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Differences in the water-balance components of four lakes in the southern-central Tibetan Plateau

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Abstract. The contrasting patterns of lake-level fluctuations across the Tibetan Plateau (TP) are indicators of differences in the water balance over the TP. However, little is known about the key hydrological factors controlling this variability. The purpose of this study is to contribute to a more quantitative understanding of these factors for four selected lakes in the southern-central part of the TP: Nam Co and Tangra Yumco (increasing water levels), and Mapam Yumco and Paiku Co (stable or slightly decreasing water levels). We present the results of an integrated approach combining hydrological modeling, atmospheric-model output and remote-sensing data. The J2000g hydrological model was adapted and extended according to the specific characteristics of closed-lake basins on the TP and driven with High Asia Refined analysis (HAR) data at 10 km resolution for the period 2001–2010. Differences in the mean annual water balances among the four basins are primarily related to higher precipitation totals and attributed runoff generation in the Nam Co and Tangra Yumco basins. Precipitation and associated runoff are the main driving forces for inter-annual lake variations. The glacier-meltwater contribution to the total basin runoff volume (between 14 and 30 % averaged over the 10-year period) plays a less important role compared to runoff generation from rainfall and snowmelt in non-glacierized land areas. Nevertheless, using a hypothetical ice-free scenario in the hydrological model, we indicate that ice-melt water constitutes an important water-supply component for Mapam Yumco and Paiku Co, in order to maintain a state close to equilibrium, whereas the water balance in the Nam Co and Tangra Yumco basins remains positive under ice-free conditions. These results highlight the benefits of linking hydrological modeling with atmospheric-model output and satellite-derived data, and the presented approach can be readily transferred to other data-scarce closed lake basins, opening new directions of research. Future work should go towards a better assessment of the model-chain uncertainties, especially in this region where observation data are scarce.

1 Introduction

The drainage system of the interior Tibetan Plateau (TP) is characterized by numerous closed-lake (endorheic) basins. Because an endorheic lake basin integrates all hydrological processes in a catchment, lake-level or volume changes provide a cumulative indicator of the basin-scale water balance. While most of the lakes located in the central part of the TP are characterized by a water-level increase over recent decades (e.g., Zhang, et al., 2011; Phan et al., 2012), there are also several lakes with nearly stable or slightly decreasing water levels in the southern part of the TP. These high-elevation lakes are therefore considered to be one of the most sensitive indicators for regional differences in the water balance in the TP region (e.g., B. Zhang et al., 2013; G. Zhang et al., 2013; Song et al., 2014).

Neglecting the influence of long-term storage changes such as deep groundwater and lake–groundwater exchange, the net water balance of an endorheic lake basin with water supply from glaciers can be expressed as $\Delta V_{\mathrm{lake}} = P_{\mathrm{lake}} - E_{\mathrm{lake}} + R_{\mathrm{land}} + R_{\mathrm{glacier}}$, where ΔV_{lake} is the lake-volume change (net annual lake-water storage), P_{lake} the on-lake precipitation, E_{lake} the evaporation rate from the lake, and R_{land}

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and $R_{glacier}$ are the runoff from non-glacierized land surfaces and from glaciers (in units of volume per unit time). Under constant climatic conditions, endorheic lakes will eventually tend towards a stable equilibrium ($\Delta V_{lake} = 0$), where the several water-balance terms are balanced (Mason, 1994). Lake-level changes thus result from a shift in the water input or output.

Due to the accelerated glacier mass loss, it has been hypothesized that lake-level increases are primarily due to an increased inflow of glacier meltwater (e.g., Yao et al., 2007; Zhu et al., 2010; Meng et al., 2012). Nevertheless, glacier runoff into lakes itself should not increase the overall water-volume mass on the TP, as indicated by GRACE satellite gravimetry data (G. Zhang et al., 2013). Furthermore, numerous lakes of the TP are not linked to glaciers (Phan et al., 2013), and the water-level changes of lakes without glacier meltwater supply in the 2000s were as high as those of glacier-fed lakes (Song et al., 2014). In other studies, increased precipitation and decreased evaporation were generally considered to be the principal factors causing the rapid lake-level increases (e.g., Morrill, 2004; Lei et al., 2013, 2014). Y. Li et al. (2014) argued for the importance of permafrost degradation on recent lake-level changes. Thus, recent studies addressing the controlling mechanism of lake-level fluctuations remain controversial.

In order to explore differences in the water balance of endorheic lake basins in the TP region, recent studies emphasize the urgency of the quantification of water-balance components by using hydrological models (e.g., Cuo et al., 2014; Lei et al., 2014; Song et al., 2014). Hydrological modeling studies of endorheic lake basins in the TP region are rare (e.g., Krause et al., 2010), principally due to a lack of hydro-climatological observations and limitations in spatial and temporal coverage of available gridded climate data (Biskop et al., 2012). The paucity of spatial information of climatological variables was addressed by Maussion et al. (2014), who developed a high-resolution (up to 10 km × 10 km) atmospheric data set for the 2001–2011 period, the High Asia Refined analysis (HAR). The HAR10 data set was successfully applied in surface energy balance/mass balance (SEB/MB) modeling studies (Mölg et al., 2014; Huintjes et al., 2015) and in a hydrological modeling study in the Pamir Mountains (Pohl et al., 2015), but has not yet been used as input for catchment-scale hydrological modeling studies in the central TP. The objective of this study is the hydrological modeling of endorheic lake basins across the southern-central part of the TP in order to

– analyze spatiotemporal patterns of water-balance components and to contribute to a better understanding of their controlling factors, and

– quantify single water-balance components and their contribution to the water balance, and obtain a quantitative knowledge of the key factors governing the water balance and lake-level variability during the 2001–2010 period.

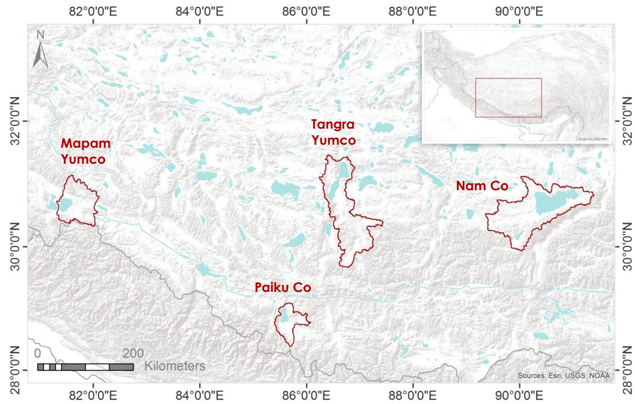

Figure 1. Location of the study region comprising four selected endorheic lake basins.

Lakes Nam Co and Tangra Yumco with increasing water levels (i.e., positive water balance) and lakes Mapam Yumco and Paiku Co with stable or slightly decreasing water levels (i.e., stable or slightly negative water balance, respectively) were selected to investigate differences in the water-balance components. The paper is organized as follows. In Sect. 2, we describe the study area and the data used. Section 3 gives details of the hydrological modeling approach and, in Sect. 4, we present the modeling results and assess similarities and differences among the basins; in Sect. 5, the results, limitations and uncertainties of this study are discussed with respect to findings from other studies. Finally, Sect. 6 highlights the principal results and concludes with remarks on future research needs and potential future model applications.

2 Study area and data

2.1 Description of the study area

The study region comprises four endorheic lake basins along a west–east (W–E) lake transect in the southern-central part of the TP between 28–32° N and 81–92° E (Fig. 1). Basic characteristics of the selected lake basins are summarized in Table 1. Climatologically, the study region encompasses a semi-arid zone and is characterized by two distinct seasons: a temperate-wet summer season dominated by the Indian Monsoon and a cold-dry winter season determined by the westerlies. The mean annual air temperature (MAAT) lies between 0 and −3 °C and the mean annual precipitation ranges between 150 and 500 mm, with 60–80 % of this total occurring between June and September (Leber et al., 1995). The study region features a climate gradient, with increasingly cooler and drier conditions in a westward direction.

Due to the semi-arid and cold climate conditions as well as the complex topography, soils in the study area in general are

Table 1. Basic information of selected basins in the study region. Data sources are described in Sect. 2.2.

