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Possible orographic and solar controls of Late Holocene centennial-scale moisture oscillations in the northeastern Tibetan Plateau

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[1] We present a record of lake-level changes from Hurleg Lake, a freshwater lake in the arid Qaidam Basin on the NE Tibetan Plateau, an area with few high-resolution paleoclimate records. The carbonate percentage and ostracode abundance show a consistent pattern with ~200-year moisture oscillations during the last 1000 years. The moisture pattern in the Qaidam Basin is in opposite relation to tree-ring-based monsoon precipitations in the surrounding mountains, suggesting that topography may be important in controlling regional moisture patterns as mediated by rising and subsiding air masses in this topographically-complex region. Cross-spectral analysis between our moisture proxies and solar activity proxy shows high coherence at the ~200-year periodicity which is similar to Chinese monsoon intensity records, implying the possible solar forcing of moisture oscillations in the NE Tibetan Plateau. **Citation:** Zhao, C., Z. Yu, Y. Zhao, and E. Ito (2009), Possible orographic and solar controls of Late Holocene centennial-scale moisture oscillations in the northeastern Tibetan Plateau, *Geophys. Res. Lett.*, *36*, L21705, doi:10.1029/2009GL040951.

1. Introduction

[2] Documenting the effect of large-scale forcing to regional climate is an important first step in understanding the underlying mechanisms of climate change. In NW China, *Chen et al.* [2008] reported opposite moisture conditions during the early Holocene between monsoonal China and arid central Asia. *Herzschuh* [2006] attributed this pattern to the topographically induced rising air on the Tibetan Plateau (TP) and sinking air north of it. The centennial-scale climate oscillations have been reported from ice cores, tree rings and lake sediments in the monsoonal areas [e.g., *Thompson et al.*, 2003; *Sheppard et al.*, 2004; *Ji et al.*, 2005; *Shao et al.*, 2005] and attributed to solar variations [*Ji et al.*, 2005; *Shao et al.*, 2005]. However, the regional responses to large-scale climate forcings are still poorly understood in the topographically complex NE TP. It is not clear if the topographic forcing applies to smaller areas and at shorter time scales. Also, possible solar-driven centennial-scale climate oscillations have not been investigated beyond the present-day monsoon limit.

[3] In this paper, we present a lake-level record from Hurleg Lake located in the Qaidam Basin on the NE TP. The objectives of this study are to document the late Holocene moisture history of the arid NW China, to compare the moisture pattern at different elevations, and to infer potential climate forcings.

2. Study Site and Methods

[4] Freshwater Hurleg Lake is located in the NE Qaidam Basin on the NE Tibetan Plateau, surrounded mostly by Quaternary lacustrine and alluvial deposits [*Yi et al.*, 1992] (Figures 1a and 1b). Today, the lake is outside the East Asian summer monsoon influence [*Tian et al.*, 2001], in an arid desert climate with a mean annual temperature of ~4°C, a mean annual precipitation of ~160 mm, and a mean annual evaporation of ~2000 mm (Delingha station, ~30 km northeast of the lake) [*Yi et al.*, 1992]. Hurleg Lake is fed by Bayin River (with a mean discharge of 1.9×10^8 m^3/year), Balegen River (prior to reservoir construction; 0.1×10^8 m^3/year) and groundwater from the mountains to the north, and drains through Lianhu River into Toson Lake (Figure 1b). The lake has a mean residence time of ~1 year. The lake is surrounded by dense common reeds. Aquatic plants, such as *Chara*, *Potamogeton*, and *Ruppia* dominate the submergent vegetation, which covers much of the lake bottom.

[5] An 85-cm-long core HL06-2 (N37°17.689′, E96°52.725′; Figure 1b) was obtained in the summer of 2006 under 7.6 m water depth using a plastic tube fitted with a piston to preserve the sediment-water interface. The core was subsampled shortly after collection in the field in continuous 0.5-cm slices for the top 30 cm and in 1-cm slices from 30 to 85 cm totaling 115 samples. Chronology was controlled by ^{210}Pb and ^{137}Cs analyses at 19 levels and ^{14}C dates from aquatic plant macrofossils at five intervals [*Zhao et al.*, 2009]. The five ^{14}C dates are in order but were corrected by a 2758 ^{14}C year carbon reservoir effect based on the difference between paired ^{14}C and ^{210}Pb dates at 12–13.5 cm. The corrected dates were then calibrated using the Calib 5.0.1 program based on IntCal04 dataset [*Reimer et al.*, 2004]. The age model was constructed using a modified 2nd-polynomial curve (Figure 2a).

