1 Magnetostratigraphy of Cenozoic **deposits** in the western Qaidam Basin and its
2 implication for the surface uplift of the northeastern margin of the Tibetan Plateau

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13 **Abstract**

14 Thick Cenozoic deposits in the northwestern Qaidam Basin record erosion of the
15 Altyn Tagh and high terrain west of that basin and presumably the concurrent growth
16 of the northeastern Tibetan Plateau. A detailed magnetostratigraphic study of the
17 Huatugou section, northwestern Qaidam basin, reveals that this section spans the
18 period from ~30 to ~11 Ma. Magnetostratigraphic and sedimentological studies
19 indicate that the accumulation rate abruptly increased near ~15 Ma. The acceleration
20 in sedimentation rate suggests enhanced tectonic deformation in the Qaidam basin
21 since 15 Ma that may have begun simultaneously with accelerated deformation along

the Altyn Tagh, Kunlun, and Haiyuan faults, which contributed to the growth history of the Qaidam basin and its surroundings since ~15 Ma.

Keywords: Qaidam Basin; Altyn Tagh fault; tectonic uplift; magnetostratigraphy; Huatugou section; accumulation rate

1. Introduction

Two end-member views of how Tibet has grown outward have prevailed over the past decades: (1) continual outward, northward and northeastward growth, either steadily [e.g. England and Houseman, 1986] or in steps [e.g., Tapponnier et al., 2001] or (2) widespread deformation over virtually the whole of Tibet beginning at or shortly after the time of collision (Yin et al., 2008; Clark et al., 2010, Duvall et al., 2011). In fact, evidence for both seems to exist, with widespread deformation at or just after the time of collision [Wang et al., 2008; Yuan et al., 2013; Staisch et al., 2014], but a later acceleration of deformation on the northern margin and continued expansion of the dimensions of Tibet. When that later phase of deformation began, however, remains unresolved, with evidence from northeastern Tibet suggesting a date near 8-10 Ma [e.g., Métivier et al., 1998; Jolivet et al., 2001; Wang et al., 2003; Zheng et al., 2006, 2010; Lease et al., 2007; Yuan et al., 2011], but farther west along the Kunlun, cooling ages suggest rapid exhumation earlier, beginning near 15 Ma [Duvall et al., 2013]. The Qaidam basin, which has undergone relatively mild Cenozoic deformation, lies between these regions and offers insights into **the onset of** deformation.

The Qaidam basin marks a transitional region where elevations decrease

stepwise toward the north and northeast from approximately 5 km over Tibet to 3 km over the basin, and finally to approximately 1-1.5 km in Tarim Basin north of the Altyn Tagh and in the Hexi corridor in the foreland of the Qilian Shan (Figure 1a). Similar features, with vast closed sedimentary sub-basins that form relatively flat surfaces surrounded by mountain ranges in the Qaidam basin and higher Tibetan Plateau, suggest that understanding the outward growth of deformation since Eocene time in the Qaidam basin may allow us to unravel how the current morphology of the northern margin of Tibet has developed (e.g., Métivier et al., 1998; Zhou et al., 2006). The thrust faulting and crustal shortening in the northwest part of the basin has formed narrow ranges. The thick Cenozoic sedimentary sequence in the Qaidam basin should record the growth of structures within the basin, as well as the timing of growth of the surrounding mountains (Métivier et al., 1998; Wang et al., 2010; Cheng et al., 2014).

We present new magnetostratigraphy of Cenozoic sediments in the northwestern part of the basin, and then using a compilation of published data, we examine the timing of tectonic deformation of the basin.

2. Regional Geology

The triangle-shaped Qaidam basin is characterized by high average elevation and low relief that are similar to morphological features on the higher Tibetan Plateau. It is the largest topographic depression inside the Tibetan Plateau, and it covers an area of 120,000 km² and lies at an average elevation of 3000 m. Morphologically the Qaidam Basin is bounded by the Altyn Tagh to the northwest, the Qilian Shan to the

66 northeast, and the Kunlun to the south (Figure 1b). Tectonically, it is bounded by the
67 Altyn Tagh Fault to the northwest, the North Qaidam thrust system to the northeast
68 and several thrusts related to Kunlun fault to the south (Bally, 1986; Song and Wang,
69 1993; Jolivet et al., 2003; Yin et al., 2007; Cheng et al., 2014).

70 Cenozoic stratigraphic units in the Qaidam basin were deposited mainly in a
71 fluvial-lacustrine environment. Even if the same units were assigned different ages
72 (Table 1), **there** is consensus that the depocenters of the Qaidam basin have
73 consistently shifted eastward along the axis of the basin since the Eocene (Song and
74 Xie, 1993; Métivier et al., 1998; Wang et al., 2006; Yin et al., 2007). Analysis of
75 sandstone and mudstone provenance for Cenozoic basin fill suggests that the
76 surrounding mountains are the sources of the basin sediment, especially the Altyn
77 Tagh (Rieser et al., 2005).

