Chronology and tectonic controls of Late Tertiary deposition in the southwestern Tian Shan foreland, NW China

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ABSTRACT

Magnetostratigraphy from the Kashi foreland basin along the southern margin of the Tian Shan in Western China defines the chronology of both sedimentation and the structural evolution of this collisional mountain belt. Eleven magnetostratigraphic sections representing \sim 13 km of basin strata provide a two- and three-dimensional record of continuous deposition since ~ 18 Ma. The distinctive Xiyu conglomerate makes up the uppermost strata in eight of 11 magnetostratigraphic sections within the foreland and forms a wedge that thins southward. The basal age of the conglomerate varies from 15.5 ± 0.5 Ma at the northernmost part of the foreland, to 8.6 ± 0.1 Ma in the central (medial) part of the foreland and to 1.9 ± 0.2 , ~ 1.04 and 0.7 ± 0.1 Ma along the southern deformation front of the foreland basin. These data indicate the Xivu conglomerate is highly time-transgressive and has prograded south since just after the initial uplift of the Kashi Basin Thrust (KBT) at 18.9 ± 3.3 Ma. Southward progradation occurred at an average rate of $\sim 3 \,\mathrm{mm \, vear^{-1}}$ between 15.5 and 2 Ma, before accelerating to ~ 10 mm year ⁻¹. Abrupt changes in sediment-accumulation rates are observed at 16.3 and 13.5 Ma in the northern part of the foreland and are interpreted to correspond to southward stepping deformation. A subtle decrease in the sedimentation rate above the Keketamu anticline is determined at \sim 4.0 Ma and was synchronous with an increase in sedimentation rate further south above the Atushi Anticline. Magnetostratigraphy also dates growth strata at $<4.0, 1.4\pm0.1$ and 1.4 ± 0.2 Ma on the southern flanks the Keketamu, Atushi and Kashi anticlines, respectively. Together, sedimentation rate changes and growth strata indicate stepped migration of deformation into the Kashi foreland at least at 16.3, 13.5, 4.0 and 1.4 Ma. Progressive reconstruction of a seismically controlled cross-section through the foreland produces total shortening of 13-21 km and migration of the deformation front at 2.1–3.4 mm year⁻¹ between 19 and 13.5 Ma, 1.4–1.6 mm year⁻¹ between 13.5 and 4.0 Ma and 10 mm year⁻¹ since 4.0 Ma. Migration of deformation into the foreland generally causes (1) uplift and reworking of basin-capping conglomerate, (2) a local decrease of accommodation space above any active structure where uplift occurs, and hence a decrease in sedimentation rate and (3) an increase in accumulation on the margins of the structure due to increased subsidence and/or ponding of sediment behind the growing folds. Since 5-6 Ma, increased sediment-accumulation $(\sim 0.8 \text{ mm year}^{-1})$ and gravel progradation $(\sim 10 \text{ mm year}^{-1})$ rates appear linked to higher deformation rates on the Keketamu, Atushi and Kashi anticlines and increased subsidence due to loading from both the Tian Shan and Pamir ranges, and possibly a change in climate causing accelerated erosion. Whereas the rapid (~ 10 mm year⁻¹) progradation of the Xiyu conglomerate after 4.0 Ma may be promoted by global climate change, its overall progradation since 15.5 Ma is due to the progressive encroachment of deformation into the foreland.

INTRODUCTION

Although many foreland basins contain upward-coarsening stratigraphic successions (e.g. Covey, 1984; Burbank & Raynolds, 1988; Heller *et al.*, 1988; Jordan *et al.*, 1988; DeCelles, 1994; Bullen *et al.*, 2001; Homke *et al.*, 2004; Jones *et al.*, 2004; Uba *et al.*, 2006), the controls on the coarsening trends are commonly uncertain. An influx of gravel into such basins could result from (1) uplift and erosion of the adjacent mountain front (Armstrong & Oriel, 1965; Burbank *et al.*, 1988), (2) enhanced precipitation or a more erosive climatic environment (Molnar, 2001; Zhang *et al.*,

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2001) that causes an increase in both the size and flux of sediment into a foreland, (3) continued erosion in the face of tectonic quiescence and reduced subsidence within the basin (Heller et al., 1988; Jordan et al., 1988; Flemings & Jordan, 1990; Burbank, 1992; Paola et al., 1992) or (4) a change in either the lithologic resistance of the source area during unroofing or the distance of the depocentre from the active deformation front (DeCelles et al., 1991a; Paola et al., 1992; Jones et al., 2004; Carroll et al., 2006). Adding to the complexity of sedimentation in foreland basins are the multiple ways in which syntectonic folds and faults cut, uplift and often cause re-working of previously deposited basin sediments (Colombo, 1994). Such structures reduce subsidence rates above actively growing folds, may cause ponding of Piggyback Basin (PBB) sediments on their hinterland flanks and create new source areas for basin deposits (Burbank et al., 1996). Thus, lithofacies changes observed at any single location within a basin may only indicate local structural influences and not basin-wide changes.

Within a foreland basin, conglomeratic facies are often localized near the mountain front where transverse streams debouch into the foreland basin and lose much of their sediment-transport capacity. In the simplest scenario, grain size should naturally decrease into the foreland as a result of the decreasing river gradient within the basin and as abrasion reduces the clast sizes (Paola, 1988; Sklar et al., 2006). However, variability in climate (discharge), subsidence rate or source lithology during the evolution of the foreland may cause the conglomerate facies to prograde irregularly into the foreland on different time scales (e.g. Paola et al., 1992). Thus, the geometry of the conglomerate facies is often described as a timetransgressive wedge that thins towards the foreland centre and may be perturbed by growth of structures within the foreland. If dispersion of gravel can be linked to initiation and growth of specific structures, or if structural shortening and gravel progradation rates match, they

suggest a tectonic coupling that is independent of climatic influences.

Both northern and southern margins of the Tian Shan in Central Asia contain a late-stage basin-capping conglomerate. In the southern Tian Shan, the Xiyu conglomerate caps the foreland sequence and has been inferred to represent either the Plio-Pleistocene initiation of deformation of the southern Tian Shan (e.g. Huang & Chen, 1981) or a change in either climate (Molnar et al., 1994) or a climate-affected erosion rate (Zhang et al., 2001). Although all of these factors may contribute to conglomerate around the Tian Shan, the age and primary depositional control of the Xiyu conglomerate remains controversial. Ages between 4.8 and 1.0 Ma have been assigned to the Xiyu conglomerate at different localities around the Tian Shan based on magnetostratigraphy, fossil assemblages and climatic interpretations for the time period (Huang & Chen, 1981; Liu et al., 1996; Burchfiel et al., 1999; Chen et al., 2002, 2007; Sun et al., 2004; Charreau et al., 2005). This age uncertainty of almost 4 Myr obscures the exact timing of rock uplift or climate change, if any, that has driven conglomerate deposition. Moreover, a time-lag commonly exists between a tectonic event in the hinterland and the arrival in the foreland of coarse-grained facies related to that event (Jordan et al., 1988; Jones et al., 2004). In addition to the simple rate of progradation, further lags can result from decreasing proportions of gravel fraction within the sediment supply and variable subsidence rates farther from the thrust front (Paola et al., 1992; Jones et al., 2004). As a consequence, the distribution of conglomerate facies across the foreland is typically time-transgressive, and any single date from an intrabasinal conglomerate is commonly a poor indicator of basin-wide history. Instead, a suite of ages that document gravel progradation are needed to test different causative depositional models.

Located at the southwest corner of the Chinese Tian Shan, the Kashi Basin contains >6-km-thick Tertiary strata (Fig. 1). The arid modern climate, when combined



Fig. 1. Digital elevation model of the Tian Shan region. (a) Shaded relief map of Central Asia showing the proximity of the intracontinental Tian Shan with India and Tibet. (b) Shaded relief map with the Kashi Basin highlighted in white at the northwest corner of the Tarim basin. The study area (boxed region, Fig. 2a) and major geographic locations are labelled.

with active deformation along the southern margin of the Chinese Tian Shan, has exposed numerous sections of foreland basin strata, making it an excellent locality to examine relationships between sedimentation and tectonic deformation. We studied 14 sections throughout the basin, 11 of which contain the Xiyu conglomerate (Fig. 2a). We define how the age, thickness and lithology vary at each section. We also use magnetostratigraphy to date growth strata on the flanks of folds, thereby providing a comparison of structural deformation with lithofacies variation and the appearance of conglomerate across the foreland. We use detailed mapping and magnetostratigraphy of the foreland strata within the Kashi Basin to define: (1) the onset of sedimentation; (2) the distribution of lithofacies; (3) the ages of lithostratigraphic units; (4) temporal and spatial relation of lithofacies to the initial formation and subsequent growth of specific structures within the foreland and (5) the relationship of the Xiyu conglomerate progradation to shortening along the southern margin of the Tian Shan. These conclusions are combined with palinspastic reconstruction of a foreland cross-section to compare the migration of the deformation front and structural uplift with facies migration across the foreland.

BACKGROUND

Western China has a complex geologic history of tectonics and erosion. Major collisional events occurred during the Late Devonian to Early Carboniferous and Late Carboniferous to early Permian, resulting in the complete amalgamation of the Tian Shan region of Central Asia (Windley et al., 1990; Avouac et al., 1993; Carroll et al., 1995; Yin et al., 1998; Dumitru et al., 2001). At least three deformational events occurred during Mesozoic time as a result of the accretion of continental blocks to the South Asian margin (Hendrix et al., 1992). From the late Cretaceous to the Oligocene, the region was an area of tectonic quiescence, with sporadic shallow-marine incursions (Allen et al., 1991). A regional erosion surface was beveled across the Mesozoic and Palaeozoic Tian Shan rocks during the early Tertiary (Burbank et al., 1999; Abdrakhmatov et al., 2001), and this unconformity is present in the Kashi Basin and across the Kyrgyz Tian Shan north of the study area.

Tertiary growth of the Tian Shan was a delayed response to the continental collision between India and Eurasia (Molnar & Tapponnier, 1975; Burchfiel *et al.*, 1999). Structural analyses from different margins of the range have revealed that the magnitude of shortening varies spatially from the east to the west and from the north to the south across the range, as do the timing and rates of deformation (Yin *et al.*, 1998; Allen *et al.*, 1999; Abdrakhmatov *et al.*, 2001; Thompson *et al.*, 2002; Scharer *et al.*, 2004). The total Cenozoic shortening for both the eastern and western Tian Shan may be as much as 124 ± 30 and 203 ± 50 km, respectively (Avouac *et al.*, 1993). Recent geodetic (GPS) shortening rates indicate that the western Tian Shan is shortening at 21 mm year⁻¹, while the eastern Tian Shan is shortening at only 8 mm year^{-1} (Abdrakhmatov *et al.*, 1996; Reigber et al., 2001; Wang et al., 2001). Much of the Cenozoic deformation has likely localized along pre-existing structural boundaries within the Palaeozoic and Mesozoic rocks (Windley et al., 1990), and commenced during the late Oligocene-early Miocene in the eastern and southwestern Tian Shan $(80^{\circ}-87^{\circ} \text{ east longitude})$, as inferred from thermal-cooling data (Hendrix et al., 1994; Sobel & Dumitru, 1997; Sobel & Strecker, 2003), the cessation of marine sedimentation and influxes of coarse-grained strata (Yin et al., 1998). In contrast, initial deformation in the Kyrgyz Tian Shan (\sim 74–80° longitude) began in the middle Miocene (12–16) Ma based on cooling histories, structural analysis and basin sedimentation (Sobel & Dumitru, 1997; Burbank et al., 1999; Bullen et al., 2001; Thompson et al., 2002).

KASHI BASIN STRATIGRAPHY

The Kashi foreland basin in NW China contains a thick (>6 km) section of Tertiary strata that unconformably overlie the Palaeozoic and/or Cretaceous bedrock (Bally et al., 1986) and provide a record of sedimentation since the onset of Neogene deformation in the Tian Shan. Although this entire Tertiary section is not exposed at any one location, we reconstruct the complete foreland stratigraphy using 14 sections (Fig. 2a) totalling more than 15.5 km (~12.5 km of magnetostratigraphy) of stratigraphic thickness. Encroachment of deformation on the foreland has caused uplift and erosion of the northernmost foreland strata, resulting in exposure of the oldest syntectonic sediments in the north, whereas the youngest beds are still accumulating south of the active deformation front. For simplicity, we have separated the Kashi foreland into four regions based on geographic location, bounding structures and geologic similarities. From north to south, we delineate the (1) hinterland, (2) northern foreland, (3) medial foreland and (4) southern foreland (Fig. 2a). The hinterland is defined as the region north of the Kashi Basin Thrust (KBT). The northern foreland consists of the region south of the KBT but north of the Tashipishake Anticline. The medial and southern foreland are located south of the northern foreland basin and separated by the syncline between the Keketamu and Atushi Anticlines (Fig. 2b). Stratigraphy from each region is synthesized based on (1) distinctive formation boundaries, (2) the unconformity along the base of the Tertiary strata and (3) magnetostratigraphic correlation between sections.

