$  S gradient of the modern river (Figure 6B). South of the bend the terrace surface is sub-horizontal, implying a backrotation that we attribute to changes in underlying fault dip. In outcrop, the Central Naryn fault dips  $37\pm2^{\circ}$  N, whereas Neogene strata exposed in the hangingwall dip ~10° N (Figure 6A). If the tilting in the hangingwall is due to changes in the dip of the Central Naryn fault, a 37° N dip at the surface flattens to dip ~10° N at the kink, approximately parallel to the dip of Neogene strata.

Surveyed points close to the scarp define the terrace surface in the hangingwall and the footwall (Figure 6C). Because the fault location and dip are known, we project the hangingwall and footwall lines onto the fault to determine the amount of fault slip. Calibrated radiocarbon ages of three charcoal samples collected from silt and sand overlying river gravel in the terrace show that the QIII(2) terrace formed 13.5-15.5 cal ky B.P. These parameters yield a late Quaternary slip rate for the Central Naryn fault of 2.2  $\pm 0.4/-0.3$  mm/yr (Figure 6C).

### Oinak-Djar fault, At-Bashi basin

The south-vergent Oinak-Djar fault ("Quaternary fault" of Makarov (1977)) accommodates most of the late Quaternary shortening across the western and central At-Bashi basin and dips under the mountains that separate the Naryn and At-Bashi basins (Figure 2). For most of its 80 km strike length the fault is manifested at the surface by an abrupt scarp that separates hills composed of steeply dipping Neogene sedimentary rock and large, late Quaternary fans (Figure 7). The thrust fault is "blind" to the east, however, where slip is transferred into growth of a fault-propagation fold. Right-stepping en-echelon fold axes in the hangingwall of the fault and deflected drainages across the fault trace indicate a component of left-lateral shear (Makarov, 1977).

A 20 m-long trench across the Oinak-Djar fault, where the fault ruptured the head of an alluvial fan repeatedly during the late Quaternary Period (Figure 7B), revealed four thrust faults that displace a series of colluvial and alluvial layers (Figure 7C). We correlate two deposits across the fault zone that contain radiocarbon-dated charcoal. The older deposit (layer C) is the lower of two layers of silty sand interstratified with alluvial sand and fine gravel. The younger deposit (layer A) is a layer of sand and fine gravel alluvium that has a laminar silt layer at its base and is overlain by silty sand colluvium.

The cumulative amount of offset of the base of layer A (i.e., the laminar silt layer) across each fault strand constrains displacement across the Oinak-Djar fault. In order to extrapolate through folding and rotation of the beds close to the faults, we assume that the original slopes of the deposits were similar to the slopes away from the fault. The dip of the head of the alluvial fan provides the original hangingwall surface ( $\sim 8^\circ$ ) and the dip of continuous units and the fan surface represents the footwall ( $\sim 5^\circ$ ). The minimum and maximum amounts of slip across the fault zone are determined by assuming that in between faults, each section of the laminar silt layer dip between 8° and 5°, respectively.

Layers A and C have similar vertical displacement, indicating that they have been deformed during the same ground-rupturing earthquakes. Radiocarbon ages of charcoal samples indicate that these units were deposited between ~10.5 and ~9.0 cal ky B.P. (Figure 7C). In the slip-rate calculations, this range represents the duration over which slip occurred. We add additional uncertainty to the slip rate calculations due to the earthquake cycle: the magnitude of stored elastic strain at the time these units were laid down and at the present are not known. The calculations yield a Holocene slip rate of the Oinak-Djar fault of  $0.9\pm0.3$  mm/yr.

### Summary of Late Quaternary slip rates

We have calculated slip rates of eight faults using methods similar to the ones described above, and in some cases we have measured slip in several places along the length of a single fault to quantify the variability of slip rate (Table 1). On the basis of our geologic mapping, these eight faults have the highest late Quaternary slip rates along a ~north-south transect that crosses the portion of the central Tien Shan in Kyrgyzstan (Figure 2). Rates of dip-slip faulting on individual faults range from ~0.2 mm/yr to ~3.1 mm/yr, and accommodate large components of north-south shortening, although our methods measure dip-slip components of faulting only. The highest rates of shortening in the late Quaternary occur across the Kochkor and Naryn basin margins, which lie in the middle of the Tien Shan. The lowest rate of shortening occurs across the southernmost intermontane basin, the Aksay basin, which also lies at the highest elevation.

# Geologic cross-sections

We have constructed geologic cross-sections across several range-basin margins in order to determine the geometry of deformation that has produced the large faulted and folded blocks across the mountain belt, and to constrain the total amount and partitioning of crustal shortening since the onset of Cenozoic deformation. We assume that slip on faults accommodates virtually all of the shortening, and produces all of the folding, in the upper crust.

Different theories have been advanced that relate folds in crystalline rock to underlying fault slip, but all generally rely on the fundamental relationship between relative vertical movement, *u*, dip of an underlying fault,  $\delta$ , and amount of fault slip, *s*, such that  $u = s \sin \delta$  (Figure 8). Broad, gently dipping blocks of crystalline basement in the Eastern Rocky Mountains of the western U.S.A. and the Sierras Pampeanas of western Argentina are inferred to result from slip on curved thrust faults that penetrate much of the crust (Erslev, 1986; Jordan and Allmendinger, 1986; Erslev and Rogers, 1993) (Figure 8A). Several curved thrust faults in basement rock have been imaged with reflection seismology and penetrated by wells (Gries, 1983). Other researchers have interpreted abrupt folds in basement rock in the Eastern Rocky Mountains to be related to sharp bends between planar fault sections (Narr and Suppe, 1994) (Figure 8B). Basement blocks that are bounded on both sides by faults or sharp folds are interpreted to have formed by back-thrust splays off major crustal faults (e.g., the "structural wedge" in Figure 8B).

