8-2014

Evaluation of Evaporative Drivers and Hydrologic Influences using Energy and Water Budget Techniques at a Shallow Saline Lake in the Western Sandhills of Nebraska, USA

Colin Peake
University of Nebraska-Lincoln, cspeake87@gmail.com

Follow this and additional works at: http://digitalcommons.unl.edu/natresdiss

Part of the Environmental Monitoring Commons, Natural Resources and Conservation Commons, and the Water Resource Management Commons


This Article is brought to you for free and open access by the Natural Resources, School of at DigitalCommons@University of Nebraska - Lincoln. It has been accepted for inclusion in Dissertations & Theses in Natural Resources by an authorized administrator of DigitalCommons@University of Nebraska - Lincoln.
EVALUATION OF EVAPORATIVE DRIVERS AND HYDROLOGIC INFLUENCES
USING ENERGY AND WATER BUDGET TECHNIQUES AT A SHALLOW,
SALINE LAKE IN THE WESTERN SANDHILLS OF NEBRASKA, USA

by

Colin S. Peake

A THESIS

Presented to the Faculty of
The Graduate College at the University of Nebraska
In Partial Fulfillment of Requirements
For the Degree of Master of Science

Major: Natural Resource Sciences

Under the Supervision of Professor Diego Riveros-Iregui

Lincoln, Nebraska

August, 2014
The western Sandhills of Nebraska contain many shallow lakes and wetlands that interact strongly with groundwater and the overlying atmosphere. The region is semi-arid, and most of the lakes are saline, supporting a wide range of ecosystems. Water levels and salt concentrations are highly sensitive to variations in precipitation, evaporation, and groundwater fluxes, making the Sandhills an excellent laboratory for examining the effects of climate on the water balance of interdunal lakes. In this study, we investigate the atmospheric controls on evaporation rates, as well as the water balance of Alkali Lake, one of the more saline lakes in the western Sandhills. The Bowen ratio energy balance and mass-transfer methods are applied over a three-year period (2007-2009) to quantify summer evaporation rates. Daily evaporation is found to vary widely, but averages around 5-6 mm/day during the summer. Evaporation rates are largely controlled by solar radiation on a seasonal basis and by variations in wind and vapor pressure gradient at shorter timescales. Adjustments for salinity affected the mass transfer method more than the energy budget method, with a root mean squared error of 0.49 mm/day, but only 0.09 mm/day for the energy budget method. Evaporation dominated the water balance during the summer months, exceeding precipitation by a factor of 3.2, on
average. The lake water balance also indicates that evaporation exceeds the sum of all water inputs during summer months, causing lake levels to decrease in summer (but rebound during the winter). Net groundwater inflow is the largest source of water into the lake and averages 2.5 mm/day. Occasional negative net groundwater values indicate complex interactions between the coupled lake and groundwater systems. Lake reactions to precipitation inputs suggest that short term groundwater flow reversals are possible.
ACKNOWLEDGEMENTS

I would like to thank my advisor Diego Riveros-Iregui for seeing my potential and helping me grow into a better scientist. I would also like to thank John Lentes who put countless hours into guiding and checking my work. I am also grateful to my other advisors, Vitaly Zlotnik, David Wedin, and Tala Awada for their support and helpful suggestions. I am grateful for the numerous works previously done by UNL students and professors. Specifically I would like to thank John Ong for his work on lake-groundwater interactions at and around Alkali Lake, and his previously unpublished work that allowed for the calculation of lake salinity. Also, a special thanks to Nathan Healey for his work modeling the sediment heat flux. I would also like to thank the numerous other people who helped on some aspect of this project, no help is too small when doing field science.

Thanks to my parents for their love and support not only during my graduate career, but also throughout my entire life. Thanks to my fellow students that made life here in Lincoln more fun. I would like to thank the staff here in the School of Natural Resources for their help and understanding. Your work for the graduate students does not go unnoticed. Lastly I would like to thank my wonderful wife Maggie who was willing to move to Lincoln, NE so I could complete my M.S. Your love and encouragement kept me going and has made me the person I am today.
# Table of Contents

List of Figures .................................................................................................................. i
List of Tables ..................................................................................................................... iii
List of Abbreviations ......................................................................................................... iv

CHAPTER 1: Multi-year energy balance, variability, and climatic drivers at seasonal and interannual timescales ........................................................................................................ 1

1.1 INTRODUCTION ........................................................................................................ 1
1.2 SITE DESCRIPTION ..................................................................................................... 7
1.3 METHODS .................................................................................................................. 12
  1.3.1 Bowen Ratio Energy Balance (BREB) ................................................................. 12
  1.3.2 Sediment Heat Flux Model .................................................................................. 15
  1.3.3 Effects of Salinity ............................................................................................... 17
  1.3.4 Mass Transfer Estimates of Sensible and Latent Heat Flux ............................... 18
1.4 DATA COLLECTION AND QA/QC ......................................................................... 19
  1.4.1 Instrumentation and Data Sampling .................................................................. 19
  1.4.2 Shortwave and Longwave Radiation .................................................................. 23
  1.4.3 Lake Level and Salinity .................................................................................... 23
  1.4.4 QA/QC of Latent and Sensible Heat Flux .......................................................... 25
  1.4.5 Winter Estimates of Evaporation ....................................................................... 30
1.5 RESULTS .................................................................................................................... 34
  1.5.1 Distribution of Values ....................................................................................... 34
  1.5.2 Seasonal Variability ......................................................................................... 38
  1.5.3 Interannual Variability ....................................................................................... 43
1.6 DISCUSSION ............................................................................................................... 48
1.7 SUMMARY AND CONCLUSIONS ............................................................................. 52

CHAPTER 2: Daily energy budget variations and associated meteorological influences ................................................................................................................................. 54

2.1 INTRODUCTION ........................................................................................................ 54
2.2 RESULTS .................................................................................................................. 58
  2.2.1 De-Seasonalized Daily Variability ....................................................................... 58
  2.2.2 Salinity’s Influence on Mass Transfer and Energy Balance Methods .......... 70
2.3 DISCUSSION .............................................................................................................. 74
List of Figures

Figure 1.2.1  Alkali Lake geographic and instrument location……………………………………8
Figure 1.2.2  Annual climatological data from Alliance North…………………………………10
Figure 1.2.3  JAS climatological data from Alliance North……………………………………12
Figure 1.4.1  Data availability at Alkali Lake for 2007-09………………………………………20
Figure 1.4.2  E and H mass transfer regressions before and after QC’ing……………………26
Figure 1.4.3  Winter mass transfer relationship and QC’ing……………………………………32
Figure 1.5.1  Histograms of daily energy fluxes………………………………………………35
Figure 1.5.2  Histograms of daily meteorological variables……………………………………36
Figure 1.5.3  Timeseries of energy budget evaporation rates with winter estimates………38
Figure 1.5.4  Seasonal variability of energy balance components…………………………40
Figure 1.5.5  Seasonal variability of other meteorological variables………………………42
Figure 1.5.6  Annual JAS averages for E, H, R_{net}, Total ΔS, Lake levels, and Q_{sed}……45
Figure 1.5.7  Annual JAS averages for other climatic variables……………………………..47
Figure 2.2.1  Deseasonalized 5-day running means of E and H……………………………..60
Figure 2.2.2  Daily relationships between evaporation and sensible heat…………………..61
Figure 2.2.3  Explores the covariances between wind and vapor pressure………………….63
Figure 2.2.4  Relations between E or H with wind speeds…………………………………….65
Figure 2.2.5  Relationships between meteorological variables and air temperature……67
Figure 2.2.6  Regressions between E or H with other energy terms………………………68
Figure 2.2.7  Balance between the main evaporative drivers…………………………………70
Figure 2.2.8  Salinity’s effects related to mass transfer evaporation rates…………………..72
Figure 2.2.9  Effects of salinity on EBBR evaporation rates.................................74
Figure 3.3.1  Timeseries of average lake water levels with cumulative (P-E).............93
Figure 3.3.2  JAS averages for each water budget component..................................95
Figure 3.3.3  Daily values of net groundwater............................................................96
Figure 3.3.4  Daily relationships between water balance components.....................98
List of Tables

Table 1  List of Instrumentation used at Alkali Lake.................................22
Table 2  Relation of specific electrical conductivity to lake water properties......25
Table 3  JAS averages and uncertainty for each energy term.........................46
Table 4  JAS averages and uncertainty for climatic terms...........................48
List of Abbreviations

BREB  Bowen Ratio Energy Balance
MT    Mass Transfer
R_{net}  Net Radiation (W m^{-2})
Q_{sed}  Sediment Heat Flux (W m^{-2})
A_{net}  Net Advection (W m^{-2})
A_{GI}  Advection from groundwater influx (W m^{-2})
E    Latent Heat Flux (W m^{-2})
H    Sensible Heat Flux (W m^{-2})
\Delta S  Lake Heat Storage (W m^{-2})
B    Bowen Ratio
U    Wind Speed (m s^{-1})
T_{a}  Air Temperature (°C)
T_{s}  Lake Surface Temperature (°C)
e_{a}  Vapor Pressure of the air (kPa)
e_{s}  Vapor Pressure of the lake surface (kPa)
CHAPTER 1: Multi-year energy balance, variability, and climatic drivers at seasonal and interannual timescales

1.1 INTRODUCTION

Free-surface evaporation from lakes and reservoirs plays an important role in the hydrologic cycle and is an important factor in water resources, irrigation, and ecosystem management. Evaporation determination via the energy balance technique can be difficult and time consuming, but much uncertainty exists where less costly empirical methods have been used. While there are numerous ways to estimate evaporation, the Bowen ratio energy balance (BREB) method is considered one of the most accurate (Lenters et al., 2005; Winter et al., 2003). The energy balance approach determines rates of energy inputs and outputs for all non-negligible components, and then distributes the energy into latent and sensible heat fluxes by the experimentally estimated Bowen ratio. This involves measuring variables such as solar radiation, advection, and the rate of change in energy stored in the lake water and underlying sediments. The residual heat flux in the energy balance equation then determines the amount of energy available for sensible and latent heat flux, which is partitioned according to the Bowen ratio. While the energy balance is not a direct estimate of evaporation, it is a preferred technique for long term evaporation measurements (Assouline and Mahrer, 1993; Winter et al., 2003). The energy budget method has been widely used for freshwater lakes (Gallego-Elvira et al., 2010; Lenters et al., 2005; Winter et al., 2003) and, to a lesser extent, for saline lakes (Lensky et al., 2005).
Saline lakes mainly form in arid or semi-arid regions where evaporation exceeds precipitation, resulting in net moisture loss (Meybeck, 1995). As a result, they are typically located in closed basins that are supported by groundwater seepage (Zlotnik et al., 2010). In dry regions, evaporation is a large component of the water balance, but it also tends to be one of the least understood components. Saline lakes are important signals of climate change and human disturbance, because they often respond rapidly to changes in water balance variables (Williams, 2002). Determination of evaporation rates from saline lakes is, therefore, an important consideration for water management in arid and semi-arid regions, where water is scarce. Evaporation rates from saline lakes are not well measured, because they commonly receive far less attention than freshwater lakes (Fritz et al., 2001).

Whereas freshwater evaporation is controlled by meteorological variables and lake physical characteristics, saline lakes present an additional challenge in that the effects of salinity must also be accounted for (Oroud, 2001). Salinity is an important consideration, because it changes the thermophysical properties of water and reduces the saturation vapor pressure above a saline lake. This effectively reduces the vapor pressure gradient between the lake surface and overlying air. It has been well documented that evaporation from a saline water source is smaller than that of a freshwater source under the same meteorological conditions (Harbeck, 1955; Oroud, 1995; Salhotra, 1985). On the other hand, Oroud (1997) found that hypersaline conditions lowered evaporation rates by 40% in the summer months in small shallow ponds near the Dead Sea. Further complexity occurs when saline lakes undergo significant lake level changes. Since salinity is sensitive to changes in water level, evaporation rates (e.g., in mm/day) cannot
be considered constant when water level changes occur (Oroud, 2001). Furthermore, large changes in water level can lead to changes in lake area that also affect the total evaporative flux (e.g., in m$^3$/s).

Shallow lake evaporation typically follows the cyclic behavior of energy available from solar radiation on daily and seasonal time scales (Brutsaert, 1982). On the other hand, if the lake is so shallow that it occasionally dries up – as in a playa – the rate of evaporation can also be limited by the amount of available water (Jacobson and Jankowski, 1989). Evaporation has been found to vary with available energy and water in shallow groundwater discharge playas using the energy balance approach (Menking et al., 2000). In contrast, deeper lakes have the ability to store and release much larger amounts of energy, while also having a much larger supply of available water. The former effect often causes a significant lag between net radiation and evaporation, up to several months after the peak in solar radiation (Brutsaert, 1982). On a freshwater lake in Ontario, Canada, Yao (2009) found a lag time of about 2 months between net radiation and peak evaporation rates. While there have been evaporation studies on some of the freshwater lakes and wetlands, to the author’s knowledge, no previous studies have used the Bowen ratio energy budget (BREB) method to calculate evaporation rates from any of the saline lakes in the Nebraska Sandhills.

The Nebraska Sandhills, with an area of 58,000 km$^2$, is the largest vegetated dune field in the Western Hemisphere. The dune field is currently stabilized by vegetation but was extensively active during the Holocene (Hanson et al., 2009; Mason et al., 2004). Recharge into the well-drained dune field during the Pleistocene raised the groundwater levels and formed numerous interdunal lakes (Loope et al., 1995). Over 2000 lakes are
found in the Sand Hills (Bleed and Flowerday, 1998) and have concentrations of total dissolved solids ranging from freshwater (~0.3 g/L) to saline (100g/L) (McCarraher, 1977). Nearly 75% of the annual precipitation occurs from April to September, with about 50% occurring when warm moist air from the Gulf of Mexico propagates northward, interacting with persistent low pressure systems from May to July (Loope et al., 1995). Annual mean precipitation in the Sandhills ranges from roughly 400 mm/year in the west to roughly 700 mm/year in the east. The generally eastern sloping landscape of the Sandhills is interrupted on the western side by a region of minimal slope, creating endorheic (closed-basin) lakes (Bleed and Flowerday, 1998). These lakes have very unique water chemistry and salinity, because mineral deposition from groundwater seepage becomes concentrated through evaporation. Evaporation has been shown to be a large driver in the solute balance of the saline Sandhills lakes (Zlotnik et al., 2012, 2010). As the western Sandhills is classified as a semiarid region, the presence of such high salinity suggests that groundwater is a key source of water for these lakes.

A wide range of research has been performed on the Sandhills, which sits atop the Ogallala and High Plains Aquifers. These aquifers offer a wealth of groundwater resources that are mainly used for irrigation and local water supply. This has led to interest from researchers in investigating the regional evapotranspiration patterns in order to determine groundwater recharge rates and soil water fluxes (Chen and Chen, 2004; Gosselin et al., 2006; Sridhar et al., 2006). This research has highlighted the need for water management to determine the best sustainable usage to maintain the groundwater resources. A previous global study of land-atmospheric interactions has also shown that the region encompassing the Sandhills may influence the local climate by supplying
atmospheric moisture for this semi-arid region (Koster et al., 2004; Sridhar and Wedin, 2009). Thus, the Sandhills region may act as a “hot spot” for land-atmosphere interaction. Many of these previous studies relied on remotely sensed data or empirical estimates from meteorological data to compute evaporation rates over large swaths of land, wetlands, and lakes. A study of land surface evapotranspiration in the Sandhills using the BREB method was also done by Billesbach and Arkebauer, (2012). They determined that annual evapotranspiration was highest in subirrigated meadows (i.e., between dunes where water availability was highest). This highlights the disproportionate loss of water through lake, wetlands, and wet interdunal areas that cover an otherwise small portion of the Sandhills.