[6] Loss-on-ignition (LOI) analysis was carried out every 1-cm interval along the core, with the loss between 550 and 1000°C to estimate carbonate content. Ostracode shells were identified and counted from 0.2–7.6 g of dry sediment at 1-cm intervals and are presented as concentrations (# shells/gram). We also performed the spectral analysis on the carbonate percentage using the program REDFIT [*Schulz and Mudelsee*, 2002], and cross-spectral analysis between carbonate percentage and residual Δ^{14}C [*Reimer et*

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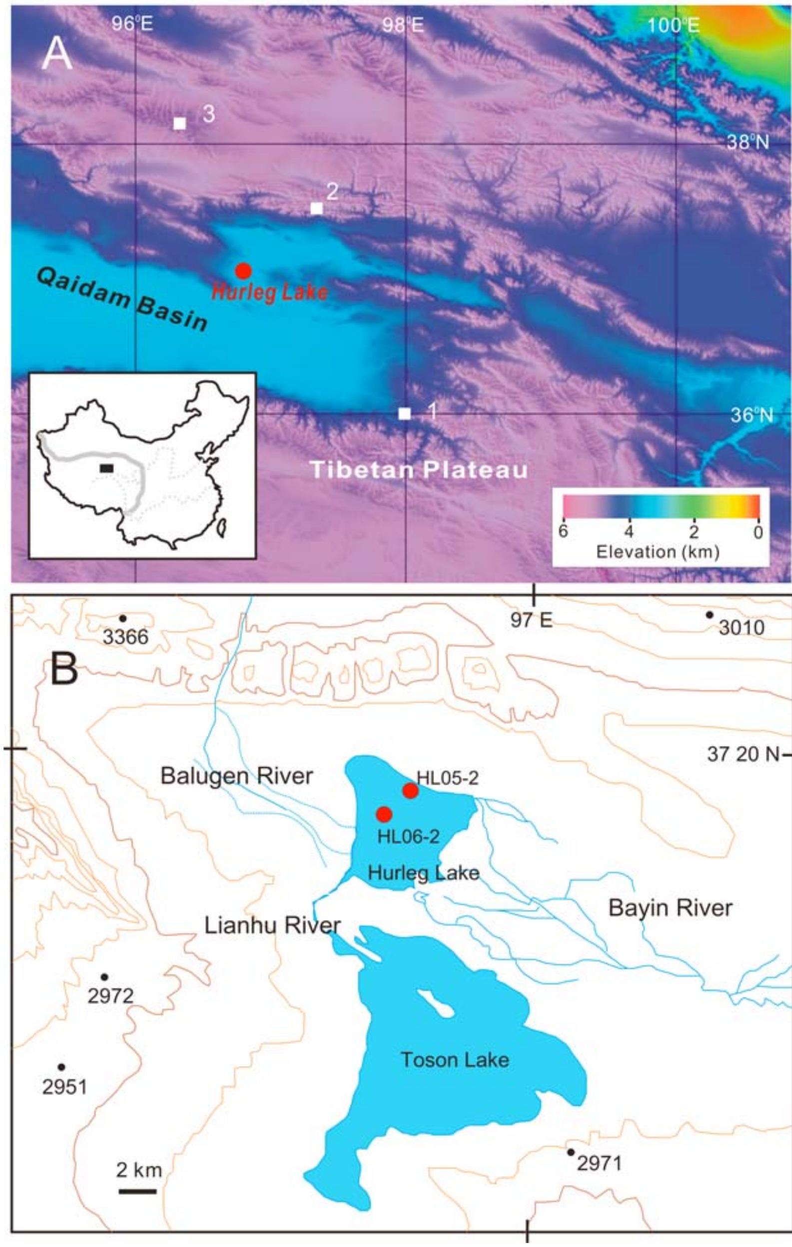

Figure 1. (a) Digital elevation model of the NE Tibetan Plateau, showing the locations of Hurleg Lake (red circle) and other sites (white squares): 1-Dulan tree-ring site, 2-Delingha tree-ring site, and 3-Dunde ice core. The inset indicates the location of study region in Qaidam Basin (the black rectangle). (b) Figure shows topography and the location of rivers and Toson Lake in relation to Hurleg Lake. Coring locations are indicated by red circles.

al., 2004] with AnalySeries version 2.0.4 using a Bartlett window.