An extensive, low-relief surface developed on highly deformed bedrock allows us to test and apply these models in the Tien Shan. Often called the "preorogenic surface" or "erosion surface" by Russian and Kyrgyz geologists, it is widespread within the high bedrock ranges and ubiquitous wherever the base of the Tertiary stratigraphy is exposed. This has led to consensus that there was a time prior to the uplift of the current Tien Shan when the entire region was occupied by a gentle, low relief and implicitly low elevation bedrock surface that has now been uplifted to elevations greater than 5 kilometers and depressed under some basins to kilometers below sea level. Thus, many geologists (e.g., Sadybakasov, 1972; Makarov, 1977; Chediya, 1986; Sadybakasov, 1990) have attempted to use this surface as a marker for reconstructing the deformation in the Tien Shan, and we have followed their lead.

We have constructed a map of the current elevation of the erosion surface with the aid of a digital topographic model (DEM) sampled at a spacing of 90 m that allows a relatively objective rendering of the surface (Figure 9). To make this map we digitized the outcrop extent of remnants of the erosion surface (Figure 9A). To this we added the outcrop extent of the large Tertiary basins and depth contours of the thickness of sediment within the basins that we digitized from the published atlas of Kyrgyzstan and other published maps. By registering these data with the digitized topography in a Geographical Information System (GIS) database, we extracted the elevations of all points within outcrop remnants of the erosion surface, and determined the elevation of the surface wherever there was a thickness contour in the basins. To extend the surface over areas of high, rugged topography where there are no preserved remnants of the erosion surface, we added the elevations of high, smooth ridge crests that lie at least 5 km from the local drainage network. This criterion eliminates many smooth ridge crests that are inset in major valleys. Inclusion of ridge crests is motivated by the observation that in many places the highest elevations are either planar remnants of the erosion surface or unusually smooth ridge crests of similar elevation.

Extrapolation between outcrops of the erosion surface, all major ridge crests, and Tertiary basin margins results in a final, smoothed reconstruction of the erosion surface (Figure 9B). We chose not to smooth the 200 m contours within the basins so that we could clearly see where the surface was buried by significant thickness of sediment. While it is clear that this surface is an approximation, we believe that in many areas it at least crudely mimics the shape of the actual surface. In some areas this approach fails, such as the north slope of the Kyrgyz Range where it does not "smooth out" the major north-south drainages into the range that are young (Figure 9B). Elsewhere it produces reasonable patterns, such as the east-west faults and folds south of Issyk-Kul and the high, gently dipping plateau on the Kyrgyz/China border. Because we had tentative depth contours for the Tarim basin in China, the deep hole shown on the southeast margin of the map is only a schematic representation.

The evidence for fault-related folding in crystalline basement from the Eastern Rocky Mountains has influenced our interpretation of the deformed erosion surface. As an example, Figure 10 shows a small portion of the erosion surface map on the southeast side of the Kyrgyz Range (Figure 2). Here, the erosion surface ranges in elevation from 2 to more than 4 km. The mapped Shamsi-Tunduk thrust fault separates the relatively low relief, high elevation regions in the central and southwest part of the map from the low elevation region in the northeast. A tilted part of the erosion surface lies across the central part of the map: it strikes ~east-west and dips south, and appears to extend beneath the adjacent Kochkor basin. The tilted section can be traced as far west as Suusamyr basin, where it is broken by a south-vergent fault.

A north-south profile across the erosion surface in Figure 10 shows how the backlimb of this fold consists of two relatively planar sections, one dipping ~8-9° S and the other nearly horizontal across the crest of the range, separated by an abrupt bend (Figure 11). The preservation of original remnants of the erosion surface across the abrupt bend rules out surface faulting as the cause of this deformation.

We interpret the tilted section in the erosion surface as the backlimb of a fold that forms the southern limit of the Kyrgyz Range and is produced by displacement along the underlying Shamsi-Tunduk thrust fault (Figures 10 and 11). Two end-member cases of fault-related folding may explain the development of the backlimb. A curved fault would produce a backlimb with a dip that is related to the amount of displacement and the radius of fault curvature (Figure 8A). A sharp, upward bend in a fault produces a backlimb with a dip that approximately equals the difference in the dips of the fault beneath the backlimb and farther south ("Fault-bend Syncline" in Figure 8B). In this case, the width of the backlimb approximately equals the amount of crustal shortening produced by fault displacement. Thus detailed analysis of folds in the erosion surface may indicate whether underlying faults are planar or curved, and this information can in turn be used to quantify crustal shortening.