Another previous study examined the water balance of four wetland sites, with one of them being Island Lake, a freshwater flow-through lake in the Sandhills (Winter et al., 2001). The results from this study show the importance of groundwater influx in maintaining the Sandhills lakes and wetlands, as well as their susceptibility to climate change. Winter et al. (2001) showed in the multi-year study that evapotranspiration (lake evaporation plus surrounding plant transpiration) was about 30% of the water budget. Other estimates have shown that surface water evaporation from lakes in the Sandhills may exceed precipitation by as much as 600-750 mm annually (Winter, 1990; Winter et al., 2001). On an open water portion of a Sandhills freshwater wetland, the BREB method was used to determine evaporation rates over a single summer (1994), and the values averaged around 4.1 mm/day from May to October (Burba et al., 1999).

On a diurnal timescale Burba et al. (1999) found that open water evaporation was largely affected by stable overlying air created by temperature inversions occurring over
the water surface. Temperature inversion can create negative atmospheric buoyancy reducing water vapor transport away from the lake surface. Low wind speeds and negative atmospheric buoyancy over saline lakes has been shown to impact latent heat flux across the lake surface-atmospheric boundary (Assouline and Mahrer, 1993). Suppression by temperature inversion caused evaporation to not follow the daily pattern of net radiation the largest control on evaporation. Other terms in the energy balance such as the water column heat storage component were found to be a sink of energy during the day and source at night. However, on a longer time scale the heat storage term was small and sensible heat flux was a minor component over the entire study. While shallow, freshwater lakes typically have lower temperatures than the overlying air during most of the daytime hours (Brutsaert, 1982; Stannard and Rosenberry, 1991), highly saline lakes often have water temperature that remain near the air temperature for most of the diurnal cycle (Oroud, 1997). This reduced atmospheric stability over the lake surface could enhance evaporation from the saline lake, whereas a freshwater lake under the same conditions may have its evaporation suppressed somewhat by the temperature inversion.

While these studies provide insight into regional energy and water fluxes, particularly those in semi-arid regions, they were not intended to fully describe the unique collection of Sandhills saline lakes. The objective of this study, therefore, is to use the BREB method to quantify and analyze evaporation rates for Alkali Lake, a saline lake in the Nebraska Sandhills. This study provides new insights into the climatic factors that drive the lake energy balance (and associated evaporation rates) at short-term, seasonal, and interannual temporal scales. Although numerous studies have focused on evaporation
from saline lakes, to the best of our knowledge no previous study has similarly addressed any saline lakes in the Nebraska Sandhills region, particularly at this level of complexity.

1.2 SITE DESCRIPTION

Alkali Lake, located at 41.82° N, -102.60° W, is a shallow, groundwater - fed lake located near the western margin of the Sandhills (Figure 1.2.1). It has a surface area of roughly 50 hectares (but widely varying, depending on water supplies) and an average depth of 0.3 meters. Negligible overland flow enters the lake because of the high permeability of the sandy soils. The lake water is dominated by sodium (Na) and potassium (K) anions, has a pH of 10.4, and salinity ranging from 36.9 – 78.9 parts per trillion (McCarraher, 1977). A piper diagram in Ong, (2010) shows the linear increase in Na+K and HCO₃ resulting from evaporative deposition in and around Alkali Lake.
Figure 1.2.1 Locations of the main hydrogeologic features of the study area, the High Plains Aquifer, the Nebraska Sandhills, and Alkali Lake. Dots indicate water level sensors, and the diamond indicates the location of the instrumented buoy and associated water temperature/conductivity sensor. This figure was created by John Ong used with permission. The left diagram was adopted from Ong, (2010).

Although groundwater hydrology and chemistry around Alkali Lake has been previously studied, the temporal variability of groundwater interactions with the lake have not been quantified (Befus et al., 2012; Ong et al., 2010). Alkali Lake is thought to be almost exclusively discharging groundwater across the entire lake bed, but how the groundwater dynamics change over time is relatively unknown (Befus et al., 2012). Since Alkali Lake is saline, along with many others in the region, electrical resistivity has been
used as a tool for mapping groundwater discharge and direction (Befus et al., 2012; Ong, 2010; Ong et al., 2010). Salinity was found in the groundwater using electrical resistivity on east and south-eastern sides of Alkali Lake (Ong et al., 2010). It was also determined that there could be a possible minor seepage outflow through the lakebed on the eastern side of the lake (Befus et al., 2012). These findings suggest that Alkali Lake may not be wholly a discharge lake at all times, but rather can fluctuate over time. Alkali Lake also exhibits a solute deficit over the past 700 years, as evidenced from a solute budget that considers advective and diffusive fluxes (Zlotnik et al., 2012, 2010). Additional solute losses likely occur from eolian deflation, periods of reduced solute influxes, and free convection (Zlotnik et al., 2012, 2010).

Other than the instrumented buoy that was installed on site for the purposes of this study, the closest weather station to Alkali Lake is located in Alliance, Nebraska (42.18 N, -102.92 W). This site is known as “Alliance North”, which is an Automated Weather Data Network (AWDN) approximately 50 km north of the lake. The station has roughly 25 years of continuous data, allowing for some general observations about climatic variability in the region (Figure 1.2.2). Of specific interest are the years 2007-2009, when data from Alkali Lake was collected. 2007 was warmer and drier than 2008 and 2009, with some of the highest annual precipitation totals falling in 2009.
Figure 1.2.2 Annual mean climatological data from Alliance North, an AWDN station ~50 km from Alkali Lake. Top to bottom: air temperature, relative humidity, wind speed, and total annual precipitation. Red lines indicate the 25-year mean.

Much of this study will focus on the summer months due to data availability, but this is also the time of year when evaporation rates are largest. Data from the Alliance North station were, averaged over the July, August, September (JAS) period to create 3-month “summer” means. Figure 1.2.3 shows the JAS averages and sums for the same meteorological variables as in Figure 1.2.2. Comparing the annual and JAS averages, the
JAS period is generally warmer, drier, and less windy. Precipitation during the JAS months varies widely over the 25 years of data, ranging from 0 - 66% of the annual total. From 1989 – 2013, an average of 38% of the annual precipitation fell during JAS.
Figure 1.2.3 JAS averages of climatological data from Alliance North, an AWDN station located ~50 km from Alkali Lake. Top to bottom: air temperature, relative humidity, wind speed, and total JAS precipitation. Red lines indicate the 25-year average.

1.3 METHODS

1.3.1 Bowen Ratio Energy Balance (BREB)
The energy balance for a lake can be described by equation 1:

\[
\overline{R}_{\text{net}} + \overline{Q}_{\text{sed}} + \overline{A}_{\text{net}} - (\overline{E} + \overline{H}) = \Delta \overline{S}
\]  

(1)

where \( \overline{R}_{\text{net}} \) is net radiation, \( \overline{Q}_{\text{sed}} \) is the sediment heat flux, \( \overline{A}_{\text{net}} \) is net heat advection, \( \overline{E} \) is the latent heat flux, \( \overline{H} \) is sensible heat flux, and \( \Delta \overline{S} \) is the change in heat storage. The units of all terms are W m\(^{-2}\), with the overbars indicating averages over a daily timescale. Similar to Lenters et al. (2005), we assume that the advection terms are much smaller than the other components of the energy budget at Alkali Lake, and so \( A_{\text{net}} = 0 \). As Alkali Lake is a groundwater discharge lake, the majority of error caused by this assumption likely to come from neglecting cold, groundwater seepage. Advection from groundwater influx can be calculated from \( A_{\text{GI}} = \rho_w * c_w * F_{\text{GI}} * \Delta T_{\text{GI}} \), where \( \rho_w \) is the density of water (1000 kg m\(^{-3}\)), \( c_w \) is the specific heat of water (4186 J kg\(^{-1}\) C\(^{-1}\)), \( F_{\text{GI}} \) is the groundwater influx (m s\(^{-1}\)), and \( \Delta T_{\text{GI}} \) is the difference in temperature between the lake water and inflowing groundwater (Lenters et al., 2005). Average groundwater temperatures near Alkali Lake are \( \sim 12 \) °C, while summer lake temperatures range from 10.0 °C to 28.3 °C. A simple water balance (Chapter 3) reveals that net groundwater averages 2.5 mm/day into Alkali Lake. Using the maximum temperature difference between lake water and groundwater of 15.0 °C, this would lead to an average error of 1.81 W m\(^{-2}\) per day due to advection from groundwater inflows. An extreme outlier of 73.4 mm/day was observed for a single day which would lead to an error of 53.3 W m\(^{-2}\). The next largest net groundwater value was less than half this value at 31.0 mm/day, which corresponds to an error of 22.5 W m\(^{-2}\). While these values are large they are also rare. Removal of the
advection terms has been done in shallow waters and been shown to have little effect (Parkhurst et al., 1998). Neglecting the advection terms and rearranging the energy balance, we arrive at the equation used to calculate daily evaporation rates:

\[
\overline{E} = \frac{\overline{R_{net}} - \Delta \overline{S} + \overline{Q_{sed}}}{1+B},
\]

(2)

where B is the Bowen Ratio. The Bowen ratio is the ratio of sensible to latent heat and is calculated as in dos Reis and Dias (1998):

\[
B = \frac{\overline{H}}{\overline{LE}} = \gamma \frac{U(T_\text{s} - T_\text{a})}{U(e_\text{s} - e_\text{a})}
\]

(3)

Here, \( \gamma \) is the psychrometric constant where \( \gamma = (c_{pa} * P)/(0.622 L_v) \), \( c_{pa} \) is the specific heat of air, \( P \) is atmospheric pressure, \( L_v \) is the latent heat of vaporization, \( U \) is wind speed, \( T_\text{a} \) is air temperature, \( T_\text{s} \) is the water surface temperature, \( e_\text{s} \) is the saturation vapor pressure of the water surface, and \( e_\text{a} \) is the saturation vapor pressure of air. \( \overline{R_{net}} \) is calculated from:

\[
\overline{R_{net}} = (\overline{R_{swd}} - \overline{R_{swu}}) + (\overline{R_{lwd}} - \overline{R_{lwu}})
\]

(4)

where \( \overline{R_{swd}} \), \( \overline{R_{swu}} \), \( \overline{R_{lwd}} \), and \( \overline{R_{lwu}} \) are incoming shortwave, reflected shortwave, incoming longwave radiation, and outgoing longwave radiation, respectively. Outgoing longwave radiation, which is comprised of both emitted and reflected components, is
calculated from \( \bar{R}_{lwu} = \varepsilon \sigma T_s^4 + (1 - \varepsilon)\bar{R}_{lwd} \), where \( \varepsilon \) is the emissivity of water (0.97) and \( \sigma \) is the Stefan-Boltzman constant (5.67 x 10^{-8} \text{ Wm}^{-2} \text{ K}^{-4}). The daily rate of change in lake heat storage is calculated from the following equation:

\[
\Delta \bar{S} = \rho_w * c_w * \bar{h} * \frac{\Delta \bar{T}}{\Delta t}
\]  

(5)

where \( \rho_w \) is the density of water, \( c_w \) is the specific heat of water, \( \bar{h} \) is the daily mean lake depth, \( \Delta \bar{T} \) is the daily change in mean lake temperature (midnight to midnight), and \( \Delta t \) is 86400 seconds (i.e. one day). Although it is usually necessary to apply the BREB method on weekly or longer timescales, due to uncertainty in the lake heat storage term (e.g., Lenters et al., 2005), the very shallow depth of Alkali Lake (~0.3 m) allows us to implement the BREB method on much shorter timescales. Nevertheless, to minimize uncertainty in the daily heat storage term, hourly mean water temperatures were first smoothed to a 3-hour running mean before calculating the daily change in lake temperature. An extensive error analysis was also undertaken to fully assess the level of uncertainty in the energy budget (see section 1.5.2).

### 1.3.2 Sediment Heat Flux Model

While sediment heat flux is often neglected, especially for deep lakes (dos Reis and Dias, 1998; Oroud, 1997), or simplified to represent the average annual cycle (Lenters et al., 2005; Winter et al., 2003), it can often be important for shallow lakes in a seasonally varying climate. Therefore, the sediment heat flux into and out of the water
column for Alkali Lake was explicitly modeled in this study to determine its significance in the lake energy balance.

One-dimensional heat flux by conduction through sediments can be calculated as follows (Keshari and Koo, 2007; Núñez et al., 2010; Smith, 2002):

$$\frac{\partial T_s}{\partial t} = \alpha_s \frac{\partial^2 T_{sed}}{\partial z^2}$$

(6)

In this equation, sediment temperature is driven by the lake water temperature and is denoted by $T_{sed}$, while thermal diffusivity of the sediment is represented by $\alpha_s$ (4.0 * 10$^{-6}$ m$^2$ sec$^{-1}$, which was calculated from the quotient of thermal conductivity and volumetric heat capacity of sediments), $t$ is time (here, one hour), and $\delta z$ is thickness of each sediment layer. To solve equation 6, an explicit finite difference scheme was utilized with upper and lower boundary conditions. The first assumes that the uppermost sediment temperature is driven by hourly mean lake temperatures (over the full annual cycle), while the second boundary condition assumes zero heat flux below 12 meters. The model was initialized by running ~13 years of spin-up, with the first 10 years forced by 2005 water temperature (i.e., 10 years in a row; water temperature was estimated via a regression against local air temperature for 2007-2009), followed by January 2005 – June 2007 water temperatures, and then finally by observed water temperatures from June 2007 – November 2009. At the start of the simulation, the entire sediment matrix began at a temperature of 11.64°C, which is the average sediment temperature that the model converged to during the 2005 spin-up period. Using a finite difference depth of 6 cm for each soil layer and a constant thermal conductivity of 1.2 W m$^{-1}$ K$^{-1}$ (determined from in
situ measurements at various locations throughout the lake bed), the model was found to be stable during the spin-up and "actual" observation periods.

Having simulated the hourly mean sediment temperatures, $T_{sed}$, for the period 2007-2009, the total sediment heat flux into/out of Alkali Lake was then calculated by integrating vertically across the 12-m soil column:

$$Q_{sed} = \rho_{sed} c_{sed} \sum \frac{\partial T_{sed}}{\partial t} \partial z$$

(7)

where $\rho_{sed}$ is the soil density (kg m$^{-3}$), and $c_{sed}$ is the specific heat (J kg$^{-1}$ K$^{-1}$) of the sediments. The quantity $\rho^*c_p = 3.1 \times 10^6$ J m$^{-3}$ K$^{-1}$ is the volumetric heat capacity of the sediments and – similar to the thermal conductivity – was determined from direct measurements using a Decagon KD2 Pro.

### 1.3.3 Effects of Salinity

In order to include the effects of salinity when calculating the water-air vapor pressure gradient, the activity of water was applied to the calculation of saturation vapor pressure. The water activity, $a_w$, is defined as the ratio of the saturation vapor pressure of water over a saline surface ($e_s$) compared to that over fresh water ($e_s^*$) and is always less than 1 (Oroud, 2001; Salhotra, 1985):

$$a_w = \frac{e_s}{e_s^*} = \gamma_w \cdot X_w$$

(8)
Equivalently, and as noted in Equation 8, $a_w$ is the product of the activity coefficient of water $\gamma_w$, and the mole fraction of water, $X_w$, containing a given solute. The mole fraction can be calculated from:

$$X_w = \frac{m_w}{m_w + \sum_{i=1}^{N} m_i}$$  \hspace{1cm} (9)

where $m_w$ is the molality of water and $m_i$ is the molality of the solute. For dilute solutions obeying Raoult’s Law, $\gamma_w = 1$ and $a_w \approx X_w$ (Garrels and Christ, 1965). The salinity-corrected saturation vapor pressure, which is used in calculating the Bowen ratio and mass-transfer estimates of latent heat flux, can then be calculated from:

$$e_s^* = a_w \cdot e_s^*$$ \hspace{1cm} (10)

It should be noted that while $a_w$ is being used as a coefficient, it is not the same as $\gamma_w$, the activity coefficient of water.

