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The influence of irrigation water on the hydrology and lake water budgets of two small arid-climate lakes in Khorezm, Uzbekistan

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SUMMARY

Little is known regarding the origins and hydrology of hundreds of small lakes located in the western Uzbekistan province of Khorezm, Central Asia. Situated in the Aral Sea Basin, Khorezm is a productive agricultural region, growing mainly cotton, wheat, and rice. Irrigation is provided by an extensive canal network that conveys water from the Amu Darya River (AD) throughout the province. The region receives on average 10 cm/year of precipitation, yet potential evapotranspiration exceeds this amount by about 15 times. It was hypothesized that the perennial existence of the lakes of interest depends on periodic input of excess irrigation water. This hypothesis was investigated by studying two small lakes in the region, Tuyrek and Khodjababa. In June and July 2008, surface water and shallow groundwater samples were collected at these lake systems and surrounding communities and analyzed for δ^2H, $\delta^{18}O$, and major ion hydrochemistry to determine water sources. Water table and lake surface elevations were monitored, and the local aquifer characteristics were determined through aquifer tests. These data and climate data from a Class A evaporation pan and meteorological stations were used to estimate water budgets for both lakes. Lake evaporation was found to be about 0.7 cm/day during the study period. Results confirm that the waters sampled at both lake systems and throughout central Khorezm were evaporated from AD water to varying degrees. Together, the water budgets and stable isotope and major ion hydrochemistry data suggest that without surface water input from some source (i.e. excess irrigation water), these and other Khorezm lakes with similar hydrology may decrease in volume dramatically, potentially to the point of complete desiccation.

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1. Introduction

Located in Central Asia and forming part of the border between Kazakhstan and Uzbekistan, the Aral Sea is a closed lake basin that receives almost all of its surface input from two rivers, the Syr Darya and Amu Darya (AD). In 1960, the sea was the 4th largest inland body of water in the world. However, by 2003, consumptive use of river water to annually irrigate millions of hectares of cotton, rice, and wheat in Central Asia in part resulted in approximately 90% loss of sea volume (Micklin, 2007). Since the days of the Soviet Union, Uzbekistan has maintained its position as one of the top four cotton-exporting countries in the world (ZEF, 2009). One of the most productive and heavily irrigated regions of Uzbekistan is the Khorezm province.

Khorezm has an area of about 6250 km^2 and is located approximately 225 km south of the present remainders of the Aral Sea (Conrad et al., 2007). The AD forms part of the eastern boundary of the province, which lies within the river's historic floodplain (Fig. 1). Approximately 2600 km^2 can be irrigated in Khorezm with AD water carried by about 16,000 km of mostly unlined canals (Conrad et al., 2007; Ibrakhimov et al., 2007). The commonly used practice of flood irrigation has caused extensive water and soil salinization in this arid and flat landscape (Micklin, 2007). Scattered among, and often adjacent to, the many agricultural fields in Khorezm are hundreds of small shallow lakes.

The research presented here focuses on two of these lakes, Tuyrek (TUY; 41.32126627°N, 60.57582789°E) and Khodjababa (KHO; 41.4371638°N, 60.28848502°E). These and other Khorezm lakes are the subject of a larger study to determine both their potential utility for fish production and tourism and the impacts of anthropogenic activities on sustaining the health of these aquatic ecosystems. These goals are complemented here by investigating the influence of irrigation water on the water budget of TUY and KHO, which are taken to be representative of other shallow lakes in Khorezm. Specifically, the current study investigates the hypothesis that the perennial existence of TUY and KHO depends on periodic input of excess irrigation water.

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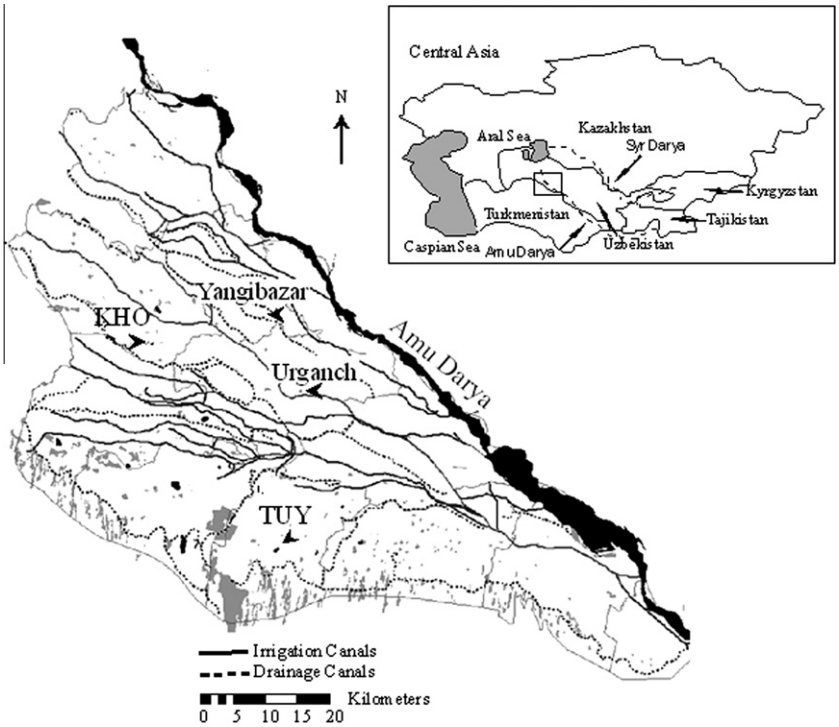

Fig. 1. Map of the Khorezm province and its irrigation and drainage canals. Labeled are the two study lakes, the AD, and the two weather station locations.

As is the case in other arid environments, in Khorezm there is an urgent need to reduce water demand by increasing water use efficiency in the agricultural sector (ZEF, 2009). If TUY and KHO depend on excess irrigation water input for perennial existence, then increasing water use efficiency in Khorezm could have the unintended result of removing or reducing a major lake input.

The research hypothesis was evaluated through the analyses of water sample major ion hydrochemistry and stable isotope (^{2}H and ^{18}O) data as well as lake water budgets for both TUY and KHO. Researchers have demonstrated the usefulness of major ion hydrochemistry and stable isotope data in studies of sources and movements of water in lake and wetland systems (e.g., Krabbenhoft et al., 1990; Herczeg et al., 1992; Marimuthu et al., 2005). Similarly, lake water budgets have been used to identify major lake inputs and outputs (e.g., Lee and Swancar, 1997; Mann and McBride, 1972; Winter, 1981). To the best of our knowledge, the study presented here is the first to examine the hydrology of the small lakes in Khorezm, a region with water resource concerns shared by other arid and heavily irrigated agricultural landscapes.

2. Study site description

Air temperatures in Khorezm annually range from −20 to 45 °C and summer relative humidity is low (ZEF, 2009). The region receives on average 92 mm of precipitation annually, of which less than five mm occurs in the summer, on average (Wehrheim et al., 2008). This contribution to both surface and groundwater is considered minimal because it is exceeded about 15 times by potential evapotranspiration (Conrad et al., 2007). Most of the hundreds of lakes scattered throughout this arid agricultural landscape receive irrigation water.

TUY and KHO have approximate surface areas of 80,900 and 208,000 m^2 and range in depth between 0.5–3.0 m and 0.5–2.0 m, respectively. A narrow band of emergent macrophytes grows along lake perimeters (Fig. 2A and B). Small spurs from larger canals deliver water when available to farms adjacent to TUY and KHO. Flow is generally gravity driven (Conrad et al., 2007), although pumps are used occasionally.

Potential surface water contributors to the lakes include excess irrigation water drainage from nearby agricultural fields and spring leach water. Non-point runoff may also contribute to lake water budgets. Leaching usually begins in late February when water is released into and flows down the canals from the AD and is directed onto agricultural fields by farmers for removal of salts accumulated in the soils from the past growing season. Once the accumulated salts have been leached from the soil, non-infiltrated water is removed by drain canal or drained into the lakes. Later, during the growing season, fields are flood irrigated when water becomes available. Farmers dig ditches to deliver some or all excess water into the lakes. Thus, surface inputs from leach water and excess irrigation water do not constitute permanent perennial surface water contributions. Farmers report reducing lake stage when input threatens to overfill the lake by digging an outlet to a drain canal; however, to the best of our knowledge, this did not occur during the study period at TUY or KHO.

Depending on lake stage and local water table elevation, lakes could be gaining, losing, or both. The aquifers interacting with the lakes in Khorezm are unconfined. Shallow aquifer material near the lakes of study was found to be mostly sand mixed with small amounts of silt and mud, which is consistent with the general knowledge of soil texture in the region. Prior to the 1950s, the depth to the water table was relatively large (Ibrakhimov, 2009), and recharge sources were likely from winter precipitation and occasional flooding of the AD. Between 1954 and 1960, the average depth to groundwater from six monitoring wells in Khorezm decreased from 15 to 10 m because of increased irrigation with AD water and subsequent infiltration (Ibrakhimov, 2009). Today, the water table is much higher. For example, between 2000 and 2006, average depth to groundwater in the region near the lakes of study fluctuated between 3.5 m during dry periods to 1.2 m or less during the growing season (Ibrakhimov, 2009).