Lake name	Elev.	Lake center		Basin area	Lake area	Land cover (%)				
	(m a.s.l.)	Lat	Long	(km^2)	(km^2)	Lake	Glacier	Grassland	Wetland	Barren land
Nam Co	4725	30°42	90°33	10 760	1950	18	2	39	8	33
Tangra Yumco	4540	31°00	86°34	9010	830	9	0.96	31	0.04	59
Paiku Co	4585	28°55	85°35	2380	270	10	6.5	43	0.5	40
Mapam Yumco	4580	30°42	81°28	4440	420	10	1.5	64	2.5	22

poorly developed and vegetation throughout the study area is generally sparse. The growing period lasts approximately 5 months, from late April or early May to late September or mid-October (B. Zhang et al., 2013). The highest mountain regions are covered by glaciers and permanent snow. Among all basins, the Paiku Co catchment exhibits the largest glacier coverage (6.5 % of the basin area). The areas covered by glaciers in the Nam Co, Tangra Yumco and Mapam Yumco basins accounts for 2, 1 and 1.5 % of catchment area. The lake area in the several basins corresponds to 18 % (Nam Co), 11 % (Mapam Yumco), 9.5 % (Paiku Co) and 9 % (Tangra Yumco). Based on GLAS/ICESat data, the lake levels for Nam Co and Tangra Yumco rose by approximately 0.25 m yr^{-1} between 2003 and 2009, whereas the lake levels for Paiku Co and Mapam Yumco slightly decreased by around -0.05 m yr^{-1} (Zhang et al., 2011; Phan et al, 2012).

2.2 Data used

Because of limited availability of climatological data in the TP region, we used a new atmospheric data set for the TP, the HAR (Maussion et al., 2014) as input for the hydrological model. The HAR data sets were generated by dynamical downscaling of global-analysis data (final analysis data from the Global Forecasting System; data set ds083.2), using the Weather Research and Forecasting (WRF) model (Skamarock and Klemp, 2008). A detailed description of this procedure is given in Maussion et al. (2014). HAR products are freely available (http://www.klima.tu-berlin.de/HAR) in different spatial (30 km × 30 km and 10 km × 10 km) and temporal (hourly, daily, monthly and yearly) resolutions. In this study, we used the daily HAR10 data. In the WRF model version 3.3.1, which was used for the generation of the HAR10 data, the lake-surface temperature is initialized by averaging the surrounding land-surface temperatures. By analyzing the influence of the assimilation of satellite-derived lake-surface temperatures, Maussion (2014) found that the standard method of WRF leads to a much cooler lake than observed, which in turn has a strong influence on local climate. Therefore, the HAR10 data points over water surfaces were not included for hydrological modeling purposes.

The HAR10 precipitation output was compared to rain-gauge data and to Tropical Rainfall Measuring Mission (TRMM) satellite precipitation estimates by Maussion et al. (2014). They concluded that HAR10 accuracy in comparison to rain gauges was slightly less than TRMM; however, orographic precipitation patterns and snowfall were more realistically simulated by the WRF model. HAR10 temperatures in the summer months are closer to ground observations than in winter (Maussion, 2014). Despite the winter cold bias, the overall seasonality is well reproduced (Maussion, 2014). The cold bias effect on the accuracy of the hydrologic-modeling results is assumed to be low, because hydrological processes governing lake-level changes are more critical during the other three seasons of the year.

Lake-surface water temperature (LSWT) estimates from the ARC-Lake v2.0 data products (MacCallum and Merchant, 2012) served as additional input for the hydrological modeling in the Nam Co and Tangra Yumco basins. ARC-Lake v2.0 data products contain daytime and nighttime LSWT observations from the series of (advanced) along-track scanning radiometers for the period 1991–2011. Daytime and nighttime MODIS land-surface temperature (LST) 8-day data at 1 km spatial resolution (MOD11A2) were averaged after a plausibility check to obtain mean daily LSWT time series for Paiku Co and Mapam Yumco, where no ARC-Lake v2.0 data were available.

Shuttle Radar Topography Mission (SRTM) 90 m digital elevation model (DEM) data (Farr et al., 2007) were retrieved from the Consortium for Spatial Information (CGAIR-CSI) Geoportal (http://srtm.csi.cgiar.org). We used the SRTM Version 4 data for derivations of catchment-related information such as catchment boundary, river network, flow accumulation and flow direction, as well as terrain attributes (slope and aspect).

For the Nam Co and Tangra Yumco basins, land-cover classifications were generated using Landsat TM/ETM+ satellite imagery. The land-cover classifications consist of five classes used for this analysis: water, wetland, grassland, barren land and glacier. For the Paiku Co and Mapam Yumco basins, land-cover information could be obtained from the Himalaya Regional Land Cover database (http://www.glcn.org/databases/hima_landcover_en.jsp). The Himalaya land-cover map was produced as part of the Global Land Cover Network – Regional Harmonization Program, an initiative to compile land-cover information for the Hindu Kush–Karakorum–Himalaya mountain range using a combination of visual and automatic interpretation of recent Landsat ETM+ data. The land-cover classes were reclassified according to the five classes mentioned above. Classes with similar characteristics (e.g., vegetation type, degree of vegetation cover) were consolidated into a single class.

Lake-level observations from 2006 to 2010 for Nam Co were provided by the Institute of Tibetan Plateau Research (ITP), Chinese Academy of Sciences (CAS), and used for model validation. However, lake-level values during the freezing (wintertime) periods are missing, because the lake-level gauge was destroyed by lake ice, and therefore rendered inoperable each winter. Thus, data are only available for the ice-free period (May/June–November/December). Unfortunately, the lake-level observation data contain an unknown shift between the consecutive years.

Due to the absence of continuous lake-level measurements, we obtained satellite-based lake-level and water-volume data for the four studied basins from the HydroWeb database (http://www.legos.obs-mip.fr/en/soa/hydrologie/hydroweb/) provided by LEGOS/OHS (Laboratoire d'Etudes en Geodesie et Oceanographie Spatiales (LEGOS) from the Oceanographie, et Hydrologie Spatiales, OHS) (Crétaux et al., 2011). LEGOS lake-level and water-volume data for the lakes included in this study were available for different time spans (see Table 2). The start and end date of each time series were taken from the same season (as far as available) in order to make lake levels or volumes comparable. Water-volume data calculated through a combination of satellite images (e.g., MODIS, Landsat) and various altimetric height level data (e.g., Topex/Poseidon, Jason-1) (Crétaux et al., 2011) were used for model calibration (see Sect. 3.3). The mean annual lake-level changes derived from LEGOS data for Nam Co, Tangra Yumco, Paiku Co and Mapam Yumco (0.25, 0.26, -0.07, $-0.05\,\mathrm{m\,yr^{-1}}$) are close to the change rates estimated by Zhang et al. (2011) (0.22, 0.25, -0.04, $-0.02\,\mathrm{m\,yr^{-1}}$) and Phan et al. (2012) (0.23, 0.29, -0.12, $-0.04\,\mathrm{m\,yr^{-1}}$) using GLAS/ICESat data (2003–2009) (Table 4, lower part).

MODIS snow-cover 8-day data of Terra (MOD10A2) and Aqua (MYD10A2) satellites at a spatial resolution of 500 m served for validation of the snow modeling. As proposed in the literature (e.g., Parajka and Blöschl, 2008; Gao et al., 2010; Zhang et al., 2012), we combined Terra and Aqua data on a pixel basis to reduce cloud-contaminated pixels. The cloud pixels in the Terra images were replaced by the corresponding Aqua pixel. For the period of time before the Aqua satellite was launched (May 2002), this combination procedure was not possible, and we used the original MODIS/Terra snow-cover data. After the combination procedure the cloud-cover percentage was on average less than 1–2 % for all basins.

3 Methods

3.1 Hydrological model concept and implementation

The challenge for hydrological modelers is to balance the wish to adequately represent complex processes with the need to simplify models for regions with limited data availability (Wagener and Kollat, 2007). Therefore, we selected a semi-distributed conceptual model structure, primarily following the J2000g model (Krause and Hanisch, 2009). The J2000g model is a simplified version of the fully distributed J2000 model (Krause, 2002). The main differences with J2000 are that complex process descriptions (e.g., soil-water dynamics) are simplified, leading to a reduced number of land-surface and calibration parameters in the J2000g model, and lateral flow processes between spatial model units and streamflow routing are not accounted for by the J2000g model. The J2000g was successfully applied for hydrological predictions in data-scarce basins (e.g., Deus et al., 2013; Knoche et al., 2014; Rödiger et al., 2014; Pohl et al., 2015), including a previous modeling study in the Nam Co basin (Krause et al., 2010).