78 **3. Stratigraphy and sampling**

79 We studied the Huatugou Section (from 38°25.6'N, 90°53.8'E to 38°22.0'N,
80 90°52.9'E), which lies at the southwest flank of the Youshashan anticline (Figure 1c).
81 Here the thickness of the late Oligocene to late Miocene sediment, which includes the
82 Shangganchaigou, Xiayoushashan, and Shangyoushashan Formations, is >4360 m
83 (Song and Wang, 1993; Xia et al., 2001; Rieser et al., 2005; Gao et al., 2009; Cheng et
84 al., 2014).

85 The Shangganchaigou Formation, 895-m thick, consists of gray, well-sorted fine
86 sand, siltstone, and pisolitic micritic carbonates intercalated with poorly graded
87 clast-supported conglomerate and gypsum layers, most of which was deposited in a

88 fluvial or marginal lacustrine environment. The presence of Ostracoda assemblages
89 such as *Mediocypris*, *Mediocypris Candonaeformis*, *Cyprinotus* and *Hemicyprinotus*
90 sp. in the adjacent area of the northwest Qaidam basin suggests an Oligocene age
91 (Sun et al., 1999; Yang et al., 2000; Sun et al., 2005). Two suspected fish fossils were
92 found at levels of 448.8 m and 762 m, respectively, but we could not identify the
93 species or ages of them. Nevertheless, two layers with fish discovered elsewhere in
94 the Shangganchaigou Formation (Wang et al., 2007) are also consistent with an
95 Oligocene age.

96 The Xiayoushashan Formation, 2803-m thick, is characterized by well-sorted
97 gray to red-brown silty clay, siltstone, and fine-grained sandstone intercalated with
98 sparse conglomerate layers. Its lower part contains numerous thin marl layers and
99 several layers with abundant fossils including fish pieces and leaves. The upper part
100 of this Formation (2712 – 3698 m) exhibits an upward coarsening succession, with
101 average grain size coarsening from siltstone to sandstone. Other studies have
102 interpreted the Xiayoushashan Formation to record deposition in an upwardly
103 shallowing lacustrine environment (Ritts et al., 1999; Hanson, 2001; Rieser et al.,
104 2005). The brackish-water species *Qaidamucythere ninae* occurs in a nearby section,
105 suggesting that it can be assigned to early-middle Miocene time (Sun et al., 1999;
106 Yang et al., 2000). The Xiayoushashan Formation conformably overlies the
107 Shangganchaigou Formation, and conformably lies beneath the Shangyoushashan
108 Formation; there are no obvious hiatuses in the stratigraphic section.

109 The Shangyoushashan Formation, >662-m thick, is characterized by gray to

brown, poorly sorted, clast-supported conglomerate intercalated with sparse sandstone and siltstone, which shows a **general** upward coarsening sequence. In some places, thicker sandstone intercalated with conglomerate contains large-scale planar cross-stratification. This formation is interpreted to record deposition in a fluvial environment. In this formation, there was a sudden appearance of *Cyprideis* sp., with occasional intercalated layers preserving *Limnocythere* sp. and *Eucypris obtusa* F. Yang and Sun (Sun et al., 1999; Yang et al., 2000).

We collected **782** block samples from the Huatugou section (depth span from 5.5 m to 4216 m, but with no samples between 4216 m and 4360 m because of the coarse-grained lithology) from the claystone, siltstone, and fine-grained sandstone layers. The sampling intervals commonly varied between 3 and 9 m, depending primarily on the lithology. The block samples were oriented by magnetic compass in the field. For paleomagnetic analysis, cubic specimens of 2.0 cm × 2.0 cm × 2.0 cm were obtained from block samples in the laboratory.

4. Methodology

All of the oriented specimens were subjected to stepwise thermal demagnetization using a MMTD-60 thermal demagnetizer, and remanence was measured using a 2G Enterprises mModel 755 Cryogenic magnetometer installed in a magnetically shielded space (< 300 nT) at the Institute of Earth Environment, Chinese Academy of Sciences. Most samples were heated to a maximum temperature of 680 °C with 20-50 °C temperature increments.

Measurements of rock magnetism, including temperature-dependent magnetic

susceptibility (χ -T) curves, and isothermal remanent magnetization (IRM) acquisitions were conducted on selected samples. The χ -T curves were determined in an argon atmosphere under a frequency of 976 Hz from room temperature up to 700 °C and back to room temperature using a Kappabridge magnetic susceptibility meter (Model MFK1-FA) equipped with a CS-3 high-temperature furnace (AGICO Ltd, Brno, Czech Republic). IRM acquisitions were measured using an ASC IM-10-30 pulse magnetizer up to saturation IRM (SIRM) at a maximum applied field of 2500 mT and AGICO JR-6 for remanence measurements.