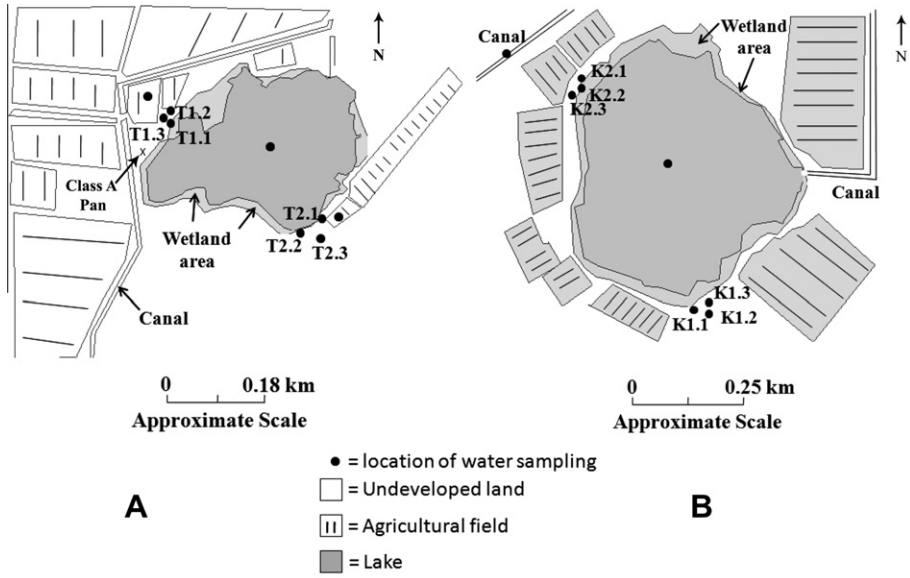

Fig. 2. (A) Location at TUY of water sampling, evaporation pan, canals, and agriculture fields. (B) Location at KHO of water sampling, canals, and agriculture fields. Spur canals near the lakes are rough ditches that can both drain into the lake and move water between fields.

3. Methods

3.1. Data collection and fieldwork

The majority of fieldwork was conducted from June 1 to August 1, 2008; however, some preliminary water sampling occurred as early as June 2006, while water table elevation measurements were made as late as December 2008. In June 2008, an array of three piezometers was installed on opposite sides of each lake for water table elevation monitoring and groundwater sampling (Fig. 2A and B). The PVC piezometers were located between 15 and 40 m from the lake shore. The distance from one piezometer to another in an array was between 10 and 30 m. Pipes were installed in approximately 3 m deep bore holes and were screened from 1.5 to 2.9 m. This depth was chosen to capture the estimated annual groundwater elevation minimum. A fine synthetic screen covering the slotted PVC prevented the piezometer from filling with sediment. The depth to water in all piezometers was measured weekly. Lake elevation was monitored nearly continuously using a HOBO 9.1 m (30-foot) depth data logger (±5.0 mm accuracy; Onset©, 2008) installed beneath the lake surface. A single HOBO pressure transducer located at TUY was used to monitor barometric pressure continuously and correct lake stage measurements at both TUY and KHO. Slug and pump tests were performed on piezometers at each lake to determine aquifer properties. Piezometers and lake surface elevations were surveyed using laser survey equipment (with an accuracy of ±1.5 mm for every 50 m) to relate water elevations to a local datum.

Water samples were taken from the lakes, irrigation canals, flooded fields, and local groundwater (Fig. 2A and B). Samples from irrigation canals and flooded fields are hereafter referred to as "surface water." Surface water and lake samples were collected using grab-sampling techniques, and groundwater samples were pumped from piezometers after three well volumes were removed using a 12-V submersible pump with PVC tubing. Groundwater samples were also collected from open community water supply wells located throughout the region, generally 1–30 km distant from the lakes of study. These wells are about 0.5 m in diameter, 3–4 m deep, and used daily by residents of the area. Production is minimal, as they are bucket drawn, not pumped. Samples were taken during March, June, and July, 2008. Water samples were also collected from the AD monthly between June 2006 and 2007. One river sample was collected in August 2008. One Class A evaporation pan was installed near TUY to measure evaporation. Logistical constraints resulted in the installation of this pan 2 m above the lake surface on the roof of a building 10 m from the shore of the lake. It was not possible to install an evaporation pan at KHO. Meteorological data covering December 2006 to 2008 were retrieved from meteorological stations (ZEF, 2009) located in Yangibazar and Urgench, within 30 km of both TUY and KHO (Fig. 1). Lake evaporation estimates were derived using data from the evaporation pan and the two meteorological stations.

3.2. Laboratory work

Water samples collected in acid-washed sealed 1-L plastic bottles were filtered through standard 0.45 μm Whatman GF/F filter papers in the laboratory. The pH of water samples was measured in both the field and the laboratory. Prior to freezing the filtered 1-L water samples collected between March and August 2008, 4 mL of filtered water were taken from every sample and sealed in a borosilicate glass vial with Teflon faced rubber lined caps. All water samples were then frozen and archived for subsequent analyses. Frozen water samples were later analyzed using standard laboratory techniques for major cation and anion hydrochemistry (Semenov, 1977). Analytic error in these measurements, as estimated from replicate samples, was found to be ±5%. The 4-mL samples were analyzed for stable isotopic composition (hydrogen (δ^2H) and oxygen (δ^{18}O)) with a Micromass IsoPrime stable isotope ratio mass spectrometer at the University of Nevada, Reno Stable Isotope Laboratory. Hydrogen stable isotopic analyses were performed using the continuous flow Cr reduction method of Morrison et al. (2001), with an uncertainty of ±1‰ (1 standard deviation). Oxygen stable isotopic analyses were performed using the CO_2–H_2O equilibration method of Epstein and Mayeda (1953), with an uncertainty of ±0.1‰ (1 standard deviation). Delta (δ) notation is used and indicates that the measurements are expressed in relation to Vienna Standard Mean Ocean Water (VSMOW). Water samples collected between June 2006 and 2007 were analyzed only for major ion concentrations.

Analysis of variance (ANOVA) was performed on δ^2H data from KHO and TUY. Tukey's Multiple Comparison Test was used to

compare the mean δ^2H values of ground, lake, and surface water samples at a $p < .05$ level of significance. All statistical analyses were performed using R software package.

3.3. Hydrochemistry modeling

The US Geological Survey's geochemical model PHREEQC (Parkhurst and Appelo, 1999) was used to perform inverse modeling, where the observed major ion hydrochemistry of lake water was numerically matched by evaporating and reacting AD water until the calculated hydrochemistry matched observed lake water hydrochemistry. The inverse model included the processes of water evaporation, gas exchange, and mineral-water reactions. The modeled minerals included calcite, gypsum, and halite. A 5% difference between modeled and observed aqueous solution concentrations was allowed. Two models were created, one for each lake. In each case, averages of the major ion concentrations from 29 different river water samples were used for the AD aqueous solution. Similarly, TUY and KHO aqueous solutions were created by using the average major ion concentrations from all available lake water samples.

3.4. Water budget methods

TUY and KHO water budgets for June through August 2008 can be described by equation:

$$\Delta S \pm e_{\Delta S} = I \pm e_I + G \pm e_G + E \pm e_E \quad (1)$$

where ΔS is lake storage flux (m^3), I is net surface water flux (m^3), G is net groundwater flux (m^3), E is net evaporation (m^3), and e is uncertainty in each flux component. Precipitation is not considered because there was no recorded precipitation during the budget time frame. Similar water budget equations have been used by Winter (1981), Lee and Swancar (1997), Mann and McBride (1972), and Li et al. (2007). Eq. (1) was applied to specific time intervals bounded by the dates when depth to water measurements were taken on both sides of the lakes. It was solved in terms of volumetric fluxes by multiplying evaporation and measured changes in water elevations (m) by the surface area of the lake (m^2), as estimated using satellite imagery collected in September 2006 and Google Earth's® polygon tool. This conversion assumed vertical lake sides, which was necessary because bathymetric information was unavailable.

In these lakes, time periods, magnitude, and location of channeled surface inflow varies and depends on farmer discretion. Overland flow may also contribute to input. For these reasons, a water budget approach was chosen to provide a satisfactory estimate of surface input. Surface input for these lakes was thus estimated indirectly by solving Eq. (1) for I, after accounting for all other components. Calculation of Eq. (1) components is described below.

3.4.1. Groundwater

Groundwater, G, was estimated by calculating the groundwater flux, Q, in m^3/day to and from the lake. Q was calculated using the Dupuit equation (Eq. (2)), applied across the water table and lake surface elevation gradient:

$$G = Q = \frac{Kw(h_1^2 - h_2^2)}{2L} \quad (2)$$

where K is hydraulic conductivity of the aquifer material (m/day), w is width through which groundwater flow occurred (m), h_1 and h_2 are heights above the bottom of the aquifer (m), and L is distance between points h_1 and h_2. This equation assumes that Darcy's law is valid, the hydraulic gradient is equal to the slope of the water table and that the groundwater flow streamlines are horizontal, the equipotential lines are vertical, and the aquifer is shallow and unconfined. While this equation only allows for horizontal flow into the lake, some near-vertical flow lines must have occurred near the lake. Errors resulting from violating these assumptions were thought to be small relative to uncertainty in determining K and other parameters, so vertical flow was ignored, and it was assumed that all water flowing horizontally toward the lake within some radial distance eventually discharged into the lake (Lee and Swancar, 1997).

