Long-Term Changes of Lake Level and Water Budget in the Nam Co Lake Basin, Central Tibetan Plateau

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ABSTRACT

Long-term changes in the water budget of lakes in the Tibetan Plateau due to climate change are of great interest not only for the importance of water management, but also for the critical challenge due to the lack of observations. In this paper, the water budget of Nam Co Lake during 1980–2010 is simulated using a dynamical monthly water balance model. The simulated lake level is in good agreement with field investigations and the remotely sensed lake level. The long-term hydrological simulation shows that from 1980 to 2010, lake level rose from 4718.34 m to 4724.93 m, accompanied by an increase of lake water storage volume from 77.33 $\times 10^9$ m$^3$ to 83.66 $\times 10^9$ m$^3$. For the net lake level rise (5.93 m) during the period 1980–2010, the proportional contributions of rainfall–runoff, glacier melt, precipitation on the lake, lake percolation, and evaporation are 104.7%, 56.6%, 41.7%, 22.2%, and 80.9%, respectively. A positive but diminishing annual water surplus is found in Nam Co Lake, implying a continuous but slowing rise in lake level as a hydrological consequence of climate change.

1. Introduction

In most regions of the world, climate change is expected to significantly impact water resources. The form and magnitude of the impact, however, varies significantly from region to region (Solomon et al. 2007). It has been reported that water resources in Asia could be seriously affected by climate change, especially in the Tibetan Plateau (TP; Barnett et al. 2005; Immerzeel et al. 2010). The TP is the called “the Third Pole” of the earth and “the Water Tower of Asia” (Qiu 2008). The numerous inland lakes in the TP are important sources of water and indicators of regional climate change. In the literature, researchers have addressed the importance of water resource changes in the TP and the effects of climate change on rivers and lakes inside and nearby.
the TP (e.g., Zheng et al. 2009; Kang et al. 2010). Because of the harsh physical conditions (remoteness, high altitude, inclement weather, etc.) and lack of in situ observations, water levels and water storage, not to mention the water budget of lakes, in the TP are still poorly known. Some studies have focused on monitoring changes of lake surface area using remote sensing images (e.g., Ye et al. 2007; Chu et al. 2008; Yang et al. 2008; Liu et al. 2009), but only a few systematic water budget analyses were conducted for the lakes (Ding and Liu 1995; Zhang et al. 2003; Qi and Zheng 2006). To understand the water cycle of the region, most recently, more research efforts have been focused on investigating the bathymetry, surface area, water storage, and water budget of these lakes (Wang et al. 2009; Wu and Zhu 2008; B. Zhang et al. 2011; Zhu et al. 2010) and also on simulating the long-term water budget of the lakes (Krause et al. 2010).

For a lake basin with sparse or no gauge measurements, satellite remote sensing data are a valuable source of information for the investigation and simulation of the lake water budget. The Ice, Cloud, and Land Elevation Satellite/Geoscience Laser Altimeter System (ICESat/GLAS) level 2 altimetry product (GLA14) is one of the available datasets, which provides surface elevation of land with laser footprint geolocation and reflectance, as well as geodetic, instrument, and atmospheric corrections for range measurements (Zwally et al. 2003). ICESat elevation data over water surface in southern Egypt, the United States, East Africa, and central Asia have been examined by numerous studies and have shown accuracy of better than 10 cm (Urban et al. 2008; Swenson and Wahr 2009). It has also been confirmed as a reliable dataset for lake level interpretation in the TP (Phan et al. 2012; G. Q. Zhang et al. 2011). Based on the dataset, the satellite altimetry approach has been well developed for inland lake and river monitoring (Berry et al. 2005; Medina et al. 2008; Kropáček et al. 2012).

The purpose of this paper is to investigate the long-term changes of lake level and water budget in the Nam Co Lake, which is located in the central TP. A dynamic monthly water balance model is developed for this purpose and is calibrated using lake level data derived from the ICESat altimetry dataset. Based on the dynamic simulation, long-term changes of lake water budget are investigated with respect to climate changes in the TP.