Stratigraphic thickness and descriptions, clast counts and palaeocurrent data were collected from the measured foreland strata. Sections were measured with tape-andcompass or with the Jacob staff and the Abney level. Clast counts in conglomerates (matrix or clast lithology, colour and *b*-axis diameter, Table 1) were made every 10 cm along transects perpendicular to bedding to avoid bias due to sorting within individual beds. Data from ~50 counts per site are shown as the percentage of limestone or



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				Clast con		Mean					
Site	Basin location	Palaeomagnetic section	Stratigraphic height (m)	%Lime- stone	%Sand- stone	%Other	'Other' clast type	Total clast counts	Total matrix counts	clast size (cm)	Standard deviation
2Y	Northern	Ayakeqiana	3	6	88	6	ch	32	16	2.0	2.0
4L	Northern	North Qigailike	230	17	84	0		30	15	1.8	1.6
1Y	Northern	Ayakeqiana	275	70	25	5	sh	44	24	3.3	2.9
5NK	Northern	North Keketamu	340	32	62	2	qtz	37	11	5.6	4.3
3Y	Northern	Ayakeqiana	700	80	6	14	sh, ba, ch, gr	50	18	3.9	4.2
6NK	Northern	North Keketamu	790	83	17	0		41	9	5.1	6.0
9K	Medial	Reservoir	1800	60	40	0		47	3	4.6	3.8
1K	Medial	Reservoir	2600	55	38	7	ch, ba	55	19	3.3	2.9
2K	Medial	Reservoir	2600	46	46	9	qtz, meta.	57	13	3.2	2.5
10K	Medial	Reservoir	2700	48	30	22	gr, sh	46	4	4.1	3.9
600K	Medial	Reservoir Bypass	3400	60	40	0		75	25	2.6	3.3
4K	Medial	ASW	3440	86	14	0		44	6	2.4	1.2
15K	Medial	ASE	4600	93	7	0		45	6	2.6	2.8
16K	Medial	ASE	4600	91	7	2	gr	46	6	2.3	2.4
14K	Medial	ASE	4800	62	38	0		50	5	3.1	3.3
4A	Southern	Boguzihe	0	28	58	13	qtz, sh	53	30	1.8	1.2
3A	Southern	Boguzihe	200	12	72	16	ba, qtz, sh	81	42	2.6	1.6
1A	Southern	Boguzihe	500	51	40	9	ch, gr	35	9	2.5	1.8
2A	Southern	Boguzihe	500	25	54	21	gr	43	23	2.6	2.8
5A	Southern	Middle Atushi	550	52	38	10	ba, ch	42	9	2.1	1.6
13E	Southern	Kashi West	600	58	29	13	ch, sh, ba, gr	55	0	3.0	3.2
7A	Southern	Boguzihe	600	42	56	2	ch	52	5	4.5	3.3
11AC	Southern	Active channel	NA	85	15	0		46	5	4.2	2.5
12AC	Southern	Active channel	NA	76	24	0		50	0	4.3	3.4
2BU	PBB	PBB	300	38	72	0		34	13	3.5	4.8
1BU	PBB	PBB	320	51	44	5	gr	44	4	1.7	1.2

Table 1. Clast count data

Clast type definitions: qtz, quartz; sh, indurated shale; ch, chert; ba, basalt; gr, granite; meta., undifferentiated metamorphic; PBB, Piggyback Basin; ASW, Ahu Sisters West; ASE, Ahu Sisters East.

sandstone clasts (Figs 3–5). Palaeocurrent measurements were obtained from channel-axis directions, imbricated clasts and cross-bedding. Typically, at each site, an average direction was calculated from >5 indicators. These data are synthesized across a broad range of outcrops within each stratigraphic lithofacies and are included with the stratigraphic logs (Figs 5–7).

Previous work from the western Tarim basin divided the Tertiary strata into more than nine formations that provided the basis for this study (Fig. 6a, Huang & Chen, 1981; Bally *et al.*, 1986; Zhou & Chen, 1990; Liu *et al.*, 1996; Jia *et al.*, 2004). The stratigraphy within the Kashi foreland, however, differs geometrically and lithologically from strata described elsewhere in the western Tarim, and therefore we avoid using the existing nomenclature unless a formation can be identified with certainty (e.g. Pakabulake Formation, Fig. 6). Moreover, this study does not attempt to describe the detailed sedimentology of the Kashi Basin, although we do provide brief lithofacies descriptions and depositional environment interpretations (Table 2). Instead, we focus on the stratigraphic geometry and lithofacies boundaries, and place magnetostratigraphic age limits on each formation. Overall, this study defines seven Neogene and three pre-Neogene formations: undifferentiated Palaeozoic, Cretaceous, Palaeogene, Wuqia Group (Units A–C, Pakabulake Formation), Atushi Formation, Xiyu conglomerate and PBB strata. These data provide a context for understanding the basin evolution, particularly since the early Neogene.

Fig. 2. Landsat image and geologic map of the Kashi foreland. (a) Landsat image with the locations of measured stratigraphic sections (bold black and white line segments) within the Kashi foreland. In this study, the region is divided into northern, medial and southern basins (shown on left of image) based on relative distances south of the Kashi Basin Thrust. Seismic line TA9402 (shown as dotted line) used for constructing the regional cross-section cuts NE–SW across most of the structures. Outline of the image area is shown in (b). (b) Geologic map of the study area. Major faults and folds referred to in the text are indicated, and the distribution of the Xiyu conglomerate throughout the basin is shown. Location of cross-section A–A' (Fig. 10c) is shown. Note the change from primarily south-vergent thrust faults in the northern half of the study area to folds in the southern half of the region.



Fig. 3. Northern foreland stratigraphy. The three sections are all located in the footwall of the Kashi Basin Thrust (KBT) (Fig. 2a). A sharp unconformity is present between the Tertiary and Cretaceous or Palaeozoic strata at the base of each section. The Xiyu conglomerate caps each section and is truncated on the north by the KBTat North Qigailike and North Keketamu and by a splay fault of the KBTat the Ayakeqiana section (shown). Thus, these sections represent the minimum thickness for the Xiyu conglomerate at these locations. Lacustrine and fluvial strata below the Xiyu conglomerate are interpreted as coarse-grained facies of Unit A in the lower Wuqia Group. Clast composition data are based on clast counts from conglomerate beds in each section, and show a composite unroofing sequence of sandstone- to limestone-dominated at each section (see Table 1 for clast count data).

Pre-Miocene strata

Pre-Miocene strata in the study area are confined to the hinterland, northern and locally in the medial basin regions (Fig 2a). These strata consist of Palaeozoic, Cretaceous and locally Eocene sedimentary rocks and are everywhere overlain unconformably by the Neogene strata. This unconformity varies from an angular unconformity between the Palaeozoic strata and overlying Cretaceous or Tertiary rocks (Fig. 7b), to a sub-parallel disconformity between the Cretaceous and/or Eocene rocks and the overlying Tertiary strata. This implies pre-Cretaceous deformation in the region, but no deformation between deposition of Cretaceous strata and the initiation of Neogene uplift of the Tian Shan. This unconformity is a sharp contact that acts as a marker for correlation of basin strata within the foreland. Palaeozoic bedrock in the Kashi region consists primarily of Permian and Carboniferous limestone and sandstone with rare conglomerate beds (Allen *et al.*, 1999). Within the Kashi foreland, Palaeozoic strata are exposed primarily north of the KBT (Figs 2b and 7a). Where Palaeozoic rocks are exposed south of the KBT, they are intensely deformed (folded and faulted) and consist of mélange blocks of limestone and sandstone in a shale matrix. Palaeozoic strata are interpreted as marine rocks deposited in a regressive marine setting within the northern Kashi foreland (Carroll *et al.*, 1995).

Cretaceous strata consist of distinctive pink, purple and orange siltstone, sandstone and conglomerate locally intruded by dark-grey diorite of early Eocene age (Huang *et al.*, 2006; E. Sobel, pers. comm.). These strata are ~200 m thick at the western edge of the study area but are not preserved east of ~75.7°E. These strata are likely part of the Kezilesu Formation (Jia *et al.*, 2004; Huang *et al.*, 2006) and are interpreted as a braided river system (Miall, 1985; Zhou & Chen, 1990; Hendrix *et al.*, 1992).



Fig. 4. Summary of medial foreland stratigraphy. The relative stratigraphic location and interbasin stratigraphic overlap of each section are shown on the left (see Fig. 2a for section locations). All Tertiary strata below the Xiyu conglomerate are part of the Miocene Wuqia Group, and formation names are listed adjacent to the strata column. The boundary between the Pakabulake Formation and the Xiyu conglomerate is gradational and begins $\sim 100 \,\mathrm{m}$ below the base of the massive conglomerate at 2600 m. Clast compositions show well-mixed lithology in the lower part of the Xiyu conglomerate, but the upper part is dominated by limestone clasts. Count locations are drawn at their appropriate stratigraphic levels within the Xiyu conglomerate.

In the west-central Kashi foreland $\sim 5 \text{ km}$ south of Wenguri village (Fig. 2a), <40 m of dark-green, fissile shale and fossiliferous limestone crop out disconformably between the Cretaceous and Neogene strata. The presence of *Ostrea* (Turkostrea, Jia *et al.*, 2004) within the limestone places this unit within the Eocene, and it likely represents part of the Kalataer Formation (Fig. 6a, Yin *et al.*, 2002; Jia *et al.*, 2004). The base of this limestone is brecciated and has sheared, foliated fabrics, suggesting that it may have been faulted into place. The dark-green, fissile shale may represent part of the marine Wulagen Formation found ~100 km south near Sanju (Fig. 1a, Yin *et al.*, 2002). These strata, like the underlying Cretaceous units, are everywhere sub-parallel but disconformable beneath the Neogene strata and are considered to have been deposited before Neogene foreland development in the region.

Wuqia Group

The Wuqia Group unconformably overlies the pre-Miocene strata and is separated into four distinct units within the Kashi foreland: Units A–C and the Pakabulake Formation (Fig. 6a). These units are correlated stratigraphically



Fig. 5. Stratigraphy of the southern foreland. (a) Stratigraphy from three sections along the south flank of the Atushi Anticline (Fig. 2a). Cross-bedding, ripple marks, channel-axis lineations and imbricated clasts indicate palaeocurrent directions change from west to east in the fluvio-lacustrine Atushi Formation to southward in the Xiyu conglomerate in the Middle Atushi section. (b) Stratigraphic sections correlated along the south flank of the Kashi Anticline (Fig. 2a). Although the Kashi West section is located only 13 km west along the strike from the Kashi Town section, the Xiyu conglomerate is much thinner and located higher stratigraphically at Kashi Town, illustrating the lateral variability of the conglomerate facies. (c) Clast composition data from point counts shown relative to other counts collected at different stratigraphic levels. See text for details.

via tracing of marker beds between sections, and are briefly described here.

Unit A is confined to the northern foreland and comprises 300–350 m of tabular siltstone and sandstone beds. This unit is described from the Ayakeqiana, North Qigailike and North Keketamu sections (Table 2, Fig. 3). Its base is marked by a distinctive 2–10-m-thick orange-brown, well-rounded chert pebble-conglomerate and it grades



Fig. 6. (a) Stratigraphy from the western Tarim Basin after Huang & Chen (1981), Bally *et al.* (1986), Zhou & Chen (1990), Liu *et al.* (1996), Jia *et al.* (2004). (b) Stratigraphy of the Kashi foreland, and probable correlation with western Tarim Basin stratigraphy. Formation thicknesses are based on mapping (this study).

up section into 1–2-m-thick beds of siltstone and shale with interbedded sandstone and conglomerate beds. Mud cracks and mottled pedogenic layers were observed within the siltstone and shale. At the top of the section, Unit A grades into massive cobble conglomerate. Together, these strata are interpreted as fluvial channel and overbank deposits (e.g. Miall, 1996; McCarthy *et al.*, 1997).