3.4.1.1. Width, length, and water table elevation. The width through which groundwater flow occurred was estimated by lake perimeters, which for TUY and KHO were found to be approximately 1300 and 1800 m, respectively, using satellite imagery from Google Earth taken in September 2006. From this imagery, it was judged that the lakes had similar stage in 2006 as in the summer of 2008. It was not feasible to allow lake perimeters to vary with changes in lake stage because bathymetric information was not available.

The distances between the piezometers and the lake shore were determined using the satellite imagery described above. The distances to the lake shore from three piezometers on the same side of the lake were averaged. This mean distance was used in Eq. (2) as L (Fig. 3).

This equation requires h_1 and h_2 to be heights above the bottom of the aquifer, and it was assumed that the bottom of the effective aquifer corresponded to the bottom of the deepest part of the lake. Thus, h_1 was calculated as the difference between the water table elevation and the elevation of the deepest part of the lake. Water table elevation measurements were averaged for an array when calculating h_1. To determine h_2, the difference in elevation between the lake surface and the lake bottom was calculated. Groundwater flow at elevations lower than the lake bottom was assumed to have no interaction with the lake.

3.4.1.2. Hydraulic conductivity. Hydraulic conductivity, K, was estimated from slug tests done on piezometer T1.1 and pump tests done on piezometers T1.1, T2.2 and K1.2. Slug test data were analyzed using the Hvorslev (1951) method. Pump test data were analyzed using the methods of Theis (1935), Theis Recovery (Theis, 1935), Cooper-Jacob Time-Drawdown (Cooper and Jacob, 1946), and Neuman (1975). Analyses were done using the software AquiferTest©.

The saturated thickness, b, of the aquifer was assumed to be the vertical distance at the time of the aquifer test between the average water table elevation and the elevation of the lake bottom. At TUY and KHO, b was estimated to be 2.5 and 1.4 m, respectively.

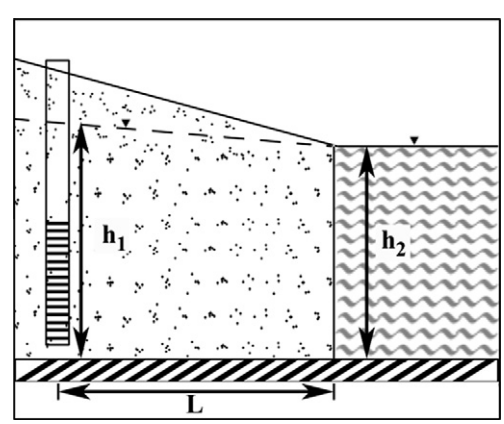

Fig. 3. Diagram showing conceptualized aquifer and lake, with components h_1, h_2, and L from Eq. (2). Aquifer and lake are resting on a horizontal impervious surface.

Hydraulic conductivity values resulting from all analyses ranged from 0.8 to 11 and 3 to 15 m/day, at TUY and KHO, respectively.

The values of K reported here fall within the range of a sandy porous media and were assumed to be representative of the aquifer surrounding the lake. This assumption was supported by the fact that during the construction of piezometers at TUY and KHO, aquifer materials were found to be visually and texturally similar in all bore holes. In Eq. (2), for TUY, K was assigned the value of the average of all T1.1 and T2.2 aquifer test estimates of K (5 m/d). For KHO, K was assigned the value of the average of all K1.2 aquifer test estimates of K (10 m/d).

3.4.2. Evaporation

Evaporation flux was estimated from climate station and Class A pan data. Temperature, precipitation, net radiation, solar radiation, relative humidity, and wind speed data were collected from two climate stations located in Yangibazar and Urgench (Fig. 1). The Yangibazar station was located 35 km from both TUY and KHO, while the Urgench station was located 26 and 39 km from TUY and KHO, respectively. These data were collected at 30 min intervals and covered June through December, 2008. To estimate the evaporative losses over the lakes of study, these data were first input to the Penman–Monteith equation for green grass cover (Finch and Hall, 2005; Allen et al., 1998). The Penman–Monteith equation produces estimates of evapotranspiration for a reference crop (grass), ET_o. To convert ET_o values into free water surface evaporation (e.g. lake evaporation), ET_o values were multiplied by 1.05, a factor empirically derived for shallow lakes up to 3 m in depth (Allen et al., 1998).

In the evaporation pan at TUY, daily change in water level in the pan was recorded using a micrometer, and the measurements were converted into ET_o values using a correction factor, k_p, which accounts for difference between pan evaporation and evapotranspiration over a green crop. This correction factor was calculated using Eq. (3) (Allen et al., 1998):

$$k_p = 0.108 - 0.0286u_2 + 0.422\ln(FET) + 0.1434\ln(RH_m) - 0.000631[\ln(FET)]2\ln(RH_m) \quad (3)$$

where u_2 is average daily wind speed (m/s), FET is green fetch, or the distance over which wind blows across a vegetated landscape before arriving at the pan (m), and RH_m is mean daily relative humidity. Wind and humidity data from the Yangibazar and Urgench weather stations were used here. Fetch was estimated to be 400 m based on field observations and Google Earth imagery. Monthly correction factors were calculated and applied to measured evaporation data (Winter, 1981). These resulting ET_o values were then multiplied by the shallow lake factor of 1.05. For both lakes, final estimates of evaporation used in Eq. (1) were average evaporation values calculated from the evaporation pan and Penman–Monteith methods.

3.4.3. Change in lake storage and surface water

Changes in lake storage were estimated by multiplying the difference between lake elevations at the end and the beginning of the time interval of interest by the surface area of the lake. Due to logistical issues, it was not possible to quantitatively measure surface inflows (I), so when each of the components above were accounted for in Eq. (1), the residual term included total flux uncertainty ($e\Delta_S + e_G + e_E$) and I.

4. Results

4.1. Tuyrek: water table elevations and stable isotopes

Irrigation water arrived via canal at TUY on June 15th, 2008. Two rice fields located about 10 m from piezometers on the north side of the lake were flooded about 1 day later. On the south side of the lake, the rice field located about 10 m from piezometer T2.1 was flooded on or around June 26th, 2008. Following the initial arrival of irrigation water, the local water table elevation was observed to increase (Fig. 4A). The water table elevation data suggest that between June 9th and July 23rd the lake was either gaining or both gaining and losing relative to the water table. Two large increases in lake stage beginning around June 26th and July 21st, 2008 corresponded to the partial draining into the lake of flooded fields via small hand-dug ditches. These two increases in lake stage were also coincident to relatively large decreases in lake stable isotope values (Fig. 4B and C), which is explained by the fact that the mean δ^2H of sampled irrigation water was significantly less than mean δ^2H of lake water ($p < 0.05$; Fig. 5A).

As a result of these inputs and beginning around July 23rd, 2008, lake stage became elevated above water table elevations on both sides of the lake, suggesting the lake was losing to the water table. This observation is supported by the stable isotope data: prior to July 21st, stable isotope values from piezometer T2.2 samples were fairly constant; however, following the onset of these losing conditions, the July 29th water sample from T2.2 reflected an increase in stable isotope value and approached the value of lake water (Fig. 4B and C). Note that only data from piezometer T2.2 reflects this reversal in groundwater flow because T1.1 was not sampled after July 21st and T2.3 was installed at such a distance from the lake edge that infiltrating lake water likely had not entered the piezometer at the time of sampling.

Surface water inputs apparently drove the reversal of the lake from a gaining to a losing situation. These inputs outweighed losses to evaporation despite the arid summer environment, as TUY lake stage was higher and lake stable isotope values lower at the end of July than in early June (Fig. 4A–C).

4.2. Khodjababa: water table elevations and stable isotopes

In contrast to TUY, no surface water inputs or outputs were observed at KHO during the sampling period. The fields immediately adjacent to KHO did not receive irrigation water throughout June or July 2008. However, water present in a small canal on the north side of the lake was sampled (Fig. 2B), and the fact that the mean δ^2H of this water was not found to be statistically different from mean piezometer groundwater δ^2H ($p = 0.746$; Fig. 5B) was taken to reflect the infiltration of irrigation water.

Infiltration was likely responsible for maintaining an elevated water table, which allowed for input to the lake between June 23rd and July 19th, 2008 (Fig. 6A). During that time, lake stage remained relatively constant or declined slightly, indicating that evaporation was nearly balanced by groundwater input. However, the water table elevation decreased and approached lake surface elevation around July 23rd, presumably following the cessation of infiltration of surface water. Evaporation was likely responsible for the relatively constant lake stage decrease that followed (Fig. 6A).

Despite input to KHO of low δ^2H groundwater, evaporation caused lake water δ^2H and δ^{18}O to increase throughout the sampling period (Fig. 6B and C). With little or no nearby irrigation and no excess surface water input to the lake, evaporation was apparently the dominant component of KHOs water budget.