2. Study area and data

a. Study area

Nam Co is a closed, semi-brackish lake (Wang et al. 2009) located in the central part of the TP (30°30′–30°55′N, 90°16′–91°03′E) with a catchment area of 10,610 km². It is the largest lake in the TP and one of the highest lakes in the world (Fig. 1). The elevation of the Nam Co Lake is 4718 m, with an area of about 1920 km² measured in 1979 (Guan et al. 1984) and a maximum depth over 90 m measured in 2005 (Wang et al. 2009). The glacier fraction in the Nam Co Lake basin is about 0.015. Significant fluctuation of the lake area in past decades has been detected by the interpretation of

![Fig. 1. Location of meteorological stations around Nam Co Lake.](image-url)
remote sensing images (Wu and Zhu 2008; Zhu et al. 2010; B. Zhang et al. 2011). The changes of the lake area could be a result of global warming. However, as a solitary cold region, the lack of observed evidence makes it difficult to understand the hydrological cycle of the region under climate change.

b. Data

No hydrometeorological observations are available in the basin before 2005. Though some investigations have been conducted to measure water depth, precipitation, and temperature of the basin recently (Wang et al. 2009), data are limited for long-term water balance research. In this case, in addition to a field investigation, we try to conduct our research based on the remotely sensed lake level data and long-term meteorological records from observations near the basin.

1) Bathymetric Survey of the Lake

The water storage capacity of a lake relies on its morphometry. To improve the estimation of water storage in the lake, in September 2005, September 2006, and August 2007 bathymetric surveys were conducted to digitize the underwater morphology of the lake. From the surveys, the available data points reached 305,721 with the accuracy of 0.01 m (Wang et al. 2009). The dataset is then used to determine the relationship between lake depth and its storage capacity. The approximate function of the relationship can be depicted as

\[ V = 0.5156H + 0.1574H^2 - 0.0006H^3 \]

for \( 0 < H < 100 \),

where \( H \) (m) is lake level and \( V \) is water storage (10⁹ m³). The expression is statistically significant at the level of 1%, which enables reliable conversion between water storage and lake level.

2) Remotely Sensed Lake Level

The ICESat launched in January 2003 was designed to determine geocentric elevations of the earth’s surface (Zwally et al. 2008). Fully calibrated ICESat data over various surfaces achieve an absolute accuracy of 2–7 cm and a precision of 2–3 cm (Urban et al. 2008). Though the primary task of the ICESat is to measure elevation changes of the polar ice sheets, the high-quality elevation measurements can be used to estimate lake level in the TP as well (Urban et al. 2008; G. Q. Zhang et al. 2011; Phan et al. 2012). To obtain lake levels of Nam Co, we use GLA14 (release 33) from the U.S. National Snow and Ice Data Center (NSIDC).

The GLAS instrument worked from February 2003 to October 2010. It measured the mean elevation of flat surfaces at the 70-m footprint spaced 172 m apart along the track. It collected elevations in 20 designated campaigns. Nam Co Lake was crossed by 43 ground tracks; each track of ICESat intersecting with Nam Co Lake was extracted for further data processing, which included procedures like conversion of data formats, conversion of spatial references (Bhang et al. 2007), and identification and rejection of outliers. Afterward, lake levels of 27 months during the period 2003–09 were obtained for subsequent calibration in hydrological simulation.

3) Meteorological Data

The routine observation records of 11 meteorological stations around Nam Co Lake basin from 1971 to 2010 were used in this research (Table 1; Fig. 1). The dataset included daily mean temperature, minimum temperature, maximum temperature, rainfall, sunshine duration, wind speed, and water vapor pressure. The kriging interpolation and the thin-plate spline (TPS) method (Hutchinson 1998) were adopted for spatial interpolation of temperature, precipitation, and the calculated potential evapotranspiration with the consideration of the terrain effect.

3. Hydrological simulation

a. Monthly water balance model

Zhang et al. (2008) proposed a monthly water balance (MWB) model based on the Budyko hypothesis (Budyko...
which assumes that actual evapotranspiration \(E_a\) is a function of aridity index \(\phi\) as

\[ E_a = P[F(\phi)], \] (2)

where \(P\) is precipitation and the aridity index is defined as the ratio of the potential evapotranspiration \(E_0\) over \(P\), that is, \(\phi = E_0/P\). The variable \(F(\phi)\) can be presented as

\[ F(x, \alpha) = 1 + x - [1 + x^{1/(1-\alpha)}]^{1-\alpha}, \] (3)

where \(x\) represents the ratio between water demand and water supply and \(\alpha\) is a parameter. The variable \(x\) is equivalent to \(\phi\) when \(E_0\) and \(P\) are assumed as water demand and water supply, respectively. The principle equations of the MWB model are listed in Table 2.