### 1.3.4 Mass Transfer Estimates of Sensible and Latent Heat Flux

The mass-transfer method (MT) relates evaporation to the processes affecting the removal of water vapor from the boundary layer above the air-water interface at the surface of a lake (Lee and Swancar, 1997). In general, higher wind speeds above the lake surface cause larger amounts of water vapor to be transported away from the lake, causing the near-surface vapor pressure gradient to increase. Therefore, evaporation is generally related to wind speed and the vertical vapor pressure gradient, measured
between the lake surface and a fixed reference height (typically 2 m). Mass-transfer derived evaporation rates can then be calculated as follows:

\[ \bar{E}_{\text{MT}} = N_E \times \frac{U}{\bar{U}} (\bar{e}_s - \bar{e}_a) \]  

(11)

where \( E_{\text{MT}} \) is the mass transfer evaporation estimate, and \( N_E \) is the transfer coefficient for latent heat. Estimation of sensible heat flux from the mass transfer method uses a similar equation, but replaces the vapor pressure gradient with the lake-air temperature gradient to produce:

\[ \bar{H}_{\text{MT}} = N_H \times \frac{U}{\bar{U}} (\bar{T}_s - \bar{T}_a) \]  

(12)

where, \( H_{\text{MT}} \) is the sensible heat calculated by the mass transfer method, and \( N_H \) is the corresponding transfer coefficient for sensible heat. The best estimates of \( N_E \) and \( N_H \) in each equation come from a calibration between mass transfer estimates of \( E \) and \( H \) and similar estimates using the more accurate BREB method. More specifically, \( N_E \) and \( N_H \) are calculated from the slope of the best-fit linear regression. An example of this can be seen in Lee and Swancar, (1997), where the best-fit line produces a \( N_E \) value of 0.0114 for a seepage lake in Florida.

1.4 DATA COLLECTION AND QA/QC

1.4.1 Instrumentation and Data Sampling
The Bowen ratio energy balance equation (Eq. 2) was used to calculate the mean daily evaporation rate. Figure 1.4.1 shows when the BREB method could be calculated, and when partial data is available. It also depicts the times of the year when data from all three years overlap (grey box). Periods when only one year of full BREB data are available are called “tails”. Data from 2008 comprises the spring “tail”, and data from 2009 makes up the fall “tail”.

**Figure 1.4.1** Data availability for the study period of 2007-09. Solid lines indicate full data coverage, and the grey box indicates the time period when all three years of data are available. The dashed line indicates the time period when radiation data was unavailable,
but all other instruments were operational. The full data coverage in spring 2008 and fall 2009 when only one year of full data is available are referred to as “Tails”.

A buoy made from the hull and mast of a catamaran sailboat was deployed near the center of Alkali Lake in June 2007 and remained on the lake for the duration of this study (2007-2009), aside from occasional times when strong wind events would blow the station to shore. Many variables were measured at the buoy, including downward and upward shortwave radiation, downward and upward longwave radiation, air temperature and relative humidity, water conductivity and temperature (both bulk and skin), barometric pressure, wind speed and direction, and rainfall rate. The instrumentation used to measure these components is listed in Table 1. In addition, two pressure transducers were placed near the eastern and western portions of the lake to measure water level and temperature. The locations of the instrumented buoy and pressure transducers are shown in Figure 1.2.1. Most of the meteorological variables were sampled every 10 s, while water level and bulk water temperature were sampled every 20 minutes. Bulk water temperatures were created by using a three hour running mean from all three (east, west, and buoy) water temperature sensors to create a lake-wide mean. All other variables were averaged to hourly and daily means.
Table 1. List of variables measured at Alkali Lake and associated instrumentation.

<table>
<thead>
<tr>
<th>VARIABLE</th>
<th>INSTRUMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Buoy/raft structure</td>
<td>Hobie Bravo catamaran sailboat (hull and mast)</td>
</tr>
<tr>
<td>Shortwave radiation</td>
<td>Kipp &amp; Zonen CMP21 pyranometer</td>
</tr>
<tr>
<td>Longwave radiation</td>
<td>Kipp &amp; Zonen CGR4 pyrgeometer</td>
</tr>
<tr>
<td>Air temperature/RH</td>
<td>Vaisala HMP45C</td>
</tr>
<tr>
<td>Wind speed/direction</td>
<td>RM Young Wind Monitor 05106</td>
</tr>
<tr>
<td>Barometric pressure</td>
<td>Campbell Scientific CS100</td>
</tr>
<tr>
<td>Water level</td>
<td>HOBO U20 titanium (pressure), and Campbell Scientific SR50 (ultrasonic)</td>
</tr>
<tr>
<td>Water temperature</td>
<td>HOBO Pro V2 (bulk), and Apogee SI-111 (skin)</td>
</tr>
<tr>
<td>Conductivity/salinity</td>
<td>YSI 600R multi-parameter sonde</td>
</tr>
<tr>
<td>Rainfall rate</td>
<td>Texas Electronics TE525MM</td>
</tr>
<tr>
<td>Datalogger</td>
<td>Campbell Scientific CR100</td>
</tr>
</tbody>
</table>

Some of the quantities in Equations 2 and 3 involve cross products of certain variables (e.g., wind speed and vapor pressure gradient), and unless the covariances among variables is small, the products must be calculated prior to averaging (Brutsaert, 1982; Jobson, 1972; Kondo, 1972; Webb, 1964, 1960). This is particularly important for the diurnal cycle, since diurnal covariances between wind speed and air temperature, for example, can be substantial (Jobson, 1972; Kondo, 1972; Webb, 1964, 1960). To avoid these systematic errors, one can simply compute the required quantities at short periods before averaging (Hage, 1975). In this study, we account for covariances in the diurnal cycle by calculating all products at the hourly mean timescale, before averaging to longer periods such as daily. It is also important to note that quantities such as albedo and the Bowen ratio require that both the numerator and denominator be averaged prior to calculating the ratio of the two. (In other words, the ratio of the means is not the same as the mean of the ratios.) Again, this is the procedure used in the current study.
1.4.2 Shortwave and Longwave Radiation

Upward and downward radiation (both shortwave and longwave) were measured from the south-facing side of the buoy, but slight variations in the tilt of the sensors can lead to erroneous values of the shortwave albedo. To correct for these errors, the hourly data were divided into sunny and cloudy measurements based on transmissivity values. Theoretical maximum and minimum albedo values were then determined from a functional relationship with sun angle, which was constructed based on data collected by Payne (1972). Using the theoretical maximum and minimum bounds of albedo, the observed albedo measurements could be screened for erroneous values. In instances where the albedo values exceeded the bounds, they were reset to the theoretical maximum or minimum values. Roughly 16% of the daytime albedo values were reset using this technique. Additionally using the relationship between albedo and sun angle from Alkali Lake itself further outliers could be determined for each hour and reset to the maximum or minimum bounds. This resulted in another 14% of daytime albedo values being reset for a total of around 30% of all daytime values. As would be expected, the majority of adjusted albedo values occurred in the morning and evening, due to the low sun angle at those times of day. The impact of these corrections on daily mean net shortwave radiation is not expected to be large, since incoming shortwave radiation is relatively low at these times of the day.

1.4.3 Lake Level and Salinity
Lake level was measured by three separate instruments – a Campbell Scientific SR50 placed on the buoy and two HOBO pressure transducers located on the east and west sides of the lake. The SR50 was not installed until June 29<sup>th</sup> 2009, whereas the HOBO pressure transducers were deployed for the entire study period. Periodic staff gauge measurements were also available over the 3-year study period. During times of overlap, the SR50 and HOBO sensors compared well, with an $R^2 = 0.98$ and an occasional bias in the pressure transducer at the west end of the lake. After careful examination of all available data, including staff measurements, the SR50, two pressure transducers, and salinity-inferred water level variations, an average of the two HOBO pressure transducers (with the westernmost sensor adjusted to correct for the known bias) was used to assess water level variations over the 3-year period. This curve was also adjusted upward by 2 cm to match the staff gauge measurements, and the final water level curve compares well with all of the other available measurements.

Direct measurements of salinity (more precisely, conductivity) began in May 2008 using a YSI probe located on the buoy and deployed at a depth of roughly 10-20 cm. The conductivity measurements continued through November 2009, covering more than a full annual cycle (including winter). The conductivity / salinity measurements allowed for a more precise estimate of water level variations, since water pressure is dependent on density and, therefore, the concentration of solutes. To account for salinity when data were not readily available (e.g., 2007) an empirical relationship between water level (assuming freshwater) and absolute salinity was developed ($R^2 = 0.89$). This relationship allowed one to estimate salinity (and density) from water level, when more direct measurements were not available. The density calculations were then used to
provide a slight correction to the aforementioned freshwater lake level values. Periodic measurements of water samples from Alkali Lake were analyzed in March 2009, March 2010, and October 2010 to assess the water chemistry and verify the YSI sonde’s estimates of conductivity and salinity. Measurements of specific electrical conductance (SEC) during these three periods ranged from 40 to 79 mS/cm. Water samples from neighboring lakes, wells, and streams were also collected to characterize a broader range of SEC values. John Ong (unpublished data) created several regression equations ($R^2 = 0.98$ or higher) relating the SEC of Alkali Lake (which was also measured at the YSI sonde) to water activity, density, salinity, and absolute salinity, which are all shown in Table 2. These regressions were used to calculate the final values of water density, specific heat of water, and activity coefficient from SEC measurements at the raft YSI sonde (or inferred from lake level when SEC measurements were absent).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Regression equation</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Activity of water</td>
<td>$y = -0.00057617x + 1.000$</td>
<td>0.990</td>
</tr>
<tr>
<td>salinity (g/L)</td>
<td>$y = 1.255 x$</td>
<td>0.987</td>
</tr>
<tr>
<td>absolute salinity (g/Kg)</td>
<td>$y = 1.186225 x$</td>
<td>0.989</td>
</tr>
<tr>
<td>density (kg/L)</td>
<td>$y = 0.000963916 x + 0.996993$</td>
<td>0.990</td>
</tr>
</tbody>
</table>

Table 2. Regression equations relating activity of water, salinity, absolute salinity, and density (all the y values) to SEC (the x value; mS/cm; n=27).

1.4.4 QA/QC of Latent and Sensible Heat Flux

In order to assist in evaluating the BREB-derived sensible and latent heat flux estimates, such as instances of $B \sim -1$ and/or errors in the heat storage term, the mass
transfer method was used for comparison. Daily values of the BREB-derived evaporation rate (eq. 2) were regressed against the daily mean mass transfer product of wind speed and vapor pressure gradient. Figure 1.4.2 shows both the original and QC’d data, with the latter being instances where outliers were removed. Outliers were determined by first calculating the difference between the BREB-derived values and the original mass transfer regression (with only large, visual outliers removed), and then ranking the differences by percentile. The top and bottom five percent outliers were then rigorously examined to assess whether the large differences resulted from errors in the mass transfer or energy balance approaches (or both, or neither).
Figure 1.4.2 Final regressions and mass-transfer equations for a) $E$ and b) $H$, determined from the corresponding mass transfer products of wind and vapor pressure/temperature gradients. Values on the y-axis were determined from the BREB method. Grey dots represent original, unaltered data, while the colored dots show the final QC’d $E$ and $H$ values.

One of the most obvious potential problems with the BREB method lies in the equation itself; that being the occurrence of Bowen ratios near -1. This causes the
denominator of the energy balance (eq.2) to be very near zero, creating unrealistically large evaporation (or condensation) values. Previous studies sometimes create fixed intervals within which to disregard data, such as Bowen ratios within the bounds of $B < -0.75$ or $-1.3 < B < -0.7$ (Ortega-Farias et al., 1996; Unland et al., 1996). This can result in large swaths of data being discarded, sometimes 30% or more (Gavilán and Berengena, 2006; Ohmura, 1982; Unland et al., 1996). Others have proposed defining intervals according to the vapor pressure gradient and temperature accuracy (Ohmura, 1982; Perez et al., 1999).

Another inherent problem with the BREB method is the potentially large uncertainties that can arise in the heat storage term, which can be sensitive to short-term changes in lake temperature (e.g., internal waves), particularly for deep lakes. Because Alkali Lake is so shallow and the climate is windy, significant mixing takes place throughout the day, creating largely isothermal conditions within the shallow lake water column. Similarly, significant horizontal mixing can lead to relatively homogenous temperatures throughout the entire lake, leading to relatively low errors in the heat storage term, even at daily timescales. Nevertheless, since the rate of heat storage in the lake water column (less so in the sediments) is the most uncertain energy balance term, instances where it significantly enhances or suppresses evaporation should be examined as suspect, and potentially due to incomplete mixing. Lastly, we note a few rare instances where the lake at least partially dried up, causing exceedingly high surface temperatures (e.g., dry sand $> 40^\circ$C). Although the BREB method is fairly robust to such large changes, the mass transfer method is not, since it assumes a wet surface, and so derived saturation vapor pressures become excessively large. Altogether, the potential problems
noted above were identified in a total of 40 days (i.e., 10.2% of the data), with the most appropriate method then being used for the final values of sensible and latent heat flux. While the mass transfer method is, in general, less accurate than the BREB approach, the mass transfer results yielded much more reasonable values in certain instances (e.g., where B ~ -1). In cases where the BREB values were replaced with those from the mass transfer method, the energy budget was also adjusted to balance. This was accomplished by calculating the rate of heat storage as a residual from the energy balance equation, since it is deemed to be the most uncertain term in the energy budget.

Of the 40 days identified in the above QA/QC procedure as being problematic, 19 instances were deemed to have suspect values for both the BREB- and MT-derived values of latent heat flux. In such instances, a “hybrid” approach was used, which takes advantage of the fact that the MT-derived sensible heat flux values (Figure 1.4.2b; RMSE = 3.29 W/m²) show considerably less scatter than the MT-derived latent heat flux data (Figure 1.4.2a; RMSE = 21.7 W/m²). In other words, we assume the MT-derived values of $H$ to be correct and then calculate the latent heat flux as a residual from the energy balance. Subsequent “corrected” values of the Bowen ratio are also calculated. This hybrid approach was used in 19 instances (4.9% of the data), when the BREB-derived evaporation rate was deemed unreliable and use of the hybrid method reduced the overall uncertainty in the latent heat flux. Note that the hybrid approach results in a changed Bowen ratio (compared to the BREB approach) but an identical heat storage term, while the use of MT-derived $H$ and $E$ values results in an identical Bowen ratio, but an altered heat storage term. Thus, the former is most commonly applied in instances where $B$ is negative, and especially when it approaches -1, while the latter is a better approach when
the heat storage term contains large uncertainties. Overall, the MT method replaced BREB latent heat estimates for 21 days, and sensible heat for 39 days. Determining the “best” approach, however, can often be somewhat subjective, which is why the above correction procedures were only applied to the most outlying 10% of the data.

1.4.5 Winter Estimates of Evaporation

Due to the removal of the radiometers for calibration in the fall of 2008, radiation data were unavailable from October 2008 to June 2009. The dashed line in Figure 1.4.1 shows when the radiation data gap occurred. All other meteorological and lake variables continued to be measured, however, during the winter of 2008-2009. This allows for at least an approximate calculation of sensible and latent heat fluxes during this time period, albeit with slight modifications. Winter estimates of evaporation from Alkali Lake present two main problems. The first is the formation of ice, which is complicated by the fact that the lake is saline, causing a freezing point that is colder than that of freshwater (and one that also changes with salinity). Secondly, the mass transfer method that was previously calibrated during the summer period must be reassessed to determine its appropriateness across a broader (and assumed linear) range of winter values. Complications have been found, for example, where seasonality in the mass transfer relationship may require a nonlinear regression (Lee and Swancar, 1997; Sturrock et al., 1992).