5. Results

5.1 Magnetic mineralogy

The dependence of magnetic susceptibility on temperature (χ -T) is sensitive to mineralogical changes in samples during thermal treatment, and such changes can provide primary indicators about magnetic mineral composition (Dunlop and Özdemir, 1997; Deng et al., 2004). Figure 2 illustrates the χ -T curves of samples from the Huatugou section. Every χ -T heating curve shows a distinct decrease near 580 °C, with a higher magnetic susceptibility before 400 °C than that after 580 °C, which implies that magnetite is the primary contributor to the magnetic susceptibility in the sediment (Figure 2). Figure 2a shows a significant increase above ~400 °C, and a peak at ~510 °C. The absence of this feature in the cooling curve suggests that it is not the Hopkinson peak, but may be the **result of** neoformation at moderately high temperature (Deng et al., 2001). The fact that all magnetic susceptibilities increase

during the successive cooling after heating to 700 °C also indicates that neoformation of fine-grained magnetite developed via annealing of iron-containing paramagnetic minerals (Deng et al., 2001; Ao et al., 2010). Many samples also show a gradual decrease of susceptibility around 680 °C (Figure 2b, c, d, e, f), which implies that hematite is in some of the sample group. Because the bulk susceptibility of magnetite is ~1000 times greater than most rock materials (Collinson, 1983), this thermomagnetic behavior suggests that magnetite is one of the dominant magnetic minerals.

All of the IRM acquisition curves indicate that these samples are not saturated at 2.5 T (Figure 3), which is consistent with a significant hematite contribution. Progressive removal of the SIRM by applying reversed fields shows that remanent coercivities of the SIRM are approximately 50 and 220 mT, indicating that highly coercive magnetic minerals, such as hematite, are present in the sediment. This result coincides with rock magnetic characteristics from similar sedimentary sequences in the Tarim Basin (Chang et al., 2012a) and the Suerkuli basin in the Altyn Tagh (Dupont-Nivet and Butler, 2003; Chang et al., 2012b).

5.2 Magnetostratigraphy

Progressive demagnetization successfully isolated characteristic remanent magnetization (ChRM) components for most of the samples after removing a viscous component of magnetization. We used Orthogonal diagrams to evaluate demagnetization results (Zijderveld, 1967), principal component analysis to determine ChRM directions (Kirschvink, 1980), and Fisher statistics to calculate site and group

mean directions (Fisher, 1953). A low temperature component (LTC) was separated from temperature around NRM – 250 °C. High stability of ChRM was separated between 400 and 680 °C for most samples, consistent with magnetite and hematite being the dominant remanence carriers (Figure 4). A high-stability ChRM component extends to a maximum temperature of 580 °C for some samples revealing magnetite to be the dominant remanence carrier.

Specimens not included in our magnetostratigraphic analysis were eliminated based on three criteria. (1) ChRM directions could not be ascertained because of ambiguous or noisy orthogonal demagnetization diagrams. (2) Obtained ChRM directions have maximum angular deviation below 16° (Van der Voo, 1990). (3) Specimens displayed magnetizations with virtual geomagnetic pole (VGP) latitude values less than 30°. Finally, a total of 642 samples (82%) gave ChRM directions. After tilt corrections, these ChRM directions plot with antipodal normal and reversed orientations, suggesting a primary record of the Earth's magnetic field (Figure 5a). A statistical bootstrap technique has been used to test whether the distributions of ChRM vectors are possibly non-Fisherian, and to characterize the associated uncertainties for both normal and reversed ChRM directions (Figure 5b) (Tauxe, 1998). In the plots, the reversed polarities have been flipped to their antipodes. Because the 95% confidence bounds from the two sets of data overlap in all three components, the means of the reversed and normal modes cannot be distinguished at the 95% level of confidence; thus, they pass the bootstrap reversal test (McFadden and McElhinny, 1990; Tauxe, 1998). Because normal and reversed directions are all within 45° of the

mean normal and reversed directions respectively, they are suitable for defining magnetostratigraphic polarity zones. 479 representative site-mean ChRMs from different formations of the section were employed for a calculation by the method of McElhinny (1964), which shows a positive fold test ($k_s / k_g = 1.26 > F(2*(n_2-1), S*(n_1-1)) = 1.14$) at the 95% confidence level (Figure 5c). Fold tests reveal the improved grouping after tilt correction.

A Jackknife technique (Tauxe and Gallet, 1991) was used to quantify the reliability of the magnetostratigraphy of the section. The jackknife parameter (J) calculated for the accepted specimen-mean directions has a value of **-0.2301** (Figure 6), which falls within the range of 0 to -0.5 recommended by Tauxe and Gallet (1991) for a robust magnetostratigraphic data set, indicating that sampling of the section has recovered more than 95% of known polarity intervals.

At least four successive ChRM directions were required to define polarity zones. Results suggest that the Huatugou section recorded 33 normal (N1-N33) and 34 reversed (R1-R34) polarity zones. Although relatively high accumulation rates occurred in northwest Qaidam basin since late Eocene (Sun et al., 2005; Zhou et al., 2006; Yin et al., 2008; Gao et al., 2009; Cheng et al., 2014), angular or erosional unconformities are absent, and no major hiatus can be recognized in the Huatugou section. Nevertheless, we cannot correlate the recorded polarity zones unambiguously with the Geomagnetic Polarity Time Scale. Accordingly, we offer two correlations of recorded polarity zones to the Geological Time Scales 2012 (GTS2012) (Gradstein et al., 2012) in Figure 7.