### Table 2. Expressions of the monthly water balance model. Note that \(\alpha_1, \alpha_2, S_{\text{max}}, d, g\) and \(d\) are parameters to be calibrated, \(g\) is the fraction of glacier, and \(t\) represents a time interval.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Expression</th>
</tr>
</thead>
<tbody>
<tr>
<td>(X): catchment</td>
<td>(X(t) = P(t) \left[ \frac{E_0(t)}{P(t)} \right] \alpha_1 )</td>
</tr>
<tr>
<td>(X_0): demand</td>
<td>(X_0(t) = E_0(t) + S_{\text{max}} - S(t-1) )</td>
</tr>
<tr>
<td>(R_d): direct runoff</td>
<td>(R_d(t) = P(t) - X(t) )</td>
</tr>
<tr>
<td>(W): water availability</td>
<td>(W(t) = X(t) + S(t-1) )</td>
</tr>
<tr>
<td>(Y): evapotranspiration</td>
<td>(Y(t) = W(t) \left[ \frac{E_0(t) + S_{\text{max}}}{W(t)} \right] \alpha_2 )</td>
</tr>
<tr>
<td>(R_c): recharge to groundwater storage</td>
<td>(R_c(t) = Y(t) - W(t) )</td>
</tr>
<tr>
<td>(E_a): actual evapotranspiration</td>
<td>(E_a(t) = W(t) \left[ \frac{E_0(t)}{W(t)} \alpha_2 \right] )</td>
</tr>
<tr>
<td>(S): soil water storage</td>
<td>(S(t) = Y(t) - E_a(t) + SM(t) )</td>
</tr>
<tr>
<td>(G): groundwater storage</td>
<td>(G(t) = (1-d)G(t-1) + R_b(t) )</td>
</tr>
<tr>
<td>(R_b): base flow</td>
<td>(R_b(t) = dG(t-1) )</td>
</tr>
<tr>
<td>(R_t): total runoff</td>
<td>(R_t(t) = gR_{\text{glacier}} + (1-g)(R_d + R_b + R_{\text{snow}}) )</td>
</tr>
</tbody>
</table>

The estimation of the portion \(f\) is essential for modeling the mass balance of monthly snow cover (Rohrer 1989) and for accurate runoff model performance (WMO 1986). The usual method for determining the division of precipitation is to set a threshold ambient temperature above which all precipitation is assumed to be rain and below which is snow. Herein, the simple linear relationship between average monthly air temperature \(T_{\text{air}}\) and the observed percentage of precipitation falling as snow proposed by Sevruk (1983) is as follows:

\[ f = \begin{cases} 0, & T_{\text{air}} > 12.22 \\ -4.5T_{\text{air}} + 55, & -10 \leq T_{\text{air}} \leq 12.22 \\ 100, & T_{\text{air}} < -10 \end{cases} \] (6)

The rate of snowmelt depends on the availability of energy to the snowpack and is usually dominated by net radiation (Marks and Dozier 1992). The snowmelt can be simulated by an energy balance model, which requires reliable measurements like air temperature, incoming solar radiation, vapor pressure, and wind speed and involves intensive calculation (Marshall and Oglesby 1994). It is recognized that energy balance snowmelt models are effective for short-term prediction for small catchments. In the absence of detailed observation of energy fluxes and snow characteristics, to estimate meltwater over larger spatial and temporal scales, simpler approaches such as temperature indices are better able to model the average conditions (Kuusisto 1984; Ferguson and Morris 1987; Vehviläinen 1992; Kuchment and Gelfan 1996). The snowmelt–runoff models that incorporate a degree-day or temperature index routine are commonly used in operational hydrology and have been successfully verified worldwide over a range of catchment sizes, physical characteristics, and climates (WMO 1986; Bergström 1992; Rango 1992). The basic form of the degree-day approach is
SM = C_s (T_{air} - T_s) \quad \text{and} \quad (7)

GM = C_g (T_{air} - T_g), \quad (8)

where C_s and C_g are melt rate factors of snow cover and glacier, respectively (mm °C\(^{-1}\) day\(^{-1}\)); T_{air} is the daily air temperature; and T_s is the threshold melt temperature. The critical melt temperature is often set to 0. The measured C_s averages in Finland are 2.4 and 3.5 mm °C\(^{-1}\) day\(^{-1}\) for forested and open areas, respectively (Kuusisto 1984). A typical range for old melting snow is 3.5–6 mm °C\(^{-1}\). Rain falling on snow is assumed to percolate through the snowpack to the soil surface.