Seasonality in the MT method has been found in a number of previous studies (Lee and Swancar, 1997; Parkhurst et al., 1998; Sturrock et al., 1992). A clear consensus has not been reached as to the reasons for this nonlinearity, but Lee and Swancar, (1997)
discuss a wide range of possible causes. Seasonality in the mass transfer coefficient may be due to errors in either method, changes in atmospheric stability, or simply an incorrect assumption that the coefficient is constant. Regardless, it is evident that some nonlinearity in the mass transfer coefficient exists at Alkali Lake, particularly at low evaporation rates (Figure 1.4.3a). Ideally, the linear relationship would have a y-intercept of 0 W m\(^{-2}\), however it crosses at 19.8 W m\(^{-2}\). This means that the MT method may be overestimating evaporation in instances of weak vapor pressure gradients, which can be common during winter months. To correct for this seasonal bias, a 3\(^{rd}\)-order polynomial was used to describe the mass transfer relationship in instances where it began to diverge from the linear regression (i.e., \(U(es-ea) < \sim 1.7 \text{ kPa m s}^{-1}\)). It was found that this 3\(^{rd}\)-order polynomial has a y-intercept very near zero (\(~1.7 \text{ W m}^{-2}\)), making it a more physically intuitive estimate of evaporation when the mass transfer product is especially small.
a) Evaporation vs. U(es-ea) MT relationship

\[ R^2 = 0.865, \ p < 0.0001 \]

\[ y = 32.9x + 19.8 \]

\[ R^2 = 0.869, \ p < 0.0001 \]

\[ y = 0.34x^3 - 4.6x^2 + 51.0x + 1.7 \]

b) Evaporation vs. Ts - Tf

\[ R^2 = 0.659, \ p < 0.0001, \ y = 6.2x + 20.0 \]

\[ R^2 = 0.618, \ p < 0.0001, \ y = 7.5x \]
Figure 1.4.3 Graphs illustrating the main two problems in estimating wintertime evaporation using the mass transfer method: a) Mass transfer relationship for latent heat flux. The black line represents the linear regression, and the red line is a 3rd-order polynomial fit. b) Evaporation regressed against water temperature, relative to the freezing point ($T_f$). Black line shows the linear regression, while the red line is the linear regression forced through zero.

Potential presence/absence of ice at Alkali Lake was estimated using a formula for the freezing point of seawater from the United Nations Educational, Scientific, and Cultural Organization handbook (Fujino et al., 1974; Millero and Leung, 1976):

$$T_f = -0.0575*S + 1.710523^{-3*S^{3/2}} - 2.154996^{-4*S^2} - 7.53^{-4*p}$$

In this equation, $T_f$ is the freezing point of water in °C, $S$ is the practical salinity, and $p$ is the atmospheric pressure in decibars. During the winter period, salinities were higher than that of seawater (~35 g/kg) with daily absolute salinities varying from 65-102 g/kg. By comparing the freezing point and water surface temperature, an assessment of ice presence/absence can be obtained. Figure 1.4.3b shows the relationship between evaporation and the water surface temperature, relative to the freezing point. Ideally the regression would go through zero, since evaporation is usually assumed to cease when ice cover is present (or at least fully covering the lake). The y-intercept is relatively small, however (20 W m$^{-2}$), and so the regression was simply forced through zero, resulting in a 21% higher slope. Based on these very similar results, and plausible physical reasoning,
we simply assume that when the surface water temperature is lower than the calculated freezing point, the lake is likely ice covered, and therefore not evaporating. Evaporation rates on days when this occurred were set to zero.

1.5 RESULTS

1.5.1 Distribution of Values

Histograms of the daily energy balance components are shown in Figure 1.5.1. They depict the JAS period from 2007-2009 (n = 276). BREB-derived evaporation rates show a wide range of values from near zero to over 300 W m\(^{-2}\), with mean and median values near 145 W m\(^{-2}\). In comparison, sensible heat fluxes are largely distributed around zero, with a smaller standard deviation (19.0 W m\(^{-2}\)) than evaporation (48.8 W m\(^{-2}\)). The mean and median of net radiation is similar to evaporation, with a difference of only 5 W m\(^{-2}\). However maximum values of R\(_{\text{net}}\) and E are quite different, suggesting other influences on evaporation. The total rate of heat storage (water plus sediments) is similar to sensible heat flux, in that it also averages around zero. For both total heat storage and sensible heat flux, roughly 45% of the data falls within -10 to +10 W m\(^{-2}\).
Figure 1.5.1 Histograms of the main energy balance components at the daily timescale. Daily values are from JAS period (2007-2009). Top Left: Latent heat flux, Top Right: Sensible heat flux, Bottom Left: Net radiation, Bottom Right: Total rate of heat storage. All units are in W m$^{-2}$.

Histograms of other meteorological variables are shown in Figure 1.5.2. They represent the daily values of air temperature, water-air temperature difference, relative humidity, and wind speed for the JAS period (n = 276). Air temperature averages around 19.8 °C, but is skewed toward colder temperatures. The temperature difference between the lake surface and the air is slightly positive (water warmer than air), but largely
centered about zero, with a standard deviation of 2.0°C. The distribution of daily mean relative humidity values reflects the dry climate of the region, with a mean value of 65%, and relatively few days with relative humidity exceeding 90%. Daily mean wind speeds have a median value of 3.6 m s\(^{-1}\) and a small standard deviation of 1.5 m s\(^{-1}\). Wind speed has a fairly broad tail toward higher values, indicating that some days can be very windy.

**Histograms of daily meteorological variables**

![Histograms of daily meteorological variables](image)

**Figure 1.5.2** Daily mean values of meteorological variables during the JAS period (2007-2009) Top left: Air temperature, Top right: Water-air temperature difference, Bottom left: Relative humidity, Bottom right: Wind speed.
A timeseries of all available BREB-derived evaporation rates (and winter MT estimates) is shown in Figure 1.5.3. Evaporation rates generally show the largest values during midsummer, and lowest values in January. It is also evident that day-to-day variations are quite high. MT-derived latent heat flux estimates from November 1st, 2008 to June 27th, 2009 show a range of -17 to +294 W m$^{-2}$. A total of 34 days were set to zero when ice was estimated to be present on the lake. A few days of negative latent heat flux indicate condensation, which is certainly possible, given the downward vapor pressure gradient.
Figure 1.5.3 Daily mean BREB-derived latent heat flux values (black dots), with winter mass transfer estimates shown in red (primarily during the winter and spring). Lines represent 4\(^{th}\)-order polynomial fits, and all units are in W m\(^{-2}\).

1.5.2 Seasonal Variability

Figure 1.5.4 illustrates the seasonal cycles of the various energy balance components, with the data organized into bi-weekly periods. Because of the limited data coverage during spring and fall (Figure 1.4.1), conclusions during these “shoulder
seasons” should be taken with a degree of caution, whereas data coverage during the JAS period includes all 3 years. Each two-week period includes a maximum of 42 days (i.e., 14 days each year for three years) and a minimum of 8 days.

Evaporation shows an average seasonal cycle that starts out low in the spring and gradually increases until late June, decreasing gradually thereafter (Figure 1.5.4a). On average, the highest daily values of evaporation occur during the four-week period June 23rd to July 21st, with a mean value of 187 W m$^{-2}$. Considerable day-to-day variability in evaporation is present within each of the bi-weekly periods, except for the fall. The last two bi-weeks of Oct-13 and Oct-27, however, largely reflect the year 2009 and have much lower variability, so the limited variability may be at least partly due to the low sample size. Sensible heat flux shows limited seasonal variability, but significant daily variations that average near zero (Figure 1.5.4b). Maximum and minimum values of sensible heat flux are 66 W m$^{-2}$ and -105 W m$^{-2}$, respectively, with an overall mean of 2 W m$^{-2}$ and median of -0.4 W m$^{-2}$.

Net radiation shows seasonal variability similar to that of evaporation (Figure 1.5.4c), with values increasing until late June, then declining through the rest of the year. Variability remains relatively large and somewhat constant throughout the study period, with the exception of the two week period of May 26, which shows very large day-to-day variability. Lastly, the total rate of heat storage shows minimal seasonal variability, similar to that of sensible heat flux (Figure 1.5.4d). There is a slight tendency toward positive values in spring and early summer (i.e., warming of the lake), followed by negative values in late summer. Daily heat storage rates vary greatly, particularly in the
spring and fall, with a maximum of 170 W m$^{-2}$ and a minimum of -124 W m$^{-2}$.

Summertime values show a more limited range of variability.

**Figure 1.5.4** Seasonal patterns in the various energy balance components for 2007-2009, as depicted by bi-weekly box-and-whisker plots, including a) latent heat flux, b) sensible heat flux, c) net radiation, and d) total rate of heat storage. Blue dots (red lines) represent the mean (median), plus symbols are outliers, and blue boxes denote the interquartile range. All units are in W m$^{-2}$ and n-values at the bottom denote the total number of days included in a given 2-week period.
Similar box-and-whisker plots for other important atmospheric and lake variables are shown in (Figure 1.5.5). The vapor pressure gradient, both with and without wind speed as a multiplier, has a seasonal cycle similar to that of evaporation, as would be expected (Figure 1.5.5a/c). Highest values of es-ea (and its variability) occur in late June and early July, while minimum values are very near zero in the fall (but still positive). Similar to sensible heat flux, both the lake-air temperature gradient and its product with wind speed have little to no seasonal pattern (Figure 1.5.5b). Wind speed shows a moderate seasonal cycle (Figure 1.5.5e), with lower wind speeds in summer and higher values in spring and (especially) autumn. Relative humidity ranges, on average, from 60-70% in spring and summer, with higher values of 70-90% in autumn (Figure 1.5.5f).
Figure 1.5.5 Seasonal patterns in the various meteorological components for 2007-2009, as depicted by bi-weekly box-and-whisker plots, including a) Wind speed times vapor pressure gradient (kPa m s\(^{-1}\)) b) Wind speed times temperature gradient (°C m s\(^{-1}\)) c)
vapor pressure gradient (kPa) d) temperature gradient (°C) e) Wind speed (m s\(^{-1}\)) f) Relative humidity (%). Blue dots (red lines) represent the mean (median), plus symbols are outliers, and blue boxes denote the interquartile range. n-values at the bottom denote the total number of days included in a given 2-week period.

1.5.3 Interannual Variability

The common time period for which data are available during all three years is roughly July-September (JAS; Figure 1.4.1), so we focus on this 3-month period to examine interannual variability in the energy and water balance of Alkali Lake (Figure 1.5.6). Tables 3 and 4 also show the mean values of each variable during the JAS period, along with their associated estimated uncertainties. In most cases, the daily variability within the JAS period is much larger than the variability among years (Figure 1.5.6). Lake level is the exception, showing an increase from 2007 to 2008, followed by a much larger increase into 2009. It is important to note that – although the JAS period includes some degree of seasonal variability in the energy balance components (e.g., Figure 1.5.4), removing the seasonal cycle had a limited impact on the overall conclusions regarding interannual variability.

JAS evaporation rates were similar among the three years (Figure 1.5.6a) with 2007 having the highest mean latent heat flux of 155 W m\(^{-2}\) and 2009 showing the lowest (141 W/m\(^2\)). These values are within the range of estimated uncertainty (Table 3), however, suggesting that one cannot demonstrably conclude any real difference in evaporation rates among years. Sensible heat flux, on the other hand, shows a slight upward trend that exceeds the much smaller bounds of uncertainty (Figure 1.5.6b; Table
3). Similar to evaporation, both net radiation and the total rate of heat storage show limited interannual variability that is within the bounds of uncertainty (Figure 1.5.6c-d; Table 3). Heat storage rates in the lake sediment are similar to those of the total heat storage rate, but with much smaller daily variability (Figure 1.5.6f).
Figure 1.5.6 Box-and-whisker plots illustrating the daily and interannual variability within the JAS period for each year of the study. Shown is the a) latent heat flux (W m$^{-2}$), b) sensible heat flux (W m$^{-2}$), c) net radiation (W m$^{-2}$), d) total rate of heat storage (W m$^{-2}$), e) lake level (m), and f) sediment heat storage (W m$^{-2}$).
2), e) lake level (m), and f) rate of heat storage in the sediments (W m⁻²). Blue dots represent averages for that year, red lines are medians, and plus symbols indicate outliers.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>2007</td>
<td>155.9</td>
<td>-7.7</td>
<td>152.1</td>
<td>0.1</td>
<td>3.9</td>
<td>3.8</td>
</tr>
<tr>
<td>2008</td>
<td>144.6</td>
<td>-0.3</td>
<td>148.2</td>
<td>0.0</td>
<td>3.9</td>
<td>3.9</td>
</tr>
<tr>
<td>2009</td>
<td>141.7</td>
<td>11.2</td>
<td>151.6</td>
<td>-3.8</td>
<td>-1.3</td>
<td>2.5</td>
</tr>
<tr>
<td>Uncertainty</td>
<td>21.6</td>
<td>3.4</td>
<td>25.7</td>
<td>10.0</td>
<td>10.2</td>
<td>2.0</td>
</tr>
</tbody>
</table>

**Table 3.** Mean JAS values for each component of the energy balance, along with their estimated uncertainty (based on assumed instrument error). All units are in W m⁻².

Figure 1.5.7 illustrates the interannual variability in other important climatic parameters, while Table 4 shows the mean values and uncertainties. Similar to the energy balance terms, daily variations in most meteorological parameters greatly exceed the interannual variability. Despite this, some interannual variability is present. This includes a higher lake-air temperature gradient in 2009 (and its product with wind speed; Figures Figure 1.5.7c and Figure 1.5.7d), which is consistent with what was found for sensible heat flux (Figure 1.5.6b). Vapor pressure gradient was also higher in 2009 (Figure 1.5.7a), but the relative uncertainty is larger as well, and the difference disappears when combined with wind speed (Figure 1.5.7c). This is consistent with the limited interannual variability that was found for latent heat flux (Figure 1.5.6a). Both relative humidity and
wind speed show rather unremarkable variations, year to year (Figures 1.5.7e and Figure 1.5.7f).

Figure 1.5.7 Box-and-whisker plots illustrating the daily and interannual variability within the JAS period for each year of the study. Shown is the a) Vapor pressure
difference (kPa) b) Temperature gradient (°C) c) Wind speed times vapor pressure difference (kPa m s\(^{-1}\)) d) Wind speed times temperature gradient (°C m s\(^{-1}\)) e) Relative humidity (%) f) Wind speed (m s\(^{-1}\)). Blue dots represent averages for that year, red lines are medians, and plus symbols indicate outliers.

<table>
<thead>
<tr>
<th></th>
<th>Ts-Ta (\circ\text{C})</th>
<th>es-ea (kPa)</th>
<th>U(es-ea) (kPa m s(^{-1}))</th>
<th>U(Ts-Ta) (\circ\text{C} \text{m s}^{-1})</th>
<th>U (m s(^{-1}))</th>
<th>RH (%)</th>
<th>Lake level (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2007</td>
<td>-0.5</td>
<td>0.93</td>
<td>3.9</td>
<td>-3.4</td>
<td>3.9</td>
<td>63.5</td>
<td>0.06</td>
</tr>
<tr>
<td>2008</td>
<td>0.5</td>
<td>0.95</td>
<td>4.0</td>
<td>-0.1</td>
<td>4.0</td>
<td>65.5</td>
<td>0.10</td>
</tr>
<tr>
<td>2009</td>
<td>2.4</td>
<td>1.12</td>
<td>4.0</td>
<td>5.3</td>
<td>3.7</td>
<td>67.1</td>
<td>0.30</td>
</tr>
<tr>
<td>Uncertainty</td>
<td>0.7</td>
<td>0.4</td>
<td>3.4</td>
<td>4.3</td>
<td>0.5</td>
<td>5.0</td>
<td>0.03</td>
</tr>
</tbody>
</table>

Table 4. Mean JAS quantities for each year, along with the estimated uncertainty (based on assumed instrument error).