Unit B is described from only the medial foreland at the Qigailike section (Fig. 4), but can be traced laterally for more than 100 km across the entire study area. The base is marked by the same <10-m-thick conglomerate unit as below Unit A, but quickly grades up into 300-400 m of tablular, 1–2-m-thick, dark brown shale and siltstone beds. Bidirectional ripples and trace fossils are abundant. Up to 5-m-thick gypsum beds occur locally near the base, and gypsum veins are present within the shale throughout the section. Ripples and trace fossils, tabular siltstone beds and the presence of gypsum are interpreted as recording a lake fringe to shallow lacustrine environment in an arid setting (Magee et al., 1995; Ortí et al., 2003). These strata grade laterally into Unit A to the north, implying that these units were deposited contemporaneously, despite the different grain size and depositional environment. Units A and B may correlate with the Keziluoyi Formation described from other regions of the Tian Shan (Fig. 6, Yin et al., 2002; Jia et al., 2004)

Unit C consists of 200 m of interbedded shale and gypsum (Fig. 4) and is found in the medial foreland (Fig. 2a). The lower 100 m consist of green and grey shale beds interbedded with gypsum beds up to 0.5 m thick. Shale beds are typically fissile and 0.5-1.5 m thick. Overlying these strata are ~ 100 m of red and brown shale and gypsum, and bedding is typically < 0.5 m thick. These strata are laterally continuous and can be traced through the Qigailike and Keketamu sections (Fig. 6) as well as along strike for >100 km. These beds, however, are not found in the northern foreland, where their stratigraphic equivalent is the Xiyu conglomerate. The Unit C lithofacies is similar to the Anjuan Formation within the Wuqia Group (Fig. 6a, Bally et al., 1986; Zhou & Chen, 1990; Jia et al., 2004), but is only ~ 200 m thick in the Kashi foreland whereas it is described as 600 m thick elsewhere in the Tarim basin. These distinctive strata are used as a marker unit to correlate laterally between sections within the medial foreland. The shale and gypsum beds are interpreted as ephemeral lacustrine and playa deposits. The variation between gypsum and shale deposition is likely the response to changes in the lake level within the enclosed basin or distance from the lake edge (Vandervoort, 1997; Carroll & Bohacs, 1999), but the absence of mud cracks and erosion surfaces suggest that this formation was deposited in a sub-aqueous environment.

The Pakabulake Formation makes up the upper part of the Wuqia Group, and is the only unit that can be unambiguously tied to the existing western Tarim basin nomenclature (Figs 4 and 6, e.g. Bally *et al.*, 1986). The formation



Fig. 7. Photos of the Kashi Basin stratigraphy. All views are towards the east. (a) Proximal basin showing the Kashi Basin Thrust (KBT) placing the Palaeozoic strata above the Xiyu conglomerate. The thin piggyback Basin strata (PBB) lie unconformably above the Palaeozoic stratigraphy in the KBT hanging wall. Field of view is approximately 3 km. (b) Angular unconformity between the Palaeozoic shale and Unit A of the Wuqia Group. Bedding is shown as the dashed lines. (c) Steeply dipping (70°) outcrop of Xiyu conglomerate with a limestone megaclast, which is \sim 300 m long \times 50 m wide. The clast is surrounded by proximal facies Xiyu conglomerate. Field of view is approximately 1 km. (d) Gradational contact of the Xiyu conglomerate with the Atushi Formation on the south flank of the Atushi Anticline in the distal basin. Note the lighter-coloured siltstone interbedded with the darker conglomerate. (e) Panoramic view of the medial basin deposits. The base of the Keketamu magnetostratigraphic section starts near the core of the Keketamu anticline and runs north. Solid lines indicate formation contacts, and dashed lines are bedding planes. The resistant Xiyu conglomerate forms the high topography in the core of the Keketamu syncline and along the southern flank off the Keitamu anticline. The stratigraphic distance (white arrows) between Wuqia Group Unit C in the core of the Keketamu anticline to the base of the Xiyu conglomerate is much greater along the southern flank than the northern flank, demonstrating the wedge geometry of the Xiyu conglomerate. Field of view is \sim 4 km.

consists of > 2000 m of dark brown and grey siltstone and shale with 3–5-m-thick, tabular sandstone beds throughout the section. The contact with the underlying Anjuan Formation is defined by the first distinctive 5–10-m-thick, fine-to-medium-grained sandstone bed above Unit C. Tabular sandstone beds contain 1-m-amplitude epsilon cross-beds. These sandstone layers within laminar siltstone and shale are interpreted to represent fluvial channels within an aggrading floodplain (Miall, 1996; McCarthy *et al.*, 1997; Bridge, 2003).

Atushi formation

The Atushi Formation conformably overlies the Wuqia Group and consists of <4000 m of yellow–brown and tan sandstone, siltstone, mudstone, gypsum (thin interbeds <10 cm) and rare pebble conglomerate (increasing near top of section). These strata are similar in appearance to the Pakabulake Formation within the Wuqia Group, but are distinguishable by the lighter yellow–brown colour of

the silt and sand grains within the Atushi Formation (also called Artux Formation at other locations around the Tian Shan). This formation is interpreted as a low-energy meandering stream depositional environment. Onemetre-amplitude epsilon cross-beds within discontinuous tabular sandstone beds are channel facies that swept across the shale floodplain deposits (Miall, 1996; McCarthy *et al.*, 1997; Bridge, 2003). Mud cracks and interbedded gypsum and shale support sub-aerial deposition with ephemeral lakes. The Atushi Formation is defined as Plio-Pleistocene in age (Yin *et al.*, 2002; Jia *et al.*, 2004), but has a gradational contact with the underlying Pakabulake Formation. The basal part of the unit is gypsiferous and may act as the uppermost structural detachment layer within the Tertiary strata.

Xiyu conglomerate

Eleven of 14 measured sections are capped or composed entirely of the Xiyu conglomerate (Fig. 2a). The Xiyu con-

Table 2. Kashi Basin Str	atigraphy		
Formation (~thickness)	Geographic occurrence	Lithologic description	Interpretation
Piggyback Basin strata: PBB Piggyback Basin: PBB (< 300 m)	Hin terland, northern	<i>Shale, siltstone and conglomerate.</i> Basal conglomerate 10–20 m thick: angular clasts, clast supported. Light-brown and green shale with interbedded sandstone beds. Hardpan carbonate layers are interbedded in these massive siltstone beds. All beds are laterally discontinuous with conglomerate lithofacies. The top of these strata are massive clast-supported nebhle and cohble conclomerate	Flucio-lacustrine, medge-top deposition. Braided channel and fan deposition with localized lacustrine environments
Xiyu conglomerate (0–3000 m)	Northern, medial and southern	Massive pebble-cobble conglomerate. Clast-supported and well-sorted grains. Grain size decreases towards south in basin. Trough cross-beds and imbricated clasts are abundant. Lensoidal conglomerate beds are typically 2–10-m-wide, <2-m-thick and truncate underlying strata	<i>Fluvial</i> : braided stream and alluvial fan deposits
Atushi Formation (3000–4000 m)	Southern	Shale and siltstone with interbedded sandstone. Moderate yellow brown tabular siltstone beds. Interbedded gypsum and ripples abundant. Discontinuous tabular sandstone <5 m thick contains 1 m amplitude cross-beds	<i>Fluvial:</i> meandering stream system within and aggrading floodplain. Mud-cracks and gypsum are evidence for deposition in an arid environment
Pakabulake Formation (<2000m)	Medial	Shale and silstone with interbedded sandstone. Brown and gray siltstone and shale: 1–2 m beds with interbedded gypsum, laminations, and ripples. Up to 4–m-thick tabular sandstone beds contain cross-beds at their base and grade up into laminar beds. Erosive base with <0.5 m relief	<i>Fluvial and Lacustrine:</i> Fluvial channel system within an aggrading floodplain
Wuqia Group: Unit C (250 m)	Medial	<i>Gysferous shale:</i> Green-grey and red shale with $\sim 40-50\%$ gypsum. 1–10 cm fissile shale beds interbedded with 1–5-cm-thick gypsum and silt beds. Silt is finely (1mm) laminated. Lower half of formation is greenish-grey and contains five to eight beds of gypsum from 0.5 to 2.0 m thick. Upper half is brownish-red and gypsum is interbedded within the shale in 1–2 cm beds, except for within the lower 10 m of section where three to five beds < 2.0 -m-thick are observed	<i>Shallow lacustrine/playa deposis within</i> an enclosed basin and arid environment
Wuqia Group: Unit B (300 m)	Medial	<i>Fine-medium sandstone and siltstone.</i> Dark red-brown micaceous siltstone and shale: 0.1–1.5 m-thick, fissile, trace fossils abundant, continuous planar beds, minor gypsum, ripples and mud-cracks along top surface of beds	Lacustrine: Near-shore, shallow lacustrine environment
Wuqia Group: Unit A (200–250 m)	Northern	Silistone and sandstone winterbedded conglomerate: Poorly consolidated, $1-2-m$ -thick siltstone beds with interbedded <1m tabular sandstone beds. Occasional lenticular pebble conglomerate. Mud cracks and $2-3$ cm mottled soil peds occur within the siltstone	<i>Fluvial:</i> channel deposits in an aggrading floodplain or ephemeral lake margin
Palaeogene strata	Medial	<i>Limestone and shale</i> : Fossiliferous (ostracods) limestone breccia unconformably below bright green, fissile shale	Marine: shallow and deep marine facies
Cretaceous: Kezilesu (0–500 m)	Hinterland, northern, medial	Sandstone, siltstone and pebble conglomerate: Orange and red, 0.5–1.0–m-thick beds of siltsone. Lenticular, discontinuous sandstone and conglomerate. Distinctive, well-rounded 1–2 cm chert pebbles are found within the conglomerate	<i>Fluvial:</i> Anastamozing stream channels within an aggrading floodplain
Undifferentiated Palaeozoic strata	Hinterland, northern	<i>Limestone, sandstone, shale and wnglomerate.</i> Fossiliferous dark- and light-grey marine limestone. Yellow-brown, well-indurated sandstone with interbedded conglomerate. Dark green and red shale. Occurs as an intensely deformed mélange	Transgressive marine strata

Tian Shan foreland gravel progradation

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glomerate makes up a distinct, mappable unit that can be traced across foreland structures from north to south, and laterally from east to west, across the basin (Fig. 2b). Stratigraphic thickness varies from at least 500 m in the northern foreland, to > 2500 m in the medial basin and decreases to $<100 \,\mathrm{m}$ in the southern foreland (Fig. 8a). Furthermore, lithofacies within the Xiyu conglomerate vary from north to south across the foreland. At its northern limit, the Xiyu conglomerate contains large Palaeozoic limestone megaclasts (< 100 m diameter, Fig. 7c) and the lithofacies are dominated by poorly sorted, matrix-supported conglomerate with angular and sub-angular, pebble-to-boulder clasts (Fig 8b). In the medial basin, pebble- and cobble-sized clasts are typical, and boulders are absent (Fig. 8c). The Xiyu conglomerate in the medial basin contains well-sorted 0.5-1.0 m lenticular beds that

are normally graded with abundant trough and planar cross-stratification. A decrease in grain size and increase in rounding are observed in the southern foreland (Fig. 8d), and the stratigraphic thickness varies from \sim 500 m at Middle Atushi (Fig. 5) to <100 m along the southern edge of the Kashi Anticline (Figs 5b and 8a). Furthermore, strata pinch out laterally in the southern foreland into the fine-grained Atushi Formation. At all locations, the contact between underlying strata and the Xiyu conglomerate is gradational, beginning with a few, 1-5-m-thick conglomerate beds interbedded within finer grained strata that change within 100 m to massive conglomerate with sparse siltstone lenses (Fig. 7d). Conglomerate beds consist of pebble-sized, well-rounded, well-sorted, clast-supported strata sorted into channels between 0.1 and 0.5 m deep (Fig. 8d).



Fig. 8. Stratigraphic architecture and Xiyu conglomerate lithofacies variability across the Kashi foreland. (a) Idealized block model of the Kashi foreland. Structural deformation has not been considered, and the stratigraphic sections are placed in their actual positions relative to each other along the east–west transect. Stratigraphic sections are pinned at the base of the Tertiary strata for the northern and medial forelands, and at the surface in the southern foreland. Surface geology of the block model shows the generalized geology of the foreland, and the locations of gravel and evaporite deposition in the active alluvium. (b) Photo of typical Xiyu conglomerate within the northern foreland, with large, angular clasts. (c) Xiyu conglomerate outcrop in the medial basin showing 0.5–1.0 m beds with sub-rounded cobbles. (d) Smaller clasts and improved clast sorting within the Xiyu conglomerate in the southern foreland.