c. Lake water balance

The original model proposed by Zhang et al. (2008) with the snow and glacier module can be used to model the water balance of the catchments and provide the inflow of the lake. To simulate the water storage or water balance of the catchments and provide the inflow of the lake, an additional water balance module of the lake is needed, which can be represented as

\[
\Delta H = \frac{A_c}{A_1} Q_{in} + P_t - E_t - Q_{out} = R_t / r_a + P_t - E_t - Perc,
\]

where \(H\) is water level of the lake (mm), \(P_t\) and \(E_t\) are precipitation (mm) and evaporation (mm) over the lake, and \(Q_{in}\) and \(Q_{out}\) are inflow (mm) and outflow (mm) of the lake. The \(Q_{in}\) is equal to runoff from the catchment \(R_t\) and consists of direct runoff \(R_{dir}\), base flow \(R_b\), snow melt flow \(R_{snow}\), and glacier melt flow \(R_{glacier}\). The variables \(A_c\) and \(A_1\) are area (km\(^2\)) of catchment and Nam Co Lake, respectively, while \(r_a\) represents the area percentage of the lake to the catchments and is set to 18.85% in this study according to the remote sensing images. Because Nam Co is a closed lake without direct runoff out of the lake, the only outflow loss is the percolation to the groundwater Perc, which is the net exchange between lake and groundwater. The percolation loss can be assumed to be a function of water level \(H\) (e.g., Kebede et al. 2006; Sene 1998). Herein, the relationship is represented by a linear model expressed as

\[
Perc = kH,
\]

where \(k\) is the percolation coefficient.

d. Evaporation

The potential evapotranspiration for rainfall–runoff simulation is estimated according to the Penman–Monteith method recommended by the Food and Agriculture Organization (FAO; Allen et al. 1998), which is presented as

\[
E_0 = \frac{0.408\Delta (R_n - G) + \gamma \frac{900}{T + 273} U_2 (e_s - e_d)}{\Delta + \gamma (1 + 0.34 U_2)}, \quad (11)
\]

where \(\Delta\) represents the curve slope of saturated water vapor pressure (kPa °C\(^{-1}\)) at temperature \(T\); \(R_n\) represents the solar net radiation at top layer (MJ m\(^{-2}\) day\(^{-1}\)); \(G\) represents soil-pass heat (MJ m\(^{-2}\) day\(^{-1}\)); \(\gamma\) is the dry–wet constant (kPa °C\(^{-1}\)); \(T\) is the daily mean temperature (°C); \(U_2\) is the wind speed at the height of 2 m (m s\(^{-1}\)); and \(e_s\) and \(e_d\) are the saturated and actual water vapor pressure (kPa), respectively, at temperature \(T\).

Equation (11) can be used to estimate potential evapotranspiration in the land surface of the catchments. When it is applied to calculate potential evaporation of the lake, however, change of heat storage in the lake has to be considered. Lake surface temperature is required to estimate the heat change of a shallow lake (e.g., Keijman 1974). For deep lakes, it is necessary to conduct thermal surveys consisting of temperature profiles with depth, measured ideally at a sufficient number of stations to produce a good average (e.g., Anderson 1954; Sturrock et al. 1992). Because the lake surface temperature or the temperature profiles of the Nam Co Lake are not available, \(E_0\) is used to represent evaporation of the lake in the nonfrozen months. In the frozen months, the lake evaporation equals 0. The calculated \(E_0\) is found to be in good agreement with evaporation observed in 2008 using a pan with a diameter of 20 cm \((R^2 = 0.84)\) and comparable with that observed via pan E601-B (Ren et al. 2002).

e. Model inputs and calibration

The inputs of the monthly water balance described above are monthly mean temperature \((T_{air})\), monthly totals of precipitation, and potential evapotranspiration \((E_0)\). The parameters of the model are \(\alpha_1, \alpha_2, S_{max}\), and \(d\) in the rainfall–runoff module; \(C_s\) and \(C_g\) in the snow and glacier module; and \(k\) in the lake water balance module. The seven parameters are calibrated using the global optimization method particle swarm optimization (PSO) to maximize the Nash–Sutcliffe efficiency (NSE) coefficient (Nash and Sutcliffe 1970):

\[
\text{NSE} = 1 - \frac{\sum_{i=1}^{n} (H_{obs,i} - H_{act,i})^2}{\sum_{i=1}^{n} (H_{obs,i} - \bar{H}_i)^2}, \quad (12)
\]

where \(H_{obs,i}\) and \(H_{act,i}\) are the remotely sensed and simulated water level, respectively; \(\bar{H}_i\) is the arithmetic mean of the
remotely sensed water level; and \( n \) is the sample number, that is, the total number of months from 2003 to 2009.