3. Results

[7] The age model shows that the core goes back ~1700 years (Figure 2a). Despite the use of constant correction factor, chronology is adequate for our discussions because (1) the old carbon correction was based on the age difference of the same sediment slice dated by both ^{14}C and ^{210}Pb analysis (AD 1925); (2) the same plant materials from *Ruppia* were used for radiocarbon dating; and (3) the derived age model shows an almost identical trend over the last 1700 years from ^{210}Pb and ^{14}C age determinations.

[8] Core HL06-2 is mainly composed of fine-grained (<250μm), light colored minerals such as sand and endogenic calcite. LOI analysis shows the core is dominated by carbonate and silicate both varying between 30 and 60% (Figure 2b). Between 300 and 1000 AD, carbonate percentage stays relatively high (>40%). After 1000 AD, carbonate percentage shows large-magnitude oscillations between 30% and 60% with generally lower values over the last 300 years. Ostracode abundance, dominated by *Candona neglecta* (Sars, 1887) and *Limnocythere inopinata* (Baird, 1843), shows inverse relation to carbonate percentage (Figures 2b–2e and 3a). Although we didn't count the full assemblage, we observed rare broken shells and more juveniles than adults. Thus the ostracodes are likely *in situ*

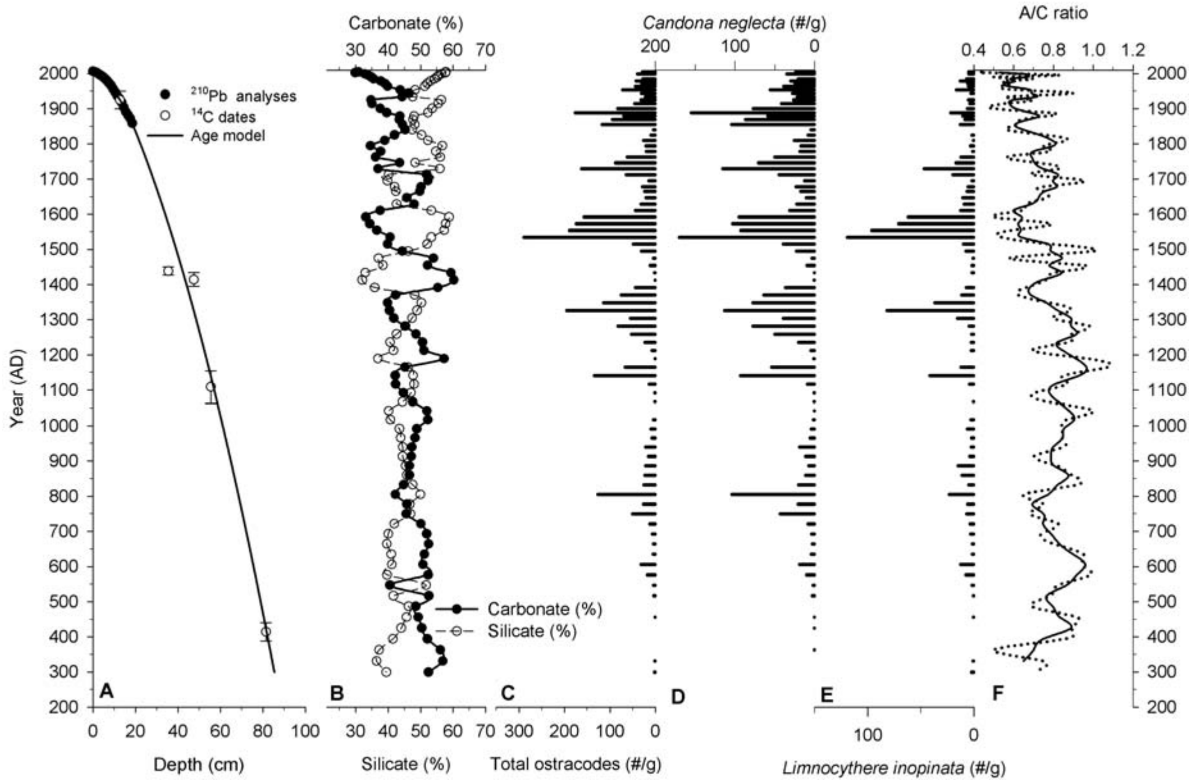

Figure 2. Lake-level proxy records from Hurleg Lake core HL06-2. (a) Age model derived from a modified 2nd-polynomial curve based on ^{210}Pb (n = 19) and ^{14}C age determinations (n = 5). (b) Carbonate (solid circles) and silicate (open circles) percentage. (c) Total concentration of ostracode shells (# shells/gram). (d) Concentration of *Candona neglecta* (# shells/gram). (e) Concentration of *Limnocythere inopinata* (# shells/gram). (f) Pollen A/C ratio: raw data (dotted) and 3-point moving average (solid) are shown [*Zhao et al.*, 2009].

and not transported from elsewhere. Both species and the total abundance show a similar pattern, except during the last 300 years. The spectral power of carbonate percentage and the coherence between carbonate percentage and residual Δ^{14}C data indicate a ~200-year periodicity during the last 1000 years (Figures 3d and 3e).

4. Discussion

4.1. Late Holocene Lake-Level History From Hurleg Lake