4.3. Water budget

After G, E, and ΔS were accounted for, Eq. (1) was solved for the net residual term, which included the surface water flux, I, and the total component uncertainty. However, at KHO, where no surface water was observed to enter or exit the lake, the net residual included only the uncertainty term.

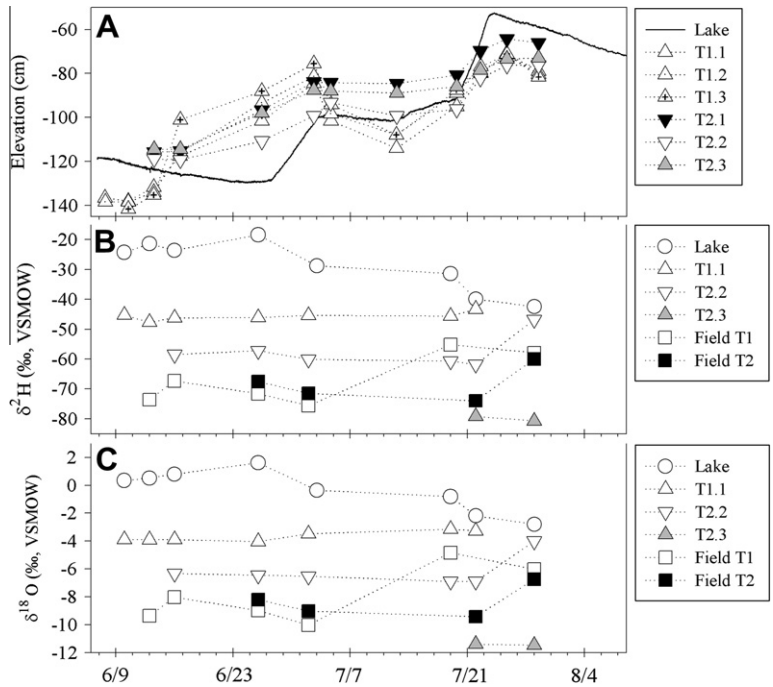

Fig. 4. (A) TUY and water table elevations with respect to a local datum. (B) δ^2H and (C) δ^{18}O of surface, lake, and ground water for the summer of 2008 at TUY.

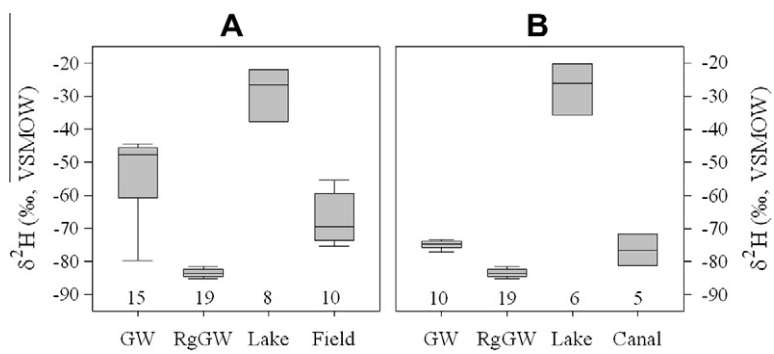

Fig. 5. δ^2H data from TUY (A) and KHO (B) grouped by piezometers (GW), regional ground water (RgGW), lake water, and surface waters. Numbers below boxes indicate sample size.

The water budgets indicated that groundwater contribution to both lakes was relatively small and evaporation relatively large (Fig. 7A and B). Evaporation dominated KHOs relatively simple water budget during the sampling period. Evaporation at both lakes was estimated to be approximately 0.8, 0.7, and 0.7 cm/day during June, July, and August, respectively. An uncertainty of ±30% may be reasonable for these data (Winter, 1981). For irrigated cotton fields in Khorezm, Conrad et al. (2007) reported cotton-specific reference ET rates to range from 0.3 to 0.7 cm/day from early June to late July, respectively, indicating that our evaporation rates are reasonable.

4.4. Stable isotopes and major ion hydrochemistry

When δ^2H and δ^{18}O data for water samples taken throughout the summer at both lake systems and throughout Khorezm were plotted against each other, a linear deviation from the Global Meteoric Water Line (GMWL) at a slope of about 5 was observed (Fig. 8A and B). This evaporative trend line suggests that all the sampled water bodies share a common original source (Clark and Fritz, 1999). The AD water sample lies directly on the GMWL, nearly at the base of the evaporative trend line, suggesting that the AD was the source water for all the water bodies sampled. If numerous AD water samples were taken over time, it is anticipated that their average isotope chemistry would plot very close to the intersection of the evaporative trend line and the GMWL. Marimuthu et al. (2005) similarly determined that δ^2H and δ^{18}O data of water samples collected at a study site in Western Australia defined an evaporative trend line where groundwater recharge was the recharge source for their study lake.

Groundwater collected from open community wells appeared to be relatively unevaporated AD water. The similarity in stable isotope composition of these samples and the fact that they were collected at various locations around Khorezm indicates the presence of regional groundwater (RgGw) that is recharged by AD water (Ibrakhimov et al., 2007). Major ion analyses of these samples further reflect AD recharge, as both RgGw and AD water samples had similar relative proportions of the major ions, indicating a common hydrochemical source (Fig. 9A and B). See Table A1 in the Appendix section for hydrochemistry data.

The observed stable isotope chemistry of piezometer water samples also can be explained within the conceptual framework where the lakes interact with a regional groundwater system recharged by the AD. The stable isotope composition of samples from

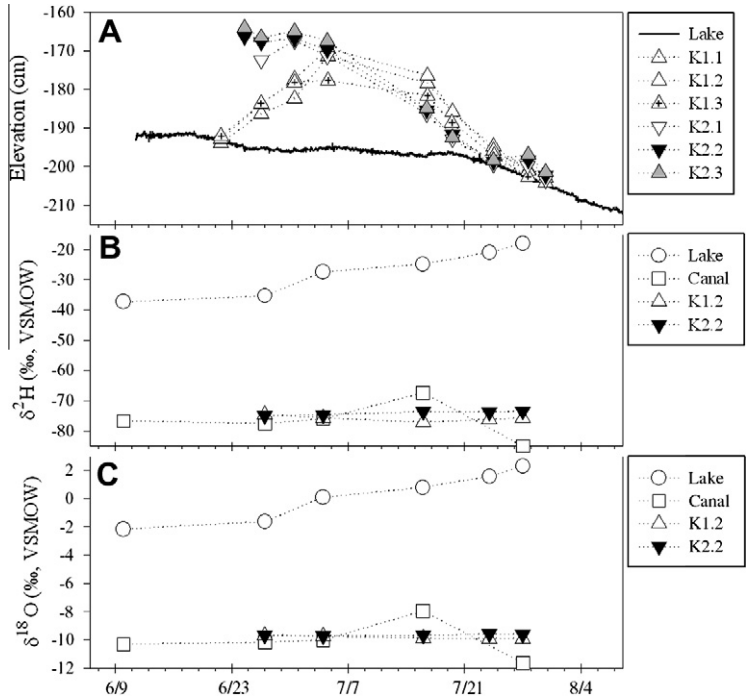

Fig. 6. (A) KHO and water table elevations with respect to a local datum. (B) δ^2H and (C) δ^{18}O of surface, lake, and ground water for the summer of 2008 at KHO.

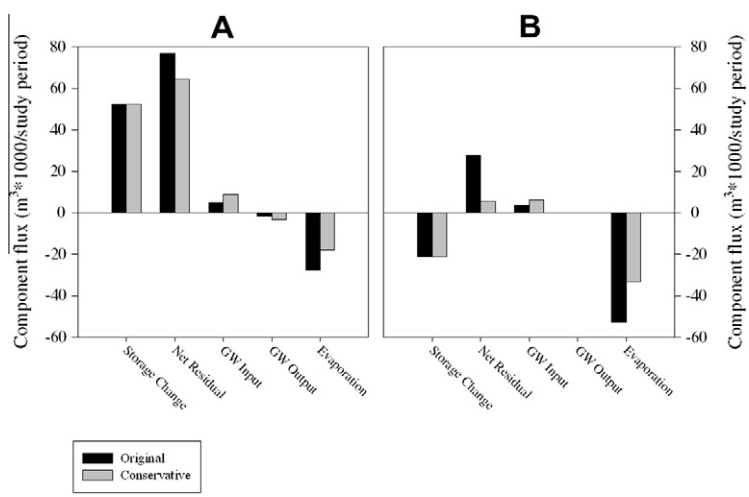

Fig. 7. Calculated water budget components for TUY (A) and KHO (B) during intervals June 13th to July 29th and June 26th to July 30th, 2008, respectively. Original water budget results are compared to the results of the conservative water budget (Section 5.3).

piezometers adjacent to the lakes suggests mixing between the lakes and the RgGw, to varying extents. For example, at TUY, samples from near-shore piezometers T1.1 and T2.2 had stable isotope values more similar to lake water than to RgGw, while water from piezometer T2.3, about 40 m distant, was isotopically similar to RgGw (Fig. 8A). At KHO, the stable isotope data suggest less mixing between the RgGw and the lakes than at TUY. This observation is supported by the water table elevation data, which indicated that KHO was gaining throughout the study period, while TUY was both losing and gaining (Sections 4.1 and 4.2).