The conceptual model presented here was realized within the Jena Adaptable Modelling System (JAMS) framework (http://jams.uni-jena.de/). An overview of JAMS, especially the JAMS software architecture and common structure of JAMS models, is given in Kralisch and Fischer (2012). Primarily, JAMS was developed as a JAVA-based framework for the implementation of model components of the J2000 model. During recent years, a solid library of single easily manageable components has been developed by implementing a wide range of existent hydrological-process concepts as encapsulated process modules and developing new model modules as needed. Due to the modular structure, the J2000g model could be easily adapted and extended according to the specific characteristics of endorheic lake basins in the TP region.

Meteorological data requirements for this study were daily times series of precipitation, minimum, maximum and average air temperature, solar radiation, wind speed, relative humidity and cloud fraction obtained from daily HAR10 data. Daily LSWT data served as additional input for the calculation of the long-wave radiation term over the lake surface. Process simulations were grouped into the following categories: (i) lake, (ii) land (non-glacierized) and (iii) glacier. A schematic illustration of the model structure and a detailed description of the model components are given in the Supplement.

In brief, we used the regionalization procedure implemented in J2000g for the interpolation of the HAR10 raster points (centroid of the raster cell) to each hydrological response unit (HRU). This combines inverse distance weighting (IDW) with an optional elevation correction. Net radiation was calculated following the Food and Agriculture Organization of the United Nations (FAO) proposed use of the

Table 2. Lake-level and water-volume changes derived from LEGOS data for the four studied lakes.

Lake name	Start date	Start volume (km^3)	Start level (m)	End date	End volume (km^3)	End level (m)	Δ Lake volume ($km^3\,yr^{-1}$)	Δ Lake level ($m\,yr^{-1}$)
Nam Co	27 Sep 2001	1.3	4722.683	1 Oct 2010	5.3	4724.697	0.44	0.22
Tangra Yumco	7 Oct 2001	0	4533.997	25 Oct 2009	1.7	4535.987	0.21	0.25
Paiku Co	2 Jun 2004	0	4578.067	4 Mar 2008	−0.08	4577.768	−0.02	−0.07
Mapam Yumco	30 Oct 2003	0.02	4585.551	21 Nov 2009	−0.1	4585.231	−0.01	−0.05

Penman–Monteith model (Allen et al., 1998). We adapted the long-wave radiation part of the FAO56 calculation to the special high-altitude conditions on the TP, according to the recommendations of Yin et al. (2008), and implemented the commonly used approach for calculating net long-wave radiation over water surfaces (e.g., Jensen, 2010).

Potential evapotranspiration (PET) from land and snow surfaces (sublimation) is calculated based on Penman–Monteith (Allen et al., 1998). For the estimation of open-water evaporation rates from large lakes, we modified the Penman equation through the addition of an empirical estimation of the lake heat storage (Jensen et al., 2005). As suggested by Valiantzas (2006), we used the reduced wind function proposed by Linacre (1993) for the estimation of evaporation from large open-water body surfaces.

The simple degree-day snow modeling approach of the standard J2000g model version was replaced by the J2000 snow module that combines empirical or conceptual approaches with more physically based routines. This module takes into account the phases of snow accumulation and the compaction of the snow pack caused by snowmelt or rain on the snow pack. For a detailed description, see Nepal et al. (2014). The glacier module calculates ice melt according to an extended temperature-index approach (Hock, 1999).

Soil-water budget and runoff processes are simulated using a simple water storage approach (Krause and Hanisch, 2009). The storage capacity is defined from the field capacity of the specific soil type within the respective modeling unit. Actual evapotranspiration (AET) is calculated depending on the saturation of the soil-water storage, PET and a calibration parameter. The J2000g model generates runoff only when the soil-water storage is saturated. The partition into surface runoff and percolation depends on the slope and the maximum percolation rate of the respective modeling unit that can be adapted by a calibration parameter. The percolation component is transferred to the groundwater storage component. The groundwater module calculates base flow using a linear outflow routine and a recession parameter (Krause and Hanisch, 2009).

The lake module calculates the net evaporation (lake evaporation minus precipitation over the lake's surface area). The lake-water storage change is the sum of (i) direct runoff and base flow from each modeling unit of the non-glacierized areas and (ii) glacier runoff (snowmelt and ice melt, and rainfall over glaciers) from each glacier HRU minus lake net evaporation. For simplicity, the terms land runoff, glacier runoff and net evaporation are used to refer to several water-balance components. Because the J2000g model does not account for water routing and thus time delay of the discharge, the model is not fully suited for providing continuous and precise estimates of lake-water storage changes.

3.2 Delineation of spatial model entities

In order to provide spatially distributed information on landscape characteristics for the hydrological modeling, we applied the hydrological response unit (HRU) approach (Flügel, 1995). Using ArcGIS software, HRUs with similar hydrological behavior were delineated by overlaying topographic-related and land-cover information. Soil and hydro-geology information were not included in the overlay analysis, due to a lack of detailed data. The distribution concept applied represents the landscape heterogeneity with a higher spatial resolution in the complex high mountain areas (a large number of small polygons) than in the relatively flat terrains at the lower elevations (smaller number of large polygons). The total number of HRUs varies between 1928 (Paiku Co) and 8058 (Nam Co).

3.3 Model-parameter estimation and model evaluation

The J2000g model requires the definition of spatially distributed land-surface parameters describing the heterogenic land surface and the estimation of calibration parameters. Land-surface parameters were derived from field studies or literature values. The field capacity was derived as a function of the soil types obtained from our own field surveys. Due to the limited availability of soil information for the TP, soil parameters were distributed according to different land-cover and slope classes (Table 3).

Parameter-optimization procedures are difficult to apply in data-scarce regions such as the TP (e.g., Winsemius et al., 2009). Moreover, various parameter set combinations may yield equally acceptable representations of the (often limited) calibration data, which are referred to as the equifinality problem (e.g., Beven, 2001; Beven and Freer, 2001). Due to a lack of calibration data, we used default settings or parameter values given in the literature (see Table S1 in the Supplement).

Table 3. Soil parameters used as input for hydrological modeling.

Combination land cover–slope	Soil depth (cm)	Field capacity Total (mm)	0–1 dm (mm dm^{-1})	1–2 dm (mm dm^{-1})	2–3 dm (mm dm^{-1})	3–4 dm (mm dm^{-1})	4–5 dm (mm dm^{-1})	5–6 dm (mm dm^{-1})	6–7 dm (mm dm^{-1})
Wetland	70	236	60	60	60	14	14	14	14
Grassland < 15°	70	120	18	18	18	18	16	16	16
Grassland > 15°	40	68	18	18	16	16	–	–	–
Barren land < 5°	20	14	7	7	–	–	–	–	–
Barren land > 5°	10	7	7	–	–	–	–	–	–

Following Mölg et al. (2014), we implemented a precipitation-scaling factor as an additional model parameter to account for (i) HAR10 precipitation overestimation related to atmospheric-model errors and/or (ii) sublimation of blowing or drifting snow that was neglected in the model. Due to the high uncertainty in the range of the precipitation-scaling factor in various regions of the TP (Huintjes, 2014; Mölg et al., 2014; Pohl et al., 2015), we performed model runs with precipitation-scaling factors varying between 0.3 and 1.0 with a 0.05 increment. Because the precipitation-scaling factor was judged to be the parameter that contributes the most to uncertainties in model results, all other climate forcing variables and model parameters were held constant. We compared simulated mean annual lake-volume changes of each model run with water-volume changes derived from remote-sensing data (Fig. 2). The dotted line in Fig. 2 indicates the lake-volume changes derived from LEGOS data (see Table 2). The model run with the minimum difference between modeled and satellite-derived lake-volume change was defined as a reference run and thereby was used for an assessment of model results. The "best" match between simulated and satellite-derived lake-volume change was achieved by applying the following precipitation-scaling factors: 0.80 (Nam Co), 0.75 (Tangra Yumco), 0.85 (Paiku Co) and 0.50 (Mapam Yumco). We discuss the possible reasons for the lower parameter value for the Mapam Yumco basin in Sect. 5.2.