[9] The carbonate percentage and ostracode abundance at Hurleg Lake reflect changes in depositional environment and in particular water depth. This is supported by (1) carbonate percentage of surface sediment samples along a transect increases from 21.1% at 1.1 m water depth, to 43.6% at 6.3 m and 45.4% at 8.6 m; (2) the down-core stratigraphy of core HL05-2 from shallow water (Figure 1b) shows that low carbonate corresponds with high ‘soil-textured’ clay content and abundant plant root remains [*Zhao et al.*, 2007]; (3) total ostracode abundance decreases from 66 shells per gram of dry sediment at 6.3 m water depth to 8 shells per gram at 8.6 m. None of the shells were found at 1.1 m which is likely to be above the wave base and thus too high energy for ostracodes and their preservation.

[10] The relation between our proxies and water depth can be explained by the dilution of endogenic carbonate by fine-grained silicates delivered from surrounding deserts with sparse vegetation cover. Occasional flash floods coursing down the alluvial plain to the west bring detrital materials into the lake, and most of them would be trapped in marginal areas by aquatic vegetation. We observed one flash flood in 2005 and how much sediment it carried off. It is common to see higher ostracode abundance near lake margins rather than in deeper part. Minerals from aeolian sources can also dilute the endogenic carbonate, but they are likely to affect the entire basin equally. Large riverine input is unlikely because the inflowing rivers are surrounded by dense marsh and the coring site is far away from the river mouths. The breakdown of the relation between carbonate percentage and ostracode abundance, as well as between *L. inopinata* and total abundance over the last 300 years could be influenced by human activities. The increase in green algae *Pediastrum* colonies after 1700 AD has been attributed to human-induced lake eutrophication [*Zhao et al.*, 2009]. *Cadona neglecta* is more tolerant of lower dissolved oxygen than *L. inopinata* so its continued high abundance is consistent with eutrophication [*Meisch*, 2000].

[11] The changes in lake levels at Hurleg Lake can be caused by changes in river inflow and/or evaporation. Higher lake level due to increased inflow will result in lower detrital mineral and ostracode abundance at the coring location. However, even though Bayin River samples typically had slightly higher concentrations of HCO_3^- and Ca^{2+} (HCO_3^-: 3.1–4.7 mmol/L; Ca^{2+}: 1.2–1.6 mmol/L, n = 6) than the lake (HCO_3^-: 2.0–2.7 mmol/L; Ca^{2+}: 0.7–1.1 mmol/L, n = 5) (unpublished data), the ratio of HCO_3^- to Ca^{2+} in both waters is nearly identical (≈2.5), resulting