4. Results

a. Model performance

As shown in Fig. 2, the simulated lake level matches well with the remotely sensed one with NSE = 0.889 and \( R^2 = 0.894 \). The absolute difference between the mean simulated lake level and remotely sensed one is 1 mm. The largest difference between the simulated and remotely sensed lake level is on March 2006, when the simulated lake level is 0.011% overestimated. In November 2004, however, the simulated lake level is 0.006% underestimated. The simulation result also shows high consistency with field investigations conducted by the Chinese Academy of Sciences (CAS) in July 1979, which concluded that the elevation of Nam Co is 4718 m (Guan et al. 1984). Compared with the simulated lake level in January 1980 (4718.512 m), the relative difference between the investigated lake level and the simulated one is 0.01%.

b. Intra-annual variation of lake level and water budget

As shown in Fig. 3, which displays the intra-annual variations of lake level, the highest lake level of Nam Co is 4722.97 m in September, while the lowest one is 4722.66 m in May, which suggests that the lake level rises 0.31 m from May to September. The result is very similar to the study of Zhou et al. (2006), where it is reported that lake level rose 0.36 m from May to mid-September, according to their observations during ice-free periods (May–October) in 2005. However, one should notice that the variations of both lake level and water storage are not concurrent with precipitation or inflow (Fig. 3). The maximum monthly precipitation and inflow are found to be in August. The lag of lake level rise could be because of the catchment storage capacity, which depends on water storage in the catchment and gradual release to the river and lake later in the dry months. It is also evident that the lake level does not depend only on the water gain, but it also relies on water surplus (gain minus loss). For a negative water surplus, water level, as well as water storage, would decrease, but positive water surplus could result in the increase of water level and water storage even in the month with rather small inflow.

Figure 3 shows that the water gain of Nam Co Lake in January is only 2.11 mm, which is mainly from the precipitation on the lake, but reaches to 253.96 mm in August. The lower water loss of the lake from winter to early spring (January–March) mainly consists of percolation with little evaporation loss. Owing to lake evaporation loss, water loss from late spring to autumn...
(May–October) is much higher. The evaporation loss accounts for 86.21% of the total water loss of the lake. The lake water budget shows a deficit in October–May, but surplus in June–September, with the highest net water gain of 131.2 mm in August. Subsequently, water storage increases from April to September then decreases until next March, leading to the observation that the maximum and minimum lake water storages appear in September (around 84.56 ± 10^9 m³) and April (around 84.06 ± 10^9 m³), respectively.

c. Interannual changes of lake level and water budget

As shown in Table 3, the long-term mean lake level is 4722.79 m for the period of 1980–2010, while mean water storage of the lake is around 84.27 × 10^9 m³, which is very close to 84.24 × 10^9 m³ estimated by B. Zhang et al. (2011). A significant increasing trend of lake level is detected by the nonparametric Mann–Kendall test (Mann 1945; Kendall 1975), showing that the rising rates of mean, minimum, and maximum annual lake levels are all around about 0.180 meters per annum (m a⁻¹). In different decades, however, the rising rate of lake level and water storage are not the same. For the three periods concerned, that is, 1980–90, 1991–2000, and 2001–10, the mean lake levels are 4720.94, 4723.02, and 4724.59 m, and the rising rates are 0.024, 0.009, and 0.014 m a⁻¹, respectively (Fig. 4). Meanwhile, as shown in Table 4, the mean water storage values are 81.35 ± 10^9, 84.63 ± 10^9, and 87.12 ± 10^9 m³, with increasing rates of 0.45 ± 10^9, 0.17 ± 10^9, and 0.26 ± 10^9 m³ a⁻¹, respectively.