Similar to the calibration process, data scarcity limited the establishment of rigorous and systematic validation tests. Because water-level measurements from Nam Co provide consistent time series between the months of June through November for the years 2006–2010, we chose this period for validation. Given the fact that water routing is not considered in the model, we compared mean monthly, instead of daily, water-level simulations and measurements. For the calculation of monthly average lake levels, the lake-level value of 1 June was set to zero in each year and the subsequent values were adjusted accordingly to make the lake-level changes during the June–November period of the years 2006–2010 comparable.

For an independent assessment of the snow model capabilities, we compared modeled snow-water equivalent (SWE) simulations with MODIS snow-cover data (see Sect. 2.2). Because MODIS data provide no information about the amount of water stored as snow (i.e., SWE), this comparison was only possible in an indirect way by comparing the percent or fraction of snow-covered area (SCAF) derived from the model simulation and MODIS data. Any given spatial model unit was considered to be snow-covered on days when the amount of SWE was larger than a specific threshold (i.e., 1, 10, and 50 mm).

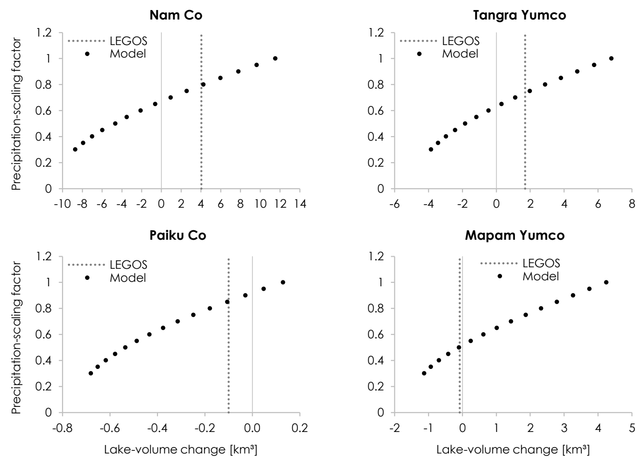

Figure 2. Model-simulated lake-volume changes for Nam Co, Tangra Yumo, Paiku Co and Mapam Yumco for the time periods given in Table 2 using precipitation-scaling factors varying between 0.3 and 1.0. The dotted line indicates lake-volume changes derived from remote-sensing data provided by LEGOS. The point where model dots are closest to the dotted line was taken as the precipitation-scaling factor for each basin.

4 Results

Section 4.1 contains the comparison of simulated and measured water levels of lake Nam Co (Sect. 4.1.1) and of simulated snow-cover dynamics with MODIS for all four study basins (Sect. 4.1.2). Section 4.2 deals with the assessment of the modeling results regarding spatiotemporal variations of water-balance components (Sect. 4.2.1) and their contributions to each basin's water balance (Sect. 4.2.2) during the 2001–2010 period.

4.1 Model evaluation

4.1.1 Comparison of simulated and measured water levels of the Nam Co lake

Lake-level observations of Nam Co indicate a distinct seasonal dynamic with continuously increasing lake levels during the months of June through September caused by runoff from the non-glacierized land surface and glacier areas, a lake-level peak in September and decreasing lake levels from October on primarily caused by lake evaporation. The overall seasonal dynamic during the June–November period is well represented by the J2000g model ($r = 0.81$) (Fig. 3). However, the model overestimates the lake level for the month of November, except for the year 2006.

In general, the magnitude of the lake-level evolution is less well simulated than its timing. The comparison reveals a non-systematic pattern (Fig. 3). In 2006, the model is not able to reproduce the observed increase in lake levels. The substantial lake-level rise of Nam Co in 2008 simulated by the model compares well with observed data. However, the lake-level increase in 2009 is slightly overestimated. The absolute deviation between observed and simulated relative changes of monthly averaged lake levels during the June–November period ranges between −0.31 m (2006) and 0.30 m (2009). The simulated relative lake-level change during the June–November period averaged over the years 2006 to 2010 is 0.41 m, which is approximately 0.05 m higher than the measured one (0.36 m).

4.1.2 Comparison of the simulated snow-cover dynamics with MODIS

The comparison of mean monthly values of modeled snow-covered area fraction (SCAF) (SWE > 1 mm) and MODIS indicates that the model captures seasonal variability quite well. However, there are large deviations in the magnitude. The modeled SCAF (SWE > 1 mm) is generally greater than the MODIS SCAF, with higher deviations in the Mapam Yumco and Paiku Co basins (Nam Co: 30 % vs. 22 %; Tangra Yumco: 17 % vs. 8 %; Mapam Yumco: 54 % vs. 28 %; Paiku Co: 49 % vs. 20 %; Fig. 4). During the winter months November through April, the overestimation by the model (up to a factor of 2) is generally higher than during the summer season. During the months May through October, the modeled SCAF (SWE > 1 mm) in the Nam Co and Tangra Yumco basins is even approximately 50 % lower compared to MODIS. Figure 4 indicates how sensitive the results are by using different thresholds for the amount of SWE to depict an area as snow-covered in the model. The use of higher thresholds (SWE > 10, 50 mm) for derivations of SCAF from the model reduces the overestimation, but also leads to an underestimation of the SCAF in early winter in most basins (Fig. 4). A threshold larger than 10 mm seems to be inappropriate for deriving SCAF from the SWE simulations. It is

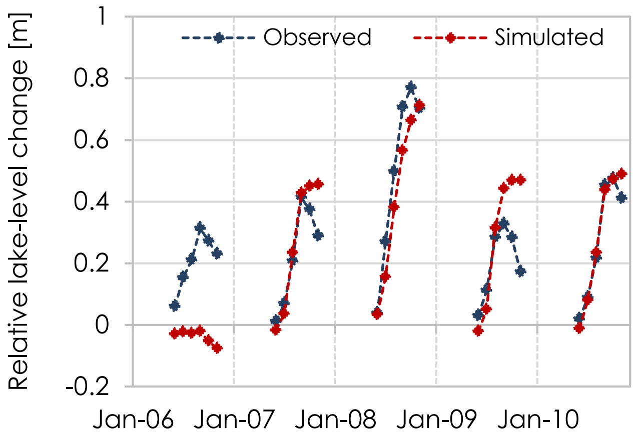

Figure 3. Monthly averaged lake-level observations from Nam Co (blue) vs. simulated lake levels (red) for the June–November period of the years 2006 through 2010.

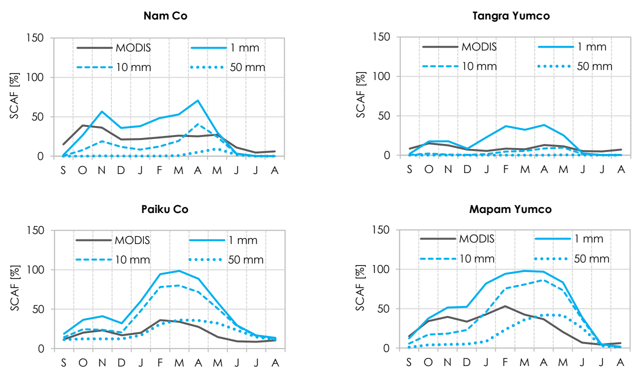

Figure 4. Mean monthly modeled–derived SCAF (blue) using SWE > 1 mm (solid line), > 10 mm (dashed line), and > 50 mm (dotted line) vs. SCAF derived from MODIS (black) for the four study basins.

more likely that the J2000g model overestimates SCAF. This will be discussed later in Sect. 5.2.

4.2 Comparative analysis of the four selected lake basins

4.2.1 Spatiotemporal patterns of hydrological components

The percentage of the precipitation occurring during the wet season (June through September) is more than half of the annual precipitation in all basins. Specifically, June-through-September precipitation is approximately 80 % of the annual total in the Nam Co and Tangra Yumco basins and around 60 % in the Paiku Co and Mapam Yumco basins (Fig. 5a). This indicates a higher influence of the westerlies in the Paiku Co and Mapam Yumco basins. As simulated by the model, snow accumulation in the basins generally occurs beginning in mid-September, reaching a first smaller peak be-

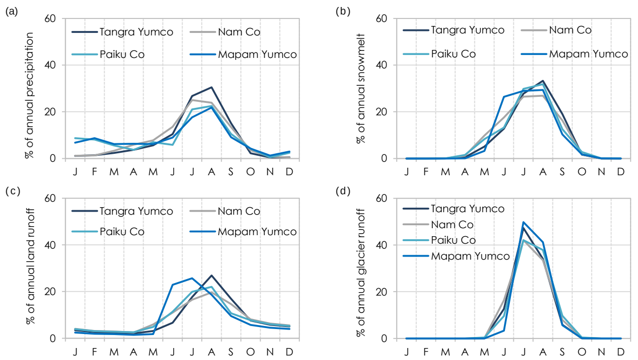

Figure 5. **(a)** Monthly percentage of annual precipitation, **(b)** snowmelt from non-glacierized land areas, **(c)** runoff from non-glacierized land areas, and **(d)** glacier runoff for the four studied basins.