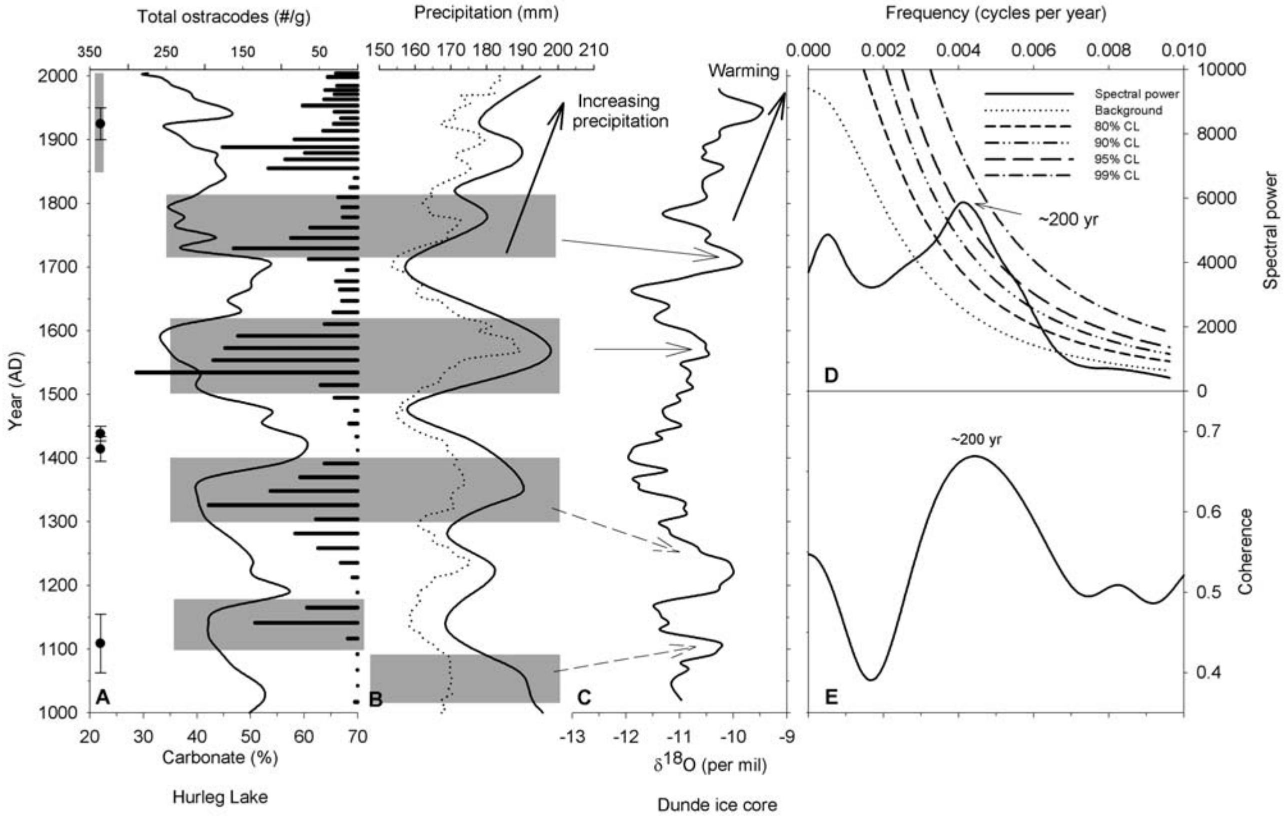

Figure 3. Regional climate correlation and forcings. (a) Carbonate percentage and total ostracode abundance at Hurleg Lake. The 4 corrected and calibrated ^{14}C ages and 2-sigma ranges are shown on the left, along with ^{210}Pb-dated interval (gray bar at 1850–2006). Note: the scale of ostracode concentration is reversed. (b) Reconstructed annual precipitation from tree-ring data in the surrounding mountains near Delingha (dashed) [*Shao et al.*, 2005] and near Dulan (solid) [*Sheppard et al.*, 2004]. The gray bands highlight the corresponding dry intervals at Hurleg Lake but wet intervals at tree-ring sites. (c) The δ^{18}O record from Dunde ice core [*Thompson et al.*, 2003]. (d) Spectral power of percent carbonate (%) at Hurleg Lake. The legends of spectral power, spectral background and different confidence levels (CL) are indicated. (e) The coherence from cross-spectral analysis of carbonate (%) and residual Δ^{14}C [*Reimer et al.*, 2004].

in little difference in calcite saturation index. Moreover, hydrologic investigations of this region indicated that Bayin River flow is rather stable due to groundwater buffering [*Yi et al.*, 1992]. Furthermore, the lake-level proxies and the precipitation data reconstructed from the mountains where the river originates [*Sheppard et al.*, 2004; *Shao et al.*, 2005] show a general inverse relationship, especially during the last 1000 years (Figures 3a and 3b). Thus, increased river flow is an unlikely mechanism.