Both correlations share the same assignments for the top 1600 m, from ~11.5 back to ~21 Ma. At the top, the most distinctive pattern includes two long reversed zones (R1 and R3) separated by two shorter normal intervals containing one reversed zone (N1-N2). Consistent with late Miocene fossils in the Shangyoushashan Formation in the northwest Qaidam basin, the only reasonable fit to this polarity pattern is provided by the C5r.1r – C55.2r of the GTS2012 (Figure 7). Below, two long normal intervals (N3-N4) interrupted by a short reversed zone (R4) correlates to the C5An.1n - C5An.1r - C5An.2n sequence of the GTS2012. Additional support for this correlation comes from a similar pattern of long reversals recorded in the stratigraphic equivalent of the Shangyoushashan Formation in nearby Xichagou section and in the eastern basin Dahonggou section (Gao et al., 2009; Lu and Xiong, 2009).

Below this reversal sequence, three normal intervals (two short followed by one relative long) (N5-N7) provide a reasonable correlation with C5Ar.1n to C5AAn of the GTS2012. Further below, the three long normal zones (N8-N10) separated by two shorter reversed intervals appear to correlate to C5ABn to C5ADn. The pattern from R11 to R17 is tentatively correlated to C5ADr to C5Dr.1r in agreement with early to middle Miocene fossils reported elsewhere from the Xiayoushashan Formation (Sun et al., 1999; Yang et al., 2000), assuming that one short reversed interval is missing. This match is imperfect, because the relatively long normal interval of C5Cn.1n and reversed interval between C5Cn.3n and C5Dn must be correlated with the proportionally shorter N11 and R12 and because there are two extra reversals between

N11 and R14 as compared to the GTS.

The pattern in the lower part of the Xiayoushashan Formation is characterized by two relatively long normal intervals (N17 and N18) followed by two relatively short intervals separated by three reversals. A reasonable fit to this polarity pattern is provided by the C5En to C6An.2n.

Below this sequence, the two correlations differ. For correlation 1, the pattern in the lower-most part of the Xiayoushashan Formation is characterized by five short normal intervals (N21-N25) that separate long reversal zones (R22-R26), which can be tentatively correlated to C6An.2r to C6Bn.2r of the GTS2012.

The pattern in the upper Shangganchaigou Formation is characterized by two long (R28-R29) and several short reversal intervals between normal intervals (N28-N32). We tentatively correlate this pattern to C6Cn.1n to C8n.2n, in agreement with fossils reported from the Shangganchaigou Formation near the section in the northwest and eastern Qaidam basin (Sun et al., 1999; Yang et al., 2000; Sun et al., 2005; Wang et al., 2007). The pattern in the lower-most part of the section, characterized by one normal interval (N33) between two reversed intervals (R33 and R34), can be correlated to C8n.2n to C9r. This includes a correlation of the thickest recorded normal polarity zone N32 with C9n.

On the basis of this correlation, the age of the top of the Shangganchaigou Formation is determined to be ~23 Ma. **Sun et al. (2005) assigned Shangganchaigou Formation to 35.5 – 26.5 Ma. The difference in the age of the top of the formation between our analysis and that of Sun et al. (2005) likely reflects a time**

transgressive change in deposition (from deposition to erosion) in the late Oligocene due to uplift of the Altyn Tagh. This inference requires that deposition of the Shangganchaigou Formation not be fully represented in the Hongshan section. Although the sampling resolution of the lower part of the Xiayoushashan Formation is not sufficient to detect a few chrons of short duration, two long reversal intervals (R11-R12 and R21-R23) appear to correlate with C5ADr - C5Bn.2r and C6An.2r – C6Aar.2r. At the top of the Xiayoushashan Formation, the representative pattern from R5-N11 is reasonably correlated to C5Ar.1r to C5ADn. If this analysis for Shangyoushashan Formation is correct, the age of its top reaches 12.44 Ma. These age spans of the Shangganchaigou and Xiayoushashan Formations from our magnetostratigraphy agree with results from basin-wide stratigraphic correlations via outcrop and seismic reflection profiles, fossil assemblages, and magnetostratigraphy elsewhere in the Qaidam basin (Sun et al., 1999; Yang et al., 2000; Xia et al., 2001; Sun et al., 2005; Yin et al., 2008; Gao et al., 2009; Lu and Xiong, 2009).