1.6 DISCUSSION

Seasonality in evaporation shows large day to day variation with the overall trend of peaking very near the summer solstice. This is consistent with nearby studies of shallow wetlands (Burba et al., 1999; Parkhurst et al., 1998), and largely a reflection of the energy available to evaporation common with shallow lakes (Brutsaert, 1982). In contrast larger lakes reach peak evaporation rates near late summer. Variability is highest in the spring through mid-summer, and with the exception of the two week period of Sep-29, declines into the fall. The periods of Oct-13 and Oct-27 mainly account for one year of data (2009); therefore, it is difficult to say with certainty that fall variability is less than spring or summer. These shortcomings aside, it does appear that evaporation has a pattern of
increasing then decreasing over a year, but that the large intraseasonal variability would prevent the generalization of an average seasonal cycle. This result is similar to a study done on an open water wetland in the prairies of North Dakota, where evaporation responded rapidly to changes in solar radiation making year to year seasonal cycles vary significantly (Parkhurst et al., 1998). An example of this can be seen by comparing Figure 1.5.4a to Figure 1.5.4c during the two week period of May 26th where only one year of data was available. This period has lower evaporation than the surrounding weeks, while net radiation varied greatly and on average was much lower. There are also connections between evaporation and the vapor pressure difference with and without wind (Figure 1.5.5a/c). The increased variability in vapor pressure differences in the summer can be attributed to different combinations of temperature gradient and relative humidity. The inclusion of wind somewhat dampens this and causes the product of wind speed and vapor pressure gradient to more closely resemble the seasonality of evaporation. This shows that net radiation is the main source of energy for evaporation, but that other atmospheric drivers are also important in controlling seasonal evaporation cycles. This could have implications for the potential of using the less costly and time consuming mass transfer method at the seasonal timescale.

Seasonally, sensible heat flux varies around zero, but has larger variability in the spring and fall (Figure 1.5.4b). The relatively minor role of seasonal sensible heat has been found in previous studies from shallow freshwater wetlands where it also varied around zero (Burba et al., 1999; Parkhurst et al., 1998). Shallow lakes in semiarid regions usually have lower temperatures than surrounding air resulting in a downward temperature gradient (Brutsaert, 1982; Stannard and Rosenberry, 1991). However, saline
lakes have different thermodynamic properties due to salinity and should have values nearer atmospheric temperatures (Oroud, 1997). The higher spring and fall variability is likely due to weather events such as frontal passage resulting in rapid air temperature changes. Evidence of this is shown in the seasonality of the temperature gradient, but is especially strong in the product of wind speed and temperature gradient that essentially mirrors sensible heat (Figure 1.5.5b). Similarly total storage varies around zero and appears to vary more in the spring and fall periods (Figure 1.5.4d). Its variation is also linked to weather events where large temperature swings may cause large positive heat storage one day and negative the next. This also shows that while heat storage is a minor component of the energy balance on most days; it is very influential to individual days and possibly more so in spring and fall.

As can be seen from Figure 1.5.6a, there is relatively little variation in evaporation year to year during this time span. This is not particularly surprising because other long term studies with better yearly coverage show most variation occurring in the spring or fall months (Lenters et al., 2005; Robertson and Barry, 1985; Winter et al., 2003). The low interannual variability in evaporation is mirrored by net radiation the largest energy source for evaporation (Figure 1.5.6c). This again emphasizes that at long time scales evaporation is largely driven by net radiation. It also shows that weather patterns during this time of the year were relatively stable across all three years, as net radiation also reflects the measure of cloud cover. Figure 1.5.6b shows that there is interannual variability in sensible heat transfer well outside estimated uncertainty with 2009 being highest. These differences are indicative of a varying temperature gradient between the lake surface and the atmosphere and evidenced by Figure 1.5.7b/d. It should be noted the
covariation between the temperature gradient and wind appears to matter little, as they have the same pattern despite the inclusion of wind in Figure 1.5.7d. Uncertainty limits for the temperature gradient, with and without wind speed, also point to differences between years (Table 4). These temperature differences between years also have implications for the heat storage terms. Figure 1.5.6d shows the interannual variability between total heat storage. Total heat storage has a negative average in 2009 indicating an overall loss in total heat storage as the lake cooled, while 2007 and 2008 are positive for the same JAS period. Since total heat storage is made up of water and sediment storages, Figure 1.5.6f shows the small interannual variability for sediment heat storage. This shows the variability in the total heat storage is mainly due to changes in water heat storage with minor influences by sediment heat storage. While temperatures determine directionality in the heat storages, they do a poor job in describing the larger magnitude of variability in 2009. Figure 1.5.6e shows the variability in lake level for the JAS period over the three years with 2009 being much higher than 2007 and 2008. The higher lake water levels in 2009 significantly increased the magnitude of variability in the lake heat storage term thereby causing the large variability.

Lake level fluctuations are instrumental in the interannual differences seen mainly for 2009 in temperature dependent terms. Physically this makes sense as more water will be capable of storing or releasing larger amounts of energy. This is especially evident in total heat storage which experiences larger variability in 2009 (Figure 1.5.6d). It is easy to see how increased lake levels would increase the variability in water heat storage because daily lake levels are a main component in its calculation. Interestingly, increased lake levels also make Alkali Lake have some characteristics of a deeper lake. For
example sensible heat is much more positive in 2009 (Figure 1.5.6b), suggesting that the lake is capable of retaining larger amounts of energy and dispensing it into the atmosphere over a longer time period than the previous two years. Oddly, this appears to have no effect on evaporation rates as evaporation and sensible heat typically fluctuate in tandem (Parkhurst et al., 1998; Robertson and Barry, 1985).

Other lake-atmospheric interactions such as the vapor pressure gradient appear slightly different in 2009, though differences are less than the estimated uncertainty (Figure 1.5.7a, Table 4). That being said when combined with wind speed Figure 1.5.7f, the years become much similar as shown in Figure 1.5.7c. The main cause for this is the likely covariation between vapor pressures and wind speeds; however, since the differences are within uncertainty limits, more significant trends may emerge at shorter timescales. Similarly to net radiation, relative humidity and wind speed both show very little difference between years, again this is likely due to the similarities in weather patterns.

1.7 SUMMARY AND CONCLUSIONS

Alkali Lake is a shallow, saline lake located in a unique part of the Sandhills of western Nebraska. This is a semi-arid region with numerous endorheic lakes, making it an excellent laboratory for field studies of evaporation from saline lakes. This study has examined the seasonal and interannual variability in lake evaporation over a 3-year period using the Bowen ratio energy balance method. The majority of data were collected during the months of July, August, and September, when evaporation rates at Alkali Lake
averaged 5.6 mm/day across all three years (2007-09). Calculations of evaporation using
the mass transfer technique were also made to aid in the QA/QC process and to provide
approximate evaporation estimates during the winter of 2008-2009.

We conclude that seasonal variations in evaporation are larger than interannual
variations and are primarily controlled by similar seasonal changes in net radiation. This
is consistent with other studies of shallow water bodies in the area (Burba et al., 1999;
Parkhurst et al., 1998) and is also a common finding of shallow lakes in general
(Brutsaert, 1982). Alkali Lake also displayed other similar characteristics to those
generally observed in shallow lakes, such as relatively small seasonal variability and
small mean values of sensible heat flux and the rate of heat storage (both in the water and
sediments).

Though Alkali Lake exhibits a fairly pronounced seasonal cycle in evaporation, very
large daily variability was also observed, suggesting that no two years are identical, in
terms of their general, seasonal pattern. Similar conclusions were reached in other multi-
year lake evaporation studies, where it has been noted that large intraseasonal variation
can cause individual years to be very different from long-term averages (Lenters et al.,
2005; Parkhurst et al., 1998). Seasonality in other meteorological variables is also evident
at Alkali Lake, with the mass transfer product of wind speed and vapor pressure gradient
closely matching that of evaporation. This shows the potential merit for using the less
costly and time-consuming mass transfer method in describing seasonal evaporation rates
at Alkali Lake in future studies.

Interannual variations in evaporation rate were much more limited, with mean
differences between years being less than 15 W m$^{-2}$, which is well within the estimated
uncertainty of 22 W m\(^2\). Similar conclusions were reached for net radiation, where as year-to-year changes in sensible heat flux were more noticeable, relative to their uncertainty (and backed up by similar differences in the lake-air temperature gradient). This included higher sensible heat fluxes and lake-air temperature gradients in 2009, when larger variations in heat storage rate were also observed. Both of these effects can likely be attributed to higher lake levels observed in 2009, which caused the lake to store more energy, release it over a longer time period, and generally have higher water temperatures relative to the ambient air temperature. These are all characteristics of deeper lakes, but they have been observed at the smaller scale of Alkali Lake, despite what might otherwise be considered inconsequential changes in water level (i.e., roughly 25 cm).

**CHAPTER 2: Daily energy budget variations and associated meteorological influences.**

**2.1 INTRODUCTION**

Free surface evaporation can have critical implications in the hydrologic cycle, and is an important factor in water resources, irrigation, and ecosystem management. Evaporation determination via the energy balance technique is a difficult and time consuming therefore, much uncertainty still exists where less costly empirical methods have been used. While there are numerous ways to estimate evaporation, the Bowen ratio energy balance (BREB) method is considered one of the most accurate methods (Lenters et al., 2005; Winter et al., 2003). The energy balance approach determines rates of energy inputs and outputs for all non-negligible components; then distributes the energy into latent and sensible heat fluxes by the experimentally estimated Bowen Ratio. This
involves measuring variables such as solar radiation, advection, and the energy stored in different substrates such as lake water and sediment. How these different energy fluxes interact with a water source of interest will control the energy available to evaporation. While the energy balance is not a direct estimate of evaporation, it is the preferred technique for long term evaporation measurements (Assouline and Mahrer, 1993; Winter et al., 2003). The energy budget method has been widely used for freshwater lakes (Gallego-Elvira et al., 2010; Lenters et al., 2005; Winter et al., 2003), but to a much lesser extent on saline lakes (Lensky et al., 2005).

Whereas freshwater evaporation is controlled by meteorological variables and lake physical characteristics, saline lakes present an additional challenge in that salinity/density must also be accounted for (Oroud, 2001). Salinity is an important consideration because it changes the thermophysical properties of water and reduces the saturation vapor pressure above a saline lake. This effectively reduces the vapor pressure gradient between the lake surface and surrounding air. It has been well documented that evaporation from a saline water source is smaller than that of a freshwater source under the same meteorological conditions (Harbeck, 1955; Oroud, 1995; Salhotra, 1985). Further complexity occurs when saline lakes exhibit significant lake level changes. Salinity is sensitive to changes in lake water levels meaning evaporation cannot be considered constant when water level changes occur (Oroud, 2001). This makes depth an important factor in determining evaporation in both saline and freshwater lakes. Shallow lake evaporation typically follows the cyclic behavior of solar radiation mainly on a daily and seasonal time scale (Brutsaert, 1982). Therefore it would be expected that the seasonal evaporation rates closely follow the energy available to evaporation.
Evaporation was found to vary with available energy and water in shallow groundwater discharge playas using the energy balance approach (Menking et al., 2000). In contrast, deeper lakes have the ability to store and release larger amounts of energy. This may cause a lag in available energy for evaporation up to several months after the peak in solar radiation (Brutsaert, 1982). On a freshwater lake in Ontario, Canada, Yao (2009) found that a lag time existed between net radiation and peak evaporation rates using the energy budget technique. No studies have calculated evaporation from any saline Sandhills lakes; however, Oroud (1997) found that hypersaline conditions lowered evaporation rates by 40% in the summer months in small shallow ponds near the Dead Sea. Despite the widespread applications of the energy budget method it has not been applied to any of the numerous saline lakes in the Nebraskan Sandhills.

The Nebraska Sandhills, with an area of 58,000 km², is the largest vegetated dune field in the Western Hemisphere. The dune field is currently stabilized by vegetation but was extensively active during the Holocene (Hanson et al., 2009; Mason et al., 2004). Recharge into the well-drained dune field during the Pleistocene raised the groundwater levels and formed numerous interdunal lakes (Loope et al., 1995). Over 2000 lakes are found in the Sand Hills (Bleed and Flowerday, 1998) and have total dissolved solids ranging from freshwater (~0.3 g/L) to saline water, that is greater than 100g/L (McCarraher, 1977). The generally eastern sloping land characteristics of the Sandhills has been interrupted on the western side creating closed (endorheic) basins (Bleed and Flowerday, 1998). These closed basin lakes have very unique water chemistry and salinity aspects because mineral deposition from groundwater seepage becomes concentrated by evaporation. Evaporation has been shown to be a large driver in the
solute balance of the saline Sandhills lakes (Zlotnik et al., 2012, 2010). As the western Sandhills is classified as a semiarid region, the presence of such high salinities suggests that groundwater interactions are a key water source for these lakes.

Evaporation rates from any of the Sandhills lakes are still poorly understood, and require further study to determine their primary controls and variability. Despite the useful information provided by previous energy balance studies many are somewhat limited in scope, and an extensive analysis of the primary drivers has not been done in the Sandhills. For example, Parkhurst et al. (1998) provide an analysis of the seasonal trends in energy balance evaporation, but focus on energy components with little explanation of other atmospheric influences. This is a trend with many energy balance studies (Robertson and Barry, 1985; Winter et al., 2003). One of the most intensive analyses examining the climatic mechanisms on energy balance evaporation at multiple temporal scales was done by Lenters et al. (2005). They acknowledge that the energy budget is limited in understanding climatic mechanisms because evaporation is a driver of lake heat storage and a response to radiation. However, one may also argue that the vapor pressure difference between the lake surface and the air are more fundamental drivers of evaporation as they depend on temperature, humidity, and wind speed (Lenters et al., 2005). This makes climatic factors potentially useful in understanding the causes of variability in energy balance evaporation rates at numerous temporal scales. For example the vapor pressure difference explained about 46% of the variation in intraseasonal evaporation rates, which includes the effects of temperature and relative humidity (Lenters et al., 2005).
The objectives of this study are to determine daily evaporative rates and the daily controls on evaporation. This study will examine the causes of daily variability in energy balance evaporation rates from one of the many Sandhills lakes. It is likely that available energy will be an important factor, but that other meteorological factors such as wind speed and vapor pressure gradients will also play a role in daily evaporative variability. The study takes place at Alkali Lake a typical saline lake located in the western Sandhills of Nebraska. Please see chapter 1 for a complete site description.

2.2 RESULTS

2.2.1 De-Seasonalized Daily Variability

Daily evaporation was calculated using the same methods and dataset as in chapter 1. The daily temporal scale was analyzed by removing the seasonal cycle so that only interannual and intraseasonal effects remained. To do this the seasonal cycle (4th order polynomial from the seasonal analysis) was subtracted from each day creating daily anomalies. It should be noted that a certain year or years will be more influential outside the common time period of all three years (Jun. 29 – Oct. 11). The general pattern across all variables was that the polynomial fit improved up until the 4th order with orders above offering little improvement. After removing the seasonality, the grand mean (average from the common period across all three years) was then added back into the anomalies to create a deseasonalized dataset. The daily anomalies were then regressed against each other to find which variables were the most influential to evaporation and other energy balance components.
The most interesting short term relationship is between evaporation and sensible heat. A timeseries of the deseasonalized 5-day running means for both E and H shows the potential for a negative relationship (Figure 2.2.1). This appears to be consistent across all data and years. This is unusual because E and H tend to vary in tandem, and can be contrasted with an energy budget study done on an arctic thermokarst lake (Potter, 2011). By examining the drivers of this relationship it also becomes easier to describe the relationships between daily evaporation rates and the other energy terms. The relationship between E and H is indeed negative in both the raw and deseasonalized data, although the deseasonalized has a much stronger relationship (Figure 2.2.2a/b). The relationship between deseasonalized evaporation and the temperature gradient yield a negative relationship of $R^2 = 0.273$, whereas sensible heat has a positive relationship with the temperature gradient at $R^2 = 0.806$ (Figure 2.2.2c). Lastly, the regression between the vapor pressure and temperature gradients shows no relationship and is insignificant (Figure 2.2.2d).
Figure 2.2.1 Deseasonalized 5-day running means for each year of energy balance data.