A preliminary cross-section between Son-Kul and the Chu basin illustrates the use of the erosion surface to interpret the deep geometry of faults, and to estimate total crustal

shortening (Figures 12 and 13). We emphasize that the cross-section includes simplifications, but that the general relationships between fault geometry and fold geometry are useful for understanding the structural style and interpreting the amount of shortening across parts of the range. Below  $\sim 10$  km, two south-dipping thrust faults with large displacement occupy the cross-section in Figure 13: the South Kochkor fault and the Shamsi-Tunduk fault. The broad Son-Kul region has been elevated by displacement on the South Kochkor fault (marked "1" on Figure 13). A bend in the erosion surface at the northern margin of Son-Kul marks a bend in the South Kochkor fault. The northdipping thrust fault that marks the northern boundary of Tulek basin is a thrust splay of the South Kochkor fault ("2" on Figure 13) that accommodates about 2.5 km of shortening. Along this cross-section, the South Kochkor fault does not intersect the surface along the margin of Kochkor basin (Figure 12). A ~3.5 km-thick section of Tertiary strata has a vertical dip north of a gentle syncline at the basin margin; the section of vertical dips implies that the erosion surface was folded above the South Kochkor fault ("3" on Figure 13) as the fault propagated northward. Eventually the fault broke through the fold and propagated to the surface. Displacement has subsequently occurred primarily on the gently dipping fault beneath the Akchop Hills ("4"). Fault-propagation folding accommodated about ~3.5 km of shortening, and displacement on the low angle Akchop Hills fault in the basin has accommodated an additional ~3 km. We draw attention to the inferred relationship between the Akchop Hills fault at the surface and the more steeply dipping South Kochkor fault at depth: they are essentially the same fault. Farther east, where the South Kochkor fault crops out at the surface (Figures 3 and 12), the Akchop Hills fault is a gently dipping splay that connects with the South Kochkor fault south of the range front. Thus the high slip rate on the Akchop Hills fault implies a similar high rate of slip on the South Kochkor fault at depth.

The  $\sim 9^{\circ}$  south dip of the erosion surface on the south side of the Kyrgyz Range (Figure 11) projected to the south side of the Kochkor basin yields a depth consistent with the measured thickness of Tertiary sediment. This geometry suggests that the erosion surface has been titled uniformly for ~35 km ("5" on Figure 13). This dipping surface is underlain by the Shamsi-Tunduk fault, which we infer to be curved as in the example in Figure 8A. The axial surface separating the top of the long, tilted section and the flat top of the Kyrgyz Range indicates a transition from a curved fault to a planar fault ("6"). Slip on the Shamsi-Tunduk fault, where it is exposed at the surface, accommodates ~4 km of shortening. Additional displacement across the Shamsi-Tunduk fault at depth occurs on a gently dipping splay that extends northward to the margin of the Chu basin. Two structural wedges (Figure 8B) with thrust faults that dip north accommodate an additional ~3 km of shortening ("7"). This low-angle splay of the Shamsi-Tunduk fault is the equivalent to the Issyk-Ata fault to the west, where it dips south at the surface and cuts Tertiary and Quaternary strata (Figures 2 and 12). The eastward transition to a region with north-dipping thrust fault splays and the involvement of pre-Cenozoic rocks in the hangingwall occurs where the Chu basin becomes narrower and shallower (Figure 12). The total ~7 km of shortening across the Kyrgyz Range in this cross-section is about <sup>1</sup>/<sub>2</sub> of the shortening estimated from cross-sections near the longitude of Bishkek (Bullen et al., in press). The difference between the amount of total shortening across the Kyrgyz Range in Figure 13 and that estimated farther west may be accommodated by shortening

across the range to the north that separates the Chu and Ili basins ("8" in Figure 13, and Figure 2).

The methods for determining  $\sim 16$  km of total north-south shortening along the crosssection in Figure 13 have been applied across other ranges to estimate total Cenozoic shortening between the Chu and Aksay basins. Near the longitude of Bishkek, shortening across the Kyrgyz Range and Issyk-Ata fault appear to be ~10-25 km (Bullen et al., in press). We use this range in our total because at that longitude the Kyrgyz Range is the northern margin of the Tien Shan, while along the cross section in Figure 13 there is an unknown additional amount of shortening in the range to the north. Crosssections between Kochkor and Son-Kul basins indicate ~7-15 km of shortening. depending on the dips inferred for the South Kochkor, Tulek, and other faults. We infer about 5-10 km of shortening across the Moldo-Too and northern Naryn basin. Crosssections between the At-Bashi and Naryn basins across the Kara-Too indicate ~6-15 km of shortening (Burbank et al., 1999; McLean, 1999). We infer ~7-15 km of shortening between the At-Bashi and Aksay basins, with most of the total shortening at the surface taking place across the southern margin of the At-Bashi basin (although the slip rate data suggest that currently the northern margin may be more active). The sum is ~35-80 km. The largest contributors to the uncertainties are the dip of the faults at the basin margins, and the relative amounts of folding and underthrusting of the erosion surface at the basin margins. Much of the shortening we see at the surface involves coherently folded Tertiary strata, indicating that a large fraction of the total shortening occurred after deposition of these units.

### Cenozoic stratigraphy of the intermontane basins

The thick sequences of Cenozoic deposits overlying the erosion surface record the onset and history of deformation across the Tien Shan. All of the units we have mapped and sampled for paleomagnetic analysis have been described in detail by earlier authors (see summaries by Makarov (1977) and Sadybakasov (1990)) and have been given separate formation names in the various basins. Commensurate with the broad scope of this summary progress report, we group formations of rock with similar lithology but from separate basins into four units. We call these units "groups" to distinguish them from their formal formations.