The Xiyu conglomerate represents a time-transgressive wedge that prograded southward and was derived from the uplifted range front in the north. The northern foreland deposits are interpreted as proximal alluvial-fan deposits, with debris flows and landslides interbedded with braided channel deposits, likely derived locally from the hanging wall of the KBT (e.g. DeCelles et al., 1991a, b). Uncommon large-scale, gravity-driven failures are interpreted to deliver these megaclasts into the proximal foreland from locally high-relief topography (e.g. Blair & McPherson, 1999). Clast composition in the northern foreland changes upsection from sandstone to limestone dominated (Fig. 3) and is interpreted to record the unroofing of Cretaceous and Palaeozoic strata, beginning with the deposition of rounded Cretaceous pebbles in the lower part of the conglomerate and changing upward to limestone pebbles derived from the Palaeozoic strata (e.g. Colombo, 1994). In contrast to the northern foreland, the thick pebble and cobble, clast-supported conglomerate that makes up the medial foreland is interpreted to encompass alluvial fan and braided stream facies that represent the downstream equivalents of the northern basin proximal facies (Miall, 1985). Clast composition varies from mixed (50% sandstone, 50% limestone) near the base of the section to limestone-dominated near the top of the section (Fig. 4). The conglomerate facies in the southern foreland are also interpreted as braided channel deposits; however, the smaller clast size and increased rounding suggest deposition more distal from the mountain front (Paola et al., 1992). Clast composition changes up-section from 60 to 70% sandstone clasts at the base of the Xiyu to approximately equal proportions at the top of the section, in contrast to the active channel of the Boguzihe River where the river clasts are \sim 75% limestone and 25% sandstone (Table 1). Interestingly, granite clasts are only found in the uppermost parts of the Xiyu conglomerate in the medial and southern foreland and within the active channel (Table 1), suggesting recent unroofing of granitic basement within the source area.

PBB strata

Less than 400 m of conglomerate, sandstone and siltstone lie unconformably on top of the Palaeozoic strata north of the KBT (Fig. 7a). Locally these strata overlie the steeply dipping Xiyu conglomerate in angular unconformity and are termed PBB strata. Green and light-brown siltstone in the lower half of the section grades up section into a massive pebble–cobble conglomerate (Fig. 9).

The PBB strata are interpreted as restricted lacustrine and fluvial deposits. Lacustrine deposits are laterally discontinuous, and in some places the entire section consists of massive pebble and cobble conglomerate. These strata are unique in the Kashi Basin in that they constitute an unfaulted sequence of strata that post-dates most deformation in the northern foreland. Strata dip locally $> 60^{\circ}$ north near their basal contact with the Palaeozoic bedrock in the south, but dips decline abruptly upward to $< 10^{\circ}$ at



Fig. 9. Piggyback Basin stratigraphy and magnetostratigraphy from the Upper Boguzihe section (Fig. 2a). The section is continuous from the Palaeozoic angular unconformity at the base to the top of the section, where the conglomerate onlaps the Palaeozoic basement at a buttress unconformity. Magnetostratigraphy shows only two polarity reversals: \sim 200-m-thick reversed zone surrounded by normal polarity zones.

the top of the section. The observed thin stratigraphic successions, local lacustrine facies and a buttress unconformity that separates the PBB from underlying strata along the northern, more proximal basin margin, are consistent with PBB sedimentation in a wedge-top setting (e.g. De-Celles & Giles, 1996). Clast composition within the pebble conglomerate is 40-50% limestone, 45-60% sandstone and <5% granite (Table 1). These PBB conglomerates may be temporally related to the Pleistocene Wusu Conglomerate that unconformably overlies the Xiyu conglomerate in other areas along the margins of the Tarim basin (Sun et al., 2004; Sun & Liu, 2006). Sparse palaeocurrent data indicate west-to-east current directions, parallel to structural strike and contrast with the north-to-south palaeocurrent directions within the Xiyu conglomerate from all other parts of the basin (Fig. 9).

Basin geometry

Together, our stratigraphic descriptions from 14 sections within the Kashi foreland allow us to construct a model for the basin's sedimentary architecture (Fig 8a). By pinning the sections to the unconformity above the pre-Tertiary rocks at the base, and placing each section in its

present-day position south of the KBT, the overall stratigraphic geometry becomes evident. The Wuqia Group forms the stratigraphically lowest part of the Tertiary strata, and grades up-section into the Atushi Formation. The Xiyu conglomerate forms a time-transgressive, southward-prograding wedge that grades up-section into the active alluvial fan surfaces or laterally into the southern foreland within the silt, sand and evaporites south of the deformation front (Fig. 8a). By correlating the Wuqia Group between the northern and medial basin sections, and by tying the Boguzihe and Keketamu sections together by tracing bedding across structures, the combined Tertiary stratigraphic depth of the Kashi foreland is \sim 6 km. Our observation of continuous, uninterrupted stratigraphic sections without obvious hiatuses from the basal disconformity throughout the entirety of each section implies that these strata were deposited before deformation and are foredeep deposits. In contrast, the PBB are likely wedge-top facies deposited after deformation of the foredeep strata.

STRUCTURE OF THE KASHI BASIN

The overall structural geometry of the Kashi Basin can be broadly described as a 'wedge thrust' triangle zone (Banks & Warburton, 1986; Couzens & Wiltschko, 1996; Jones, 1996; Chen et al., 2004), where back-thrusts and detachment folds form in front of foreland-migrating fault-bend folds and duplex structures. The hinterland of the foldand-thrust belt has been uplifted and thrust south above the more distal foreland features. This style of deformation has caused the oldest and deepest part of the foreland strata to be exposed along the northern limit, whereas the youngest strata are observed along the flanks of the detachment folds confined to the Tertiary stratigraphy along the southern deformation front (Scharer et al., 2004; Chen et al., 2007). Although detailed structural descriptions of the Kashi foreland are beyond the scope of this paper, here we summarize structures observed in the northern, medial and southern foreland as well as in the hinterland.

The South Tian Shan, Muziduke and Talas Ferghana faults are all present north of the KBT (Fig. 2b), but are not directly in contact with Tertiary deposits in the Kashi area. We focus our study on structures that deform Tertiary strata, although the presence of reset apatite fission-track ages within the hanging walls of the South Tian Shan and Muziduke faults and their absence from the immediate footwalls of these structures indicate these faults were active between 20 and 25 Ma (Sobel *et al.*, 2006).

The KBT is the major structure that cuts the northern foreland region. It juxtaposes Palaeozoic strata with the Xiyu conglomerate along a sub-vertical thrust fault for at least 80 km along strike (Fig. 2b). Thick (> 500 m) foreland-basin strata are only found south of the KBT. One apatite fission-track cooling age from ~100 m north of the KBT suggests initiation of faulting and rock uplift at 18.9 \pm 3.3 Ma within the hanging wall (Sobel *et al.*, 2006).

The Tashipishake Anticline parallels the KBT ~10 km to the south. This anticline strikes east–west and has Palaeozoic strata exposed in its core and Cretaceous and Neogene strata along its flanks. The region between its southern flank and the KBT defines the northern foreland (Fig. 2a). The anticline is south vergent with a steep-tooverturned south limb and ~50°-dipping north limb. We interpret the Tashipishake Anticline as a fault-bendfold (e.g. Suppe, 1983) with its upper detachment within the gypsiferous units at the base of the Tertiary. Farther south pre-Tertiary strata are rare, and the lower Wuqia Group strata are exposed in the cores of anticlines, suggesting that the basal Tertiary gypsiferous horizons may act as detachment levels within this thin-skinned style of deformation.

The distal basin contains the Atushi and Kashi anticlines: two active structures deforming Plio-Quaternary strata in the southernmost part of the basin. The Kashi anticline is a well-developed detachment fold above a gypsum-rich detachment horizon at between 3 and 6 km depth (Scharer *et al.*, 2004; Chen *et al.*, 2007). The Atushi Anticline, located ~10–15 km north of the Kashi anticline, is also a box-like anticline with steep limbs. Both the Kashi and Atushi folds show a distinct north vergence with steeper limbs on the north flanks: a striking contrast with the south vergence of the faults and folds in the medial and proximal basin.

Surface mapping (Fig. 2b) is combined with seismic line TA9402 (Figs 2a and 10b) to construct a cross-section (A-A': Fig. 10c) across the central part of the Kashi foreland. The time-migrated seismic line image was provided to us by PetroChina Ltd. (Beijing, China), but due to its proprietary nature, we can only present a line-drawing here (Fig. 10b). Approximate depth conversions for the seismic sections rely on a velocity profile for the Kashi foreland provided by PetroChina (Fig. 10a). Velocities above any given travel time were averaged and then multiplied by that time to determine the approximate depth. Depthconverted values were used to construct the balanced geologic cross-section. Despite errors in our depth calculations or in the velocity model, as well as unknown seismic migration errors over which we have no control, these seismic lines provide valuable constraints on subsurface geometry that would be otherwise impossible to infer based only on data at the surface.

Undeformed, sub-horizontal Tertiary strata are clearly observed below the southern half of seismic line TA 9402 (Fig. 10b). We interpret a strong reflector at 4.8–5.0 s (\sim 6.2–6.5 km depth) to represent the velocity contrast between the weak, gypsiferous shale of the basal Tertiary section and the underlying Palaeozoic limestone or Cretaceous sandstone. The calculated depth \sim 6.5 km to this reflector is consistent with our composite basin depth as determined from our measured sections (Fig. 8a), giving us confidence that this horizon represents the base of the Tertiary foredeep strata.

Line-length shortening across all the major mapped structures yields a total, minimum shortening of 13–21 km



Fig. 10. (a) Seismic velocity profile from the Kashi Basin (courtesy of PetroChina). (b) Line drawing of the time-migrated PetroChina seismic line TA9402 shown in Fig. 2a. (c) Geologic cross-section A-A' based on field mapping and depth migration of seismic line TA9402 using the velocity model for the Kashi Basin. Stratigraphic boundaries are based on outcrop locations within the basin, and interpreted stratigraphic depths from the cumulative basin section. The locations of the magnetostratigraphic sections are shown as the boxes above the cross-section. The base of the Tertiary shows up as a strong reflector on the seismic line at ~ 6.5 km depth within the basin, consistent with our totalled measured stratigraphic thickness for Tertiary deposits. (d) Palinspastic reconstruction of deformation within the Kashi foreland south of the Kashi Basin Thrust (KBT). (e) Line-length shortening estimates based on surface mapping for each structure observed south of and including the KBT. Shortening ranges are due to uncertainties in the fault cut-offs and in the sub-aereal geometry of the folds. Details on shortening calculations are in Heermance (2007).

(Heermance, 2007). This estimated shortening range is due both to unknown geometry of the folds that have been eroded and to a general lack of hanging wall cutoffs across faults. These estimates nevertheless allow us to reconstruct the basin architecture (Fig. 10d).

MAGNETOSTRATIGRAPHY

We present magnetostratigraphic results from 11 of 14 measured sections spaced throughout the Kashi Basin and spanning > 12 500 m of stratigraphic thickness (Fig. 2a, Table 3). New magnetostratigraphic data from seven sections [Qigailike, Reservoir, Reservoir Bypass, Ahu Sisters West (ASW), Ahu Sisters East (ASE), Ayakeqiana and PBB] and the lower 1600 m of the Boguzihe section are described here, and data from the Kashi West, Kashi Town, Ganhangou and the upper half (1800 m) of the Boguzihe sections are described in detail elsewhere (Chen *et al.*, 2002, 2007). The new magnetostratigraphic synthesis presented here provides a detailed temporal framework in three-dimensions for late Tertiary deposition and deformation of the Kashi foreland basin. The spatially dense array of dated sections and the total stratigraphic thickness spanned by them define this as one of the most extensive chronologic data sets for an actively deforming foreland basin that has been published to date.

Sampling methods and analysis

Palaeomagnetic samples were collected from 1441 sites (sedimentary layers) spaced throughout the 11 different sections (Table 3). Sample spacing averaged 10 m overall and was less where suitable sampling sites were present, such as at Boguzihe (7.6 m), Ayakeqiana (8.1 m) and Qigailike (4.8 m). Sites were more widely separated within poorly indurated and friable strata, e.g. the Reservoir section, and within massive conglomerates, e.g. Reservoir Bypass and

tation lard 1 Myr ⁻¹)														
Sediment 1 rate stand error $(\pm 2\sigma, m$	20 NA	130	NA	30	10	10	70	20	Na	NA	NA	30	NA	NA
 Sedimentatior rate (m Myr ^{- 1})	720–820 NA	600 520	NA NA	130 - 430	430	430	270	90–250	130 - 430	NA	NA	70–280	100-800?	NA
Standard error $(\pm 2\sigma, Myr)$	2.0 0.9	1.2	NA	2.9	1.4	0.9	0.6	2.1	3.7	NA	NA	1.6	0.6	NA
Calculated timespan (Myr)	4.3 0.8	1.5	NA	10.1	2.4	1.0	2.1	5.0	15.8	NA	NA	3.0	0.4	NA
Interpreted timespan (Myr)	4.0 1.1	2.0	NA	6.8	1.6	1.3	3.3	4.7	12.2	NA	NA	3.5	0.5-4.0?	NA
Magneto-zones	19 4		NA	39	9	5	3	22	64	NA	NA	13	3	134
Sample spacing standard deviation (m)	9.10 NA	17.70	NA	6.3	24.6	18	33.7	1.9	10.5	NA	NA	5.4	3	NA
Average sample spacing (m)	7.6 20.4	10.7	NA NA	11.5	18.1	19.8	41.7	4.8	9.8	NA	NA	8.1	4.7	10.3
Total quality samples	449 41	139	NA NA	286	48	29	28	189	515	NA	NA	92	62	1425
Total quality d sites	403 40	117	NA	202	38	29	28	189	421	NA	NA	81	60	1238
c Total sites collecte	425 44	124	85 85	247	46	73	53	215	544	NA	NA	88	74	1441
Total stratigraphi thickness (m)	3225 815	1250	972.5	2328	688	573	1168	905	3800	480	527	655	290	12 692
pmag	Yes	Yes	No No) Yes	Yes	Yes	Yes	Yes	e Yes	No	No	Yes	Yes	Yes
Site	Boguzihe Ganhangou*	Kashi-West	Middle Atushi	Reservoir (Keketamu	Reservoir Bypass	Ahu Sisters West	Ahu Sisters East	Qigailike	Keketamu Composite	North Qigailike	North Keketamu	Ayakeqiana	Upper Boguzihe	PMAG only
Basin location Southern basin						Modiathacin	INICULATUASIII				Northern basin		Hinterland	Totals

Column heading descriptions are described here. Nawdata from Bowrihe represents the lower 1600 m

New data from Boguzihe represents the lower 1600 m of the section. The upper 1625 m of stratigraphic thickness and the Ganhangou data are presented in Chen et al. (2002) Calculated after Johnson and McGee (1983), average time of magentozones; 210 ka (Neogene); 230 ka (Pliocene).