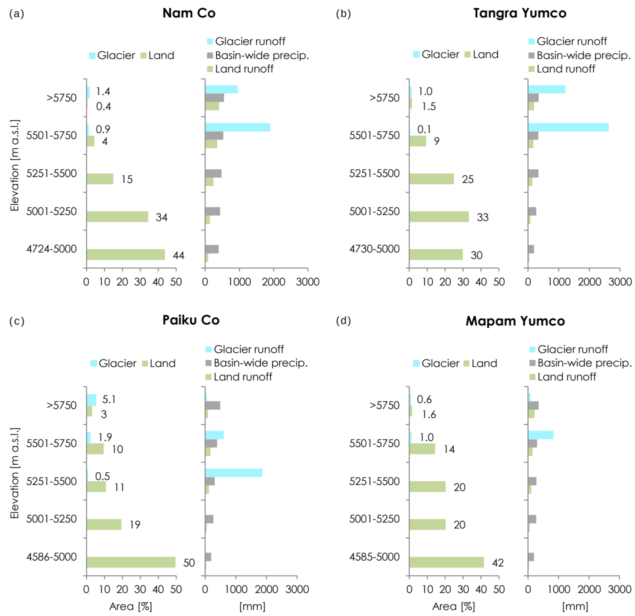

Figure 6. (**a–d**, left panels) Hypsometry of glacier and non-glacierized areas based on mean elevations of respective model entities for the four studied basins. (**a–d**, right panels) Variability of precipitation and runoff from glacier and non-glacierized areas related to altitude for the four studied basins.

tween October and November and the maximum peak between April and May, followed by a rapid decrease in snow between May and June and a slower rate of decrease until September. In the Mapam Yumco basin, simulated snowmelt starts later and occurs over a shorter time period compared to the other basins (Fig. 5b). This can be explained by lower air temperatures in this basin.

About 80 % of simulated annual terrestrial actual evapotranspiration (AET) occurs during the growing season (May–October). Modeled AET has its maximum in July when the availability of soil water and energy is highest. The seasonal cycle of modeled lake evaporation is influenced by seasonal heat-storage changes in the lakes. The released heat in fall acts as an energy source for evaporation. Thus, the evaporation is higher in fall than in spring.

Approximately 70 % of annual precipitation is released to the atmosphere through AET and does not contribute to the runoff in all basins. Discharge from non-glacierized land areas is concentrated during the wet season (∼ 80 % of annual runoff occurs during May through October). Runoff starts to increase in spring with the beginning of snowmelt. The land runoff peak in the Mapam Yumco basin occurs 1 month earlier (between June and July) compared to the other basins (Fig. 5c), because of a higher contribution of snowmelt to the discharge. Glacier runoff occurs during June through September in all basins (Fig. 5d), but with a later beginning and a shorter duration of the melt season in the Mapam Yumco basin due to the colder climate conditions.

Table 4 (upper part) summarizes annual means of modeled water-balance components for the 2001–2010 period for each basin. The annual mean of the model-simulated lake evaporation rates varies between 700 and 900 mm yr^{-1} for the four basins averaged over the 10-year study period. Because of unlimited water availability, the modeled mean annual lake evaporation is substantially higher than the land AET (see Table 4, upper part). Due to higher precipitation amounts in the eastern part of the study region, the simulated mean annual AET is higher in the east (∼ 290 mm in the Nam Co basin) than in the west (∼ 170 mm in the Mapam Yumco basin) (Table 4, upper part).

Impacted by the decreasing precipitation gradient spatially from east to west, the model-simulated mean annual land runoff in the Nam Co basin (∼ 130 mm) is estimated to be more than twice that in the Mapam Yumco basin (∼ 60 mm) during the study period (Table 4, upper part). The combination of various influencing variables such as local climate, topography, land cover, soil and hydro-geological properties results in a spatially heterogeneous pattern of runoff generation within the catchments. Figure 6 illustrates the altitudinal dependence of the mean annual basin-wide precipitation total and runoff from glaciers and non-glacierized land areas, as computed by the J2000g model. The area–altitude relation (hypsometry) for glacier and non-glacierized areas is based on mean elevations of the respective model entities. Larger precipitation amounts in the high mountainous and hilly headwater areas result in higher land runoff estimates compared to lower elevation areas (Fig. 6). Indeed, the increase in land runoff with altitude is higher than the elevation-dependent increase in precipitation. The non-glacierized high-elevation areas characterized by sparse vegetation, poorly developed soils, steep topography and lower air temperatures indicate smaller soil-water contents and lower AET rates compared to lower elevation bands, resulting in higher runoff rates.

Table 4. Mean annual water-balance components, water budget and lake-level changes for the four studied lake basins for the study period 2001–2010 derived from the reference run. The variation ranges of the mean annual water-balance components correspond to model runs with precipitation-scaling factors ±0.05.

		Western basin → eastern basin			
		Mapam Yumco	Paiku Co	Tangra Yumco	Nam Co
Water-balance components (mm yr^{-1})					
Land	Precipitation	230 (±24)	250 (±15)	300 (±20)	420 (±27)
	AET	170 (±9)	180 (±5)	210 (±7)	290 (±8)
	Land runoff	60 (±14)	70 (±8)	90 (±12)	130 (±18)
Glacier	Precipitation	330 (±33)	480 (±28)	330 (±22)	560 (±35)
	Glacier runoff	600 (±8)	320 (±4)	1320 (±12)	1320 (±4)
Lake					
	On-lake precipitation	90 (±9)	140 (±8)	150 (±10)	290 (±18)
	Lake evaporation	710 (−)	910 (−)	840 (−)	770 (−)
	Net evaporation	620 (±9)	770 (±8)	690 (±10)	580 (±18)
Water budget (km^3 yr^{-1})					
Water gain	Land runoff (% of total basin runoff)	0.23 (85)	0.14 (70)	0.70 (86)	1.15 (81)
	Glacier runoff (% of total basin runoff)	0.04 (15)	0.06 (30)	0.11 (14)	0.27 (19)
Water loss	Net evaporation	−0.26	−0.22	−0.57	−0.95
Net water budget	Lake-volume change	0.01	−0.02	0.24	0.47
Lake level (m yr^{-1})	Simulated	0.02	−0.07	0.29	0.24
	Zhang et al. (2011) (GLAS/ICESat 2003–2009)	−0.02	−0.04	0.26	0.25
	Phan et al. (2012) (GLAS/ICESat 2003–2009)	−0.043	−0.118	0.291	0.230
	LEGOS*	−0.05	−0.07	0.25	0.22

* Mean annual lake-level rates for the studied basins correspond to the following time periods: Nam Co – 2001–2010; Tangra Yumco – 2001–2009; Paiku Co – 2004–2008; Mapam Yumco – 2003–2009.

In all studied basins, the runoff from glacier areas located in lower-elevation zones (< 5750 m a.s.l.) significantly exceeds the land runoff in the same elevation zones (Fig. 6), due to high ice-melt rates in the ablation areas. Because of lower temperatures and higher snowfall rates at higher elevations, the modeled glacier runoff decreases with altitude. The modeled mean annual glacier runoff averaged over all glacier HRUs in the Nam Co and Tangra Yumco basins (∼ 1300 mm) is considerably higher than in the Paiku Co (∼ 300 mm) and Mapam Yumco (∼ 600 mm) basins. This is judged to be caused by lower air temperatures (∼ 2 °C less) in the glacier areas of the Paiku Co and Mapam Yumco basins.