The lowest unit, typically tens of meters thick, but in places up to about 100 m thick, is the Kokturpak group. Where present, it nonconformably overlies poly-phase deformed crystalline rock. Scattered fossils and a few K/Ar dates in basalt (Sobel and Arnaud, 2000) indicate that the Kokturpak formed over a very long period of time, and likely at different times in different places, from late Cretaceous to Miocene time. Most workers view this unit as having formed by deep weathering and accumulation of sediment in very shallow, long wavelength basins on a broad bedrock surface not unlike the Kazakh Platform to the north of the Tien Shan. We view this unit as preorogenic, and because of its limited thickness and apparently episodic and spatially variable deposition, we did not sample it for paleomagnetic analysis. Conformably overlying the Kokturpak group are formations that we generalize as the Shamsi group (after the prominent and representative Shamsi Formation). Where the Kokturpak group is not preserved, the Shamsi group is deposited on the erosion surface. The Shamsi group varies in thickness from several hundreds of meters to about a kilometer thick where we have mapped and sampled it. While opinions vary on how the many Shamsi-like formations correlate, we use the term to apply to about the lower quarter to third of the Tertiary section that appears reddish, is easily eroded (and thus outcrops poorly), and is less mature (well sorted and mineralogically high in quartz and feldspar) than the main basin-filling Chu group. In some places Shamsi units are not red, like the thick green siltstone and mudstone in the Kochkor basin, or are quite coarse and well indurated, like some of the red sandstone and conglomerate exposed along the At-Bashi River valley.

We refer to the main Tertiary basin-filling formations that gradationally overlie the Shamsi group as the Chu group. Compared to the Shamsi group it is more mature, generally lighter and more variable in color, with many green, white and tan beds, and is often better indurated. The Chu group is frequently well bedded and contains large tabular to trough-shaped sandstone and conglomerate lenses that represent river channels of moderate size. The Chu group also includes lacustrine sediment and evaporites in the intermontane basins—these generally occur higher in the section. The maturity and variability of the Chu group suggest deposition by large, integrated drainage systems.

Overlying the main Tertiary basins are coarse conglomerate deposits, often called the Sharpyldak Formation, and Pleistocene fluvial gravel and glacial deposits. The Sharpyldak group is usually tens to hundreds of meters thick, but occasionally may be on the order of a kilometer thick. In many places it gradationally overlies the uppermost Chu group, often laterally interfingering with and vertically succeeding lacustrine deposits, suggesting a sudden progradation of coarse material well out into the large intermontane basins. Elsewhere the Sharpyldak group unconformably overlies Tertiary deposits or even the erosion surface. Because of the coarseness of the material we did not sample the Sharpyldak group for paleomagnetic analysis, although the magnetostratigraphy of the underlying Chu group locally constrains its age.

### Magnetostratigraphy of Tertiary deposits

The Tertiary terrestrial section contains only sparse fossils, but ages are essential for determining rates of Cenozoic deformation. Thus, we carried out magnetostratigraphic analyses to date them. Currently, we have only analyzed a small fraction of our samples, so our results are preliminary. We have analyzed samples for the Chu group in three basins and the Shamsi group in one (Figure 2). We have sampled but not yet analyzed the Shamsi group in the At-Bashi basin and most of the upper Shamsi and lower Chu groups in the Kochkor basin. Work is also in progress to date sections on the southern margin of the range in China.

Figure 14 shows an example of a paleomagnetic analysis from a section of Chu group collected along the At-Bashi River in the At-Bashi basin (see Figure 2 for location). Each point indicates a site where at least 3 samples were collected. All samples were

stepwise demagnetized by a combination of alternating field and thermal demagnetization until completely demagnetized and then the demagnetization results were subjected to principle component analysis to identify primary components (Kirschvink, 1980). All samples contained strong overprints that pointed in the direction of the present field. These overprints generally required thermal demagnetization to be removed, and thus are inferred to be largely due to weathering. Most samples also contained a stable primary direction oriented either north and down (normal polarity) or south and up (reversed polarity) when structurally corrected. The primary directions were combined into site means, and are plotted on Figure 14. The sites were then grouped into magnetozones that include consecutive consistent sites. The interpreted magnetic stratigraphy is presented to the right of the data (Figure 14).

The next step is to match the record of magnetic stratigraphy to a reference magnetostratigraphy that is tied to a time scale calibrated with radiometric dates or other methods (Cande and Kent, 1995) (Figure 15). The presented match is based on fossils suggesting a late Miocene to Pliocene age for the Chu group. Alternative matches may be obtained by sliding the section up or down, or by assuming different rates of sedimentation through the section. Because of simplicity, matches that are produced with constant (or slightly variable) rates of sedimentation are preferable to those that require large changes in sedimentation rate, particularly within units of homogeneous facies. Once all the sections are completely analyzed, the longer magnetic stratigraphy should help substantiate the match or confirm alternatives. Additional fossils or other independent age estimates could also help find the appropriate match.

To compare our incomplete sections, we have estimated durations of deposition for each of the four basins by extrapolating the available age control to the top or bottom of the incomplete sections (Figure 15). We extrapolated the ages up or down using the thickness of the unanalyzed portions of the units and two end member sedimentation rates. The two rates we use to bracket the age are the highest rate of sedimentation (3 kilometers in 4 million years, measured in the Chu group at the At-Bashi section), and the lowest rate (1 km in 4 million years, measured in the Shamsi group at the southern Naryn Valley section). The ranges of ages that these extrapolations produce are shown with black arrows on the sections in Figure 15. By assuming characteristic sedimentation rates for Chu group (3 km in 4 Ma) and Shamsi group deposits (1 km in 4 Ma), we determine the most probable age (indicated by a thick bar).