Latent energy (LE) is in black while sensible heat (H) is in red.
Figure 2.2.2 Relationships between evaporation (E) and sensible heat (H), as well as their regressions with the temperature gradient. a) Unaltered relationship between E and H which includes seasonal and interannual variability b) Deseasonalized regression between E and H c) E and H’s relationship to the temperature gradient d) Relations between the vapor pressure gradients with and without wind speed, and temperature gradient.

Figure 2.2.2 explores the impact of daily covariances with wind speed using the mass transfer relationships. Figure 2.2.2a shows the mass transfer relationship for
evaporation with a strong positive relationship of $R^2 = 0.805$. Similarly, a strong positive relationship is seen for the sensible heat mass transfer relationship with a $R^2 = 0.981$ (Figure 2.2.3b). Figure 2.2.3c shows the relationship between evaporation and the individually averaged components of the mass transfer relationship, with still a strong positive relationship. Figure 2.2.3d also displays individually averaged wind speed times temperature gradient for the sensible heat mass transfer regression. Figure 2.2.3c/d have worse regressions than Figure 2.2.3a/b with $R^2$ values of 0.710 and 0.917 respectively. Figure 2.2.3e has a poorer regression of only 0.171 between evaporation and the vapor pressure gradient. Finally, Figure 2.2.3f shows the relationship between sensible heat and the temperature gradient, with an $R^2 = 0.810$. 
Figure 2.2.3 Explores the covariances between wind and vapor pressures or temperature gradients. a) Evaporative mass transfer relationship with the daily averaged product of wind speed and vapor pressure gradient b) Sensible heat mass transfer relationship with the daily averaged product of wind speed and temperature gradient. c) Evaporation regressed against the individually averaged wind and vapor pressure gradient d) Sensible
heat regressed against the individually averaged wind and temperature gradient e) Evaporation vs vapor pressure gradient f) Sensible heat vs temperature gradient.

Figure 2.2.4 examines the influences of wind on evaporation and sensible heat. Figure 2.2.4a shows the regression between evaporation and wind. A small, but significant positive regression exists with an $R^2 = 0.085$. Sensible heat experiences an opposite regression with wind speeds (Figure 2.2.4b). This regression is also small and significant; however, it is negative with a $R^2 = 0.061$. The relationships between vapor pressures and wind are examined in Figure 2.2.4c. Initially the vapor pressure gradient is negatively correlated with wind speed, with an $R^2 = 0.267$. When multiplied together, without the diurnal covariances, the relationship becomes positive with a slightly smaller $R^2 = 0.197$. Results for the regression between temperature gradient and wind show that initially, a very small negative relationship exists (Figure 2.2.4d). This correlation becomes stronger when multiplied with wind, evident by the increasingly negative slope.
In order to further investigate the link between evaporation and sensible heat, Figure 2.2.5 explores their relationships with air temperature. Figure 2.2.5a shows the regressions between surface and air temperature as well as the temperature gradient and air temperature. They exhibit opposite regressions with the surface vs. air temperature.
having a positive regression of $R^2 = 0.698$ and slope of 0.613, while the temperature gradient vs. air temperature is negative with an $R^2 = 0.481$. Figure 2.2.5b indicate the different relationships evaporation and sensible heat have with air temperature. Increasing air temperature leads to higher evaporation ($R^2 = 0.252$), but lower sensible heat ($R^2 = 0.432$). Figure 2.2.5c explores the relationships between vapor pressures and air temperature. Surface and air vapor pressures, as well as their difference, experience significant positive relationships with air temperature. Figure 2.2.5d shows the relationship between relative humidity and the temperature gradient as well as air temperature. Relative humidity vs. the temperature gradient has a significant positive relationship of $R^2 = 0.305$, but the opposite is true for relative humidity’s relationship with air temperature only ($R^2 = 0.289$).
Figure 2.2.5 Relationships between numerous variables and air temperature. a) Surface and temperature gradient’s relationship with air temperature b) E and H’s relation to air temperature c) Each individual vapor pressure and its gradient d) Relative humidity’s relation to air temperature and temperature gradient.

Figure 2.2.6 shows some of the key relationships between evaporation and sensible heat with other energy terms. Figure 2.2.6a shows that both shortwave influxes (SWin) and net radiation are closely related and both have positive relationships with air temperature. Air temperature and net radiation have almost the exact same relationship
with wind speeds (Figure 2.2.6b). Both regressions are insignificant with slopes near -1. Evaporation and sensible heat both have negative regressions with lake heat storage (Figure 2.2.6c). The combination of evaporative and sensible heat fluxes also retains the negative relationship. Figure 2.2.6d shows the relationships between evaporation or sensible heat with net radiation. Evaporation has a positive relationship with net radiation, however, sensible heat has an opposite negative relationship. The combination of evaporative and sensible heat fluxes creates a positive relationship, more closely resembling that between evaporation and net radiation.

**Figure 2.2.6** Relationships between other energy terms and sensible or evaporative fluxes. a) Relation between shortwave influx or net radiation and air temperature b) Air
temperature and net radiation’s relationship with wind speed c) Regressions between evaporation or sensible heat with water heat storage d) Regressions between evaporation or sensible heat with net radiation.

The increasing variation between evaporation and net radiation was explored further in Figure 2.2.7. Evaporation was replaced with the mass transfer product and yields a very similar significant positive relationship with a $R^2 = 0.404$. The bottom (most negative) 10% of daily $H$ values are highlighted in red. This shows the balance between the main evaporative drivers, and gives evidence to where additional energy available to evaporation is coming from.
Figure 2.2.7 Balance between the main evaporative drivers. The mass transfer product regressed against net radiation yields a positive significant relationship with increased variation moving up the linear regression. Red dots indicate the bottom 10% (most negative) daily H values.

2.2.2 Salinity’s Influence on Mass Transfer and Energy Balance Methods
The effects of fluctuating salinity were examined for variables used in the calculation of the MT and EBBR methods. Figure 2.2.8 shows how changes to $a_w$ propagate into the MT estimates. Each histogram shows the difference between each variable when salinity was and was not accounted for. These are expressed as absolute and percent differences. Values of $a_w$ ranged from 0.94-0.99 making it very near freshwater values of 1. This is used as a coefficient for $e_s$ and its changes can be seen in Figure 2.2.8a. Changes in $e_s$ are generally small and less than a 6% reduction with a median value of -3.65% (Figure 2.2.8b). Reductions in $e_s$ also result in decreased values of the mass transfer product, which is the product of wind speed and vapor pressure gradient (Figure 2.2.8c/d). The median absolute reduction is 0.32 kPa m s$^{-1}$, which corresponds to a decrease of 8.94%. Occasionally, much larger percent decreases are observed when already small products become smaller. Finally, when the mass transfer product is regressed against EBBR evaporation, absolute differences in fresh and saline MT can be quite large approaching a maximum near 40 W m$^{-2}$ (Figure 2.2.8e). The percent differences are usually small with the majority being less than 10%, but much larger changes do occur (Figure 2.2.8f).
Figure 2.2.8 Histograms indicating salinity’s effects to the surface vapor pressure, mass transfer product, and mass transfer evaporation rates if it is unaccounted for. Left column indicates the absolute differences, and the right are the percent differences. Red dashed lines indicate median values. a) Absolute difference in surface vapor pressure (kPa) b) Percent difference in surface vapor pressure (%) c) Absolute difference in mass transfer products (kPa m s\(^{-1}\)) d) Percent difference in the mass transfer product (%) e) Absolute difference in mass transfer evaporation (W m\(^{-2}\)) f) Percent difference in mass transfer evaporation (%).
The effects of salinity on EBBR evaporation are slightly more complicated because salinity will change water heat storage and the Bowen ratio. Figure 2.2.9 highlights these changes and the effect on EBBR evaporation. Water heat storage changes very little with an absolute median value very near zero (Figure 2.2.9a). All changes to water heat storage are less than 3 W m\(^{-2}\) corresponding to less than a 5% increase or decrease (Figure 2.2.9b). Bowen ratios largely experience very little change with a large cluster around 0 (Figure 2.2.9c). There are a few outliers in the absolute Bowen ratio change, but they can be attributed to changes in very small numbers. Similarly, the percent change in Bowen ratios is also much higher than any other variable simply because they tend to be very near zero (Figure 2.2.9d). The absolute changes in EBBR evaporation can be seen in Figure 2.2.9e. They show a large cluster of values around zero, with a median of only -0.17 W m\(^{-2}\). Percent changes in EBBR evaporation have a similar distribution with a large number of values near a zero percent change (Figure 2.2.9f).
Figure 2.2.9 Histogram indicating salinity’s effects on water heat storage, Bowen ratio and energy balance evaporation rates. The left column indicates the absolute differences, and the right are percent differences. Red dashed lines indicate median values. a) Absolute differences for water heat storage (W m$^{-2}$) b) Percent difference for water heat storage (%) c) Absolute changes in Bowen ratios d) Percent change in Bowen ratios (%) e) Absolute changes in energy balance evaporation rates (W m$^{-2}$) f) Percent changes for energy balance evaporation rates (%).

2.3 DISCUSSION
2.3.1 Daily Variability

Examinations of daily variability using the Bowen ratio energy balance method typically are not done to this extent making this study unique. Most studies using energy balance evaporation methods focus on energy fluxes and how they influence evaporation (Parkhurst et al., 1998; Stannard and Rosenberry, 1991; Winter et al., 2003). It is rare for energy budget studies to go into depth in both energy and atmospheric components. It has been well established that incoming solar radiation is the main driver of evaporation from shallow fresh and saline waters (Brutsaert, 1982; Burba et al., 1999; Oroud, 1997; Parkhurst et al., 1998). Perhaps equally important however are the atmospheric forcings, most notably the mass transfer product. This section will discuss the energy and atmospheric forcings at the daily scale, and how they influence evaporation rates at Alkali Lake.

In this case daily variability is best described by starting with one of the most interesting relationships; the negative relationship between evaporation and sensible heat. This relationship is particularly unusual because they tend to vary in tandem as shown in Lenters et al. (2005), Potter (2011). Other examples, not specifically commented on, can be seen in numerous energy budget studies (Hostetler and Bartlein, 1990; Sturrock et al., 1992; Yao, 2009). The possibility of a negative relationship between evaporation and sensible heat has been shown in some studies of diurnal evaporation rates; however, to this author’s knowledge none have specifically commented on it or presented a detailed analysis of the relationship (Assouline and Mahrer, 1993; Tanny et al., 2008).

Figure 2.2.2 a/b shows the curious relationship between evaporation and sensible heat in both raw and deseasonalized values as suggested by the deseasonalized 5-day
running means in Figure 2.2.1. The relationship is weaker in the raw data because the seasonal variability is still included where evaporation fluctuates mainly with net radiation and sensible heat oscillates around zero. Breaking down the relationship between evaporation and sensible heat flux one finds that they have opposite regressions with the temperature gradient (Figure 2.2.2c). This suggests that air temperature is somehow partially responsible. Figure 2.2.2d shows that the negative $E$ vs. $H$ regression may also be diagnosed through the mass transfer relationship. The relationship remains negative broken down further into $U(es-ea)$ vs. $Ts-Ta$, but the removal of wind creates an insignificant relationship between vapor and temperature gradients (Figure 2.2.2d). This suggests that there are further complicating effects with wind and relative humidity. Figure 2.2.2 points to four potentially influential forces in the negative $E$ vs. $H$ relationship, air temperature, wind speed, relative humidity, and their diurnal covariances.

Figure 2.2.3 shows the diurnal covariances in the mass transfer products of $E$ and $H$. The strongest regressions for both evaporation and sensible heat are for the daily averages of the hourly products of wind speed and vapor pressure gradient for evaporation, or wind speed and temperature gradient for sensible heat (Figure 2.2.3a/b). This is not particularly surprising, since it is well known that use of long term averaging of individual variables can induce significant error (Brutsaert, 1982; Jobson, 1972; Kondo, 1972; Webb, 1964, 1960). The errors caused by long term averaging of variables before calculating the mass transfer product are due to a diurnal covariance between wind and air temperature (Jobson, 1972; Kondo, 1972; Webb, 1964, 1960). To avoid these systematic errors one can simply compute the required quantities at short period means
before averaging (Hage, 1975). It is evident by Figure 2.2.3 c/d that the diurnal covariances do significantly impact their relationships, and therefore calculation of products at hourly intervals is important in maintaining accurate results. Interestingly the removal of wind entirely creates a much weaker regression between evaporation and the vapor pressure gradient (Figure 2.2.3e). This suggests that while a larger vapor pressure gradient enhances evaporation, the transport mechanism of wind is perhaps equally important as a catalyst. Sensible heat also experiences a reduced regression when wind is removed (Figure 2.2.3f). This is a much smaller reduction than evaporation suggesting that convective forces are important, but not as critical in comparison to evaporation.

Figure 2.2.4 explores the relationship between E and H with wind speeds. Evaporation shows a slight positive relationship with wind speed in Figure 2.2.4a. Similarly, sensible heat has a small relationship with wind speed though it is negative (Figure 2.2.4b). This suggest that wind speeds have some type of relationship with the vapor and temperature gradients. The regression of es-ea vs. U shows that wind does indeed have a relationship with vapor pressure gradients and it is negative (Figure 2.2.4c). This is important for the product U(es-ea) in that small vapor pressure gradients will be greatly enhanced by wind. This is shown by the regression of the daily averaged product of wind and vapor pressure ([U]*[es-ea]) vs. U (Figure 2.2.4c). A positive relationship is created when wind speeds are multiplied with the vapor pressure gradients. This also shows that the diurnal covariances are not solely responsible for the change in regression. The relationships between U and Ts-Ta are slightly different (Figure 2.2.4d). Again a negative relationship between temperature gradient and wind speeds is seen, meaning when the air is warmer than the lake wind speeds are higher. This relationship is
heightened when they are multiplied together creating an even more negative relationship (Figure 2.2.4d). The opposite regressions between wind speeds and vapor and temperature gradients help to enhance the negative relationship seen in E vs. H.

The breakdown of the air temperature relationships helps get to the root of the negative E vs H regression. The relationship between air and surface temperatures is positive, strong, and significant as expected; however, it does show some disconnect as the slope is only 0.613 far from the 1:1 line (Figure 2.2.5a). This means that surface temperatures react rapidly trying to reach an equilibrium with air temperatures, but cannot keep up resulting in larger temperature gradients. This can be seen in Figure 2.2.5a between Ts-Ta vs. Ta. As air temperatures increase so does the temperature gradient and vice versa in that it does not cool as fast as air temperatures. Figure 2.2.5b shows how both evaporation and sensible heat react to changes in air temperature. Both have strong and significant regressions, but they are opposite each other meaning high temperatures lead to elevated evaporation rates and lake warming. While this is important to describing the negative E vs. H relationship, it does not explain why evaporation increases with air temperature. In fact, air temperatures are usually negatively related to evaporation because higher temperatures tend to lead to higher relative humidity and as such lower vapor pressure gradients (Lenters et al., 2005). Figure 2.2.5c/d explains why this is not the case. The surface and air vapor pressure gradients are both positively related to air temperature meaning the absolute amount of water in the air increases with evaporation (Figure 2.2.5c). However, there is a disconnect between surface and air vapor pressures because the difference between the two is also positive (Figure 2.2.5c). This means that surface vapor pressure increases are accompanied by air vapor pressure
increases at a much reduced rate. This leads to Figure 2.2.5d because the difference between the vapor pressures is relative humidity. Figure 2.2.5d shows that relative humidity decreases as air temperatures rise. This is the cause of positive relationship between vapor pressure difference and air temperature, meaning that the dryness of the region is helping contribute to the negative E vs. H relationship.