[12] Increased evaporation can decrease the lake level and enrich Ca^{2+} and HCO_3^- concentrations in the lake, facilitating more carbonate precipitation. The decreased lake level will move the shoreline toward the center and result in more delivery of detritals to the coring location. The increased silicate input may effectively dilute the carbonates such that carbonate percentage may actually show a decrease. We have difficulty in obtaining materials for more ^{14}C dates to derive sediment accumulation rates to strengthen our argument. However, our interpretation is supported by parallel changes in carbonate percentage and *Artemisia*/Chenopodiaceae (A/C) pollen ratio (Figure 2f) from the same core [*Zhao et al.*, 2009]. A/C ratio is a proxy for soil moisture and effective moisture in the surrounding landscape [*Zhao et al.*, 2007, 2009], with high A/C ratio representing more steppe plants and relatively moist conditions and low A/C ratio indicating drier conditions. The A/C ratio shows shorter time-scale variations, suggesting that vegetation responds more quickly to changes in effective moisture than Hurleg Lake level. The inverse relation between the lake-level proxies and the precipitation data reconstructed from the mountains [*Sheppard et al.*, 2004; *Shao et al.*, 2005] also suggests the importance of evaporation (Figures 3a and 3b). Moreover, the inverse relation between A/C ratio and mountain precipitation has existed for at least the last 1700 years [*Zhao et al.*, 2009]. Furthermore, over the last 50 years the relative humidity data from Delingha station agree with A/C ratios from another lake 50 km to the east but show an inverse relation with the Dunde ice core [*Zhao et al.*, 2008]. For these reasons, increased evaporation rather than decreased inflow is the more parsimonious cause of lowered lake levels.

[13] The lake-level proxies show an increased variability after 1000 AD. Between 300 and 1000 AD, generally high carbonate and low ostracode abundance suggest that Hurleg Lake may have been largely spilling over to Toson Lake. In addition, according to our age-model this section has a lower sedimentation rate, so that each sample integrates a longer time interval possibly averaging out short-term

changes. Hence, we will focus on the last 1000 years to examine the centennial-scale climate oscillations.

[14] During the last 1000 years, Hurleg Lake experienced four significant low lake-level periods: 1100 – 1170, 1300 – 1400, 1500 – 1620, and 1720 –1820 AD, as indicated by low carbonate percentages and abundant ostracode shells (Figure 3a). After 1700 AD, the generally low carbonate percentage and high ostracode abundance indicate a low lake level and dry climate. Overall, both proxy data show a consistent climate pattern, with ~200-year lake-level oscillations during the last 1000 years and increased aridity after ~1700 AD.

4.2. Opposite Moisture Patterns Between the NE Qaidam Basin and Surrounding Mountains During the Last 1000 Years

[15] Our lake-level record for the last 1000 years in the NE Qaidam Basin (QB) (elevation at ~2800 m asl) generally shows a negative correlation with the tree-ring reconstructed precipitation in the surrounding mountains (3700–4000 m asl) (Figures 3a and 3b) [*Sheppard et al.*, 2004; *Shao et al.*, 2005]. During the driest periods at Hurleg Lake centered at 1140, 1350, 1580, and 1760 AD, tree-ring data show peak precipitation. During the wettest periods centered at 1020, 1200, 1430, and 1690 AD, tree-ring data show the lowest precipitation. Over the last 300 years, the lake-level record shows generally low values, whereas the tree-ring data show a slightly increase in precipitation. Dunde ice core (at 5300 m asl) shows an abrupt increase in snow accumulation after 1700 AD [*Yao et al.*, 1991].

[16] The δ^{18}O values from Dunde ice cap show regional temperature changes (Figure 3c) [*Thompson et al.*, 2003]. In general, the warm periods in Dunde ice core correspond to more precipitation in the mountains and increased aridity in the NE QB and vice versa. Thus, temperature and moisture conditions in the NE QB indicate warm-dry and cold-wet associations, just the opposite of the warm-wet and cold-dry associations in mountain areas that may be affected by the summer monsoon. The slight mismatch among these records may be caused by different age models and larger dating errors at Hurleg Lake.