A simple interpretation of this magnetostratigraphy is that the Chu group was deposited from about 8 to 4 Ma, and that the Shamsi group may span a period from about 13 to 8 Ma. The broadly similar time of the Shamsi-Chu transition, and the similar ages of the tops of the sections (coinciding with the Chu-Sharpyldak transition) suggest that processes affecting these changes were broadly synchronous, differing by only a few million years, which is a small fraction of the elapsed time since deposition began. The extrapolated ages of the base of the section in various basins suggests a similar time (within a few million years) for the onset of basin formation in this part of the Tien Shan.

### DISCUSSION

Although preliminary, the various data collected along the transect motivate a simple history of Cenozoic shortening of the Tien Shan that addresses the questions stated in the Introduction. The apparently synchronous onset of accumulation of significant thickness of sediment in the northern foreland (Chu basin) and at least 2 other widely separated basins in the intermontane portion of the Tien Shan (Kochkor and Naryn/At-Bashi) suggests that deformation began along the Kyrgyz portion of the transect at essentially the same time, currently estimated to be around 12-13 Ma. Furthermore, because the Tertiary sections appear parallel to the underlying erosion surface, we infer that it was essentially planar over extensive regions and thus can be reconstructed to determine the total displacement since ~12 Ma.

Similar to previous workers (e.g., Schulz, 1948; Makarov, 1977), we infer that the large Tertiary basins are synorogenic. The criteria that we consider important to the synorogenic interpretation include 1) the Tertiary sections are so thick, typically ~5 km, 2) their current wedge-shaped cross-sections have the originally thickest portions close to large displacement reverse fault zones, and 3) their map view extent essentially mirrors the large bounding ranges that are uplifted across the large basin-bounding reverse fault zones. The lack of significant structural unconformities within the Tertiary sections and the gradational and evolutionary changes of the sediments through time suggest that entire basins (rather than some fraction) are syntectonic. In addition, several aspects of the basins require additional complexity that may contain important insights into the evolution of the range. These aspects are: 1) the overall maturity of the sediments and especially the lack of coarse marginal facies close to the bounding fault zones except high in the Tertiary sections, 2) the general preservation of the erosion surface in the surrounding ranges, making it difficult to find a local source for all of the sediment, and 3) sedimentary transport indicators, thickness variations, and facies architecture that suggest east to west transport of sediment into the large basins that we have sampled, rather than north or south transport from the local high relief across the major bounding fault zones.

Based on these observations, we suggest three possible interpretations: 1) much of the sediment in the large deep basins comes from the high central Tien Shan, east of the transect, and that little marginally derived facies was deposited in the basins until the surrounding mountains became quite high, relatively late in the Tertiary record, 2) coarse marginal facies did not develop early because the faults began as large folds (e.g., Makarov, 1977), that later broke through to form the reverse faults we see today, or that 3) coarse, locally derived marginal facies have been tectonically buried by large thrust faults on basin margins. It is possible that more than one of the processes occurred at the same time, because they are not mutually exclusive.

The first possibility is supported by west-directed paleocurrent indicators, thickness variations, and the facies architecture of the Naryn, At-Bashi, and Kochkor basins. The second possibility is supported by the presence of large monoclinal folds along many of the basin margins that look like fault propagation folds, suggesting evolution from

folding to faulting along the basin margin. Also, the general evolutionary change in sediment type seen in the Tertiary basins may support this possibility. The similarity of the basal Shamsi group to the Kokturpak group in many places indicates that the basal Shamsi group incorporates material stripped off the erosion surface as it began to be deformed and uplifted. The gradational change from the relatively immature and monotonous Shamsi group to the Chu group suggests that the basins were getting deeper, with sustained closure to produce significant lacustrine and delta deposits, and were fed by drainage networks of sufficient size and development to deliver large volumes of mature sediments.

The third possibility, that we can only see the central and distal basin facies because low angle thrusting along the active margins have hidden or buried the proximal facies under bedrock flaps, is important because it may suggest much greater offset across the faults than we currently infer. Possible support for this includes the dominance of low-angle faulting within the basins and difficulty in reconciling the bedrock deformation with the necessary connections to steeply-dipping faults beneath the bounding ranges.

Using the deformation of the erosion surface and Tertiary strata, cross-sections across the basin margins suggest total shortening by thrust faulting on the order of 35-80 km between the northern front and Aksay basin. The shortening is concentrated at ~5 major zones through the transect: the southern margin of the Chu basin, the southern margin of the Kochkor basin (or the active margins of the Djumgal and Suusamyr basins to the west), the northern margin of the Naryn basin, the ranges that separate the Naryn and At-Bashi basins, and the southern margin of the At-Bashi basin. The two sources of largest uncertainty in amounts of shortening are the dips of the major faults at depth and the extent of underthrusting of the erosion surface and Tertiary strata beneath the margins of the ranges. Intervening regions contribute only minor amounts of shortening locally, although their cumulative effect on the shortening along the entire transect may be significant.

Although the most active faults underlie the basins (Figure 2 and Table 1), these faults appear to be gently dipping splays that connect with more steeply dipping faults that underlie the ranges. This is demonstrated for the Akchop Hills fault at Kochkor basin (Figures 4 and 13), and it explains the relationship between the Issyk-Ata fault and the Shamsi-Tunduk fault in Chu basin (Figures 12 and 13). Deformed terraces imply that the Central Naryn and Kadjerty faults have gentle dips at shallow depths that project under the Moldo-Too; these faults are probably splays of the Akchatash fault that defines much of the northern margin of the Naryn basin. North of the Kadjerty and Kurtka rivers, the erosion surface dips south along the northern margin of the Narvn basin, suggesting that a fault steepens abruptly similar to the fault-bend anticline geometry in Figure 8B. This kinematic relationship appears elsewhere in the Tien Shan (Avouac et al., 1993; Molnar et al., 1994; Brown et al., 1998; Burchfiel et al., 1999) and other convergent orogens (e.g., Gries, 1983; Ikeda, 1983; Yeats and Lillie, 1991; Narr and Suppe, 1994; Benedetti et al., 2000). These observations imply that the faults with the fastest slip rates are splays that connect at upper crustal depths with the same 5 major fault zones that have accommodated most of the crustal shortening since the onset of deformation.