4.2.2 Contributions of the individual hydrological components to the water balance

Table 4 (lower part) summarizes the model-simulated mean annual lake-volume and level changes and the contribution of non-glacierized land, glacier, and lake areas to the total water budget during the 2001–2010 study period. Comparative values for the mean annual lake-level changes derived from remote-sensing data are also given in Table 4 (lower part). The contribution of glacier runoff to the total basin runoff volume in the Nam Co (19 %), Tangra Yumco (14 %) and Mapam Yumco (15 %) basins is relatively low compared to the runoff contribution from non-glacierized land areas. The glacierization in the Paiku Co basin is about 2 to 5 times larger than in the other three basins, but the glacier-melt contribution to the total basin runoff volume is only around twice as high (30 %) due to lower glacier-melt rates. Despite the generally higher glacier contribution in Paiku Co, the water balance is slightly negative during the study period (Table 4, lower part). The water loss for Paiku Co exceeds the water gain by 10 %. In contrast, the total water inflow in the Tangra Yumco and Nam Co basins exceeds the water loss by factors of 1.4 or 1.5, respectively. In the Mapam Yumco basin the water gain and loss terms tend to balance each other out (Table 4, lower part), based upon the model simulation.

In order to better predict and understand the role of glaciers in the mean annual water balance, a hypothetical scenario with ice-free conditions was evaluated through model simulations for each lake basin. Therefore, the land-cover class of all glacier HRUs was changed to barren land. In the absence of glaciers, the total runoff volumes in the Nam Co and Tangra Yumco basins would be about 13 % lower than with an ice-melt water contribution during the 2001–2010 period (compared to the reference run). Thus, the mean annual lake-level increases of Nam Co and Tangra Yumco would be reduced from 0.24 to 0.15 m and from 0.29 to 0.17 m, respectively. In the Mapam Yumco and Paiku Co basins, the total runoff volumes would decrease by approximately 30 % and the resulting mean annual lake-level changes would change from 0.02 to −0.18 m and from −0.07 to −0.25 m, respectively, under ice-free conditions. From this latter evaluation, it can be concluded that the mean annual net water budget would noticeably change without ice-melt water contribution; however, the water balance in Nam Co and Tangra Yumco remains positive.

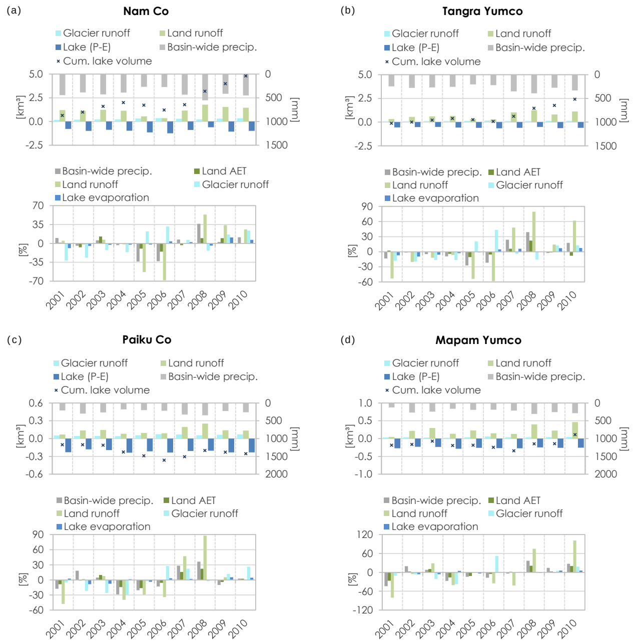

Figure 7. (**a–d**, upper panels) Cumulative lake-volume change (km^3), contribution of several water-balance components (km^3) to lake-volume change and annual basin-wide precipitation amounts ($mm\,yr^{-1}$) for the four studied basins. (**a–d**, lower panels) Annual percentage deviations from the 10-year average of several water-balance components for the four studied basins.

Based upon the J2000g modeling results, the differences in the water balance among the four studied lakes are primarily caused by relatively higher land runoff contributions in the Nam Co and Tangra Yumco basins compared to the Paiku Co and Mapam Yumco basins. This is related to relatively higher precipitation totals in the Nam Co and Tangra Yumco basins compared to the other two basins during the 2001–2010 period.

Figure 7a–d (upper panels) illustrates the yearly water contribution in km^3 of each land-cover type (land, glacier, and lake) for the 2001–2010 period. The annual percentage deviations from the 10-year average of several hydrological system components are presented in Fig. 7a–d (lower panels). Over the study period, annual relative lake-volume changes in the four basins indicate similar patterns. A relatively high correlation of lake-volume changes is found between the Nam Co and Tangra Yumco basins ($r = 0.82$). These are the basins with a higher proportion of June-through-September precipitation compared to the Paiku Co and Mapam Yumco basins.

The modeled annual lake-volume changes of all four lakes are highly correlated with inter-annual variations of land runoff ($r \approx 0.99$). The year-to-year variability of runoff from non-glacierized land surfaces, in turn, is strongly related to inter-annual variations of precipitation ($r \approx 0.92$). Inter-annual variability of lake evaporation is low in all four studied basins and not correlated with lake-level changes. Thus, lake evaporation seems to have a minor impact on inter-annual lake-level variations during the study period. There is also no correlation between annual glacier-melt amounts and lake-volume changes for the four basins. This suggests that glacier-melt runoff is not the main driving force for inter-annual lake variations during the last decade. Although the modeled annual glacier runoff is greater than the 10-year average in the year 2006 in all basins, lower precipitation amounts lead to less land runoff, causing a lake-volume decrease in this year in all basins. In contrast, the year 2008 is judged as having anomalous conditions, with modeled precipitation and land runoff substantially above average and with below-average glacier melt, resulting in a lake-volume increase in all basins. Differences in annual lake-volume changes among the basins are caused principally by regional differences in the inter-annual variations of precipitation.

5 Discussion

5.1 Comparison with other studies

5.1.1 Estimation of the water-balance components

Due to the scarcity of field measurements, model simulations of water-balance components are limited in the TP region. Evaporation over lake surfaces has been estimated for only a few lakes on the TP, based on model simulations (e.g., Morrill, 2004; Haginoya et al., 2009; Xu et al., 2009; Yu et al., 2011). Mean annual lake-evaporation estimates vary between 700 and 1200 mm. The lake-evaporation rates simulated with the J2000g model (between 710 and 910 $mm\,yr^{-1}$) are within this range (Table 4). There are only a few studies for the TP for assessing the actual evapotranspiration over alpine grassland, based on measurements and model simulations (e.g., Gu et al., 2008; Yin et al., 2013; Zhu et al., 2014). Yin et al. (2013) estimated AET over the entire TP using meteorological data available between 1981 and 2010 from 80 weather stations as model input for the Lund-Potsdam-Jena dynamic vegetation model (Sitch et al., 2003). For the southern-central TP, the simulated mean annual AET ranges from 100 to 300 mm, with generally higher values in the east and lower values in the drier regions in the west. Our simulated AET estimates for the four basins vary between 170 and 290 $mm\,yr^{-1}$, decreasing from east (Nam Co basin) to west (Mapam Yumco basin) (Table 4). This compares favorably with the study reported by Yin et al. (2013).

Using a simplified procedure, Yin et al. (2013) developed spatial patterns of the surface-water budget over the entire TP for the 1981–2010 period by estimating the difference between precipitation and AET (P-AET). The results revealed that P-AET depends on climate regimes and gradually decreases from the east ($\sim 150\,mm\,yr^{-1}$) to the west ($\sim 50\,mm\,yr^{-1}$) in the study region. Our model simulations indicate quite similar runoff patterns compared to the findings of Yin et al. (2013), with decreasing annual means from the east (~ 130 mm in the Nam Co basin) to the west (~ 60 mm in the Mapam Yumco basin). The calculated AET / precipitation ratio of around 0.7 in all basins agrees well with study results from Gu et al. (2008).

The mean annual glacier runoff of 1320 mm, simulated with the J2000g model for the Nam Co basin, compares quite well with estimated glacier-melt quantities for the Zhadang glacier in the Nam Co basin using more complex SEB/MB models (Mölg et al., 2014: 1375 mm yr^{-1}; Huintjes et al., 2015: 1325 mm yr^{-1}).