Inverse major ion hydrochemical modeling supported the conclusion that the AD was the source of water in the study lakes. Model results indicated that it was possible to create KHO and TUY lake waters by evaporating approximately 5 and 4 L of AD water down to 1 L, respectively. According to the models, calcite was precipitating along the flow path from the AD to the lakes. This scenario makes sense, as both TUY and KHO lake waters were generally over saturated with respect to calcite. The models suggest that gypsum remained approximately saturated, or slightly undersaturated, along the flow path. This agrees with calculations of lake water saturation indices, which ranged from −1.8 to 0.8 for gypsum. Concentrations averaged across different seasons and years were used for all aqueous solutions in these models, thus the degrees of concentration mentioned above are not applicable to any specific water sampled, but the results do provide evidence for the existence of conditions that allow for AD water to chemically evolve through evaporation into lake water. These model results, taken together with stable isotope observations, provide two lines of independent evidence indicating that waters in the lakes of study were predominantly made up of evaporated AD water.

Finally, it was observed that both RgGw and piezometer T2.3 stable isotope values plotted on the GMWL (Fig. 8A). One may

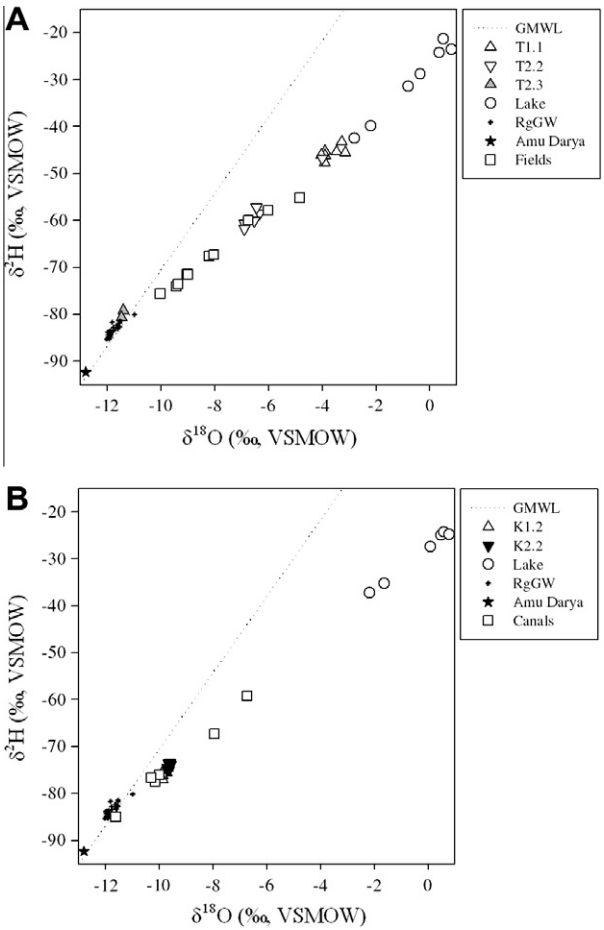

Fig. 8. Stable isotope data from regional groundwater (RgGw), Amu Darya, lakes, piezometers, and surface waters for (A) TUY and (B) KHO.

argue that these data are ambiguous in that RgGw may actually be a chemically unique original source of the water for the lakes that happens to be un-enriched. This argument does not stand when Khorezm irrigation history and major ion hydrochemistry of these waters are considered. Ibrakhimov et al. (2007) reported that after the onset of large scale irrigation with AD water in Khorezm, the average yearly depth to groundwater decreased from 15 to 3 m or less today. The proportion of major ions in AD water and RgGw sampled in Khorezm were found to be similar (Fig. 9A and B), and this supports the conclusion that AD was the original source of sampled groundwater. Taken together, a historical increase in groundwater elevation attributed to irrigation with AD water and the similarity between AD and RgGw in major ion hydrochemistry and stable isotope composition indicate that the AD, not a chemically unique regional groundwater, is the major source of water for the lakes of study.

4.5. Seasonal variability

Most of the data collected and analyzed here focused on the summer period of early June to the end of July, 2008. More data collected throughout the year would allow seasonal and yearly variability regarding the influence of irrigation water on lake hydrology to be described. To this end, additional lake stage elevation data were collected by project partners between August 1st and December 25th, 2008 at TUY and between August 1st and October 2nd, 2008 at KHO.

At TUY, large increases in stage on August 20th, 2008 and again on December 18th, 2008 provide evidence of additional irrigation water input (Fig. 10A). The latter input may be from irrigation of winter wheat, which is a known crop in the Khorezm region (Ibrakhimov et al., 2007).

At KHO, lake stage decreased during the late summer and autumn months (Fig. 10B). Visual observation indicated that by October 2nd, 2008, lake stage had decreased to the point that the pressure transducer was exposed. Observations and photos of the lake from this time indicate that lake surface area had decreased by about 50% relative to the early summer.

5. Discussion

5.1. Estimating surface water input to TUY

The evidence from TUY and KHO supports the hypothesis that these lakes depend on excess irrigation water to maintain volume. It can therefore be assumed that a basic need for a fishery and/or a tourism operation at lakes in Khorezm would be to maintain lake volume throughout the arid summer. For a lake of a similar volume, one would expect that satisfying this basic need would require input of water at least similar to the volume input to TUY during June and July, 2008.

Surface water input to TUY can be estimated by first assuming that total component uncertainty was similar at each lake. At KHO, we know that the total component uncertainty was 28,000 m^3 (residual term, Fig. 7B) because surface water input did not occur. This value can be expressed as about 35% of the absolute sum of all measured components at KHO. Using this estimate of total component uncertainty and the assumption that total component uncertainty was similar at each lake, the volume of water within the TUY residual term attributed to uncertainty can be estimated as 35% of all measured TUY components, or about 30,000 m^3. The remainder of the TUY residual term can thus be attributed to surface water input. This value is about 46,000 m^3.

5.2. Water budget uncertainty

Any water budget study must extract useful information and consider uncertainty. The uncertainty associated with each component in Eq. (1) result from (1) imperfect sampling and measurement techniques and (2) interpretation of the data (Winter, 1981). Water budget component uncertainties were assumed to be roughly similar at both lakes. Evaporation uncertainty was introduced from a variety of sources, some of which included (1) using off-site weather station data with the Penman–Monteith equation, (2) using a Class A pan without correcting for advected heat into the lake or energy storage in the pan, and (3) errors in reading and recording evaporation pan measurements (Winter, 1981). Some possible sources of uncertainty in the estimation of groundwater flow into and out of the lake include unaccounted for aquifer heterogeneities, possible lake bottom fine-sediment deposition and resulting decrease in lake bottom hydraulic conductivity, and vertical flow to the lake. Uncertainty in the change in lake storage component resulted primarily from the assumption that lake surface area remained constant at all measured stage.

5.3. The conservative water budget

Both the stable isotope data and the water budget results suggested that without surface water input, evaporation was the dominant water budget component of these lakes. To test the validity these results, a conservative water budget was calculated using component estimates that minimized evaporation losses and maximized groundwater flow. To this end, K and evaporation values from within the range of data collected in the field were chosen so that the water budget would reflect the highest groundwater

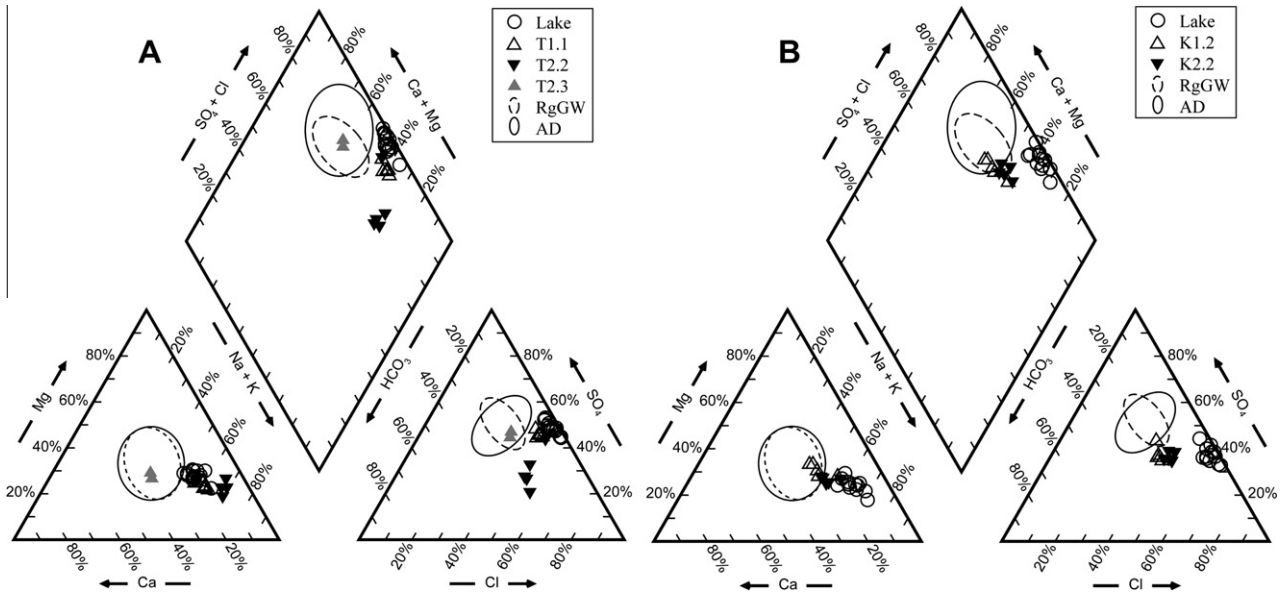

Fig. 9. Piper plots of major ion chemistry of water samples from (A) TUY and (B) KHO. Note that the range of Amu Darya (AD) and regional groundwater (RgGW) data are represented as ovals.