NA, data not applicable or available.

Basin location, relative location of magnetostratigraphic section to the Kashi Basin Thrust as defined in the text; Site, Stratigraphic section where the samples are located; Total stratigraphic thickness, stratigraphic thickness collected, total number of sites (bedding horizons) sampled within the section; Total quality sites, total number of sites yielding clear polarity directions (described in text); Total samples processed, The total number of samples samples within the section; Sample spacing standard deviation, one standard deviation from the mean distance between two adjacent samples spacing based on the population of distances between any two adjacent samples; Magne to zones, the total number of magnetozones (or partial magnetozones if located at the top or bottom of the section) determined within each section; Correlated timespan, time spanned by stratigraphic section based on our correlation within the geomagnetic polarity timescale (GPTS) (Lourens et al., 2005); Calculated timespan, expected time span calculated for each section based on the statistical method of Johnson & McGee (1983). All data termined graphically (Johnson & McGee, 1983, Fig. 6). These latter three sections have the highest standard deviation of sample spacing due to the poor quality and irregularity of sample sites within the Xiyu conglomerate. The processed at the laboratory; Total quality samples that yield unambiguous normal or reversed polarity directions; Average sample spacing, average sample spacing between the stratigraphically highest and lowest Kashi West section also has a high standard deviation of sample spacing due to two widely spaced (\sim 140 m) samples at the very top of the section within the Xiyu conglomerate. Otherwise, all sections, including most of the Kashi West section, had relatively uniform sample spacing throughout; Standard error, 95% confidence level for the calculated time span of each section after Johnson & McGee (1983); Sedimentation rate, Non-compaction between the lowest and highest magnetostratigraphic samples. Actual measured section is usually longer but the base and top of the sections were not sampled due to either faulting or poor sampling localities; Total sites rate data based on the least squares linear regression through datapoints shown in Fig. 16. Where more than one regression was used to fit non-linear data, the errors reported are the maximum value between the different are calculated based on uniform distribution of sample sites except for the Reservoir Bypass, Ahu Sisters East, and Ahu Sisters West (italicized) that were based on random sample distributions and their time-span was decorrected sedimentation rate calculated for a section based on the correlated magnetostratigraphy and stratigraphic thickness (Fig. 16); Sedimentation rate standard error, 95% confidence interval for the sedimentation regressions

Table 3. Magnetostratigraphic data from the Kashi Basin

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ASW and ASE, where samples were collected opportunistically from irregularly spaced, thin-siltstone and finesandstone beds. Two or three oriented cores per site were obtained *in situ* using a gas-powered drill with a 2.5-cmdiameter core bit. The core-plate orientation and bedding attitude of the site were determined by a magnetic compass and, when possible, a sun-compass.

Sample spacing was deemed sufficient to capture most of the palaeomagnetic reversal events preserved in the Kashi Basin strata for the following reasons. Previous work within the Kashi foreland documented Plio-Pleistocene sediment accumulation rates of 600-800 m Myr⁻¹ (Chen et al., 2002, 2007). Miocene and Pliocene sediment accumulation rates from elsewhere within the northern Tarim Basin vary between 200 and 700 m Myr $^{-1}$ (Sun *et al.*, 2004; Charreau et al., 2006; Huang et al., 2006), although the late Oligocene and early Miocene rates may have been as low as 70 m Myr⁻¹ (Huang *et al.*, 2006). The average timespan of a magnetozone during the Neogene is 0.21 Myr, but ranges from 0.01 to 1.07 Myr (calculated from Lourens et al., 2005). Thus, by assuming that most of our strata are confined to the Pliocene and Miocene and that deposition rates vary between 200 and $800 \,\mathrm{m}\,\mathrm{Myr}^{-1}$, the 10 m average site spacing would span between 0.012 and 0.05 Myr. This spacing would sample at least four times an average magnetozone timespan of 0.21 Myr. Furthermore, all zones > 0.1 Myr should be sampled at least twice.

Magnetic polarities were determined for each analysed sample using standard methods described in Appendix A. One specimen per sampling site or stratigraphic level was initially measured for natural remnant magnetization, then subjected to alternating-field (AF) demagnetization up to 100 G in 15 or 25 G steps to remove low-coercivity magnetizations, and subsequently treated with stepwise thermal demagnetization using 6-12 demagnetization steps between 50 and 680 °C, typically in (1) 150 °C steps up to 300 °C, (2) 100 or 50° steps up to 500 or 550 °C and (3) 30 or 20° steps up to 680 or 690 °C (Fig. 11). At each site, a second and third specimen were demagnetized if the first yielded unstable or overprinted directions or if a magnetostratigraphic interval of normal or reversed polarity was defined by only one specimen. Thus, all magnetozones are defined by at least two samples with consistent polarity.

Many previous magnetostratigraphic studies employ a method that ranks sites based on two or more samples per site (e.g. Johnson *et al.*, 1982; Schlunegger *et al.*, 1997; Kempf *et al.*, 1999; Ojha *et al.*, 2000; Chen *et al.*, 2002). We do not utilize this method for the following reasons: (1) all samples are run through the magnetometer right-side up and upside down for each thermal demagnetization step to determine whether or not the sample gives consistent readings; (2) the high accuracy of the computer-controlled magnetometer and consistency of the sample changer at the Caltech lab reduce much of the laboratory uncertainty in measurements that plagued early studies; (3) our method likely sampled each magnetozone at least twice (but four on average) and (4) all single-site reversals are checked for consistency with samples from both the same stratigraphic

level and the adjacent sites stratigraphically above and below. For these reasons, we did not consider it necessary to run multiple samples from each site unless the site was at a reversal boundary or defined its own magnetozone.

Magnetozones and correlation with the geomagnetic polarity timescale (GPTS)

Within the 11 sampled sections in the Kashi Basin, 134 magnetozones were identified (Figs 12 and 13, Table 3). Reversal boundaries were drawn at the midpoint between two superposed sites with opposite polarities. Thus, one source of error in the age assigned to strata associated with any individual reversal is a function of the sedimentation rate and of the stratigraphic distance between the sites defining the reversal. Correlation of the magnetostratigraphy [magnetic polarity timescale (MPTS)] of each section to the GTPS of Lourens et al. (2005) was assisted by the construction of longer composite magnetostratigraphic sections based on stratigraphic ties between individual sections. These composites provide distinctive magnetostratigraphic patterns that, given a few assumptions about the age or timespan of the strata, can be correlated with the GPTS.

Southern foreland results

Correlation of southern foreland deposits to the Pliocene and Pleistocene epochs is based on: (1) ostracods within the Boguzihe section that place them within the Pliocene (Jia et al., 2004); (2) the observation that the section grades upwards into the active Quaternary fan surfaces without obvious major unconformities or hiatuses, implying the deposits are young (Fig. 5) and (3) the Atushi Formation, which dominates the section stratigraphy, has traditionally been considered Pliocene (Sun & Liu, 2006). Following the previous correlations of Chen et al. (2002) and using the extended magnetostratigraphy of the 3700-m-thick Atushi Anticline section, we interpret the 23 magnetozones defined at the Atushi Anticline section to span from \sim 5.5 Ma to \sim 1 Ma (Fig. 12). This correlation to the GPTS appears unambiguous except for short magnetozones that are based on single sites, and thus ignored, in the upper part of each section.

Strata exposed in the Kashi Anticline correlate via cross-section A–A' (Fig. 10c) with the upper part of the Atushi Anticline strata and are, therefore, Pliocene or younger. The magnetostratigraphic sections from the Kashi Anticline (Kashi Town, Kashi West) are inter-correlated by using air photos to trace beds between them (Fig. 14, Chen *et al.*, 2007: Fig. 2a). These sections are dominated by reversed polarities and their upper parts correlate with the reversed zone between 1.6 and 0.78 Ma on the GPTS (Lourens *et al.*, 2005). The base of the Kashi West section extends to \sim 3 Ma. The Kashi Town section was sampled up to the active fan surface, where its magnetos-tratigraphy is interpreted to terminate in the Brunhes chron at <0.78 Ma (Fig. 12).



Fig. 11. Examples of palaeomagnetic results. (a–d) Results from stepwise alternating-field and thermal demagnetization of selected oriented samples from the Keketamu composite (b–d) and Qigailike (a) sections. Zijderfeld and equal area plots are shown side by side, with some of the major steps labelled. Examples of the low-temperature component (LTC) and high-temperature component (HTC) of the characteristic magnetic remanence (ChRM) are shown in (a). (e–f) Isothermal remanent magnetism (IRM) within an increasing magnetic field of the samples shown in (c) and (d), respectively. Both magnetite and hematite are interpreted as the carriers of ChRM for both samples.

Progressive limb rotation of the flanks of the Atushi and Kashi anticlines during deformation produced growth strata that display fanning dips and angular unconformities (Scharer *et al.*, 2006; Chen *et al.*, 2007) that are dated at 1.4 (Kashi West) and 1.07 Ma (Kashi Town) along the Kashi Anticline, and at 1.4 (Boguzihe) and 1.2 (Ganhangou) Ma along the Atushi Anticline (Fig. 12). Two ages of growth strata

from the Kashi Anticline provide evidence for lateral fold propagation at 40 mm year⁻¹ (Chen *et al.*, 2007). Moreover, these ages only apply to the exposed and dated parts of the fold. The growth-strata ages used to define fold initiation for both the Atushi and Kashi anticlines are the older of two separate such ages for each fold; only one growth-strata age, however, was obtained from the Keketamu anticline.



Fig. 12. Correlation of the southern foreland magnetic polarity timescale (MPTS) with the Pliocene Epoch on the geomagnetic polarity timescale (GPTS) of Lourens *et al.* (2005). Data from the upper 1600 m of the Atushi Anticline and from the Kashi Anticline are from Chen *et al.* (2002, 2007), respectively. Individual sample-site coordinates are listed in Heermance (2007). Only zones with more than one sample are used to define a zone, and black and white zones indicate normal and reversed polarity, respectively. Inset shaded relief map shows the section locations within the southern foreland region (highlighted). The contact between the Xiyu conglomerate and Atushi Formation, as well as growth-strata locations, are shown within the lithology logs adjacent to each plot of magnetostratigraphy.