For the annual water budget of the lake, according to the simulation results, the annual-mean water gain,
water loss, and water surplus are \(2.03 \times 10^9\), \(1.66 \times 10^9\), and \(0.37 \times 10^9\) m\(^3\) a\(^{-1}\), respectively (Table 4). The inflow of the lake (i.e., the runoff from the basin to the lake) accounts for 67.0% of the total water gain, outweighing the amount of precipitation on the lake. The proportions of surface runoff, base flow, and glacier melt flow to total inflow are 97.18%, 2.82%, and 0.89%, respectively, indicating that the water gain of the lake depends greatly on rainfall–runoff. As shown in Table 4, evaporation is the dominant factor in water loss, accounting for 78.46% of the total amount. Based on the budget during 1980–2010, it is estimated that the total water gain could result in a lake level increase of 12.04 m, among which rainfall–runoff, glacier melt, and precipitation on the lake account for 51.58%, 27.89%, and 20.53%, respectively. However, the water loss has led to a 6.11-m decrease of lake level, where lake evaporation and percolation account for 78.46% and 21.54%, respectively. For the net lake level rise (5.93 m), the proportional contributions of runoff, glacier melt, \(P_i\), Perc, and \(E_l\) are 104.7%, 56.6%, 41.7%, 22.2%, and 80.9%, respectively.

Figure 4 shows the long-term variation of water budget in the past decades. It is estimated that annual inflow of 1991–2000 and 2001–10 is 13.46 and 1.23 mm smaller

<table>
<thead>
<tr>
<th>Period</th>
<th>Max</th>
<th>Min</th>
<th>Mean</th>
<th>(\Delta L) (m)</th>
<th>Max</th>
<th>Min</th>
<th>Mean</th>
<th>Change rate (10^9 m^3 a^{-1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>1980–90</td>
<td>4722.64</td>
<td>4718.51</td>
<td>4720.94</td>
<td>4.13</td>
<td>84.02</td>
<td>77.59</td>
<td>81.35</td>
<td>0.45</td>
</tr>
<tr>
<td>1991–2000</td>
<td>4723.89</td>
<td>4722.41</td>
<td>4723.02</td>
<td>1.48</td>
<td>86.02</td>
<td>83.66</td>
<td>84.63</td>
<td>0.17</td>
</tr>
<tr>
<td>2001–10</td>
<td>4725.47</td>
<td>4724.59</td>
<td>4724.59</td>
<td>1.87</td>
<td>88.54</td>
<td>85.55</td>
<td>87.12</td>
<td>0.26</td>
</tr>
<tr>
<td>1980–2010</td>
<td>4725.28</td>
<td>4719.00</td>
<td>4722.79</td>
<td>6.28</td>
<td>88.23</td>
<td>78.33</td>
<td>84.27</td>
<td>0.29</td>
</tr>
</tbody>
</table>

**Fig. 4.** Interannual variations of lake level \((H)\), lake water storage \((V)\), precipitation \((P)\), potential evaporation \((E_0)\), catchment actual evapotranspiration \((E_a)\), total runoff \((R_t)\), glacier melt runoff \((R_{\text{glacier}})\), lake evaporation \((E_l)\), and air temperature \((T)\).
than the reference period (1980–90), respectively. Total water gain (annual inflow plus annual precipitation on the lake) in 1991–2000 is 100.53 mm smaller than the reference period, but it is 0.86 mm higher in 2001–10. Concurrently, the total water loss (annual percolation plus annual evaporation on the lake) of 1991–2000 and 2001–10 is 179.10 and 295.50 mm higher than the reference period (1980–90), respectively. The smaller water gain in addition to the higher water loss of the lake could result in the decrease of water surplus. Estimated water surpluses of the periods 1980–90, 1991–2000, and 2001–10 are 0.74 × 10^8, 0.18 × 10^9, and 0.15 × 10^9 m³ a⁻¹, respectively, decreasing distinctly at a rate of 15.35 mm a⁻¹ (p < 0.0001) during 1980–2010. The positive but smaller surpluses indicate that water storage as well as lake level should have been increasing, but slowing down.