In order to quantify the amount and distribution of shortening rate accommodated by the most active faults, we have converted the fault slip rates to estimates of late Quaternary geologic shortening rates (Figure 16). The conversion from slip rate on a dipping fault to horizontal north-south shortening rate requires corrections for uncertain fault geometry at depth, and uncertain northward components of slip on faults that do not strike east-west. The estimate of shortening rate is influenced by our interpretation that the faults have gentle dips in the basins (e.g., Figure 4) and intermediate dips through the upper crust (Figures 8 and 13). This general geometry has been interpreted elsewhere in the Tien Shan (as stated above) and has been shown by deep reflection seismology and bore hole data in other intracontinental mountain belts (Gries, 1983; Stone, 1993; Narr and Suppe, 1994). The cumulative late Quaternary geologic shortening rate is comparable to the modern shortening rate determined by GPS along a similar north-south transect (Figures 2 and 16). Geologic and geodetic measurements of both the total shortening between Aksay basin and Ili basin, and the pattern of shortening, with high strain rates in the middle of the belt and low strain rates in the southern region, are similar (Figure 16). This result suggests that current shortening of the Tien Shan is partitioned across a few, roughly equally spaced fault zones that separate relatively stable blocks of the upper crust. Paleoseismic data from these faults indicates that they have ruptured repeatedly during the last 10,000 years (Figure 7). Large earthquakes such as the 1992 Suusamyr event (Ghose et al., 1997; Mellors et al., 1997) or the 1911 Chon-Kemin event (Bogdanovitch et al., 1914) should be considered as prototypes for earthquakes on any of these faults. Although most of the largest earthquakes in the Tien Shan since the 19<sup>th</sup> century have occurred on or near the edge of the belt, the interior is obviously vulnerable to earthquakes of comparable magnitude.

The total estimate of Cenozoic shortening suggests, but does not require, that the rate of shortening across the Kyrgyz central Tien Shan has changed during the late Cenozoic. The late Quaternary and GPS-measured ~12 mm/yr rate of shortening and ~35-80 km of total shortening imply an onset ~3-7 Ma. Currently, the cross-sections favor the lower end of the range of shortening (e.g., Figure 13), implying a later onset. Thus the apparent initiation of shortening ~12 Ma may have occurred at a lower rate than occurs at present. An inference of slow initial Cenozoic deformation is consistent with other observations from the basin stratigraphy and from cross-sections. The lack of major growth strata or coarse marginal facies in the Tertiary strata and large displacement on the faults that deform the entire Tertiary section indicate that a significant amount of the total Cenozoic shortening across major basin margins has occurred recently.

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Figure 1. Map of the Tien Shan. Shades of gray indicate 1000, 2500, 3500, and 5500 m elevation contours. Inset map shows the location of the Tien Shan in central Asia. The box outlines the location of Figure 2. Our field work has focused on a transect approximately between Bishkek and Kashgar.



Figure 2. Active faults and folds in the central Tien Shan plotted with shaded relief. Major intermontane basins and lakes, and selected ranges, are labeled. Thick lines are faults and folds for which we have calculated late Quaternary slip rates. Stratigraphic sections sampled for paleomagnetic analysis that we discuss in this paper are indicated with triangles. GPS stations, selected from a regional network, that lie close to profile line A-B are shown (see Figure 16). Box outlines the area in Figure 12.



Figure 3. Geologic map of the Djuanarik River area, south Kochkor basin, showing nested terraces along the west bank of the Djuanarik River and strikes and dips of underlying Neogene strata. The South Kochkor fault (SKF) and Akchop Hills fault (AHF) have both been active in the late Quaternary, as indicated by deformation of the river terraces. Terrace profiles and dipping Neogene strata show that the deformation is caused by fault-related folding, and suggest that the Akchop Hills and South Kochkor faults are splays of the same fault that underlies the range south of the basin. We use the extensive QII(2) terrace in our analyses of slip rate for the Akchop Hills fault. A cross-section of line A-A' is shown in Figure 4.



Figure 4. A. Vertically exaggerated river terrace profiles along the Djuanarik River. The profiles are rotated  $\sim$ 1° to remove the gradient of the modern river. The river terraces are folded into an anticline with abrupt bends in the backlimb of the fold, marked (i) and (ii). A thermoluminescence age of silt deposited on the QII(2) terrace is shown, as are calibrated radiocarbon ages from charcoal collected on the QIII(2) and QIV(1) terrace. We infer that the QIII(2) terrace has been reoccupied by the QIV(1) terrace south of the fold. B. Cross-section with the same horizontal scale as the profiles in A, showing dips of strata measured along the Djuanarik River and terrace risers. Dips are given relative to the present-day river and therefore are  $\sim$ 1° greater than those relative to horizontal. The axial surfaces (i) and (ii) mapped in the Neogene strata coincide with the abrupt bends in the terrace profiles. We interpret the folding to be due to bends in the underlying Akchop Hills fault.