Because of changes in accumulation rates, the recorded pattern of polarity zones differs from the pattern expected from the GTS2012, and the correlation between them is not unique. We offer another possible correlation (correlation 2) in Figure 7, for sediment older than R22. The pattern in lower Xiayoushashan Formation is characterized by two short normal intervals in long reversal (R21-R23). We tentatively correlate this pattern to C6An.2r to C7n.1r, assuming that several short normal intervals are missing. The pattern in the lowermost Xiayoushan formation and the upper part of the Shangganchaigou Formation is characterized by normal

polarities interrupted by several reversals (N23-N27). We correlate this part to C7n.2n to C8n.2n. We correlate the pattern in the lower part of the Shangganchaigou Formation, from N28-N33, to C9n to C12n of GTS2012. According to the correlation 2, the bottom age of the Xiayoushashan Formation would be about 25 Ma.

According to the Jackknife test, the recorded polarities zones should cover 95% of **known** polarities zones. The ratios of the recorded to **known** in correlation 1 and 2 are 92% and 81%, respectively, making them both acceptable based on the data statistics. Both correlations are imperfect. Correlation 2 requires that several continuous polarity chrons (C6AAn - C6AAr.1n - C6AAr.2n and C6Cn.1n - C6Cn.2n - C6Cn.3n) have not been recorded, which contradicts the geologic inference that there are no obviously hiatuses, depositional changes, or angular unconformities during the part of the Huatugou section. Although the statistical evidence above suggests that correlation 1 is better than correlation 2, correlation 1 shows two more magnetic reversals between chrons R29 and N32 than given by the GPTS. Based on the imperfections in both correlations, we present both of them as viable.

Magnetostratigraphy in the Huatugou section is different to that in the Hongshanhan section collected by Sun et al. (2005). Deposits of the Hongshanhan section are composed of the Xiaganchai and Shangganchaigou Formations. The Hongshanhan section is closer to the Altyn Tagh fault than the Huatugou section. Based on correlation 1, the age of the top of the Shangganchaigou Formation is determined to be ~23 Ma. Although the sampling resolution of the lower part of the Xiayoushashan Formation is not sufficient to detect the duration of several small

chrons, two long reversal intervals (R11-R12 and R21-R23) appear to correlate with C5ADr - C5Bn.2r and C6An.2r – C6Aar.2r. At the top of the Xiayoushashan Formation, the representative pattern from R5-N11 is reasonably correlated to C5Ar.1r to C5ADn. If this analysis for Shangyoushashan Formation is correct, the age of its top reaches 12.44 Ma. These age spans of the Shangganchaogou and Xiayoushashan Formations from our magnetostratigraphy agree with results from basin-wide stratigraphic correlations via outcrop and seismic reflection profiles, fossil assemblages, and magnetostratigraphy elsewhere in the Qaidam basin (Sun et al., 1999; Yang et al., 2000; Xia et al., 2001; Sun et al., 2005; Yin et al., 2008; Gao et al., 2009; Lu and Xiong, 2009).

6. Discussion

6.1. Oligocene-early Miocene sedimentation rate change in the northeastern Tibetan Plateau

The striking outcome of this study is that high average accumulation rates (252 m/Ma in correlation 1 and 205 m/Ma in correlation 2 characterize early Oligocene to late Miocene deposition in the northwest Qaidam basin. These accumulation rates are notably higher than rates for the same time window on the margin of the Tibetan Plateau: ~40 m/Ma between 31 and 15 Ma in Xianshuihe section in the Lanzhou Basin [Yue et al., 2001]; ~17 m/Ma between 29 and 13 Ma in the Maogou section in the Linxia basin [Fang et al., 2003]; ~93 m/Ma between 20.7 and 17.6 M, in the Ashigong section in the Guide basin [Fang et al., 2005]; ~34 m/Ma between 27.5 and

11.5 Ma in the Sunan section in the Jiuquan basin [Fang et al., 2013]; ~30 m/Ma between 33 and 17 Ma, in the Xiejia section in the Xining basin [Dai et al., 2005]; 80 m/Ma between 28.5 and 11.5 in the Sikouzi section in the Sikouzi basin [Wang et al., 2011]; 102 m/Ma between 28 and 11.5 Ma in the Hualong section in the Xunhua basin [Lease et al., 2012]; 28 m/Ma between 35.5 and 16 Ma in the Tashan section in the Xining basin [Xiao et al., 2012]. We interpret this result to imply that high, steep, rapidly eroding terrain provided a nearby source of sediment in the northwestern Qaidam basin during this time span.

According to the isopach study of Yin et al. (2008), in the northern part of the Qaidam basin, the largest sub-basin enlarged continuously during deposition of the Shangganchaigou and Xiayousahshan formations. That is consistent with the increase in sedimentation rate observed in the Huatugou section and in other parts of the basin.