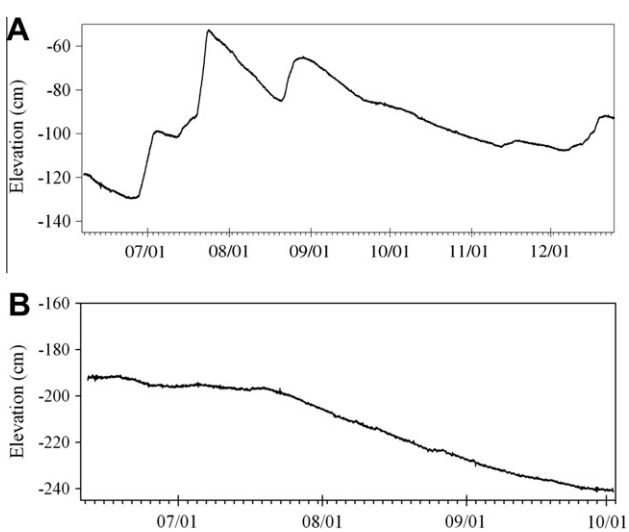

Fig. 10. Extended lake elevation data with respect to a local datum from (A) TUY and (B) KHO.

input and lowest evaporation output, within a reasonable range. For example, the evaporation component in the original water budget was calculated by averaging the evaporation pan and Penman–Monteith method estimates, whereas only the smaller of the two estimates was used for the conservative calculations. In this conservative water budget, at TUY, where surface water input did occur, total inputs approached 400% of evaporative loss (Fig. 7A). More importantly, at KHO, groundwater inputs were found to still be just a fraction of the evaporative loss (Fig. 7B). This supports the validity of the original water budget results: evaporation losses greatly exceed groundwater inputs when surface water input does not occur during the summer months. Further support of this observation lies in the fact that by October 2nd, 2008, KHO had lost much of its volume and decreased in surface area by about 50%, while TUY, which differed in that it had received surface water input throughout the summer, maintained its volume through the fall and into the winter (Fig. 10A and B).

In general, the total uncertainty in the lake water budgets is significant, and extending the temporal scope of these budgets would improve our understanding. However, taken along with this study as a whole, these lake budgets begin to demonstrate the importance of evaporation and surface water input relative to groundwater flux at KHO, TUY, and other similar Khorezm lakes.

6. Conclusion

Stable isotope and major ion hydrochemistry data analyses confirmed that the AD was the source of near-lake groundwater and lake water. Stable isotope data from these waters formed an evaporative trend line with AD water at its base. Major ion hydrochemical modeling demonstrated how AD water could evolve into lake water through evaporation and precipitation of calcite, dissolution of halite, and mostly dissolution of gypsum along the flow path from river to lake. These conclusions are consistent with the regional framework in which the presence of agricultural practices and inefficiencies result in AD irrigation water entering the lakes periodically as direct surface flow, and to a lesser extent, as groundwater flow.

The water budget and the stable isotope chemistry data indicate that, in Khorezm, evaporation can be the dominant lake water budget component when surface water is not an input to the lake. Under both relatively natural (KHO) and irrigation-influenced (TUY) conditions, the amount of groundwater transmitted to these lakes in the summer months was relatively minor when compared to evaporative losses. Beginning on June 10th, 2008 at the latest, virtually no irrigation water was input to KHO via surface inflows and the lake was nearly dry by October. If summer conditions and lake hydrology are similar from year to year, as they are expected to be, input of water to the lakes periodically during the growing season may be necessary to prevent the lakes from desiccating.

This study therefore suggests that many shallow lakes in Khorezm, like KHO and TUY, are dependent upon surface input of water, whether that input comes from excess irrigation or water otherwise allocated for lake input. Without such input, KHO, and possibly TUY, would face desiccation from the arid climate. Thus, the origin of these lakes, as they are known today in expanse and depth, is likely coincident to the onset of large scale irrigation in the region. Prior to onset, it is possible but unlikely that they

Table A1
Hydrochemistry data from lakes, surface water, the AD, and regional and local groundwater.

Date	ID (type)	pH	Cl^- (mg/L)	SO_4^{2-} (mg/L)	HCO_3^- (mg/L)	Na^+ (mg/L)	K^+ (mg/L)	Ca^{2+} (mg/L)	Mg^{2+} (mg/L)	$\delta^{18}O$ (‰)	δ^2H (‰)
08/23/2007	KHO (lake)	7.78	876	764	169	571	8.0	140	148.9	–	–
09/25/2007	KHO (lake)	–	1115	789	142	703	9.2	285	167.2	–	–
10/26/2007	KHO (lake)	7.70	1287	1243	129	968	11.0	135	203.7	–	–
11/30/2007	KHO (lake)	7.70	1383	1611	171	1253	10.0	170	173.3	–	–
01/24/2008	KHO (lake)	7.70	1801	1414	224	1268	14.0	220	237.1	–	–
02/21/2008	KHO (lake)	7.49	123	103	20	77	3.6	16	21.9	–	–
03/26/2008	KHO (lake)	–	1191	1112	208	905	14.1	188	165.3	−8.6	−68.7
04/25/2008	KHO (lake)	7.70	1311	1125	194	917	9.4	195	185.4	–	–
06/10/2008	KHO (lake)	8.06	1722	1546	88	1207	8.9	205	258.0	−2.2	−37.3
06/27/2008	KHO (lake)	7.70	1843	1299	107	1149	8.8	241	249.3	−1.6	−35.3
07/4/2008	KHO (lake)	7.81	2020	1379	78	1223	11.0	251	279.7	0.1	−27.4
07/16/2008	KHO (lake)	7.13	2233	1833	79	1536	10.0	241	316.2	0.8	−24.8
07/24/2008	KHO (lake)	8.48	2566	2525	76	2022	9.9	261	346.6	1.6	−21.0
07/28/2008	KHO (lake)	6.68	2765	2424	61	1255	11.0	371	288.8	2.3	−17.5
06/07/2006	TUY (lake)	–	1086	1500	149	852	20.3	241	194.6	–	–
07/6/2006	TUY (lake)	–	1183	1447	159	873	12.8	261	203.7	–	–
08/06/2006	TUY (lake)	–	1127	1391	181	695	14.1	321	255.4	–	–
09/06/2006	TUY (lake)	–	1361	1460	281	986	20.1	246	234.1	–	–
10/06/2006	TUY (lake)	–	1751	1949	262	923	18.8	481	376.9	–	–
11/06/2006	TUY (lake)	–	1559	2104	186	1062	20.8	376	328.3	–	–
12/06/2006	TUY (lake)	–	1645	2012	188	1036	23.9	411	325.3	–	–
02/07/2007	TUY (lake)	–	1312	1966	125	858	31.3	401	279.6	–	–
03/07/2007	TUY (lake)	–	1383	1854	130	851	31.1	356	307.1	–	–
04/07/2007	TUY (lake)	–	1368	1854	125	774	24.5	401	316.2	–	–
05/07/2007	TUY (lake)	–	1524	1822	166	876	27.9	411	310	–	–
06/07/2007	TUY (lake)	–	1336	1427	151	849	31.7	288	264.6	–	–
06/11/2007	TUY (lake)	–	–	–	–	–	–	–	–	−2.7	−39.0
08/30/2007	TUY (lake)	8.03	1010	1631	157	818	16	256	212.8	–	–
09/24/2007	TUY (lake)	7.72	1176	1710	166	888	18	276	242.9	–	–
11/2/2007	TUY (lake)	7.72	1176	1772	131	916	19	271	240.2	–	–
11/29/2007	TUY (lake)	7.72	1019	1664	139	869	20.0	230	220.4	–	–
01/23/2008	TUY (lake)	7.72	1425	2346	162	1353	26	266	255.4	–	–
02/21/2008	TUY (lake)	7.76	326	464	52	253	10	65	66.9	–	–
03/27/2008	TUY (lake)	7.72	1064	1425	137	763	28	230	212.8	−4.1	−47.0
05/01/2008	TUY (lake)	7.56	1127	1478	102	779	13	226	243.2	–	–
05/29/2008	TUY (lake)	7.72	1247	1603	78	928	12	185	262.0	–	–
06/10/2008	TUY (lake)	8.68	1333	1718	55	935	12	235	279.0	0.3	−24.3
06/13/2008	TUY (lake)	8.68	1382	1826	50	953	14	270	292.0	0.5	−21.4
06/16/2008	TUY (lake)	8.68	1467	1671	64	921	13	321	270.6	0.8	−23.6
06/26/2008	TUY (lake)	7.72	1531	1762	81	957	13	331	294.8	1.6	−18.4
07/03/2008	TUY (lake)	7.47	1113	1253	49	667	11	251	212.8	−0.4	−28.8
07/19/2008	TUY (lake)	–	–	–	–	–	–	–	–	−0.8	−31.4
07/22/2008	TUY (lake)	7.82	1382	1980	162	1052	12	311	279.7	−2.2	−39.9
07/29/2008	TUY (lake)	7.00	1488	2054	174	1153	12	276	307.0	−2.8	−42.5
06/10/2008	T1.1 (GW[a])	6.60	1354	1800	418	1173	9.1	281	234.1	−3.9	−45.3
06/13/2008	T1.1 (GW)	6.80	1234	1740	433	1103	9.2	255	231.0	−3.9	−47.7
06/16/2008	T1.1 (GW)	6.51	1361	2096	494	1315	7.4	291	249.3	−3.9	−46.2
06/26/2008	T1.1 (GW)	6.60	1283	1753	522	1194	11.0	276	275.6	−4.0	−46.1
07/02/2008	T1.1 (GW)	6.61	1305	1791	453	1149	14.0	276	234.1	−3.5	−45.4
07/19/2008	T1.1 (GW)	–	–	–	–	–	–	–	–	−3.2	−45.6
07/22/2008	T1.1 (GW)	6.61	1382	2154	494	1297	11.0	326	258.4	−3.3	−43.3
07/29/2008	T1.1 (GW)	6.46	1368	1892	422	1109	10.0	331	264.5	–	–
06/16/2008	T2.2 (GW)	6.61	995	936	678	946	11.0	95	170.2	−6.3	−58.5
06/26/2008	T2.2 (GW)	6.59	1064	840	897	956	17.0	140	140.3	−6.5	−57.2
07/2/2008	T2.2 (GW)	6.56	1113	808	883	973	21.0	83	208.5	−6.5	−60.1
07/19/2008	T2.2 (GW)	–	–	–	–	–	–	–	–	−6.9	−60.7
07/22/2008	T2.2 (GW)	6.70	1092	587	871	895	14.0	120	164.2	−6.9	−61.8
07/29/2008	T2.2 (GW)	6.49	2311	2857	636	1750	15.0	501	443.8	−4.0	−46.8
07/22/2008	T2.3 (GW)	6.70	161	307	157	111	10.0	90	48.6	−11.4	−79.2
07/29/2008	T2.3 (GW)	7.21	161	330	149	109	8.8	90	53.5	−11.5	−80.7
06/27/2008	K1.1 (GW)	6.53	706	1159	674	697	6.4	190	190.4	−9.7	−74.6
07/4/2008	K1.2 (GW)	6.30	752	946	721	585	7.5	259	183.6	−9.7	−75.6
07/16/2008	K1.2 (GW)	6.31	702	854	665	481	6.7	236	200.6	−9.9	−76.7
07/24/2008	K1.2 (GW)	7.03	667	827	674	452	5.5	241	194.6	−9.9	−76.1
07/28/2008	K1.2 (GW)	6.48	730	809	659	516	6.2	231	182.4	−9.9	−75.7
06/27/2008	K2.2 (GW)	6.61	590	732	419	511	8.2	132	132.3	−9.6	−74.9
07/4/2008	K2.2 (GW)	6.39	681	795	403	514	10.0	184	137.4	−9.7	−74.6
07/16/2008	K2.2 (GW)	6.39	624	692	468	478	7.7	185	124.6	−9.7	−73.6
07/24/2008	K2.2 (GW)	6.89	589	672	499	444	6.5	180	133.8	−9.6	−73.8
07/28/2008	K2.2 (GW)	6.76	652	675	448	443	6.9	190	139.8	−9.6	−73.4
06/11/2007	TUY canal (SW[b])	–	–	–	–	–	–	–	–	−10.8	−78.9
03/27/2008	TUY inflow (SW)	–	315	780	216	276	12.5	175	97.3	−11.1	−80.3
03/27/2008	TUY collector (SW)	–	865	934	277	571	18.0	235	148.9	−10.9	−80.3
06/13/2008	TUY field (SW)	–	499	901	228	441	6.5	152	125.2	−9.4	−73.6