5.1.2 Factors controlling the water balance and lake-level variability

Many studies emphasize the importance of glacier-meltwater contribution to the water budget of Tibetan lakes (e.g., Zhu et al., 2010). However, only a few studies have quantitatively estimated glacier-meltwater contribution to total runoff in the TP region, due to the difficulty in estimating glacier-volume changes (B. Li et al., 2014). B. Li et al. (2014) estimated a glacier-runoff contribution of 15 % of the total runoff (during 2006–2011) in a sub-basin of the Nam Co basin, the Qugaqie basin (8.4 % glacier coverage), using an energy-balance-based glacier-melt model and the Gridded Subsurface Hydrologic Analysis (GSSHA) model (Downer and Ogden, 2004). Based upon the J2000g model results, the glacier contribution ranges between 14 and 30 % in the four studied basins (1–6 % glacier coverage) during the 2001–2010 period. This range of values is higher than that computed by B. Li et al. (2014) considering also the lower percentage of the basin area covered by glaciers in our study basins.

Simulated glacier-meltwater contribution is generally lower compared to the runoff contribution from non-glacierized areas. However, glaciers make an important contribution to the water budget during the 10-year period considering the small extent of ice-covered areas in the four studied lake basins. Indeed, the water balance in the Nam Co and Tangra Yumco basins would also be positive without ice-meltwater contribution during the study period, based on the results of the ice-free scenarios (Sect. 4.2.2). Thus, the question arises why Nam Co and Tangra Yumco indicate a non-equilibrium state, whereas Paiku Co and Mapam Yumco are at a state close to the hydrologic equilibrium.

Endorheic lakes respond to climatic changes to maintain equilibrium between input and output, and to reach steady state. Due to the time lag of lakes in responding to climatic changes, this modeling study cannot confirm whether or not the shift towards a positive water balance in the Nam Co and Tangra Yumco basins or the negative shift in the water balance of the Paiku Co basin, respectively, was primarily caused by changes in precipitation, glacier melt, evapotranspiration, etc. However, inter-annual lake-level variations are highly positively correlated with precipitation and land runoff. This supports the assumption of other studies (e.g., Lei et al., 2014; Y. Li et al., 2014; Song et al., 2014) that increasing precipitation is the primary factor causing lake-level increases in the central TP (where Nam Co and Tangra Yumco are located). The relative stability or slight lake-level declines in the marginal region of the TP (where Paiku Co and Mapam Yumco are located) seem to be related to relatively stable or slightly decreased precipitation (e.g., Lei et al., 2014). Both changes in large-scale circulation systems and local circulation are assumed to be responsible for spatially varying changes in moisture flux over the TP (e.g., Gao et al., 2014, 2015). However, these factors are still under debate and further research is needed (Gao et al., 2015).

The potential evaporation decreased in most areas on the TP during 1961–2000, primarily caused by decreasing wind speeds (Xie and Zhu, 2013). A decreasing trend in potential evaporation before 2000 might have resulted in rising lake levels in the central TP. However, this factor did not prevent the lake shrinkage along the southwestern periphery of the TP, indicating that lake evaporation is not a primary factor for explaining the spatial differences of lake-level changes between the central and southern TP (Lei et al., 2014). In addition, Y. Li et al. (2014) argued that recent rapid lake expansion in the central TP cannot be explained by changes in potential evaporation, because the overall increasing tendency of potential evaporation in the TP region after 2000 (Yin et al., 2013; Y. Li et al., 2014) would negate the effect of increasing precipitation on lake levels.

Under the assumption that glacier-melt runoff increased during the last decades due to climate warming, it is very likely that glacier-meltwater supply augmented the precipitation-driven lake areal expansion in the central TP region (Song et al., 2014). In the Mapam Yumco and Paiku Co basins, glacier-meltwater discharge might have mitigated lake-level decline and acted as a regulating factor (Ye et al., 2008; Nie et al., 2012).

Y. Li et al. (2014) suggest that spatial variations in lake-level changes might be related to different distributions and types of permafrost. Most lakes in the central-northern TP with continuous permafrost are rapidly expanding, whereas lakes in the southern region with isolated permafrost are relatively stable or slightly decreasing. Thus, accelerated permafrost melting might have contributed to the rapid lake expansion in the central and northern TP subregions (Y. Li et al., 2014). The water contribution from permafrost will become limited when ground temperature remains above the melting point of the frozen soil and the water held in the frozen soil has been released (Y. Li et al., 2014). Y. Li et al. (2014) suggest that the permafrost-meltwater contribution may have already become limited in the southern TP. However, this is difficult to corroborate given the absence of observational data for the studied lake basins. The questions remain (i) how large the volume of water released due to thinning and thawing of permafrost is, and (ii) to what extent it can modulate basin runoff. These cannot be answered without adequate information about permafrost occurrence, thickness and ice content in the studied basins.

Differences in the response time or, in other words, the time required to reach an equilibrium state, could also be a reason for observed differences in lake-level changes. Based upon remote-sensing data, the lake areas of Nam Co and Tangra Yumco expanded by 4.6 and 1.8 %, respectively, between 1970 and 2008 (Liao et al., 2013). The lakeshore slopes of Tangra Yumco are steeper compared to Nam Co. Steep-sided lakes have a longer equilibrium response time, because of a lower rate of change of the lake area with volume (Mason, 1994). Based upon paleo-shorelines, the postglacial lake-level high of Nam Co and Tangra Yumco was about 29 m (Schütt et al., 2008) and 185 m (Rades et al., 2013), respectively, above the present-day lake level, supporting the assumption that Nam Co has a shorter response time to compensate for the increment in net inflow (i.e., faster and stronger reaction of its lake area). Moreover, the water supply coefficient (basin area / lake area ratio) for Nam Co is smaller than for the Tangra Yumco basin (5.5 vs. 11.0).

The lake extent of Paiku Co and Mapam Yumco decreased by 3.7 and 0.8 %, respectively, between 1970 and 2008 (Liao et al., 2013). The lakeshores of Mapam Yumco are generally flatter compared to Paiku Co. However, due to only small lake-area variations of Mapam Yumco during the last decades (Liao et al., 2013), differences in lake morphology seem not to be the reason for the relative stability of the recent water levels of Mapam Yumco. Differences in the water-supply coefficient of the Paiku Co and Mapam Yumco basin are quite low (8.8 vs. 10.6).

Based upon results of other studies in the region (e.g., Lei et al., 2013), the effects of upwelling and downwelling groundwater related to fault zones and lake–groundwater exchanges on the water balance were assumed to be negligible. However, Zhou et al. (2013) suggest that water leakage related to seepage might play an essential role in the hydrological cycle of the TP, due to the large numbers of lakes and the sub-surface fault system in the TP region. Groundwater outflow from Mapam Yumco to Langa Co (located only a few kilometers to the west about 15 m below Mapam Yumco) cannot be excluded and could be a reason for the relatively stable water levels of Mapam Yumco. In the more recent past, Mapam Yumco and Langa Co were connected by the natural river Ganga Chu having an extent of about 10 km. However, currently there is no surface outflow (Liao et al., 2013).

5.2 Limitations and uncertainties

Hydrological modeling is hindered by systematic or random model-input errors, model-parameter uncertainty and model-structure inadequacies (Sivapalan, 2003). As stated in many studies (e.g., Knoche et al., 2014; Pohl et al., 2015), precipitation input is the primary source of uncertainty in hydrological modeling studies in data-scarce regions. HAR10 data have been successfully used as modeling input in various studies (Huintjes, 2014; Mölg et al., 2014; Huintjes et al., 2015; Pohl et al., 2015). However, these studies also needed to apply a precipitation-scaling factor of less than 1. Maussion et al. (2014) could not find a systematic bias in comparison with station observations, but it is probable that overestimation of precipitation amounts occurs at high altitudes.

The precipitation-scaling factors were kept constant for the entire 10-year period, because there is no opportunity to derive varying scaling factors for individual years, due to a lack of observations in the lake basins included in this study. This may have an impact on inter-annual variations of modeling results. The non-systematic deviations between simulated and measured lake levels of Nam Co (Fig. 3) might be related to a non-systematic error pattern in the HAR10 precipitation data. The primary issue is that HAR10 precipitation cannot be validated to a sufficient degree, because available data are for stations that are located at lower elevations, and no accuracy assessment can be done for the higher-elevation zones where the study basins are located. The comparison with available station data suggests that the accuracy of the precipitation data is probably regionally dependent (Maussion et al., 2014). This makes it difficult to find a fixed precipitation-scaling factor that is applicable for different regions of the TP.