6.2. Sedimentation rate increase since middle Miocene

The most obvious change in sedimentation rate occurred at the stratigraphic height of 2443 m, where the rate increased from about 180 m/Myr (in correlation 1) or 130 m/Myr (in correlation 2) between ~28 or ~31 and 15 Ma to about 500 m/Myr since ~15 Ma (Figure 8). This accumulation rate change also coincides with an increase of the coarse sediment fraction, as described above. Sediment accumulation rates within terrestrial basins depend both on climatic conditions that affect the transport of sediment and on regional relief, which results, in part, from tectonic movements. **Though compaction will change the sedimentation rate in a thick sedimentary sequence, it is not the main cause of the sedimentation rate increase.**

1) **Changes in compaction tend to cause more gradual apparent sedimentation rate decrease from the top to the bottom of a stratigraphic succession because of the increasing compaction. The pronounced change in sedimentation rate at ~15 Ma is inconsistent with compaction as the cause (Figure 8 and 9).** 2) **Similar sedimentation rates of different sections in the Qaidam basin also do not show gradually increasing sedimentation rates from bottom to top, except increases driven by tectonic processes (Sun et al., 2005; Fang et al., 2007; Lu and Xiong, 2009).** 3) **Perhaps compaction alone cannot lead to a 2- to 3-fold increase in sedimentation rate in the Qaidam Basin.** Climate change also does not appear to be a factor to account for the ~2-fold increase in accumulation rate since the middle Miocene because of the following two lines of evidence: (1) Regional climate does not show an obvious change during the early – middle Miocene in the Qaidam Basin (Figure 9) (Wang et al., 1999; Miao et al., 2011). Furthermore, although global climate changed during the early – middle Miocene (e.g., Flower and Kennett, 1993; Zachos et al., 2001), variability is not in phase with sediment-accumulation rates recorded in the Huatugou section. (2) The sediment in the Xiayoushashan Formation is lacustrine, and the Huatugou section does not show sharp lithofacies changes in the period from 23 to 13 Ma. Lithofacies should show a change if climate change was dramatic (Allen and Allen, 2005; Leeder, 2011). Therefore, we conclude that climatic change was not the predominant factor that led to a change in the sedimentation rate.

On the other hand, several observations call attention to tectonic changes that could have affected the sedimentary environment in the northwestern Qaidam basin.

(1) Although lithofacies do not obviously change in the Xiayoushashan Formation, a largely upwardly coarsening succession in its upper part suggests that the distance from sediment source to sediment trap may have decreased to allow the same amount of runoff to transport more and coarser sediment. (2) Growth index, which is defined as the ratio of the stratal thickness on the footwall to that on the hanging wall of growth faults can reflect growth rate or activity intensity of the growth faults (Song and Xie, 1993; Xiao and Hao, 2003). Growth index studies using seismic profiles suggest that the growth faults in the northwest Qaidam basin began to be active at about 15 Ma (Wang et al., 2012). (3) The southeastward migration of the depocenter in the Qaidam basin during early – middle Miocene is also consistent with deformation of the northwestern part of basin (Wang et al., 2006). (4) **Stable oxygen isotopes studies have been used to infer that the initiation of uplift of surrounding mountains of the Qaidam basin began since early-middle Miocene time (Kent-Corson et al., 2009; Li et al., in preparation). Leaf wax stable isotope results have been used to infer topographic growth during 15 – 10.4 Ma (Zhuang et al., 2014).** (5) Provenance research has suggested that the Altyn Tagh is the primary source of sediment in the northwestern part of the Qaidam basin (Rieser et al., 2005, 2006), and research on apatite fission track and (U-Th)/He thermochronology also indicate rapid exhumation and cooling since ~15 to 16 Ma in the northern Altyn Tagh (Ritts et al., 2008). An uplifted Altyn Tagh can feed more material to the basin because of enhanced erosion and energy of the runoff. (6) Yin et al (2002) suggested that the modern Altyn Tagh fault system was established during 20 – 10 Ma even though there

has been long-lived higher topography between the Tarim basin and the Qaidam basin since the early Oligocene or possibly late Eocene. (7) The initiation of pull-part basins in the western Kunlun ranges (Jolivet et al., 2003) and the exhumations of adjacent rock in the middle segment (Duvall et al., 2013) have occurred since middle Miocene. **Potassic adakitic volcanic rocks dated between 18-15 Ma (Wang et al., 2005) in the Tibetan Plateau, derived from lithospheric mantle sources (Qu et al., 2004),** also suggest tectonic activity in Kunlun Mountains in early to middle Miocene time. (8) The NE-SW compression reconstructed according to the anisotropy of magnetic susceptibility of Eocene and Miocene sediment has been intense in the western Qaidam basin since the Miocene (Yu et al., 2014).

All of these observations suggest enhanced tectonic activity around and within the Qaidam basin since middle Miocene time. Deformation of ranges surrounding the Qaidam basin can increase the elevation different between the basin and its source areas, thereby increasing the erosional flux to the basin. This process should increase the sedimentation rate in the basin if the climate change was not significant. We suggest that tectonics was the primary reason for the increase in sediment accumulation rate in the northwest Qaidam basin since ~ 15 Ma.

7. Conclusion

The evolution of the Qaidam basin is characterized by continuously high accumulation rates since Oligocene time, and especially since ~15 Ma. The increased accumulation rate near 15 Ma suggests that tectonic activity in the northwest Qaidam

basin may have accelerated at that time. This increased sedimentation rate is approximately synchronous with inferred deformation along the Altyn Tagh, Kunlun, and Haiyuan faults, and perhaps with deformation in the whole northeastern part of the Tibetan Plateau (e.g., Ritts et al., 2004, 2008; Duvall et al., 2013; Yuan et al., 2013).