[17] As regional hydrological investigation [*Yi et al.*, 1992] and our observations suggest the precipitation in the mountains areas will take several years rather than hundred years to come to Hurleg Lake through groundwater and river recharge, the observed opposite trends in effective moisture at high and low elevations indicate that topography may have played a role in regional climate change by affecting rising and sinking of air mass. During stronger monsoon periods, the intense heating and uplift of air mass over the Tibetan Plateau leads to a strong subsidence to the northwest and north of the plateau as compensating flow, inducing a dry climate in central Asia [*He et al.*, 1987; *Broccoli and Manabe*, 1992]. This mechanism has been invoked as an explanation for the opposing early Holocene wet/dry climate pattern between monsoonal region and arid central Asia [*Herzschuh*, 2006; *Zhao et al.*, 2007; *Chen et al.*, 2008]. A similar mechanism appears to operate at a smaller scale as was shown in regional climate simulations of the QB [*Sato and Kimura*, 2005]. Such localized high pressure system will bring hotter climate to the basin due to adiabatic heating. The topographically induced occurrence of increased hot air subsidence into the QB and increased monsoon precipitation on surrounding mountains may have been responsible for the observed opposite changes in moisture condition at low and high elevations.

4.3. Possible Solar Forcing of Moisture Oscillations in the Qaidam Basin

[18] The coherence at the ~200-year periodicity between the lake-level data (carbonate percentage) from Hurleg Lake and the solar proxy (residual Δ^{14}C) indicates possible solar forcing of centennial-scale climate oscillations in the Qaidam Basin (QB; Figures 3d and 3e). Our average sampling resolution is about ~20 years, so we will not discuss any periodicities less than 100 years (>0.01 cycles per year). Also our spectral analysis shows no other significant periodicities. The solar forced ~200-year oscillations have been reported on the Tibetan Plateau, including the tree-ring precipitation reconstructions near Delingha [*Shao et al.*, 2005] and the monsoon precipitation signal indicated by iron oxide content at Qinghai Lake [*Ji et al.*, 2005]. The spectral analysis of the δ^{18}O values from Dongge Cave as a proxy of Asian monsoon strength also showed significant peaks at the ~200-year periodicity [*Dykoski et al.*, 2005, *Wang et al.*, 2005]. Higher solar output corresponds to a stronger monsoon [*Wang et al.*, 2005], which intensifies the uplift of air mass on the high Tibetan Plateau and strengthens the subsidence of air mass over the QB. The reverse is true during the period of lower solar output. Thus, high solar activity is correlated with dry climate in QB and increased precipitation in monsoonal areas.

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References

Broccoli, A. J., and S. Manabe (1992), The effects of orography on mid-latitude Northern Hemisphere dry climates, *J. Clim.*, *5*, 1181–1201, doi:10.1175/1520-0442(1992)005<1181:TEOOOM>2.0.CO;2.

Chen, F. H., et al. (2008), Holocene moisture evolution in arid central Asia and its out-of-phase relationship with Asian monsoon history, *Quat. Sci. Rev.*, *27*, 351–364, doi:10.1016/j.quascirev.2007.10.017.

Dykoski, C. A., R. L. Edwards, H. Cheng, D. X. Yuan, Y. J. Cai, M. L. Zhang, Y. S. Lin, J. M. Qing, Z. S. An, and J. Revenaugh (2005), A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China, *Earth Planet. Sci. Lett.*, *233*, 71–86, doi:10.1016/j.epsl.2005.01.036.

He, H. Y., J. W. McGinnis, Z. S. Song, and M. Yanai (1987), Onset of the Asian summer monsoon in 1979 and the effect of the Tibetan Plateau, *Mon. Weather Rev.*, *115*, 1966–1995, doi:10.1175/1520-0493(1987)115<1966:OOTASM>2.0.CO;2.

Herzschuh, U. (2006), Palaeo-moisture evolution in monsoonal central Asia during the last 50,000 years, *Quat. Sci. Rev.*, *25*, 163–178, doi:10.1016/j.quascirev.2005.02.006.