As described in Sect. 3.3, we conducted multiple model runs using precipitation-scaling factors between 0.3 and 1.0, seeking a precipitation-scaling factor that best simulates satellite-derived lake-volume changes. There may be errors in the satellite-derived water-volume data, which in turn might have affected the estimation of the precipitation-scaling factor and thereby the accuracy of model results. However, the precipitation-scaling factors obtained for the Nam Co (0.80), Tangra Yumco (0.75) and Paiku Co (0.85) basins are relatively close to the scaling factor used for the Zhadang glacier in the Nam Co basin in the study of Mölg et al. (2014). They found very good agreement between glacier mass-balance model calculations and available in situ measurements by applying a precipitation-scaling factor of 0.79. This gives us confidence that the scaling factors used in our study seemed to be within an acceptable range.

The relatively low precipitation-scaling factor of 0.50 obtained for the Mapam Yumco basin seems to be plausible when comparing HAR10 precipitation with weather station data of Burang (30°17′ N, 81°15′ E, ∼ 30 km to the south, the closest station with available data) published in Liao et al. (2013). The mean annual precipitation total of Burang is 150 mm yr^{-1} for the period 2001–2009, whereas the nearest HAR10 point gives a mean annual precipitation amount of 330 mm yr^{-1}. Huintjes (2014) also found that a reduction of the precipitation by more than 50 % leads to more reliable mass-balance results for the Naimona'nyi glacier (Gurla Mandhata, southwestern TP), which is located close to the Mapam Yumco basin.

Uncertainty also arises from the fact that the precipitation-scaling factor can compensate for not only input data errors, but also model-structure inadequacies. Blowing-snow sublimation was neglected in our modeling approach, due to the complexity of this process in complex terrain (Vionnet et al., 2014). However, wind-induced sublimation of suspended snow above the snow pack can be a significant water loss to the atmosphere (e.g., Bowling et al., 2004; Strasser et al., 2008; Vionnet et al., 2014). Vionnet et al. (2014) simulated total sublimation (surface + blowing snow) in Alpine terrain (French Alps) using a fully coupled snowpack–atmosphere model. They estimated that blowing-snow sublimation is two-thirds of total sublimation. This process is judged to be important in the study area, due to the relatively dry near-surface conditions and relatively higher wind speeds occurring during the winter months. Thus, the low values of the scaling factor applied in the Mapam Yumco basin (0.5) and in the study of Pohl et al. (2015) (0.37) might be an indication that drifting-snow sublimation plays a greater role in regions that are more strongly influenced by westerlies.

The omission of processes such as snow redistribution by wind and avalanches and snow loss by blowing-snow sublimation may affect snow-cover patterns as well as the magnitude and timing of melt runoff (Pellicciotti et al., 2014). This could also be a reason for the larger areal snow-cover extent in the model simulation during the winter season compared to MODIS (Sect. 4.1.2). Explanations for lower SCAF values of the model during the summer period could be related to the fact that the MODIS/Terra data are collected only in the morning (10:30) rather than at several times during the day. That means that MODIS indicates snow cover on days when snow was accumulated during the previous night or early morning but which might be sublimated or melted later during the day (Kropacek et al., 2010).

Given the limited data availability, further assumptions and simplifications in the model were required. The currently implemented glacier-melt model component according to Hock (1999) is a simple, robust and easy-to-use methodology that does not account for the transformation of snow into ice. Thus, simulated snowmelt amounts on glacier surfaces might be overestimated. Because glacier-volume changes are not considered in J2000g, unrealistic amounts of glacier-meltwater could be generated. However, the impact of this effect on model results is assumed to be small over the 10-year period. The consideration of glacier-volume changes would be of higher importance for long-term model simulations.

Effects of lake–groundwater interactions were neglected in the model, because the quantification of flow between aquifer systems and a deep lake is difficult (Rosenberry et al., 2014). However, it is unclear whether and to what extent intermittent (at irregular time intervals) exfiltration and infiltration processes might occur, thereby impacting water-level fluctuations. The stated values of lake–groundwater exchange rates do strongly vary within the literature by more than 5 orders of magnitude (Rosenberry et al., 2014). The lack of consideration of lake–groundwater interactions could be the reason that the observed lake-level decrease of Nam Co during the months of October and November is not well represented by the model. If lake levels rise higher than adjacent groundwater levels, lake water may move into the adjacent lakeshores' subsurface. This additional storage factor would basically have a dampening effect on lake-level dynamics. However, in view of multi-annual lake changes, lake–groundwater exchanges are assumed to be negligible.

6 Conclusions and outlook

Hydrological modeling is required to allow for a quantitative assessment of differences in the water balance and thus a better understanding of the factors affecting water balance in the TP region. Addressing this research need, we developed a modeling framework integrating atmospheric-model output and satellite-based data, and applied it to four selected endorheic lakes across the southern-central part of the TP. The J2000g hydrological model was adapted to the specific characteristics of endorheic lake basins in the TP region. The model-derived HAR10 atmospheric data and satellite-derived lake-water surface temperature (Sect. 2.2) served as input for the modeling period 2001–2010. Due to missing continuous lake-level in situ data, we used satellite-derived lake-volume changes as a model-performance criterion.

The adapted J2000g model version reasonably captured seasonal dynamics of relevant hydrological processes. Water-balance estimates of individual years should be interpreted with care, due to possible unsystematic error patterns in HAR10 precipitation. Nevertheless, uncertainties that appear to be related to the precipitation-scaling factor should not affect the overall conclusions drawn from model-application results, as discussed in Sect. 5.2.

The major outcomes can be summarized as follows.

The seasonal hydrological dynamics and spatial variations of runoff generation within the basins are similar for all lake basins; however, the several water-balance components vary quantitatively among the four basins.

Differences in the mean annual water balances among the four basins are primarily related to higher precipitation totals and attributed runoff generation in the basins with a higher monsoon influence (Nam Co and Tangra Yumco).

The glacier-meltwater contribution to the total basin runoff volume (between 14 and 30 % averaged over the 10-year period) plays a less important role compared to runoff generation from rainfall and snowmelt in non-glacierized land areas. However, considering the small part of glacier areas in the study basins (1–6 %), glaciers make an important contribution to the water balance.

Based upon hypothetical ice-free scenarios in the hydrological model, ice-melt water constitutes an important water-supply component for basins with lower precipitation (Mapam Yumco and Paiku Co) in order to maintain a state close to equilibrium, whereas the water balance in the basins with higher precipitation (Nam Co and Tangra Yumco) would still be positive under ice-free conditions.

Precipitation and associated runoff are the main driving forces for inter-annual lake-level variations during the 2001–2010 period. Both are highly positively correlated with annual lake-level changes, whereas no correlation is found between inter-annual variability of lake levels and glacier runoff or lake evaporation.

For the 10-year modeling period used in this study, it is not possible to draw definitive conclusions about the hydrological changes that might have led to imbalances in the water budgets of the four studied lakes. However, the model results support the assumption of other studies that contrasting patterns in lake-level fluctuations across the TP are closely linked to spatial differences in precipitation.

This study demonstrates the feasibility of a methodological approach combining distributed hydrological modeling with atmospheric-model output and various satellite-based data to overcome the data-scarcity problem in the TP region. The integration of readily available model-derived atmospheric and remote-sensing data with hydrological modeling has the potential to improve our understanding of spatiotemporal hydrological patterns and to quantify water-balance components, even in ungauged or poorly gauged basins. The modeling framework presented in this study provides a useful basis for future regionally focused investigations on the space–time transition of lake changes in the TP region.

Model applications in such a data-scarce region have inherent uncertainty that should be perceived as useful information rather than a lack of basic knowledge or understanding (Blöschl and Montanari, 2010). An uncertainty and sensitivity analysis that includes the assessment of spatially and temporally variable effects on model outputs will allow specific and detailed recommendations on the timing and locations of future field measurements (e.g., Ragettli et al., 2013). There is an urgent need in such studies for meteorological observations (particularly precipitation in high mountain regions) and monitoring of land-surface characteristics (vegetation, soil and hydrogeological properties), in order to reduce the model uncertainties arising from input data and land-surface parameterization.

Overall, future research should focus on model-independent data describing hydrological system components that can be used for multi-response calibration and validation purposes. Water-level and volume estimations with a higher temporal resolution are expected to be produced from new satellite-altimetry data, such as from Cryosat (data continuously available since 2012, planned until 2017), Sentinel-3 (2015) and Jason-CS (2017) (Kleinherenbrink et al., 2015), which could be used as calibration or validation data in further model applications in the future.

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