(*continued on next page*)

Table A1 (*continued*)

Date	ID (type)	pH	Cl^- (mg/L)	SO_4^{2-} (mg/L)	HCO_3^- (mg/L)	Na^+ (mg/L)	K^+ (mg/L)	Ca^{2+} (mg/L)	Mg^{2+} (mg/L)	$\delta^{18}O$ (‰)	δ^2H (‰)
06/16/2008	TUY field (SW)	–	422	985	181	299	9.1	233	132.5	−8.0	−67.3
06/26/2008	T1.1 field (SW)	–	603	860	261	486	10.0	172	172.3	−9.0	−71.6
06/26/2008	T2.1 field (SW)	–	822	1041	248	626	14.0	196	196.4	−8.2	−67.6
07/02/2008	T1.3 field (SW)	–	173	388	104	151	7.8	80	48.6	−10.0	−75.6
07/02/2008	T2.1 field (SW)	–	716	1219	219	638	13.0	206	142.3	−9.0	−71.5
07/19/2008	T1.3 field (SW)	–	–	–	–	–	–	–	–	−4.8	−55.2
07/22/2008	T2.1 field (SW)	–	667	640	198	355	8.3	180	136.8	−9.4	−74.0
07/29/2008	T1.2 field (SW)	–	1333	1638	104	1021	8.4	251	218.9	−6.0	−57.9
07/29/2008	T2.1 field (SW)	–	1093	1517	139	899	8.3	220	194.6	−6.8	−60.0
06/10/2008	KHO inflow 1 (SW)	–	1247	1315	343	1017	6.4	210	182.4	−6.7	−59.2
06/10/2008	KHO inflow 2 (SW)	–	292	703	259	273	4.6	188	74.2	−10.3	−76.6
06/27/2008	KHO canal (SW)	–	709	1613	234	670	5.1	326	325.6	−10.2	−77.5
07/04/2008	KHO canal (SW)	–	1368	1890	314	1129	8.8	406	188.5	−10.0	−76.0
07/16/2008	KHO canal (SW)	–	1098	1764	285	938	7.8	366	179.4	−8.0	−67.3
07/28/2008	KHO canal (SW)	–	822	543	166	510	4.8	110	124.6	−11.6	−84.9
06/07/2006	AD (river)	–	87	184	185	57	3.9	80	34.1	–	–
06/07/2006	AD (river)	–	106	276	137	117	5.4	72	28.0	–	–
06/07/2006	AD (river)	–	84	221	119	89	5.1	55	27.4	–	–
07/06/2006	AD (river)	–	83	178	134	62	4.3	66	26.8	–	–
07/06/2006	AD (river)	–	111	235	130	86	4.1	78	30.4	–	–
07/06/2006	AD (river)	–	247	352	127	73	4.5	100	59.6	–	–
08/06/2006	AD (river)	–	103	179	148	59	2.8	80	30.4	–	–
08/06/2006	AD (river)	–	96	181	156	60	3.6	84	26.8	–	–
09/06/2006	AD (river)	–	135	288	129	106	9.3	84	35.3	–	–
09/06/2006	AD (river)	–	135	267	124	95	5.8	84	36.5	–	–
10/06/2006	AD (river)	–	174	324	153	116	7.9	106	44.9	–	–
10/06/2006	AD (river)	–	162	347	146	114	7.8	100	49.8	–	–
10/06/2006	AD (river)	–	193	376	146	122	10.7	106	58.4	–	–
10/06/2006	AD (river)	–	170	368	162	120	7.3	104	55.9	–	–
11/06/2006	AD (river)	–	147	285	155	99	6.8	94	42.6	–	–
11/26/06	AD (river)	–	177	334	136	109	5.7	106	49.9	–	–
11/06/2006	AD (river)	–	224	373	139	117	11.9	110	66.9	–	–
11/06/2006	AD (river)	–	174	342	131	107	5.7	104	52.3	–	–
12/06/2006	AD (river)	–	166	326	160	102	8.7	116	45.0	–	–
12/06/2006	AD (river)	–	162	320	159	97	7.1	108	49.9	–	–
02/07/2007	AD (river)	–	186	357	140	110	6.0	74	79.1	–	–
02/07/2007	AD (river)	–	196	432	134	130	15.5	102	68.2	–	–
03/07/2007	AD (river)	–	244	493	168	200	10.7	96	83.4	–	–
03/07/2007	AD (river)	–	269	636	146	211	12.6	136	92.4	–	–
04/07/2007	AD (river)	–	269	448	125	178	9.7	114	66.9	–	–
04/07/2007	AD (river)	–	184	421	126	139	12.1	104	55.9	–	–
05/07/2007	AD (river)	–	123	289	100	90	7.9	82	36.5	–	–
05/07/2007	AD (river)	–	124	313	106	119	4.8	78	32.8	–	–
08/01/2008	AD (river)	–	143	254	105	95	3.9	66	43.8	−12.8	−92.3
06/11/2007	TUY well (GW)[c]	–	–	–	–	–	–	–	–	−11.8	−82.8
03/27/2008	TUY well (GW)[c]	–	176	389	250	161	6.4	136	41.3	−11.8	−81.7
03/27/2008	TUY well (GW)[c]	–	269	419	262	183	8.8	150	65.4	−11.9	−84.2
1-July/08	KHO well 1 (GW)[c]	6.19	461	619	384	342	11.0	170	109.4	−11.9	−83.6
03/26/2008	KHO well 2 (GW)[c]	7.12	198	364	157	128	9.6	112	53.5	−11.0	−80.1
07/02/2008	E1 (GW)[c]	7.05	163	447	145	183	10.0	98	40.1	−11.5	−81.4
07/23/2008	E1 (GW)[c]	7.38	165	353	138	131	7.0	122	37.7	−11.6	−82.9
07/31/2008	E1 (GW)[c]	7.18	128	341	149	91	5.3	118	40.1	−11.5	−81.8
03/28/2008	ESH well (GW)[c]	7.05	574	842	441	468	18.9	225	121.6	−11.5	−82.7
05/30/2007	ESH well (GW)[c]	–	–	–	–	–	–	–	–	−11.6	−83.0
03/25/2008	SHU well 1 (GW)[c]	7.01	674	1478	355	570	12.8	361	170.3	−11.9	−85.3
03/25/08	SHU well 2 (GW)[c]	7.31	173	380	284	155	7.4	130	49.8	−11.9	−84.5
03/25/08	SHU well 3 (GW)[c]	7.47	199	373	268	152	6.6	126	59.6	−11.9	−84.4
06/05/2008	SHU well 2 (GW)[c]	7.01	184	421	199	188	4.2	106	47.4	−11.9	−84.6
06/12/2008	SHU well 2 (GW)[c]	6.43	192	364	258	169	3.9	110	54.9	−11.8	−83.4
06/28/2008	SHU well 2 (GW)[c]	7.01	101	411	226	137	4.1	112	112.2	−12.0	−83.8
07/17/2008	SHU well 2 (GW)[c]	7.07	184	322	195	131	4.0	118	43.8	−12.0	−85.3
07/24/2008	SHU well 2 (GW)[c]	7.39	181	321	221	128	3.7	120	47.4	−12.0	−84.3
07/28/2008	SHU well 2 (GW)[c]	6.46	179	306	218	119	3.4	120	47.4	−12.0	−84.8