Ji, J. F., J. Shen, W. Balsam, J. Chen, L. W. Liu, and X. Q. Liu (2005), Asian monsoon oscillations in the northeastern Qinghai–Tibet Plateau since the late glacial as interpreted from visible reflectance of Qinghai Lake sediments, *Earth Planet. Sci. Lett.*, *233*, 61–70, doi:10.1016/j.epsl.2005.02.025.

Meisch, C. (2000), *Freshwater Ostracode of Western and Central Europe*, Spektrum Akad., Heidelberg, Germany.

Reimer, P. J., et al. (2004), INTCAL04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP, *Radiocarbon*, *46*, 1029–1058.

Sato, T., and F. Kimura (2005), Impact of diabatic heating over the Tibetan Plateau on subsidence over northeast Asian arid region, *Geophys. Res. Lett.*, *32*, L05809, doi:10.1029/2004GL022089.

Schulz, M., and M. Mudelsee (2002), REDFIT: Estimating red-noise spectra directly from unevenly spaced paleoclimatic time series, *Comput. Geosci.*, *28*, 421–426, doi:10.1016/S0098-3004(01)00044-9.

Shao, X. M., E. Y. Liang, L. Huang, and L. Wang (2005), A 1437-year precipitation history from Qilian Juniper in the northeastern Qinghai-Tibetan Plateau, *PAGES News*, *13*, 14–15.
Sheppard, P. R., P. E. Tarasov, L. J. Graumlich, K. U. Heussner, M. Wagner, H. Osterle, and L. G. Thompson (2004), Annual precipitation since 515 BC reconstructed from living and fossil juniper growth of northeastern Qinghai Province, China, *Clim. Dyn.*, *23*, 869–881, doi:10.1007/s00382-004-0473-2.
Thompson, L. G., E. M. Thompson, M. E. Davis, P. Lin, K. Henderson, and T. A. Mashiotta (2003), Tropical glacier and ice core evidence of climate change on annual to millennial time scales, *Clim. Change*, *59*, 137–155, doi:10.1023/A:1024472313775.
Tian, L., V. Masson-Delmotte, M. Stievenard, T. Yao, and J. Jouzel (2001), Tibetan Plateau summer monsoon northward extent revealed by measurements of water stable isotopes, *J. Geophys. Res.*, *106*, 28,081–28,088, doi:10.1029/2001JD900186.
Wang, Y. J., H. Cheng, R. L. Edwards, Y. Q. He, X. G. Kong, Z. S. An, J. Y. Wu, M. J. Kelly, C. A. Dykoski, and X. D. Li (2005), The Holocene Asian monsoon: Links to solar changes and north Atlantic climate, *Science*, *308*, 854–857, doi:10.1126/science.1106296.
Yao, T. D., Z. C. Xie, and Q. Z. Yang (1991), Temperature and precipitation fluctuations since 1600a provided by Dunde Ice Cap, China, in *Proceedings of the International Symposium on Glacial-Ocean-Atmosphere Interactions*, *IAHS Publ.*, *208*, 61–70.
Yi, X., D. Yang, and W. Xu (1992), China regional hydrogeology survey report—Toson Lake (in Chinese), 1:200,000, *Map J-47-[25]*, Qaidam Integrative Geol. Surv., Golmud, China.
Zhao, Y., Z. C. Yu, F. H. Chen, E. Ito, and C. Zhao (2007), Holocene vegetation and climate history at Hurleg Lake in the Qaidam Basin, northwest China, *Rev. Palaeobot. Palynol.*, *145*, 275–288, doi:10.1016/j.revpalbo.2006.12.002.
Zhao, Y., Z. C. Yu, F. H. Chen, X. J. Liu, and E. Ito (2008), Sensitive response of desert vegetation to moisture change based on a near-annual resolution pollen record from Gahai Lake in the Qaidam Basin, northwest China, *Global Planet. Change*, *62*, 107–114, doi:10.1016/j.gloplacha.2007.12.003.
Zhao, Y., Z. C. Yu, X. J. Liu, C. Zhao, and F. H. Chen (2009), Late Holocene vegetation and climate oscillations in the Qaidam Basin of the Northeastern Tibetan Plateau, *Quat. Res.*, doi:10.1016/j.yqres.2008.11.007, in press.

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