Medial foreland results

Magnetostratigraphy from the medial basin is based on five sections: Qigailike, Reservoir, Reservoir Bypass, ASW and ASE (Fig. 13). Despite 78 total magnetozones defined by these sections, correlation of any individual (< 1000 m) section with the GPTS is ambiguous due to a lack of radiogenic ages or biostratigraphy to pin the magnetostratigraphy to a specific reversal sequence. To help remedy this ambiguity, a 3800-m composite

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Fig. 13. Proximal and medial basin magnetic polarity timescales (MPTS) and correlation with the geomagnetic polarity timescale (GPTS) of Lourens et al. (2005). Section locations are shown in highlighted region of the inset digital elevation model (A). Virtual geomagnetic pole latitudes (VGP) for all samples are shown relative to their stratigraphic position and indicate normal and reversed polarity for each sample. Site coordinates are listed in Heermance (2007). Interpreted ages for the base of the Xiyu conglomerate in the Reservoir and Avakegiana sections and the age of growth strata are shown. Stratigraphic correlations tie together the sections for a complete chronology of the medial and northern foreland. Sampled lithology is shown adjacent to each VGP plot.

magnetostratigraphy containing 64 of the 78 magnetozones was constructed by tracing and correlating beds between the Reservoir, Reservoir Bypass and the ASW section (Fig. 14). The Qigailike data are correlated with the base of the Reservoir section by the distinctive marker sandstone bed at the base of the Pakabulake Formation. The Qigailike data are used in lieu of the Reservoir data here because they have a denser (\sim 5 vs. 12 m) sample-site spacing and extend to the base of the Tertiary strata. Although the stratigraphic position of the ASE section that caps the medial basin sequence is well defined (Fig. 14), it is not included in the composite section (Fig. 15) because the sample spacing through the conglomerates is too large. Correlation of the medial basin composite with the GPTS of Lourens *et al.* (2005) is based on the following three criteria: (1) The Anjuan and Pakabulake Formations within the Wuqia Group have previously been assigned to the middle and late Miocene (Yin *et al.*, 2002; Jia *et al.*, 2004). This interpretation is consistent with their position unconformably above the biostratigraphically dated Late Eocene Kalataer Formation. (2) The uppermost medial basin composite section may be traced laterally across folds into the distal basin deposits at the Boguzihe section (Fig. 15a), making them Pliocene in age (this study, Chen *et al.*, 2002). The top of the medial basin composite section, however, must be significantly older than Pleistocene, because the 1000-m-thick ASE section is located above the

Fig. 14. Aereal photo showing sample-site locations and the correlation between sections from the Keketamu composite magnetostratigraphy. Correlation between sections was completed by tracing beds along strike between sections (white dashed lines). Normal (black) and reversed (white) polarity sites are shown as the circles shown on the photo. Only some of the sites are numbered for clarity.

medial basin composite MPTS, does not include the Brunhes normal chron, is dominated by reversed polarity and contains at least two, but likely five, reversals (Fig. 13). (3) The statistical method of Johnson & McGee (1983), described below, suggests that the section is likely to span 16 ± 4 Myr, predicting an average sedimentation rate of 0.22 mm year ⁻¹. This rate is consistent with rates of 0.13- $0.43 \,\mathrm{mm\,year^{-1}}$ inferred for the middle and late Miocene from the northern and southern Tian Shan flanks (Charreau et al., 2005, 2006; Huang et al., 2006). This range of rates suggests that the total time span for the 3800-mthick medial basin composite MPTS is between 30 and 10 Myr (Fig. 15b). In sum, given the vertical frequency of samples, the pattern of reversals and an estimated duration of 12–20 Myr, we conclude that a convincing match can be made to only one part of the GPTS between 35 and 1 Ma: between \sim 18 and 5.5 Ma (Fig. 15c).

The best-fit correlation ties magnetozone K16n (Figs 13 and 15c), an anomalously long normal polarity zone that spans \sim 400 m, with the longest normal zone in the Neogene between 10.0 and 11.0 Ma. Based on this tie point, a reasonable correlation is made between \sim 18.3 and 5.5 Ma by simply matching magnetozones between the MPTS and GPTS. The long reversed magnetozone (\sim 100 m

thick) above L7n at Qigailike correlates best with the long reversed chron from \sim 16–15 Ma in the GPTS. Although this preferred correlation of the MPTS to the GPTS (Figs 13 and 15c) appears reasonable and the number and spacing of reversals are consistent between the two, some mismatches exist. For example, an 'extra' reversed magnetozone below K10n (Fig. 13) does not appear to correlate with the GPTS, and may result from unremoved remanence that added an extra normal polarity zone at K11n. The most uncertain correlations are for the base of the Qigailike section where many reversals within a short stratigraphic section imply slow sedimentation rates, similar to previous studies for the early Miocene along the southern Tian Shan foreland 1000 km east in the Kuqa Basin (Fig. 1, Huang et al., 2006). Our preferred correlation extends the base of the Qigailike section to 17.5-18 Ma, and produces a match with the GPTS that shows less time ($\sim 5 \text{ Myr}$) spanned by the upper half of the section than the lower half (~ 8 Myr), consistent with increasing sedimentation rates since the early Miocene (Huang et al., 2006).

Correlation of the ASE section located above the medial basin composite is more ambiguous due to widely spaced sites (average spacing: 40 m) within the Xiyu conglomerate, such that some reversals may not have been sampled. (b) Hypothetical magnetostratigraphic time span

Fig. 15. Correlation for the medial basin composite stratigraphy. (a) Landsat image showing correlation of the top of the medial basin with the distal basin magnetostratigraphy. (b) Hypothetical length of the magnetostratigraphy based on assumed slow $(0.13 \text{ mm year}^{-1}, \text{left})$ and fast $(0.4 \text{ mm year}^{-1}, \text{right})$ sedimentation rates. (c) Preferred magnetostratigraphy that spans ~ 16 Ma, and provides the best fit to the geomagnetic polarity timescale (GPTS). See text for complete explanation. (d, e) Alternative correlations of the medial basin composite section with the GPTS.

Moreover, three of the magnetozones are based on single sites (grey bars, Fig. 13) and are not included in our official analysis. Thus, several different correlations are possible. One correlation places the top of the ASE section, which corresponds to the location of the growth strata, at \sim 2.5 Ma (Fig. 13). If we allow the sediment accumulation rates to double to 800 m Myr⁻¹ after ~ 6 Ma (Chen *et al.*, 2002), the top of the ASE section and the growth strata would occur at \sim 4 Ma. This span of 2.5–4 Ma set reasonable limits on the likely age of the top of the section in the medial basin.

Alternative correlations of the composite magnetic section (Fig. 15d and e) are limited by assumptions for the maximum and minimum ages interpreted for the MPTS. Magnetostratigraphic studies of middle and upper Miocene strata elsewhere in the Tian Shan foreland suggest accumulation rates could range from 0.13 to 0.43 mm year $^{-1}$ (Charreau et al., 2006; Huang et al., 2006). Applying these end-member rates to our data (Fig. 15b) predicts too many or too few reversals within the calculated time to be confidently correlated anywhere to the GPTS within the Mio-Pliocene, thereby suggesting an average rate midway between these end members. Compared with our preferred match to the GPTS, attempts to produce either a younger or older correlation yield less-satisfactory matches. Alternative correlation A (Fig. 15d) places the medial basin composite section as young as possible. As discussed above, chron K4n cannot be younger than 1.78 Ma because the ASE section above the medial basin composite (Fig. 14) contains at least one normal polarity zone with more than two sites, and the top of the ASE cannot include the Brunhes chron younger than 0.78 Ma. Correlation of the magnetostratigraphic pattern back in time from 1.77 Ma requires that observed magnetozones be ignored (e.g. K9n, K11n and K13n), whereas other subchrons in the timescale are not observed in the local magnetostratigraphy. Although the same basal age of 17.5–18 Ma as in our preferred correlation is predicted for this alternative correlation, the mismatches with the GPTS suggest that this alternative correlation is less likely.

Alternative correlation B (Fig. 15e) attempts to correlate the base of the medial basin composite to the late Oligocene at ~ 25 Ma, which is the estimated age of uplift for the southern flank of the Tian Shan (Sobel & Dumitru, 1997; Yin et al., 1998). This correlation spans from \sim 30 Ma to \sim 11.5 Ma and matches the GPTS quite well. This correlation predicts, however, that the top of the medial basin composite (Fig. 15e) would be ~ 11 Ma, much older than the same stratigraphic level within the distal basin at Boguzihe (Fig. 15a). The absence of any obvious depositional hiatuses within the sections combined with our ability to trace bedding between structures strongly argues that the top of the medial basin composite corresponds with the lower Pliocene and not older, as an alternative correlation would suggest. Thus, we dismiss alternative B as a probable correlation.

Northern foreland results

Similar to the Qigailike section, the Ayakeqiana section in the proximal basin lies unconformably on Cretaceous strata. This unconformity serves as the stratigraphic marker for correlating the sections and tying the northern foreland magnetostratigraphy with the GPTS. Within the 13 magnetozones at Ayakeqiana (Fig. 13), the prominent reversed magnetozone (Y3r) is interpreted to correlate with the reversed chron between 15.1 and 16 Ma on the GPTS based on stratigraphic height above the basal unconformity (similar to the correlation at Qigailike). Matching of magnetozones below Y3r suggests that the base of the section extends to ~17.5 Ma.

PBB results

The PBB is distinct from the Xiyu conglomerate in that it lies unconformably over the sub-vertical, middle Miocene Xiyu conglomerate and is preserved only north of the Tashipishake Anticline (Fig. 2b). Only two reversals were observed in PBB magnetostratigraphy within the relatively short 400-m-thick section (Fig. 9), making correlation with the GPTS ambiguous. The PBB must be younger than ~14 Ma, the age of the Xiyu conglomerate that it unconformably overlies, but older than the bottom of the Brunhes chron at 0.78 Ma, because of the reversed polarity zone observed within the section (Lourens *et al.*, 2005).

Expected time span of section correlations

To assess our correlations against the expected duration of each section, we used the statistical method of Johnson & McGee (1983). This method uses (i) the number of reversals discovered in a stratigraphic section; (ii) the number of sample sites within which those reversals were found; (iii) a time estimate for the average duration of polarity intervals during the Neogene (210 kyr.) or Plio-Pleistocene (220 kyr) and (iv) a choice of equal or random site spacing to calculate the expected time spanned by each section. The results (Table 3) indicate excellent agreement (within error) between the statistically estimated duration of each section (Johnson & McGee, 1983) with that predicted by our correlation with the GPTS (Lourens et al., 2005) for all sections except for the Reservoir section. The calculated time spanned by the Reservoir section is 10 Myr, which is \sim 3 Myr more than our correlated result. This mismatch may be attributed to the fact that the frequency of reversals between 15 and 8 Ma is greater than that assumed in the Johnson & McGee (1983) model. Despite the Reservoir section disparity, this statistical model, in combination with microfossil constraints from the Boguzihe section and tying the Keketamu section to chron C5n within the Miocene, suggests our correlations of these magnetostratigraphic sections to the GPTS are reasonable.

Sediment-accumulation rates

This extensive magnetostratigraphy defines sediment-accumulation rates for the sampled localities in the Kashi Basin (Fig. 16). The accumulation rates are not corrected for post-depositional compaction, in part because of the unknown compaction history. Given the prevalence of carbonate in the source area and the arid to semi-arid conditions, early cementation is likely, such that standard porosity-vs.-depth curves would be inapplicable. Observed rate changes do not correlate with lithologic boundaries, and thus we interpret them as independent of compaction or lithology. Nonetheless, rates presented here should be minima that show how sedimentation rates change at specific localities relative to basin structures and geometry. From these data, we interpret how tectonic loading and deformation relate to sedimentation within the basin.

Overall, accumulation rates are fastest for younger strata such that at any given position, rates tend to accelerate from older to younger. The oldest Tertiary strata (Wuqia Group), however, have generally the finest grain size (shale) that may be more compacted and thus exhibit a slower apparent accumulation rate. Within any section, the rates may have remained steady for as much as 4 Myr. Rates before 16.5 Ma in both the medial and southern foreland are comparable and represent the slowest rates in this study (~80 m Myr⁻¹). After 16 Ma, the rates accelerated by more than three-fold, but are higher in the proximal basin where the Xiyu conglomerate first appears at ~15.5 \pm 0.5 Ma (Fig. 16). Another three-fold increase in

Fig. 16. Plot of stratigraphic depth vs. age based on magnetostratigraphic correlations for each measured section (locations shown in inset map). For visual clarity and to avoid overlapping data, stratigraphic depth measurements are relative to only the section measured and should not be used to correlate between sections. Sedimentation rates are calculated based on a minimum of five reversals from within each section. Specified uncertainties on sedimentation rates are $2-\sigma$ values from least-squares linear regression through all data as shown by the lines behind the data points. Sections have not been corrected for compaction effects, and thus the slopes calculated for each represent minimum sedimentation rates at each section. The stratigraphic location of the Xiyu conglomerate in each section is indicated with the correlated age for the base of the conglomerate within each section. Age correlation of basin strata are shown along the base of the graph.

sediment accumulation rate from 130 to $430 \,\mathrm{m}\,\mathrm{Myr}^{-1}$ occurred at 13.5 Ma at the Reservoir Section.

Pliocene accumulation rates within the Atushi Formation in the distal basin were up to twice as rapid as Miocene rates in the medial basin, despite less conglomerate deposition (Fig. 16), but are consistent with a regional increase in sedimentation rate inferred for the Pliocene (Métivier et al., 1999; Zhang et al., 2001). Rates for the Atushi and Kashi anticlines sections average 700 m Myr⁻¹, but appear dependent on position within the basin: faster rates prevail in more northerly and westerly sections. Regional subsidence during the Pliocene could have provided accommodation space for climatically influenced rapidly eroding mountains surrounding the basin during the Pliocene (Molnar, 2001; Zhang et al., 2001). The flexural rigidity of the Tarim platform is interpreted to be $\sim 10^{24}$ N m (Burov & Diamont, 1992) and as such the forebulge should be present at least 200 km from the tectonic load (Beaumont, 1981). However, the Kashi Basin lies within 50 km

of the Tian Shan mountain front and within 100 km of the Pamir mountain front, and thus is located within the flexural foredeep of two active, converging mountain ranges. Hence, the Kashi Basin is likely becoming deeper as the Tarim basin closes up at this location, providing accommodation space for an increased sediment supply during the Pliocene and Pleistocene.