In summary there are four main contributors to the negative E vs. H relationship, air temperature, relative humidity, wind speed, and diurnal covariances. First, elevated temperatures create larger vapor pressure gradients (Figure 2.2.5c). While the absolute measure of water vapor increases with air temperature, the relative humidity decreases meaning hot and dry conditions tend to occur simultaneously (Figure 2.2.5d). However, these relationships alone would not make the negative E vs. H relationship as evidence by Figure 2.2.2d showing the relationship between the vapor and temperature gradients. Wind is needed to enhance the vapor pressure gradient as seen in Figure 2.2.4c. This creates the furthest broken down state where the negative relationship still exists (Figure 2.2.2d). Figure 2.2.4c also shows how the relationship will exist even without the diurnal covariances. This does not mean that the diurnal covariances are not important, rather that they help to enhance the negative E vs. H relationship (Figure 2.2.3). All together then these main influences help create the negative evaporative and sensible heat relationship. A simple analogy for these conditions can be made by comparing Alkali Lake to a sling psychrometer. A sling psychrometer is typically used to measure relative humidity by means of differences between dry and wet bulb thermometers due to evaporative cooling. In this case Alkali Lake is acting like a wet bulb thermometer. The hot, dry and windy conditions cause high evaporative rates that limit lake surface
temperature increases because of the feedback effects of evaporative cooling. This creates a large sensible heat flux into the lake, while evaporation rates remain high.

Exploring some other relationships between meteorological variables we see that higher air temperatures are associated with larger net radiation values (Figure 2.2.6a). This means that high air temperatures, and therefore reduced relative humidity, are associated with larger energy inputs. As already explained that does not necessarily mean high evaporation rates as wind acts as a catalyst to increase evaporation. This is especially apparent in Figure 2.2.6b as both air temperature and net radiation have no significant relationship with wind speeds. Evaporation and sensible heat both show negative relationships with water heat storage (Figure 2.2.6c). This is not particularly surprising because it indicates that when energy is leaving the lake energy storage decreases. Net radiation also has an expected relationship with both evaporation and sensible heat (Figure 2.2.6d). Higher energy inputs lead to more evaporation, and increase air temperature (Figure 2.2.6a). The increase in air temperature causes a negative temperature gradient into the lake, and therefore negative sensible heat (Figure 2.2.6d). Lastly, the increasing variability between the mass transfer product and net radiation is rather curious because it suggests there is another source of energy significantly impacting evaporation (Figure 2.2.7). This is indicative of the delicate balance between meteorological forcings as net radiation and the mass transfer product are both considered large drivers of evaporation. This is not to say that other variables do not matter at the daily scale. In Figure 2.2.7, the bottom 10% of sensible heat terms are highlighted in red. This shows that the large negative sensible heat values are a source of energy available to evaporation at this time causing higher evaporation rates. Rather than saying any one
term is the primary driver of evaporation at the daily scale, these findings indicate that a collaboration of events occur to drive evaporation rates at Alkali Lake.

### 2.3.2 Influence of Salinity

It was unknown whether water activity, specific heat, density, etc. are important factors at Alkali Lake, or if it could be neglected and methodology for freshwater lakes may be used. It should be noted that Alkali Lake is not being compared to a completely freshwater counterpart. Rather, we examine the use of constant freshwater or fluctuating saline values in the EBBR and supplementary MT evaporation estimates. Numerous comparisons between EBBR and MT evaporation estimates have been done before, and a wealth of literature is available on the subject (Lee and Swancar, 1997; Rosenberry et al., 2007, 2004; Winter et al., 1995; Yao, 2009). The MT method is more susceptible to salinity changes because it is based on the regression between EBBR determined evaporation and the product of wind and vapor pressure gradient; therefore, a change in surface vapor pressure reduces the gradient and the resulting evaporation estimate. Figure 2.2.8 shows a series of histograms detailing the absolute and percent difference between when \( a_w < 1 \) (“Saline”) and \( a_w = 1 \) (“Fresh”). It also shows how a change in \( e_s \) propagates into the eventual MT evaporation estimate. If salinity is not taken into account the resulting root mean square error (RMSE) for MT evaporation is 13.8 W m\(^{-2}\). Figure 2.2.8e shows that while the majority of changes are small, but occasionally the absolute differences for individual days can be quite large approaching almost a 40 W m\(^{-2}\) difference. The range of \( a_w \) values at Alkali Lake was 0.94 - 0.99 making salinity impacts smaller than other studies usually on hyper saline lakes where percent differences in
evaporation approach as much as 60% between actual freshwater and saline surfaces under the same conditions (Oroud, 1995; Salhotra, 1985; Turk, 1970).

EBBR evaporation rates take into account the same salinity affected variables but respond differently as can be seen in Figure 2.2.9. Salinity will affect water heat storage (eq. 5) through differences in density and specific heat. While these terms vary in opposite directions they do not offset each other and cause very small changes (< 5%) in the water heat storage term (Figure 2.2.9a/b). Depending on whether energy is directed into or out of the lake, the resulting absolute and percentile changes will be either positive or negative; however, in both cases energy transfer is reduced. The absolute changes in water heat storage are very small (< 3 W m\(^{-2}\)). This makes salinity a very minor concern for water heat storage. Secondly salinity will affect EBBR evaporation through the Bowen ratio. As shown in eq. 3, the reduced saturation vapor pressure will change B. Figure 2.2.9c/d demonstrates how much changes in \(e_s\) affect B. Most changes to B are very small clustered around zero; however, occasionally larger changes occur.

Combining the salinity effects on S and B the resulting changes to EBBR evaporation are shown in Figure 2.2.9e/f. The absolute changes are small and generally less than 10 W m\(^{-2}\) with a large clustering around zero and a RMSE of only 2.5 W m\(^{-2}\). This is not surprising in that B acts as a corrective term in eq. 2 and errors in it its calculation create smaller errors in evaporation (Anderson, 1954).

Salinity had a minor influence on the determination of MT and EBBR evaporation at Alkali Lake. However, the effects of salinity on each evaporation equation vary with the MT method being more sensitive to changes in salinity. The mass transfer method also has a bias of overestimating evaporation rates that the energy balance does not. This
is evidenced by the distribution of error in each method with the energy balance being centered about zero (Figure 2.2.9e), but the mass transfer always overestimating evaporation (Figure 2.2.8e). These effects are not surprising and are due to the differences in how evaporation is calculated in each method. Overall, the absolute differences in EBBR evaporation Figure 2.2.9e show that neglecting changes in salinity would result in small errors that may justify ignoring the effects of salinity on evaporation. The MT method experiences much larger changes in evaporation due to salinity (Figure 2.2.8e). While the relative difference between using fresh and saline values is still small, occasional large differences occur on individual days making salinity a much more important variable in MT evaporation determination. In this case MT estimates have a relatively large error when compared to the EBBR already (RMSE 21.6 W m\(^{-2}\)), and neglecting salinity would only serve to increase error making the MT a poorer supplementary evaporation estimate.

2.4 SUMMARY AND CONCLUSIONS

Despite numerous energy balance studies, a detailed analysis of surface evaporation from one of the many western Nebraskan Sandhills lakes was lacking. This study successfully calculated evaporation rates at Alkali Lake for the daily timescale and examined the sources of variability. While numerous Bowen ratio energy balance studies have looked at the influences of the energy terms, very few have gone into the same amount of detail presented here. It was found that daily evaporation rates were controlled by a complex combination of factors.
Examination of daily evaporative variability was done through the negative relationship between evaporative and sensible heat fluxes. This relationship was of particular interest because it shows a negative correlation, and typically evaporation and sensible heat vary in tandem (Hostetler and Bartlein, 1990; Lenters et al., 2005; Sturrock et al., 1992; Yao, 2009). Similar to longer timescales, net radiation was found to be the largest source of energy to evaporation; however, other meteorological factors were instrumental at influencing evaporation rates. At Alkali Lake it was found that there were four main factors influencing the negative relationship between evaporation and sensible heat. Those factors were air temperature, atmospheric moisture, wind speed, and their diurnal covariances. The initial hot and dry conditions, common in this semiarid region, are the first step to creating large evaporation rates. These alone do not lead to large evaporation rates, rather wind is required as a transport mechanism to remove water vapor from the lake surface. Finally, proper calculation of the diurnal covariances further increases evaporation rates. While evaporation rates were high it was found that a negative temperature gradient was found suggesting possible effects of evaporative cooling similar to the effects seen with a sling psychrometer. These conclusions show that high daily evaporation rates at Alkali Lake result from a complex collaboration of air temperature, atmospheric moisture, wind speed, and diurnal covariances. These same factors lead to the unusual negative relationship between evaporative and sensible heat fluxes.

This study focuses on summer evaporation rates and variability simply because evaporation rates are highest. Salinity was found to influence the mass transfer method more substantially because it has biases that the Bowen ratio energy balance does not.
The mass transfer method resulted in a root mean square error of 13.8 W m\(^{-2}\), where the Bowen ratio energy balance had an error of only 2.5 W m\(^{-2}\). Neglecting salinity in the mass transfer method caused errors that consistently overestimated evaporation, while the salinity induced errors in the energy balance were centered about zero. It may be possible to neglect salinity when performing the BREB on a lake similar to Alkali, however salinity should be accounted for in MT estimates.

**CHAPTER 3: Alkali Lake water level variability and water budget interactions**

**3.1 INTRODUCTION**

The hydrologic cycle is the most basic water balance where water moves from one pool to another on a global scale in a cyclical motion. The system is based on conservation of mass where it is capable of receiving inputs, storage, and discharges of volumes of water (Dingman, 2002). A general water balance equation can be defined as

\[
P + G_{in} + R_{in} - (R_{out} + ET + G_{out}) = \Delta S
\]

where \(P\) is precipitation, \(G_{in}\) is groundwater inflow, \(R_{in}\) is stream flow in, \(R_{out}\) is stream flow out, \(ET\) is evapotranspiration, \(G_{out}\) is groundwater outflow and \(\Delta S\) is the change in storage. A water balance can be applied to a wide range of scales from global to single entities (USGS, 2012). This means that while the hydrologic cycle is applied to a global scale it is made up of many linked subsystems on regional or local scales. Water balances are used at these wide range of spatial scales to quantify movement, influences and
variability of numerous components which are usually unique to the system of interest. Combining the individual components of the water balance one can get a better understanding of the system as a whole. One of the more unique and interesting subsystems is the Nebraskan Sandhills; with many studies focusing on regional water balances because of its ability to recharge groundwater (Billesbach and Arkebauer, 2012; Ginsberg, 1987; Sridhar et al., 2006).

The Nebraska Sandhills, with an area of 58,000 km², is the largest vegetated dune field in the Western Hemisphere. The dune field is currently stabilized by vegetation but was extensively active during the Holocene (Hanson et al., 2009; Mason et al., 2004). Recharge into the well-drained dune field during the Pleistocene raised the groundwater levels and formed numerous interdunal lakes (Loope et al., 1995). Over 2000 lakes are found in the Sand Hills (Bleed and Flowerday, 1998) and have total dissolved solids ranging from freshwater (~0.3 g/L) to saline water, that is greater than 100g/L (McCarraher, 1977). The generally eastern sloping land characteristics of the Sandhills has been interrupted on the western side creating closed (endorheic) basins (Bleed and Flowerday, 1998). These closed basin lakes have very unique water chemistry and salinity aspects because mineral deposition from groundwater seepage becomes concentrated by evaporation. Evaporation has been shown to be a large driver in the solute balance of the saline Sandhills lakes (Zlotnik et al., 2012, 2010). As the western Sandhills is classified as a semiarid region, the presence of such high salinities suggests that groundwater interactions are a key water source for these lakes.

The Sandhills regional subsystem can be broken down further into smaller subsystems such as individual lakes, rivers, groundwater reservoirs etc. For example, a
water balance has been used to estimate evaporation outputs and stream inputs for individual lakes (Mohammed and Tarboton, 2012; Tanny et al., 2008). In individual lakes the water balance helps quantify inputs and outputs of water including variables such as precipitation, evaporation, groundwater exchange, and surface runoff. This knowledge can then be used to effectively manage lake ecosystems. Closed lakes are more sensitive to changes in inputs making effective management a high priority to maintain the lake system. For example the water balance of the Great Salt Lake was determined and modeled (Mohammed and Tarboton, 2012). Mohammed and Tarboton, (2012) found that the Great Salt Lake was most sensitive to changes in stream inputs and effective management of the stream flows would be needed to sustain current lake levels. The Great Salt Lake has been likened to the Aral Sea, an internationally known mismanagement of water resources (Bedford, 2009). Numerous papers have been written regarding the Aral Sea, and as such its water balance is now well understood (Glantz, 1999). The root cause of the shrinkage of the Aral Sea was the diversion of freshwater inputs for usage as irrigation (Bedford, 2009). The destruction of such a large ecosystem often serves as the prime example of a worst case scenario when water resources are mismanaged.

Of all the water balance variables evaporation is one of the most complex to accurately estimate. The most precise estimates of evaporation require knowledge of the energy balance parameters (Rosenberry et al., 2007; Winter et al., 1995; Yao, 2009). The connection between the water and energy balance occurs through the evaporative term in both balances. To quantify the evaporative loss of a lake using the energy balance latent energy needs to be calculated. Latent energy is energy going into the phase change from
water to water vapor, and through its quantification evaporation rates can be determined. While there are numerous ways to determine or estimate evaporation, the energy budget method is considered one of the most accurate methods (Lenters et al., 2005; Winter et al., 2003). Many other evaporation estimates exist but the energy budget is usually used as a benchmark that all others are compared to (Rosenberry et al., 2007, 2004; Winter et al., 1995; Yao, 2009). Evaporation rates from inland saline lakes are largely unknown because they commonly receive far less attention than freshwater lakes (Fritz et al., 2001). Whereas freshwater evaporation is controlled by meteorological variables and lake physical characteristics, saline lakes present an additional challenge in that salinity/density must also be accounted for (Oroud, 2001). It has been well documented that evaporation from a saline water source is smaller than that of a freshwater source under the same meteorological conditions (Harbeck, 1955; Oroud, 1995; Salhotra, 1985). Evaporation has not been calculated for any of the saline Sandhills lakes; however, freshwater wetlands in the region have been shown to have high evaporation rates (Burba et al., 1999; Winter et al., 2001). Once evaporation has been quantified through the energy balance a specific volume can be introduced into the water balance as one portion of the entire system.

Precipitation is the main source of water for Sandhills lakes and wetlands (Bleed and Flowerday, 1998). It affects lakes directly during rain events and indirectly through recharge of groundwater reservoirs that in turn affect lake levels. A significant precipitation gradient exists across the Sandhills where annual precipitation is more than 700 mm in the east, but only about 400 mm in the west (Szilagyi et al., 2011). Precipitation in the Sandhills is greatest from May-July when about 50% of precipitation
occurs (Bleed and Flowerday, 1998). This is because cyclonic storm paths pass over the region during this time and typically shift further northward in late summer. During July and August precipitation is mostly associated with convection or weak frontal passage (Bleed and Flowerday, 1998). This makes precipitation more variable over short distances in amount and intensity during late summer. Using a moderate resolution imaging spectroradiometer (MODIS) it has been shown the precipitation gradient contributes to recharge rates being greater in the eastern Sandhills when compared to the western (Szilagyi et al., 2011).