5. Discussion

a. Impacts of climate change

The long-term changes and variations of lake level and water storage could be the consequence of climate change in the TP. According to long-term meteorological records (1980–2010) of the stations around the Nam Co Lake, the annual-mean temperature (T) showed significant increasing trend at the rate of 0.58°C decade⁻¹ and higher rate in winter season (November–January). Compared with the period 1980–90, the annual-mean temperatures of 1991–2000 and 2001–10 are 0.29° and 1.09°C higher, respectively. Owing to the increase of temperature, it can be seen from the simulation that the evaporation of the lake (Eₜ) during 1980–2010 increased at a rate of 2.12 mm a⁻¹. For 1991–2000 and 2001–10, the annual-mean Eₜ is 27.18 and 31.01 mm more than that in 1980–90 (Table 4). The greater evaporation means larger water loss and smaller water surplus. Significant correlation (R² = 0.58) between temperature and water loss is found during 1980–2010. However, it should be noted that relations between global warming and lake level are far more complicated. Global warming could increase not only water loss (e.g., higher evaporation) but also water gain (e.g., higher glacier melting rate). Moreover, the influences of global warming can be amplified or attenuated by other environmental factors such as precipitation, wind, etc. On the one hand, precipitation can be the direct water source of the lake (Pᵢ), but on the other hand, precipitation on the land in turn becomes runoff to the lake as part of its water gain. During 1980–2010, the fluctuations in both total inflow and water gain of the lake are quite similar to that in annual precipitation, showing significant correlation between the Pᵢ and water gain (R² = 0.98) and between P and inflow (R² = 0.93) as well. In general, it can be concluded that increasing temperature could result in more water loss, while changes of water gain are mainly due to precipitation change. To figure out the potential response of lake level to climate change, however, sensitivity analysis based on the model developed would be helpful, which will be investigated in our future research.

b. Uncertainties

The results shown above have been verified by some field investigations in the literature; however, uncertainty exists because of the limited hydrological observation in the Nam Co Lake and intrinsic uncertainties of the hydrological model. For instance, the freezing and thaw process of permafrost could have an important role in the hydrological cycle (Zhang et al. 2003; Liu et al. 2009; Zheng et al. 2009), but it is not represented in the current version of the hydrological model. The model can be further improved if the relations among water level, water storage, and percolation are represented by a more complicated model when more observations are available for parameter calibration. With the limitations of observation, however, a more complicated model may not help to reduce the uncertainty. As mentioned in the calculation of lake evaporation using the Penman–Monteith method, the lake surface temperature or temperature profile is required to estimate heat storage change in the lake. Though the calculated evaporation has been adjusted according to pan observation, the neglect of the heat storage change could result at bias of the lake evaporation. Uncertainty may also come from the spatial interpolation of meteorological observations.

### Table 4. Decadal variation of lake water budget in response to climate change.

<table>
<thead>
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<th>Period</th>
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<th></th>
<th>Water loss</th>
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<th>Water surplus</th>
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<td>Rᵢ</td>
<td>Total</td>
<td>Eₜ</td>
<td>Perc</td>
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<td>2001–10</td>
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<td>1980–2010</td>
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<td>2.03</td>
<td>1.30</td>
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and the remotely sensed lake level. The meteorological inputs such as precipitation and temperature used in this study are all obtained via spatial interpolation, which inherits uncertainties in the original observation records and interpolation approaches.

6. Conclusions

The long-term change of lake water budget in the TP responding to climate change is of great research interest. However, the harsh physical conditions as well as the limitation of in situ observations has made it a challenge for researchers. In this paper, in addition to field investigation, lake level derived from the ICESat/GLAS dataset has been used to calibrate a monthly water balance model to simulate the long-term water budget of Nam Co Lake. The results show that the model performs well if evaluated according to the remotely sensed lake level.

According to the long-term water budget simulation using the monthly model, the annual-mean level and water storage of Nam Co Lake are 4722.79 m and 84.27 $\times 10^9$ m$^3$, respectively. During 1980–2010, the lake level rose from 4718.34 to 4724.93 m, accompanied by an increase in water storage from 77.33 $\times 10^9$ to 83.66 $\times 10^9$ m$^3$. It is found that annual-mean water gain and loss of the lake are 2.03 $\times 10^9$ and 1.66 $\times 10^9$ m$^3$ a$^{-1}$, respectively. The positive annual water surplus is around 0.37 $\times 10^9$ m$^3$ a$^{-1}$, indicating a possible increase in both lake level and water storage. The annual water surplus remains positive but decreases in the period 1980–2010, implying that the lake level of the Nam Co Lake continues to rise, but at a slower rate. As a consequence of climate change and variation, the surplus of water budget fluctuates with the variation of precipitation and tends to decrease with the increasing temperature.

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