DISCUSSION

Stratigraphic architecture and facies development of the Xiyu conglomerate

The Xiyu conglomerate forms a southward-thinning wedge that caps Tertiary foreland strata at almost every location within the foreland. The conglomerate can be mapped as a continuous unit (i.e. formation) across all the structures south of the KBT (Fig. 17), despite its occur-

Fig. 17. Schematic geometry of the Kashi foreland showing the magnetostratigarphic sections at their reconstructed pre-deformation basin locations with accurate stratigraphic thickness. The magnetostratigraphic correlation on the right shows the correlation with the geomagnetic polarity timescale of Lourens *et al.* (2005). Ages for the foreland stratigraphy, magnetostratgraphic sections and sediment accumulation rate changes are indicated in the figure. Spatial location and the interpreted initiation age for the major basin structures are indicated along the base of the figure. Ages for the top and bottom of each magnetostratigraphic section, and interpreted changes in sediment accumulation rate, are shown in the figure.

rence at different stratigraphic levels throughout the foreland. At its northern limit, the conglomerate appears $< 200 \,\mathrm{m}$ above the base of the Tertiary and is $> 400 \,\mathrm{m}$ thick (Fig. 3). Here the conglomerate is truncated by the KBT and thus represents its minimum thickness at this location. Within the medial foreland, the conglomerate is > 2500 m thick, and occurs approximately 2000 m above the base of the Tertiary (Fig. 4). Within the southern foreland along the flanks of the Atushi and Kashi anticlines, the Xiyu conglomerate forms a thin (<300 m) veneer above siltstone and sandstone of the Atushi Formation (Fig. 5). Here, the conglomerate forms digitate lobes that jut into the youngest foreland strata. Furthermore, the locally thickest Xiyu conglomerate at these distal locations is located where the active rivers cross folds, and these pinch out within a few kilometres laterally (e.g. Fig. 8a).

Combining all of the lithofacies and chronostratigraphic data, the picture that emerges is of laterally continuous conglomerate deposition that prograded south over time. The coarse-grained facies are localized along the mountain front, whereas at more distal locations the conglomerate lithofacies are restricted to palaeovalleys that cut southward across the basin (e.g. López-Blanco, 2002). The persistence of the active water gaps localized through the thickest, and more resistant, conglomerate facies along the southern foreland suggests that incision has outpaced uplift, and that rivers have been long lived (>1-2 Myr) across the landscape. Conglomerate deposition in the northern foreland was contemporaneous with evaporite deposition farther south (Fig. 8a), similar to the present conditions where gravel deposited within the Baishikeremu He (He is Mandarin for 'river') channel south of the Kashi anticline occurs only within $\sim 6 \,\mathrm{km}$ of the southern exposed fold flanks. South of this 'gravel front' are areas of strictly evaporite, fine-grained fluvial and aeolian deposits (Fig. 8a).

Time-transgressive Xiyu conglomerate progradation

Five magnetostratigraphic age constraints combined with the location of the southern edge of active gravel deposition provide six basal ages for the Xiyu conglomerate wedge across the Kashi foreland. These ages of $15.5\pm0.5\,\mathrm{Ma}$ (northern foreland), $8.6\pm0.1\,\mathrm{Ma}$ (medial foreland), 1.9 ± 0.2 Ma (distal foreland), ~ 1.4 Ma (not bounded by reversals, distal foreland), 0.7 ± 0.1 Ma (distal foreland) and 0 Ma allow us to illustrate the profoundly time-transgressive nature and provide insights on the apparent disparity of ages for the Xiyu conglomerate found around the Tian Shan and Tarim basin (Fig. 18a). Errors associated with each age designation are based on the time span of each reversal in which the base of the Xiyu conglomerate was observed. Only four previous magnetostratigraphic studies provide absolute dates on the age of the Xiyu conglomerate near the Tian Shan. Charreau et al. (2005) and Sun et al. (2004) interpreted different ages for the Xivu conglomerate at the Dushanzi section on the northern flank of the Chinese Tian Shan at 4.8 and 2.7 Ma. Magnetostratigraphy from the southern margin of the Tarim Basin provide ages of 3.5 Ma at Yecheng (Zheng et al., 2000) and 3.0 Ma (Sun & Liu, 2006) at Sanju, \sim 300 km southeast of the Kashi Basin. However, all of these magnetostratigraphies are based on one measured section within each foreland region, or two sections correlated along strike within the same part of the basin relative to the mountain front. Other ages given to the Xiyu conglomerate in the Tian Shan are based on assumed correlations with global climate cooling at 2.5 Ma (Avouac et al., 1993; Burchfiel et al., 1999) or its stratigraphic position above Pliocene strata (Huang & Chen, 1981; Yin et al., 1998). Furthermore, the Xiyu conglomerate has been interpreted as a chronostratigraphic unit around the Tian Shan (Huang & Chen, 1981; Liu et al., 1996; Burchfiel et al., 1999). Although these studies do not systematically study the spatial or temporal variability of the Xiyu conglomerate within any individual basin, the individual, site-specific ages are commonly used as evidence for either tectonic or climatic events since the Pliocene. In contrast, we demonstrate here that the Xiyu conglomerate in the Kashi region is a highly time-transgressive lithofacies that has continuously prograded south over > 15 Ma.

The depositional ages for the arrival of gravel at different locations within the Kashi foreland, combined with reconstructed shortening estimates across individual basin structures (Fig. 18b), allow us to reconstruct pre-deformation gravel progradation rates into the foreland (Fig. 18c). Reconstructed distances along cross-section A–A' from the initial appearance of the Xiyu conglomerate at the Ayakeqiana section to the Keketamu and Atushi Anticlines are 16–22 and 37–44 km, respectively. Based on the appearance of the conglomerate at 15.5 Ma at Ayakeqiana, 8.6 Ma at Keketamu and 1.9 Ma at Atushi, a fairly constant rate (based on these three points) of 2.5–3.5 mm year⁻¹ is calculated (Fig. 18c). In contrast, using the reconstructed distances between the Atushi and Kashi Anticlines and the furthest south location of gravel within the active river channels, and ages of 1.9, 0.7 and 0 Ma, respectively, the gravel front migrated at 10–11 mm year⁻¹ since 1.9 Ma (Fig. 18c). Furthermore, west to east migration of the gravel front along the southern flank of the Kashi Anticline occurred at ~18 mm year⁻¹ since 1.4 Ma, suggesting that gravel progradation rates are fastest towards the southeast. This three- to four-fold increase in southern gravel progradation rates occurred after ~2 Ma. Overall, the Xiyu conglomerate has prograded 54–62 km south since its initial deposition at 15.5 Ma for an overall rate of 3.5– 4 mm year⁻¹, although higher gravel progradation rates may be southeasterly towards the Tarim Basin centre.

Implications of observed changes in sedimentation rate

At least three distinct changes in sediment accumulation rates at \sim 16.3, 13.5 and \sim 4 Ma are interpreted within the Kashi foreland, and these ages are independent of lithofacies changes and formation boundaries that could be attributed to a climate shift (Fig. 16). Sedimentaccumulation rates within terrestrial basins are controlled primarily by rates of sediment supply and accommodation space. Sediment supply is a function of climate that may increase the runoff and thus available stream power for fluvial erosion (Molnar, 2004), source-rock lithology that is either easily denuded and re-deposited or resistant and thus not as readily available for re-deposition (Carroll et al., 2006), and the proximity of the sediment source (Burbank & Beck, 1991). Accommodation space is created either by tectonically induced subsidence beneath an uplifted block (Beaumont, 1981; Jordan, 1981) or by uplift that traps sedimentation within a PBB (DeCelles & Giles, 1996). Thus, an abrupt change in the sediment-accumulation rate could be the result of either a change in the flux of sediment (Molnar, 2001; Zhang et al., 2001) and/or tectonic subsidence (Paola et al., 1992). The increased sedimentation rates observed at 16.3 and 13.5 Ma occur in the footwall of the KBT and south of the Tashipishake Anticline, both locations that would experience tectonic subsidence due to shortening across each structure (Fig. 10c). Furthermore, uplift above the Keketamu anticline likely reduced accommodation space, and thus sedimentation rate, while providing a proximal source of sediment that contributed to the increased rates above the Atushi Anticline 13 km south. Based on these observations, we interpret the changes in sedimentation rates at 16.3, 13.5 and 4 Ma to be tectonically influenced phenomena due to uplift above the KBT, Tashipishake anticline and Keketamu anticlines, respectively.

Structural control on conglomerate progradation

The initiation of rock uplift above specific structures, combined with shortening estimates and location of each structure within the basin, permits calculation of the migration rate of deformation into the foreland over time.

Fig. 18. (a) LandSat photo of the Kashi foreland that shows the spatial variation of basal ages from the Xivu conglomerate. Palaeocurrent directions for the conglomerate facies are indicated by the black arrows. The numbers within the arrows are the number of data points used for palaeocurrent average. Bidirectional white arrows are average palaeocurrent directions for the siltstone facies below the conglomerate. This figure illustrates how palaeocurrents change from structure parallel to transverse as the Xiyu conglomerate progrades southward over time. (b) Pre and post-deformation distances of the structures and base of conglomerate are indicated, as well as the inferred ages for the conglomerate facies and structure initiation. (c) Plot of migration rates of the deformation front and the gravel front over time. Distance from the Kashi Basin Thrust (KBT) (x-axis) are based on reconstructed, undeformed distances to each structure. Error bars (deformation front) or circles (Xiyu Conglomerate) define the uncertainty in the reconstruction (described in text).

Deformation has proceeded south from the KBT since ~ 16.3 Ma. By pinning our basin location at the KBT, and adding the shortening range from each structure to its present-day distance south of the KBT, we can plot the pre-deformation location of incipient structures against their time of initiation. The slope of the line between any two locations within the basin is the southward migration rate of the deformation front (Fig. 18c). Errors within the calculated rates are based on the range of shortening values interpreted for each structure (Fig. 10e). After

an initial uplift above the KBT at 16.3 Ma, deformation jumped south to the Tashipishake Anticline and structures in between at 13.5 Ma at a rate of 2.1–3.4 mm year⁻¹, but slowed after that time to 1.4–1.6 mm year⁻¹ as deformation propagated south to the Keketamu anticlines by 4 Ma. After ~4 Ma, deformation propagated south at > 10 mm year⁻¹ to the Atushi and Kashi Anticlines.

We suggest that localized uplift within the northern and medial foreland caused steady progradation of the gravel front towards the Boguzihe section until 1.9 Ma. Changes in gravel progradation rate can be influenced by a change in climate (Molnar, 2001), a decrease in tectonic subsidence (Heller *et al.*, 1988; Paola *et al.*, 1992), an increase in rock uplift rate of the gravel source (Burbank *et al.*, 1988) or a change in source area lithology (DeCelles *et al.*, 1988) Carroll *et al.*, 2006). A lack of stratigraphic formation boundaries or lithofacies associated with changes in sediment accumulation rate or in gravel progradation suggests that climate perturbations are unlikely to control the gravel progradation observed. Because the thrusts are southvergent, the gravel source might move south at a rate comparable with shortening across the fault. Therefore, in the absence of climate perturbations or changes in tectonic rates across the basin, the gravel progradation rate is likely to match shortening rates.

Deformation jumped south to the Keketamu anticline at \sim 4 Ma and to the Atushi and Kashi anticlines at 1.4 Ma. After \sim 4 Ma, uplift above the distal structures reduced subsidence in the area of deformation, driving the conglomerate front basinward at rates comparable with the rate of structure migration. The ~ 2.0 Myr discrepancy between deformation at Keketamu and the appearance of gravel 14.8 km south at Boguzihe are most likely the result of a lag time for prograding gravels to arrive at distal locations (Jordan et al., 1988; Jones et al., 2004). Observed southward progradation rates for the Xiyu conglomerate are generally slower than syn-tectonic gravels in the Siwalik foothills $(35-45 \text{ mm year}^{-1})$ in the southern Himalayan foreland (Burbank et al., 1988) or in the Sevier foreland $(20 \text{ mm year}^{-1})$ in western North America (Horton *et al.*, 2004). We suggest our slower rates may be due to lower sustained shortening rates across the Kashi foreland, or lower stream power due to a more arid climate, than for the other regions.