A difficult variable to assess in a water balance is the interaction with groundwater. The proximity of the groundwater table to the surface is crucial to the existence of many Sandhills lakes (Winter, 1986). How a lake responds to changes in groundwater levels depends on the characteristics of the groundwater-lake connection. A study in the south central portion of the Sandhills found that lakes in the region were partially or wholly connected with the groundwater table and many were groundwater discharge areas (Ginsberg, 1987; Winter, 1986). Flow between groundwater reservoirs and lakes may be complex and change with time. Winter (1986) shows how changes in groundwater levels affect the groundwater-lake interactions. Winter (1986) demonstrates how groundwater mounds can form in the Sandhills disrupting normal lake discharge into groundwater reservoirs. This blockage occurs when the mound is high and the gradient between the mound and discharge area is small, causing changes in flow direction. Winter (1986) also found that the water table configuration was unstable, and further complications arise from the changes in the water table with time. Groundwater gradients may also switch direction if changes in water table level are severe enough. Since the
Sandhills lakes are hydraulically connected to the groundwater system they are highly susceptible to changes in water table levels over time. Closed basin lakes are a direct display of the water table fluctuations, in that the groundwater influx is controlled by the hydraulic head under and around the lake (Almendinger, 1990). Changes in the hydraulic head from precipitation events can change groundwater flow direction stopping or slowing normally discharging areas (Webster et al., 2008). Since groundwater recharge is some percentage of precipitation, any changes in rainfall will affect the height of the water table and therefore the height of the lake. Work by Winter et al. (2001) shows comparisons between four wetlands one of which was in the western Sandhills. Results show varied management responses are needed for each region in terms of water system management. Winter et al. 2001 found that the Sandhills wetland responded to precipitation but were moderated by groundwater inputs. This makes pumping and water exports resulting from groundwater development a major threat to western Sandhills lakes (Winter et al., 2001).

The purpose of this chapter is to determine the water balance of Alkali Lake and the relative contributions of its water variables over short and long time periods. While water balances are nothing new to the Sandhills (Billesbach and Arkebauer, 2012; Sridhar and Wedin, 2009; Webster et al., 2008), or even to individual lakes within the Sandhills (Ginsberg, 1987; Winter et al., 2001), a water balance on any of the saline lakes has not been attempted. As there are numerous such lakes in western Sandhills, this analysis will help describe the relative amounts of water movement within the Alkali Lake system. This knowledge will provide insight for management priorities and protection of the alkaline lakes and their unique ecosystems.
3.2 METHODS

The dataset available prevents a volumetric water balance analysis due to the large lake level fluctuations and shallowness making bathymetric data very difficult to obtain. This creates a water balance based on height. The water balance of Alkali Lake can be described as:

\[
GW_{\text{net}} = \Delta h_L - (P - E)
\]

Where \(\Delta h_L\) is the daily change in lake height, \(P\) is precipitation, \(E\) is evaporation determined by the energy balance in chapter 1, and \(GW_{\text{net}}\) is the net groundwater contribution calculated as the residual. A positive net groundwater value indicates flow into the lake, and negative out from the lake. The highly porous sands create negligible runoff in this area allowing for the removal of surface water inputs from the water balance. As groundwater is a difficult variable to measure here it is calculated as the residual of the water balance. This common practice results in amplification of errors, resulting from the uncertainty in other water components that can create large uncertainty (Robertson and Barry, 1985; Winter, 1981). It is difficult to determine actual errors in each term because the true values are unknown and therefore must be estimated. Uncertainty in \(E\) is estimated at 0.76 mm/day resulting from the RMSE between EBBR and MT determined evaporation. A rough estimate of 10% for \(P\) and 20% for lake level change was used based on differences between sensors and their location. These
estimates make lake level the most uncertain term in determining the net groundwater. The average daily uncertainty in net groundwater flux from these estimates is 1.5 mm/day, but may be as high as 14.6 mm/day with the largest daily lake level flux.

3.3 RESULTS

Daily values of lake level and cumulative P-E can be seen in (Figure 3.3.1) for when data was available during 2007-09. Lake levels between 2007 and 2008 are similar but higher in 2009. Lake water levels show a pattern of being higher in late spring and early summer then falling until midsummer where they remain relatively steady. Due to longer data coverage in 2009, water levels can be seen rising from mid-October until early November. Similarly the spring of 2008 show fluctuating but steady lake levels before decreasing. The cumulative P-E values for each year decline through much of the study period, as evaporation greatly eclipses precipitation, with small upward spikes related to precipitation events. Changes in the slope of the cumulative (P-E) lines occur with much smaller slopes in the spring of 2008 and the fall of 2009.
Figure 3.3.1 Timeseries of daily average lake water levels and cumulative precipitation minus evaporation for 2007-09. Solid lines indicate lake levels, and dashed show cumulative (P-E).

The annual differences of each variable can be seen in Figure 3.3.2a. The JAS totals for each year reveal relatively little difference in the magnitude of each variable, meaning that each year was similar throughout the JAS period with the exception of net groundwater. Across the three years, JAS inputs averaged 1.7 mm/day for precipitation and 2.4 mm/day for net groundwater. Evaporation, the main output, eclipsed all inputs at an average of 5.2 mm/day resulting in a mean lake water level drop of 1.1 mm/day. The only large interannual difference occurred in 2007 when net groundwater was 3.2
mm/day, about 1 mm/day higher than the following two years. Averages of each variable during JAS can be seen in Figure 3.3.2b along with their percentages. Evaporation averages 50% of the water balance at this time, and is larger than precipitation and net groundwater inputs at 16% and 23% respectively. This creates the overall decrease in lake water levels throughout the study period which is 10% of the total water budget.
Figure 3.3.2 JAS averages for each water budget component. a) JAS totals for each individual year (2007-09). Units are in mm. b) JAS averages over all three years as percentages of the entire water budget during this time.

Daily net groundwater values can be seen in Figure 3.3.3. Values were separated into two week periods to show magnitude and variability and assess any trends. Net groundwater values averaged 2.5 mm/day with a maximum of 73 mm/day and a minimum of -45 mm/day occurring on consecutive days in 2007. The maximum and
minimum days appear as large outliers in the two week period of August 19th, and have been removed from Figure 3.3.3 because they dwarf all other variability. Each two week period has a positive average and median indicating net groundwater influx to the lake. Overall variability shows that while no apparent pattern exists, there is a larger occurrence of positive values than negative.

**Figure 3.3.3** Daily values of net groundwater separated into 2 week periods. Blue diamonds represent averages for that period, and red lines are medians. The red plus signs indicate outliers. Not shown are the extreme outliers of 73 mm/day and -45 mm/day during the Aug-19 period.
The short term relationships between the water balance variables reveal what the most influential components are on lake levels. The relationships are of the raw unaltered data at the weekly scale and shown in (Figure 3.3.4). Lake water levels regressed against evaporation have a small negative relationship (Figure 3.3.4a). Lake water levels are largely dependent on the net groundwater flux with a strong significant relationship (Figure 3.3.4b). The effects of precipitation on lake water levels can be seen in Figure 3.3.4c. Precipitation shows a strong positive relationship with lake water levels. It also shows a large cluster of values just above zero with a smattering of days with larger precipitation values. Figure 3.3.4d depicts the positive relationship between net groundwater and precipitation demonstrating the effects of precipitation on net groundwater.
**3.4 DISCUSSION**

**Figure 3.3.4** Relationships between water balance components based on weekly water balance calculations. Each dot represents one week of data. a) Relationship between lake water levels and evaporation b) Regression between lake water levels and net groundwater values c) Lake water levels regressed against precipitation d) Relationship between net groundwater levels and precipitation.
Daily values of lake water levels and cumulative (P-E) in Figure 3.3.1 are indicative of an evaporation dominated system as may be expected. The lake water levels begin to decline in midsummer as rainfall becomes less frequent, though small spikes can be seen in relation to precipitation events. This pattern can be seen among all three years despite higher lake levels in 2009. The higher lake water levels in 2009 can be attributed to greater annual precipitation. While data coverage is unavailable at Alkali Lake for a complete year, a nearby automated weather data network site located near Alliance, NE shows annual precipitation was about 431 mm in 2009, but only 328 mm and 262 mm for 2008 and 2007 respectively. While this dataset does not allow for a full seasonal analysis of the water balance some hypotheses can be gathered from the spring data of 2008 and fall data of 2009 depicted in Figure 3.3.1. Because Alkali Lake is a perennial lake and does not entirely dry up most years, precipitation and groundwater inputs must become larger influences at other times of the year to maintain lake presence. This means that Alkali Lake will typically experience higher lake levels in winter months that then begin to decrease through the summer. The seasonality of evaporation, where highest values typically occur in late June and slowly decline into the fall, means it is likely most influential during lake water level declines. This means that the balance between water inputs and outputs changes in the spring and fall. Evidence for this can be seen in Figure 3.3.1. The limited data in May 2008 show relatively stable lake levels and cumulative (P-E) values. As time progresses through the months of May and June cumulative (P-E) drops off sharply and lake levels begin to decline. The reverse is seen in the fall of 2009. Beginning around the start of October cumulative (P-E) starts to flatten out and lake levels gradually increase. This shows the change from dominating evaporative influences
to precipitation and groundwater influx, where the progression is reversed in the spring of 2008. Similar results have been found throughout the Sandhills, where in general there is a moisture surplus from late fall to early spring (inputs exceed outputs) and the reverse is true the rest of the year (Bleed and Flowerday, 1998).

Comparisons between years show little differences in the magnitude of the water balance components from Figure 3.3.2a. This may because of the limitation of only comparing the JAS period. Precipitation totals only differed by about 40 mm and evaporation by about 60 mm. This likely can be attributed to similar weather patterns during this time of year, where conditions are generally hot and dry with occasional brief convective thunderstorms. For groundwater net totals, 2007 stands out as higher than the following two years. Interestingly evaporation is slightly higher in 2007 but water levels dropped the least, which is the main cause of the higher net groundwater. This indicates that the lake and groundwater systems constantly influence each other.

Net groundwater averaged 2.5 mm/day for the entire study period and remain relatively steady throughout the year Figure 3.3.3. These results are similar but slightly higher than the estimated 1.3 ± 0.1 mm/day calculated by Ong, (2010); however, it should be noted that this value is within our level of uncertainty. From these results it can be concluded that Alkali Lake is largely gaining groundwater, but occasionally lake water is lost to the groundwater system. Notably these results largely agree with the literature in that Alkali Lake is mainly a discharge lake (Befus et al., 2012; Ong et al., 2010; Zlotnik et al., 2010). Outseepage from Alkali Lake was found to be extremely low and has been considered negligible (Ong, 2010; Zlotnik et al., 2012). This has been supported by direct methods such as potentiomanometer data and groundwater wells located on the eastern
and western shores of Alkali Lake and a well in the lake itself all showing an upward groundwater gradient. Work using continuous resistivity profiling has also been done on Alkali Lake (Befus et al., 2012). These results showed a mainly uniform electrical resistivity profile across the lake indicating upward groundwater seepage into the lake. Befus et al., (2012) also found a small area in the northeast area of Alkali Lake with low electrical resistivity at depth. This suggests downward lake seepage or the presence of fine sediments, but no seepage measurements or head gradients have been done in this area. Furthermore, electrical resistivity transects to the east of the lake and located down the groundwater flow gradient show the possibility of a saline plume from Alkali Lake (Ong et al., 2010). Again it is also possible that fine sediments have created the differences in electrical conductivity in that area. Also an analysis of the solute balance of Alkali Lake has revealed a lack salinity relative to what may be expected if salts never left the lake basin. This lack of salinity was contributed to eolian processes that actively transport salt dust away from the lake or reduced solute influxes (Zlotnik et al., 2012). Combined these studies show the possibility of downward seepage from Alkali Lake that is observed in Figure 3.3.3; however, these studies are also limited to representing a snapshot in time or a localized response over longer periods.

When recharge reaches the groundwater table depends on the thickness of the unsaturated zone (Winter, 1983). Also, localized groundwater flow reversal would be necessary for negative net groundwater fluxes to take place and have been previously described in the Sandhills (Winter, 1986). It is possible that a time lag exists between when precipitation effects the lake and groundwater systems. A cross correlation analysis (not shown) revealed no lag time at the daily scale suggesting the lake and groundwater
system reach a new equilibrium in a matter of hours not days. The porous sands and high groundwater table are likely instrumental in allowing for this fast recharge. Unfortunately an hourly analysis is beyond the scope of this paper; however, this provides the base knowledge for future works.

Relationships between water balance components were based on the weekly values of each water balance component. Initially, one may think the relationship between lake levels and evaporation should be stronger given how much larger evaporation is than any other water balance variable on longer timescales (Figures 3.3.1, 3.3.2). This is likely due to the fact that evaporation is being overshadowed by the effects of net groundwater flux on lake levels. Regardless, the negative relationship does show that higher evaporation rates lead to lower lake water levels Figure 3.3.4a. The strong positive relationship between lake water levels and net groundwater shows how reliant lake levels are on groundwater fluxes (Figure 3.3.4b). The direct influence of precipitation on lake levels is shown in Figure 3.3.4c. Since precipitation directly influences lake levels the strong positive relationship is expected, but precipitation amounts were very small in most cases as evidence by the large cluster just above zero, making them a small percentage of lake flux most days. Figure 3.3.4d gives evidence to the influence of precipitation on net groundwater. A significant positive relationship exists between net groundwater and precipitation. This means that precipitation causes a larger net groundwater flux into Alkali Lake; likely due to the effects of recharge that in turn lead to higher lake levels. At a Sandhills wetland mitigation site a negative relationship between groundwater and precipitation is suggested (Webster et al., 2008). They found that excessive precipitation events caused some of their nested piezometers to
register negative groundwater fluxes, which were attributed to similar hydraulic pressure heads. This speaks to the high level of complexity required to describe Alkali Lake’s interactions with groundwater.

3.5 SUMMARY AND CONCLUSIONS

Water balances are crucial in understanding movement of water from one pool to another. The Nebraskan Sandhills has a unique hydrologic cycle largely because of its ability to recharge groundwater. This recharge led to an elevated water table that intersects the surface in between sand dunes. Coupled with a low groundwater slope, this created numerous lakes and wetlands. This study accomplished a multiyear water balance of Alkali Lake, one of the numerous shallow saline lakes located in the western Sandhills region. The relative contributions of its water balance components were quantified and analyzed at short and long timescales.

Daily values of water levels and cumulative values of precipitation minus evaporation show that Alkali Lake is largely dominated by evaporation during the summer months. In all three years lake levels declined from spring to late summer when evaporation was highest. Since Alkali Lake is a perennial lake, it is hypothesized that better data coverage would find lake inputs exceeding evaporation at other times of the year, likely during the spring or early summer. Net groundwater was found to be an important source of water to Alkali Lake averaging 2.5 mm/day into the lake; however, negative net groundwater values were also observed indicating possible groundwater flow reversal. This speaks to a highly complex relationship between Alkali Lake and the groundwater system. These
conclusions largely agree with the literature in describing Alkali Lake as mainly a discharge lake (Befus et al., 2012; Ong, 2010; Zlotnik et al., 2010).

Finally, the link between water and energy balances through evaporation opens the door to broader questions about the Sandhills. This work described evaporation rates and the water balance related to only one of the many shallow saline lakes. Using these conclusions, the framework has been laid for upscaling from local to regional scales encompassing the unique collection of lakes and their role in the regional microclimate. For instance, how important is the role of sensible heat advection from the surrounding landscape in increasing evaporation? Are the Sandhills lakes an important source of water to the atmosphere? These questions begin to lead into the relationships between land and lake evaporation with rainfall at a larger scale. A global study of land-atmospheric interactions shows that a region encompassing the Sandhills may be particularly influential on local climate by supplying atmospheric moisture in a semiarid region (Koster et al., 2004; Sridhar and Wedin, 2009). However, the role of the Sandhills lakes in local climate is poorly understood. Similar questions were raised at an alpine lake in Canada, but also whether climate change would enhance the role of lake evaporation regionally (Bello and Smith, 1990). At Alkali Lake a warmer and drier climate would likely lead to more lake evaporation while reducing recharge, and combined these factors could cause some lakes or wetlands to dry up.
REFERENCES


Unland, H.E., Houser, P.R., Shuttleworth, W.J., Yang, Z., 1996. Surface flux measurement and modeling at a semi-arid Sonoran Desert site 1923.


Webster, W., Hoagland, K., Harvey, F., Arkebauer, T., 2008. Water balance analysis for a Nebraska Sand Hills wetland mitigation site.


Winter, T., 1986. Effect of ground-water recharge on configuration of the water table beneath sand dunes and on seepage in lakes in the sandhills of Nebraska, USA. J. Hydrol. 86, 221–237.