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711 Figure 1a. Location of the Qaidam basin in the Tibetan Plateau (**based on Yin et al.,**
712 **2002**). b. Morphotectonic map of the Qaidam Basin and locations of the Huatugou
713 and other sections (**based on Yu et al., 2014**). c. and geological map in the
714 Youshashan range region (**based on our field work and Wang et al., 2012**). ATF:
715 Altyn Tagh fault, KLF: Kunlun Fault, NQT: North Qaidam thrust, QLT: Qilian thrust,
716 HSH: Hongsanhan section, DHG: Dahonggou section, XCG: Xichagou section,
717 HTTL: Huaitoutala section, HTG: Huatugou section, SKZ: Sikouzi section, XSH:
718 Xianshuihe section, MG: Maogou section, ASG: Ashigong section, HL: Hualong
719 section, XJ: Xiejia section, TS: Tashan section, SN: Sunan section. E₃S:
720 Shangganchaigou Formation, N₁¹X: Xiayoushashan Formation, N₁²S:
721 Shangyoushashan Formation, N₁³-N₂S: Shizigou Formation

Figure 2. Magnetic Susceptibility vs. Temperature (χ -T) curves from selected samples at various horizons of the Daban section. Heating and cooling cycles are indicated in red and blue respectively.

Figure 3. Isothermal remanent magnetization (IRM) acquisition curves for selected samples from the Huatugou section. Numbers at the top of the graphs show the sample number and its depth in meters.

Figure 4. Orthogonal projections of representative progressive thermal demagnetization for selected samples from the Huatugou section. The triangles and rectangles represent the vertical and horizontal planes respectively. NRM is the natural remanent magnetization.

Figure 5. (a) Equal-area projections of the Characteristic remanent magnetization (ChRM) directions, as measured on the left, and with corrections for tilting of strata on the right, and means directions (with oval of 95% confidence) for the Huatugou section. Downward (upward) directions are shown as squares and triangles. (b) Bootstrap reversal test diagram (Tauxe, 1998). Reversed polarity directions have been inverted to their antipodes to test for a common mean for the normal and reversed magnetization directions. The 95% confidence intervals for all components overlap, indicating a positive reversal test. (c) **Diagram showing a positive fold test of the 479 representative site-mean ChRM from different formations of the Huatugou section using the method of McElhinny (1964).**

Figure 6. Magnetostratigraphic jackknife analysis for the Huatugou section (Tauxe and Gallet, 1991). The obtained slopes J have values of **-0.2301** in the section, which predicts that obtained polarity zones has recovered more than 95% of the true number of polarity intervals.

Figure 7. Lithology and magnetostratigraphy of the Huatugou section in northwest Qaidam Basin, and correlations of the observed polarity zones of the Huatugou section with the geological time scale (Gradstein et al., 2012). Bold lines show identical correlations for Correlations 1 and 2, and fine lines and dashed lines show different correlations.

Figure 8. Stratigraphic level in meters versus age according to our magnetostratigraphic correlations to the GTS 2012 at the Huatugou section.

Figure 9. Comparison of accumulation rates of the Huatugou section calculated from the sediment thickness between limits of correlated chrons to $\delta^{18}\text{O}$ at **DSDP 588A (Flower & Kennett, 1993)**, average Xerophytes percentages in three boreholes (Wang et al., 1999) and KC-1 core and percentages of conifers from relatively humid environments (Miao et al., 2011) in northwest Qaidam.

Table 1 Cenozoic integrated stratigraphy of the Qaidam Basin.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

	Tilt uncorrected					Tilt corrected				
	Dec	Inc	N	K	a95	Dec	Inc	N	k	a95
(C1)	64	27.8	244	8.5	3.3	26.2	42.8	244	11.09	2.8
(c2)	241.6	-35.2	235	6.63	3.9	196.6	-47.1	235	8.54	3.3

Figure 6

Figure 7

Figure 7 displays a stratigraphic column and magnetic polarity data for the Shangyoushashan, Xiayoushashan, and Shangganchaigou Formations. The left side shows the lithology and depth (m) from 0 to 4500. The middle section shows VGP -latitude/° from -90 to 90. The right side shows correlations between the Huatugou section and the GTS2012 (Gradstein et al., 2012) polarity scale.