Figure 5. Methodology and results of the slip-rate calculation for the Akchop Hills fault. A. Schematic illustration of terrace deformation along the Djuanarik River predicted by fault-bend fold theory (Suppe, 1983). The strath terrace is an unconformity that is displaced along with underlying layered strata that deform by slip on a fault. The thick black arrows are particle paths that show material being deformed across a fault bend. Bedding parallel shear within the Neogene strata accommodates slip across the fault bend. This diagram illustrates the case for strata (dip =  $\alpha$ ) parallel to the underlying fault  $(dip = \delta)$ , an assumption we make for the Akchop Hills fault. The height of the river terrace at its time of formation  $(r(t_0))$  is indicated by the dashed line, here shown to indicate a relative rise in river level since the time of terrace formation. B. Ideal geometry used to calculate fault slip. s = amount of displacement;  $h_1$  and  $h_2 =$  heights of the terrace above the modern river up-dip and down-dip of the fault bend, respectively.  $\gamma =$  orientation of the axial surface, which in the case of layered strata is commonly the angle that bisects the bedding dip angles  $(1/2^*(\alpha_1 + \alpha_2))$ . The backlimb length and angle are  $\lambda$  and  $\psi$ , respectively. The change in height of the river since the formation of the terrace, relative to a local reference frame, is  $h_r$ . Equation (1) shows the calculation of displacement from the change in terrace height across the fault bend and the dips of the underlying strata. Also derived is the relationship between  $\psi$  and  $\alpha_1$  and  $\alpha_2$ . This relationship presents a test we can use to validate our model. C. Slip rate result using deformation of the QII(2) terrace across axial surface (i). The measured 9° backlimb angle  $\psi$  matches the peak of the predicted  $\psi$ , giving us confidence in our analysis of fault slip. TL/IRSL analyses of silt collected from the top of the QII(2) terrace suggest that the terrace formed  $\sim 128$  ka. The histogram on the right shows the probability distribution for the slip rate. Most probable and 95% confidence values are written beneath the histograms and are indicated by black bars.



Figure 6. A. Central Naryn fault exposure along the Kadjerty River. View is to the southwest. The QIII(2) terrace, preserved in the hanging wall and footwall, is offset vertically about 20 m. The exposed fault dips  $37\pm2^{\circ}$  north. Neogene strata in the hanging wall dip ~11-15° north. The Kadjerty River flows south along the base of the exposure. Notice the thicker accumulation of late Quaternary gravel in the footwall of the fault, indicating that the eroded bedrock surface has been offset more than the top of the QIII(2) terrace gravel. B. Profile of the upper gravel contact of the QIII(2) terrace across the Central Naryn fault. The gradient of the QIII(2) terrace and the modern river are similar north of the hinge, but south of the hinge the terrace surface is subhorizontal, indicating that it has been rotated due to slip on a non-planar Central Naryn fault. C. Close-up view of the profile across the fault scarp, showing the surveyed points used to define the hanging wall and footwall and the dips of those surfaces. A single point surveyed on the exposed fault surface and the measured dip are used to constrain the fault location in the slip rate calculation. Uncertainty represented as in Figure 5.



Figure 7. A. View to the north of the Oinak-Djar fault scarp. The fault (between horizontal arrows) separates the hills of steeply dipping Neogene strata in the hanging wall from the aggrading alluvial fans in the footwall. The trench site is situated in the eastern part of a wide wash in which multiple terrace levels are progressively displaced by recent slip on the fault. B. View to the west along the Oinak-Djar fault scarp. The trench and spoil pile are in the foreground. The spoil pile is about 2 m high. C. Map of the west wall of the trench across the Oinak-Djar fault. Ages of charcoal samples are in calibrated radiocarbon years before present, and represent ~95% confidence limits. A laminar silt horizon at the base of layer A is used to measure fault slip across the exposed fault zone by adding slip across individual fault strands (labeled A-D) using simplified traces of the silt contact that remove folding. Histogram shows the probability distribution of slip rate. Layers A (deposited ~9.0-9.3 ka) and C (deposited ~9.6-10.5 ka) are offset similar amounts across the fault zone, indicating that they have been deformed during the same earthquakes. Note that the laminar silt above layer A (~7.8 ka) and the overlying alluvial layers are offset by a lesser amount than layer A across faults B, C, and D, indicating that the younger units were deformation during fewer earthquakes.





Figure 8. Illustrations of folding in crystalline rock due to slip on underlying faults, based on work in the Eastern Rocky Mountains of the western United States. A. Erslev (1986) interprets long, dipping backlimbs and abrupt monoclinal forelimbs to be due to slip on curved faults. His technique for balancing such sections is illustrated in the figure. B. Narr and Suppe (1994) propose that abrupt bends in basement rock are due to changes in dip of underlying planar fault segments. A zone of shear (s) undergoing shear strain ( $\psi$ ) accommodates space problems created by changing the fault geometry. This method predicts an active axial surface a-a' associated with ongoing deformation. Three fundamental fault geometries produce different fold shapes: A fault that flattens towards the surface makes a backlimb that dips away from the direction of vergence; a fault that reverses polarity towards the surface produces a structural wedge with a monocline that dips in the direction of vergence, but with no surface manifestation of a fault.