[a] GW: groundwater.
[b] SW: surface water.
[c] Regional groundwater (RgGw).

existed as ephemeral features of the landscape. It seems that the fate of Khorezm lakes may be inextricably tied to the fate of the Aral Sea: if the volume and expanse of these lakes of study are to be maintained at their current levels, it will require continued periodic input of Amu Darya water, and will therefore come to some degree at the expense of the Aral Sea.

7. Future Work

The water budgets estimated here would benefit from future work focused on decreasing the error associated with budget components. Measuring lake surface area from a satellite image taken during the period of study and surveying lake bathymetry would

improve volumetric calculations. Two alternative data collection techniques that would further improve estimation of the water budget are using a floating weather station on the lake to improve on-site evaporative loss estimates (Winter, 1981) and installing more piezometers to increase the knowledge of near-shore aquifer heterogeneities and improve description of groundwater-lake interaction. Further, measuring the vertical component of groundwater flow would allow for increased accuracy when describing groundwater-lake interaction, particularly with regards to lake seepage (Winter and Pfannkuch, 1984). Both of these approaches would require serious consideration of issues of potential vandalism, theft, and access to equipment left in the field over an extended period of time.

The water budget indicated, and collaborative stable isotope evidence independently supported, that during the period of study, evaporation was the dominant water budget component when periodic inflow of surface water was not present, as was the case at KHO. As templates for further studies of the water budget of lakes in this region, this study and the methods applied offer great potential. Extending the period of sampling to cover the entire year and returning to the lakes during years of relative drought and plenty would allow seasonal and annual variability in lake water budgets to be assessed.

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Appendix A

See Table A1.

References

Allen, R.G., Pereira, L.C., Raes, D., Smith, M., 1998. Crop Evapotranspiration. Guidelines for Computing Crop Water Requirements. FAO Irrigation and Drainage Paper 56. FAO, Rome, p. 300.

AquiferTest v3.5, Waterloo Hydrogeologic software package.

Clark, Ian D., Fritz, Peter, 1999. Environmental Isotopes in Hydrogeology, CRC Press, New York.

Conrad, Christopher, Dech, Stefan W., Hafeez, Mohsin, Lamers, John, Martius, Christopher, Strunz, Gunter, 2007. Mapping and assessing water use in a Central Asian irrigation system by utilizing MODIS remote sensing products, Irrig. Drain. Syst. doi: 10.1007/s10795-007-9029-z.

Cooper, H.H., Jacob, C.E., 1946. A generalized graphical method for evaluating formation constants and summarizing well field history. Am. Geophys. Union Trans. 27, 526–534.

Epstein, S., Mayeda, T., 1953. Variation of O18 content of waters from natural sources. Geochim. Cosmochim. Acta 4, 213–224.

Finch, Jonathan W., Hall, Robin L., 2005. Chapter 44: Evaporation from Lakes, Centre for Ecology and Hydrology, Wallingford, Oxfordshire, UK.

Herczeg, Andrew L., Barnes, Chris J., Macumber, Phillip G., Olley, John M., 1992. A stable isotope investigation of groundwater-surface water interactions at Lake Tyrrell, Victoria, Australia. Chem. Geol. 96, 19–32.

Hvorslev, M.J., 1951. Time Lag and Soil Permeability in Ground-Water Observations, bul. no. 26, Waterways Experiment Station, Corps of Engineers, U.S. Army, Vicksburg, Mississippi.

Ibrakhimov, M., 2009. Personal Communication Regarding Studies done in Russian by Mukhammadiev (1982), Ramazanov A., Yakubov, K. (1988), and Nurmanov (1966).

Ibrakhimov, M., Khamzina, A., Forkutsa, I., Paluasheva, G., Lamers, J.P.A., Tischbein, B., Vlek, P.L.G., Martius, C., 2007. Groundwater table and salinity: spatial and temporal distribution and influence on soil salinization in Khorezm region (Uzbekistan, Aral Sea Basin). Irrig. Drain. Syst. 21, 219–236.

Krabbenhoft, David P., Bowser, Carl, Anderson, Mary J.P., Valley, John W., 1990. Estimating groundwater exchange with lakes 1. The stable isotope mass budget method. Water Resour. Res. 26 (10), 2445–245.

Lee, T.M., Swancar, Amy, 1997. Influence of Evaporation, Ground Water, and Uncertainty in the Hydrologic Budget of Lake Lucerne, a Seepage Lake in Polk County, Florida, USGS Water-Supply Paper 2439.

Li, Xiao-Yan, Xu, He-Ye, Sun, Yong-Liang, Zhang, Deng-Shan, Yang, Zhi-Peng, 2007. Lake-level change and water budget analysis at Lake Qinghai, West China during recent decades. Water Resour. Manage. 21, 1505–1516.

Mann, William B., IV, McBride, Mark S., 1972. The Hydrologic Budget of Lake Sallie, Becker County, Minnesota, Geological Survey Professional Paper 800-D.

Marimuthu, S., Reynolds, D.A., Le Gal La Salle, C., 2005. A field study of hydraulic, geochemical and stable isotope relationships in a coastal wetlands system. J. Hydrol. 315, 93–116.

Micklin, Philip., 2007. The Aral Sea disaster. Annu. Rev. Earth Planet Sci. 35, 47–72.

Morrison, J., Brockwell, T., Merren, T., Fourel, F., Phillips, A.M., 2001. On-line high-precision stable hydrogen isotopic analyses on nanoliter water samples. Anal. Chem. 73, 3570–3575.

Neuman, S.P., 1975. Analysis of pumping test data from anisotropic unconfined aquifers considering delayed yield. Water Resour. Res. 11 (2), 329–342.

Onset, 2008, Water Level Loggers. <http://www.onsetcomp.com/water-level-logger>.

Parkhurst, David L., Appelo, C.A.J., 1999. User's Guide to Phreeqc (version 2) – A Computer Program for Speciation, Batch Reaction, One-dimensional Transport, and Inverse Geochemical Calculations, Water-Resources Investigations Report 99-4259, Denver.

Semenov, A.D. (Ed.). Leningrad, Gidrometizdat, 1977. Manual on Chemical Analysis of Terrestrial Surface Waters, 541p.

Theis, C.V., 1935. The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storage. Am. Geophys. Union Trans. 16, 519–524.

Wehrheim, Peter, Schoeller-Schletter, Anja, Martius, Christopher, 2008. Continuity and Change: Land and Water Use Reforms in rural Uzbekistan, IAMO, Germany.

Winter, Thomas.C., 1981. Uncertainties in estimating the water budget of lakes. Water Resour. Bull. 17 (1).

Winter, T.C., Pfannkuch, H.O., 1984. Effect of anisotropy and groundwater system geometry on seepage through lakebeds, 2. Numerical simulation analysis. J. Hydrol. 75, 239–253.

ZEF (Zentrum fur Entwicklungsforschung), 2009, Economic and Ecological Restructuring of Land and Water Use in the Khorezm Region (Uzbekistan) <http://www.zef.de/khorezm.0.html>.