Good correlation between the structural propagation rates and gravel progradation rates suggests a causative link between the two in this foreland basin. Between 15.5 and \sim 4 Ma, the gravel prograded south at 2.5–3.5 mm year⁻¹, similar to the 1.4–3.4 mm year⁻¹ for structural migration into the basin (Fig. 18c). Furthermore, a sharp increase in the gravel progradation rate to $>10 \,\mathrm{mm\,year^{-1}}$ is observed after 2 Ma, and correlates with the same increase in structural migration rate after 4 Ma. The proximity of the gravel front to the deformation front, however, appears inversely related to the activity of specific structures. During times of more rapid shortening between 16.3 and 13.5 Ma, and after 2 Ma, the gravel front was located within a few kilometres of the active deformation front (Fig. 18c). In contrast, the gravel front migrated to over 20 km ahead of the deformation front between 13.5 and 4 Ma, when shortening rates were slower (Fig. 18c). Thus, higher shortening on specific structures at the deformation front are associated with more proximal gravel fronts, and low rates or tectonic quiescence may cause the gravel front to migrate away from the deformation front within a foreland. Whereas a climate change between 2 and 4 Ma might have accelerated prevailing rates of gravel progradation (Molnar, 2001; Zhang et al., 2001), it is unlikely that it would also affect the rate at which deformation propagated across the basin. Hence, we view climate as a secondary influence on gravel progradation, and uplift above specific structures increased local slopes such that gravel was pushed farther into the foreland.

Basin evolution

A three-fold increase in sedimentation rate is observed at 16.3 Ma at the Ayakeqiana and Qigailike sections within 1 and 6 km, respectively, south of the KBT (Figs 2a and 16). This age overlaps with the 18.9 ± 3.3 Ma apatite fissiontrack cooling from the KBT hanging wall (Sobel et al., 2006). Furthermore, the coarse-grained, debris-flow facies Xiyu conglomerate appears at Ayakeqiana at 15.5 Ma. Together, these three ages suggest that uplift of the KBT at ~ 16.3 Ma induced subsidence in the region of Ayakeqiana at that time. By 15.5 Ma, sufficient relief was produced in the hanging wall to shed debris flows into the basin that produced the proximal, coarse-grained Xiyu conglomerate. Before 16.3 Ma, deposition of fluvio-lacustrine Wuqia Group strata resulted from deposition 20-30 km south of the Maidan and Muziduke faults that were active by 20-25 Ma (Sobel et al., 2006).

These new chronologic data indicate that the overall pattern of deformation and the basin depocentre have migrated southward within the basin since at least 16.3 Ma (Fig. 18a). This age is 4–5 Myr earlier than the bedrockinvolved thrusting that defines initial growth of the Kyrgyz Range on the north side of the Tian Shan (Bullen et al., 2001, 2003), but is consistent with other estimates for initial deformation in the southern Tian Shan (Hendrix et al., 1994; Sobel & Dumitru, 1997; Yin et al., 1998; Sobel & Strecker, 2003). Moreover, near the study area, deformation likely began earlier along the Maidan and Muziduke faults (Fig. 2b) to the north (Sobel et al., 2006). Abrupt changes in sediment-accumulation rates at 16.3 and 13.5 Ma are interpreted as responses to tectonic loading due to thrusting above the KBT and Tashipishake Anticline. Steady sediment-accumulation and gravel-progradation rates suggest modest but continual thrusting between the KBT and Tashipishake Anticline, but no major southward steps of deformation until 4.0 Ma when the Keketamu anticline began to deform. After $\sim 2 \text{ Myr}$ of deformation focused on the Keketamu fold, faultrelated folding stepped south again to the Atushi and Kashi Anticlines in the distal basin (Fig. 18c).

CONCLUSIONS

Although an overall basinward progression of deformation characterizes most forelands, the specific timing of initiation of individual thrust faults and folds is commonly unknown. New chronologic data from the Tian Shan foreland yield one of the densest magnetostragraphic studies of an evolving terrestrial foreland basin system. These data provide a detailed temporal context within which to quantify the history of sedimentation and deformation of the Kashi Basin. New magnetostratigraphies from 11 measured sections define a continuous 18 Myr record of deposition, where rapid lateral and vertical facies changes occur due to pulsed episodes of southward-migrating deformation into the foreland. Using the extensive magnetostratigraphic data in the Kashi foreland, we interpret the timing of deformation based on accelerated accumulation rates in the footwalls of thrust faults and both decelerating accumulation rates and growth strata in their hanging walls.

Data from the Kashi Basin show that the Xiyu conglomerate is highly time-transgressive and spans from at least 15.5 Ma to the present. This finding contrasts with previous studies that assume the Xiyu conglomerate is either Plio-Pleistocene in age or represents a chronostratigraphic unit. Distribution of the Xiyu conglomerate is controlled by river drainage patterns over millions of years, by the proximity of conglomerate source area and by the growth of new folds that promote sedimentary bypassing. Along with bypassing, reworking of the Xiyu conglomerate from hinterland uplifts has pushed the gravel front out into the basin, whereas these same uplifts have ponded separate PBB strata behind their uplifted hanging walls. Because the age and migration rate of a single conglomerate lithofacies have rarely been so extensively constrained within a single foreland, the punctuated but sustained and long-term conglomeratic progradation seen here may serve as a catalyst for careful assessment of similar lithofacies in other basins. Rather than a steady migration of the deformation front, our data define a tectonic history that is punctuated by southward jumps in the locus of fault-related folding. Gravel progradation rates are similar to structural shortening rates, suggesting a causal linkage. The distal limit of gravel progradation lies within a few kilometres of the structural deformation front during times of active, relatively rapid shortening, but migrates basinward, away from the locus of deformation during times of little or no tectonic activity.

Our results indicate that the overall stratigraphic geometry was controlled by a long-term (>15 Myr) wave of deformation that swept across the foreland and modulated patterns of facies migration. Thus, within the Kashi foreland, tectonics is the fundamental control on conglomerate distribution, although climate likely plays a role at shorter time scales. Such findings provide an explanation for the variable timing of conglomerate deposition around the Tian Shan, illustrate the relationship of facies patterns to spatial and temporal changes in rates of subsidence and provide a well-constrained conceptual model for interpreting upward-coarsening foreland sequences from other terrestrial foreland basins.

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APPENDIX A

Palaeomagnetic analysis

Specimens from the northern and medial foreland, as well as from the Atushi Anticline and Kashi Town section, were analysed at the California Institute of Technology and Occidental College palaeomagnetic laboratories, both located in Los Angeles, CA. Both laboratories have identical sample changer and magnetometer set-ups. Remanent magnetization was measured using a three-axis DCSQUID moment magnetometer system housed in a magnetically shielded µ-metal room. The magnetometer has a background noise of $< 1 \text{ pA m}^2$ and is equipped with a vacuum pick-and-put, computer-controlled sample handling system, which can measure up to 180 samples automatically (Ward et al., 2005). Alternating field (AF) demagnetization was performed with a computer-controlled, three-axis coil system. Thermal demagnetization was performed in a commercially built, magnetically shielded furnace.

The Kashi West and the Upper Boguzihe specimens were measured at the Laboratory for Palaeo- and Rock Magnetism, GeoForschungsZentrum Potsdam using a fully automated 2G Enterprises 755 SRM DC-SQUID cryogenic long-core magnetometer with a background noise level of \sim 300–400 pA m². Despite higher background noise levels than the California labs, the samples processed in Potsdam yielded clear remanence directions.

The intensity of the natural remanent magnetism (NRM) for these specimens was $\sim 10^{-5}$ – 10^{-6} emu cm⁻³. Progressive demagnetization successfully resolved multiple components of magnetization. For most specimens, equal-area projection and orthogonal vector plots reveal two clear components: a high-temperature component (HTC) and a low-temperature component (LTC) (Fig. 11). The LTC, which is a low-coercivity, low-blocking tem-

perature component, is removed by AF demagnetization at 30 or 60 G and thermal demagnetization at 150 or 300 °C (e.g. Fig. 11b). This LTC has an *in situ* declination and inclination similar to the present day magnetic field direction of 3.56° E inclined 58.7° (Fig. Ali and j), although significant scatter observed in the LTC directions implies a more complex overprint than simply the present-day field. Typically, the LTC does not decay towards the origin (Fig. 11a–f). The HTC was completely unblocked by 680 °C (Fig.11a, b and f), indicating a hematite carrier of the magnetic remanance, although magnetite is likely present as well. At times the HTC would cluster in either a normal

Fig. A1. Stereonet plots of palaeomagnetic data. Grey (inclined down) and white (inclined up) circles indicate the $\alpha 95$ error around the Fisher mean. Fisher mean data are listed adjacent to each plot. The grey star is the location of the present-day magnetic field (decl: 3.56°, incl: 58.7°) for 39.6°N 75.9°E. (a-f) Geographic and tiltcorrected least-squares fits for the Keketamu composite (KC), Qigailike (L) and Ayakeqiana (Y) sections. (g, h) Fold test (McFadden & Jones, 1981) data showing the pre- and post-tilt correction groupings for the normal and reversed zones for KC, L and Y. (i, j) Nontilt-corrected secondary remanence directions for the KC and the combined Yand L. These secondary overprint directions are similar to the present-day field directions, and represent the viscous overprint of the deformed strata.

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(Fig. 1lc) or reversed position before decaying to the origin.

Coercivity spectrum analysis of isothermal remanent magnetism (IRM) for typical samples indicates that most IRM for the samples is acquired at low magnetizing field strengths (<2000 Oe), consistent with magnetite and titanomagnetite as the primary carrier for the magnetism (Fig lle and f). However, IRM continues to be acquired at higher field strengths, consistent with a hematite contribution to the characteristic remanence (ChRM) as well. For example, the rapid acquisition of most IRM (>50%)by 1000 Oe in sample K188.1 (Fig. 11h) is consistent with primarily magnetite as the carrier. In contrast, sample K70.3 (Fig. 1le) has a much flatter curve and a smaller percentage (<50%) of their IRM is acquired by 1000Oe, consistent with observed complete unblocking temperature of 680 °C and a hematite and magnetite HTC carrier for these samples (Fig. 11d and f).

To determine ChRM directions, principal component analysis (Kirschvink, 1980) using least-squares fits was performed for each specimen based on data from a minimum of four, and more typically five to eight, temperature steps. The data were analysed using palaeomagnetic software of Jones (2002) and N. R. Nowaczyk (pers. comm., 2004). The origin was included when stepwise demagnetization showed a progression towards it (Fig. 11c and d). Typical maximum angular deviation (MAD) was between 1 and 10°, but a MAD of 15° was accepted, if an unambiguous north or south polarity could be interpreted. When HTC components clustered in a normal or reversed polarity position, the ChRM direction was determined by forcing a line from the cluster through the origin (Fig. 11c).

Analytical tests of the palaeomagnetic data

The reversal test (McFadden and McElhinny, 1990) was used to assess whether the ChRM had likely been isolated for each site. Both the Ayakeqiana (Fig. Alf) and Qigailike (Fig. Ald) sections pass this test at C quality, but the Keketamu composite (comprising data from the Reservoir, Reservoir Bypass, ASE, and ASW) section does not (Fig. Alb). The failing test results from variation of 17 and 13° for both inclination and declination, respectively (Fig. Alb). Similar contrasts in inclination were observed at the Ayakeqiana (Fig. Alf), Qigailike (Fig. Ald), Kashi West and Kashi Town sections (Chen et al., 2007), and likely were the cause of the C-quality tests, rather than A or B quality. This difference is most likely due to an unremoved overprint (McElhinny, 1964; Quidelleur & Courtillot, 1996; Charreau et al., 2005) that, given the folding geometry and present magnetic field direction, can steepen or flatten the normal polarity directions with respect to the reversed polarity directions. Given that the normal and reversed polarity directions all plot within their respective hemispheres, that the sample populations are qualitatively antipodal and acknowledging that some overprint component has not been removed, we accept our polarity determinations despite the failure of the reversal test for the Reservoir section.

In the fold test, we utilize the multiple-limb technique from McFadden & Jones (1981) and compare the normal and reversed polarity zones from three separate fold limbs: north limb of the Tashipishake Anticline, north and south limbs of the Keketamu anticline (Fig. 2a). We kept the normal and reversed zones separate to minimize the effect of the unremoved overprint on the outcome of the test. Strata that span approximately the same time, the base of the Reservoir section (400–1400 m stratigraphic height), the entire Ayakeqiana section, and Qigailike sections (Fig. Alg-h), pass the fold test at 95% level of confidence with an observed F statistic of 2.37 and 2.28 for normal and reversed polarities, respectively ($f_{crit} = 2.38$, N = 382). Furthermore, samples from the same stratigraphic level (bed) always had consistent polarity directions, despite their collection <5m from each other, indicating the magnetozones were layer bound. The combination of the positive results from the fold and reversal tests, as well as the presence of layer-bound magnetic polarity zones indicate that these ChRM directions were acquired by the sediments at or soon after deposition. The expected declination and inclination for normal polarity samples from the late Miocene (7.9 Ma) is $D = 7.5^{\circ} I = 60^{\circ}$ (Enkin et al., 1992), similar to the present-day field of $D = 3.56^{\circ}$, $I = 58.7^{\circ}$.