Figure 9. A. Data used to generate a regional approximation of the erosion surface (the area covers the approximate extent of Figure 2). Dark gray areas are the mapped extent of preserved erosion surface that we have incorporated to date. Light gray areas are Tertiary basins that show 200 m depth contours of the buried surface taken from existing maps. Intervening areas are filled with selected topographic ridge crests. The data are incorporated in a Geographical Information System database that includes the 90 m digital topography (DEM). B. Shaded-relief map of the reconstructed erosion surface, based on the data in A. Box shows the area of Figure 10.



Figure 10. Detail of a portion of the erosion surface map near the northern margin of the Kochkor basin. The gray scale represents 300 m elevation contours of the inferred erosion surface. Preserved remnants of the erosion surface and ridge crests used in the construction are shown. Note that there is little data to constrain the surface in the NW part of the figure. Major faults, the approximate outline of Kochkor basin, and the profile line for Figure 11 are shown.



Figure 11. Profile A-B of the inferred erosion surface map north of the Kochkor basin. The different shades of symbols represent the elevation along the profile (black), the greatest elevation (light gray), and the least (darker gray) within a 10 km-wide window parallel to and centered on the profile line. The linear (~8-9°) slope between elevations 2.1 km and 4 km represents a fold limb in the erosion surface. The lower end of this limb extends beneath Kochkor basin and its extent and subsurface shape are unknown. The axial surface at the top of the fold limb marks an abrupt change in slope to a subhorizontal remnant of the erosion surface. The deformation of the erosion surface is inferred to be the result of displacement on the Shamsi-Tunduk fault and other fault splays that outcrop on the north side of the high relief.



Figure 12. Landsat MSS mosaic showing the area around Kochkor basin; the mosaic boundary cuts across the southern Kyrgyz Range. Heavy lines mark major faults. For reference, the boxes outline the areas of Figures 3 and 10. The thick north-south line A-B marks the cross-section in Figure 13. KB - Kochkor Basin; TB - Tulek Basin; IAF - Issyk-Ata fault; STF - Shamsi-Tunduk fault; AHF - Akchop Hills fault; SKF - South Kochkor fault.



Figure 13. Preliminary north-south cross-section between Son-Kul and Chu basin; the location is shown on Figure 12. Numbers in circles refer to specific features that are discussed in the text. The total shortening in this section is ~16 km. The lesser amount of shortening across the Kyrgyz Range than interpreted farther west suggests that displacement is transferred north to the thrust fault that underlies the ranges that bound the Ili basin in Kazakhstan.



Figure 14. Example of the paleomagnetic data from the At-Bashi River valley. The data shown represent the upper 3 km of the 5 km thick section we have sampled; the rest of the section has yet to be completely analyzed. Each point represents a "site" where at least 3 oriented samples were taken; the mean direction of the sites are plotted according to their stratigraphic position. Large points are inferred to be normal polarity and small reverse. To interpret the data, a normal (black) or reversed (white) magnetozone is assigned if consecutive sites have the same polarity. Inconsistent sites are colored gray and may be either normal or reverse. In portions of the section of mixed polarity, such as at ~2700 m, the dominant polarity is assigned and the anomalous sites are gray.



Figure 15. Based on the approach outlined in Figure 14, the preferred ages of the four sections we have analyzed are summarized here. We have also extrapolated to the base of the Shamsi group and top of the Chu group in each locality using a range of sedimentation rates based on the data available (thick gray areas with dashed outlines). The arrows and thin horizontal bars within the gray areas indicate the range, and the thick bar indicates the best estimate of the ages of the base or top of the Shamsi or Chu, respectively. The preferred age is determined by using characteristic sedimentation rates for the Shamsi and Chu groups. Whereas these extrapolations are highly speculative, they represent our best estimate of the age ranges for the entire Tertiary section in the four basins. Group descriptions and approximate stratigraphic boundaries of the groups are to the right of the sections.



Figure 16. Comparison of cumulative late Quaternary geologic shortening rates and current geodetic shortening rates along north-south profile A-B through the Kyrgyz central Tien Shan (GPS data courtesy of T. Herring and B. Hager). Basin names are labeled above the profile. Figure 2 shows the GPS sites used in the analysis and the location of the profile. Fault names are labeled below the profile line. Faults at the surface that are inferred to intersect at shallow depths (e.g., Akchop Hills and South Kochkor faults) are combined. Dashed lines show the minimum and maximum cumulative geologic shortening rate, with vertical jumps at the faults to indicate rigid blocks separating the faults. Total north velocity and the pattern of distributed shortening are consistent between the data sets.

Fault name	Location of measurement (basin; nearby river or	Slip rate (mm/yr)	95% confidence (mm/yr)	
	latitude, longitude)	· · · ·	+	-
Issyk-Ata	Chu; Alamedin River	2.2	1.7	0.2
Issyk-Ata	Chu; Sokuluk River	0.9	0.3	0.3
Akchop Hills	Kochkor; Djuanarik River	3.1	1.6	0.7
South Kochkor	Kochkor; Djuanarik River	0.2	0.2	0*
Kadjerty	Naryn; Kadjerty River	1.5	0.4	0.3
Kadjerty	Naryn; Kurtka River	1.2	0.8	0.7
Central Naryn	Naryn; Kadjerty River	2.2	0.4	0.3
Oinak-Djar	At-Bashi; 41°N, 75°20'E	0.9	0.3	0.3
North Kyrkungey	Aksay; Kashkasu River	0.2	0.4	0.2
South Kyrkungey	Aksay; Kashkasu River	0.2	0.2	0*

Table 1. Fault slip rates

\*uncertainty less than 0.05 mm/yr