Depth/m

Lithology

VGP -latitude/°

Correlations 1

Correlations 2

GTS2012 (Gradstein et al., 2012)

Polarity of the Huatugou section

Polarity of the Huatugou section

Shangyoushashan Formation

Xiayoushashan Formation

Shangganchaigou Formation

Legend:

- carbonate
- clay
- clayed siltstone
- siltstone
- sandstone
- conglomerate
- Gypsum
- fishs fossil
- leaves fossil

Stratigraphic Column Details:

- Shangyoushashan Formation:** Depth 4000-4500m. Lithology: clay, clayed siltstone, siltstone, sandstone, conglomerate.
- Xiayoushashan Formation:** Depth 2500-4000m. Lithology: clay, clayed siltstone, siltstone, sandstone, conglomerate.
- Shangganchaigou Formation:** Depth 0-2500m. Lithology: clay, clayed siltstone, siltstone, sandstone, conglomerate, Gypsum, fishs fossil, leaves fossil.

Magnetic Polarity Data:

- VGP -latitude/°:** Ranges from -90 to 90. Data points are plotted as black diamonds.
- Correlations 1:** Shows correlations between the Huatugou section and the GTS2012 polarity scale. Correlation lines connect R1 to N1, R2 to N2, R3 to N3, R4 to N4, R5 to N5, R6 to N6, R7 to N7, R8 to N8, R9 to N9, R10 to N10, R11 to N11, R12 to N12, R13 to N13, R14 to N14, R15 to N15, R16 to N16, R17 to N17, R18 to N18, R19 to N19, R20 to N20, R21 to N21, R22 to N22, R23 to N23, R25 to N25, R27 to N27, R28 to N28, R29 to N29, R30 to N30, R32 to N32, R33 to N33, and R34 to N34.
- Correlations 2:** Shows correlations between the Huatugou section and the GTS2012 polarity scale. Correlation lines connect R1 to N1, R2 to N2, R3 to N3, R4 to N4, R5 to N5, R6 to N6, R7 to N7, R8 to N8, R9 to N9, R10 to N10, R11 to N11, R12 to N12, R13 to N13, R14 to N14, R15 to N15, R16 to N16, R17 to N17, R18 to N18, R19 to N19, R20 to N20, R21 to N21, R22 to N22, R23 to N23, R25 to N25, R27 to N27, R28 to N28, R29 to N29, R30 to N30, R32 to N32, R33 to N33, and R34 to N34.

GTS2012 (Gradstein et al., 2012) Polarity Scale:

- Normal (N):** N1, N2, N3, N4, N5, N6, N7, N8, N9, N10, N11, N12, N13, N14, N15, N16, N17, N18, N19, N20, N21, N22, N23, N25, N26, N27, N28, N29, N30, N31, N32, N33, N34.
- Reversed (R):** R1, R2, R3, R4, R5, R6, R7, R8, R9, R10, R11, R12, R13, R14, R15, R16, R17, R18, R19, R20, R21, R22, R23, R25, R27, R28, R29, R30, R32, R33, R34.

Figure 8

Figure 9

Table 1

Formation	Age	References
Qigequan	Pleistocene	Yang et al., 2000; Rieser et al., 2005; Fang et al., 2007; Yin et al., 2008; Zhuang et al., 2011, Wang et al., 2012; Zhang et al., 2014
	Late Pliocene - Pleistocene	Yang et al., 1992; Liu et al., 1998; Wang et al., 1999; Sun et al., 2006
Shizigou	Pliocene	Yang et al., 1992; Sun et al., 1999; Wang et al., 1999; Yin et al., 2008
	Late Miocene - Pliocene	Rieser et al., 2005; Fang et al., 2007; Gao et al., 2009; Zhuang et al., 2011; Wang et al., 2012; Zhang et al., 2014
Shangyoushashan	Late Miocene	Yang et al., 1992; Yin et al., 2008
	Middle - Late Miocene	Sun et al., 1999; Wang et al., 1999; Rieser et al., 2005; Fang et al., 2007; Lu and Xiong, 2009; Zhuang et al., 2011; Wang et al., 2012
Xiayoushashan	Early - Middle Miocene	Yang et al., 1992; Sun et al., 1999; Wang et al., 1999; Yin et al., 2002; Fang et al., 2007; Lu and Xiong, 2009; Zhuang et al., 2011; Wang et al., 2012
	Late Oligocene - Middle Miocene	Rieser et al., 2005
	Early Oligocene - Early Miocene	Lu and Xiong, 2009; Wang et al., 2012
Shangganchaigou	Late Eocene - Oligocene	Yang et al., 1992; Sun et al., 1999; Rieser et al., 2005; Sun et al., 2005; Yin et al., 2008; Zhuang et al., 2011
	Middle Eocene - Oligocene	Liu et al., 1996; Wang et al., 1999
	Paleocene - Middle Eocene	Liu et al., 1996; Wang et al., 1999
Xiaganchaigou	Early - late Eocene	Yang et al., 1992;Sun et al., 1999; Rieser et al., 2005; Yin et al., 2008
	Middle - late Eocene	Sun et al., 2005; Zhuang et al., 2011
Lulehe	Early - Middle Eocene	Rieser et al., 2005; Zhuang et al., 2011; Wang et al., 2012
	Paleocene - Early Eocene	Yang et al., 1992;Yin et al., 2008
	Cretaceous - Paleocene	Liu et al., 1996; Wang et al., 1999;