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What are the key drivers of regional differences in the water balance on the Tibetan Plateau?

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Abstract

Lake-level fluctuations in closed basins on the Tibetan Plateau (TP) indicate climate-induced changes in the regional water balance. However, little is known about the region's key hydrological parameters, hampering the interpretation of these changes. The purpose of this study is to contribute to a more quantitative understanding of these controls. Four lakes in the south-central part of the TP were selected to analyze the spatiotemporal variations of water-balance components: Nam Co and Tangra Yumco (indicating increasing water levels), and Mapam Yumco and Paiku Co (indicating 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 hydrological model J2000g 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. Our results reveal that because of the small portion of glacier areas (1 to 7 % of the total basin area) the contribution of glacier melt water accounts for only 14–30 % of total runoff during the study period. Precipitation is found to be the principal factor controlling the water-balance in the four studied basins. The positive water balance in the Nam Co and Tangra Yumco basins was primarily related to larger precipitation amounts and thus higher runoff rates in comparison with the Paiku Co and Mapam Yumco basins. This study highlights the benefits of combining atmospheric and hydrological modeling. The presented approach can be readily transferred to other ungauged lake basins on the TP, opening new directions of research. Future work should go towards increasing the atmospheric model's spatial resolution and a better assessment of the model-chain uncertainties, especially in this region where observational data is missing.

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1 Introduction

The drainage system of the interior Tibetan Plateau (TP) is characterized by numerous closed (endorheic) lake 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 on the TP (e.g., Zhang, B. et al., 2013; Zhang, G. et al., 2013; Song et al., 2014).

Due to the accelerated glacier mass loss, it has been hypothesized that lake-level increases are primarily due to an increased inflow of glacier melt water (e.g., Yao et al., 2007; Meng et al., 2012). Nevertheless, glacier runoff into lakes itself should not increase the overall water-volume mass on the TP, based upon the GRACE satellite gravimetry data (Zhang, G. 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). However, potential evaporation has tended to increase, especially in the inner Plateau, during the last decade (e.g., Yin et al., 2013). Thus, recent studies addressing the controlling mechanism of lake-level fluctuations remain controversial. This raises the question: what are the key drivers of regional differences in the water balance on the TP?

To answer this question, knowledge of spatiotemporal variations of several water-balance components and their contribution to the basin water balance is needed. Some recent studies emphasize the urgency of the quantification of water-balance compo-

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nents by using hydrological models (e.g., Cuo et al., 2014; Lei et al., 2014; Song et al., 2014). However, previous studies primarily rely on simplified water-balance calculations (e.g., Zhu et al., 2010; Zhou et al., 2013; Li, L. et al., 2014). Hydrological modeling studies of endorheic lake basins are rare on the TP (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) by developing 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 (Huintjes, 2014; Mölg et al., 2014), but has not yet been used as input for catchment-scale hydrological modeling studies on the central TP. The objective of this study is the hydrological modeling and system analysis of glacierized endorheic lake basins along a lake transect across the southern-central part of the TP in order to:

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

ii. quantify single water-balance components and their contribution to the water balance, and obtain a quantitative knowledge of the components governing water-balance changes at a catchment-to-regional scale.

To our knowledge, this is the first hydrological study applying a distributed, process-oriented model in multiple lake basins on the TP. The Nam Co basin was chosen as starting point for the modeling approach applied in this study, because it is the basin with the best hydrological data availability. The Tangra Yumco, Paiku Co and Mapam Yumco basins (Fig. 1) were included in this study for the transfer of the hydrological model. 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 asses similarities and differences among the basins; in Sect. 5, the results are discussed with respect to findings from other

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studies, sources of modeling uncertainties and factors influencing long-term lake-level changes. 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 closed lake basins along a west–east (W–E) lake transect in the south-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 through 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 poorly developed and vegetation throughout the study area is generally sparse. The growing period lasts approximately five months, from late April/early May to late September or mid-October (Zhang, B. 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 area 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 the Paiku Co and Mapam Yumco slightly decreased by around $-0.05\,\mathrm{m\,yr^{-1}}$ (Zhang et al., 2011; Phan et al., 2012).

2.2 Data used

Because of limited availability of climatological data on the TP, a new atmospheric dataset for the TP, the "High Asia Refined analysis (HAR)" (Maussion et al., 2014) was used as input for the hydrological model. The HAR data sets were generated by a dynamical downscaling of global-analysis data (Final Analysis data from the Global Forecasting System; dataset 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 resolutions (hourly, daily, monthly and yearly). In this study, the HAR10 data were used. 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 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.

Lake-surface water temperature (LSWT) estimates from the ARC-Lake v2.0 data products (MacCallum and Merchant, 2012) were used as additional input for the hydrological modeling in the Nam Co and Tangra Yumco basins (see Sect. 3.1.1). 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 plausibility check to obtain mean daily LSWT time series for the 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). The SRTM Version 4 data were used 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). Land-cover information were derived from Landsat ETM/TM data classification (for the Nam Co and Tangra Yumco basin) or obtained from the "Himalaya Regional Land Cover" data base (http://www.glcn.org/databases/hima_landcover_en.jsp) (for the Paiku Co and Mapam Yumco basin).

Due to the absence of continuous lake-level measurements, satellite-based lake-level and water-volume data were obtained for the four studied basins from the HydroWeb data base (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 the four basins are close to the change rates estimated by Zhang et al. (2011) and Phan et al. (2012) using GLAS/ICESat data (2003–2009) (see Table 4).

MODIS snow-cover 8 day data of Terra (MOD10A2) and Aqua (MYD10A2) satellites at a spatial resolution of 500 m were used for validation of the snow modeling in the Nam Co basin (see Sect. 3.3). As proposed in the literature (e.g., Parajka and Blöschl, 2008; Gao et al., 2010; Zhang et al., 2012), Terra and Aqua data were combined 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 the original MODIS/Terra snow-cover data were used. After the combination procedure the cloud cover percentage was on average less than 1 %, with higher values during the wet summer season (~ 1 %) than in the drier winter season (~ 0.6 %).

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, a semi-distributed conceptual model structure, primarily following the J2000g model (Krause and Hanisch, 2009), was selected. The J2000g model is a simplified version of the fully-distributed J2000 model (Krause, 2002). The J2000g model has a smaller number of calibration parameters and does not account for lateral flow processes between spatial model units. It was successfully applied for hydrological predictions in data-scarce basins (e.g., Deus et al., 2013; Knoche et al., 2014; Rödiger et al., 2014), including a previous modeling study in the Nam Co basin (Krause et al., 2010).

Meteorological data requirements are daily times series of precipitation, minimum, maximum and average air temperature, solar radiation, wind speed, relative humidity and cloud fraction. The HAR10 data were used as climate input for the 10 year study period. Daily LSWT data were used as additional data inputs 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. Regardless of the influence of long-term storage changes, such as deep groundwater, the net water budget of the four selected closed lake basins (in terms of lake-volume change) was estimated by summing up the runoff from non-glacierized land areas (generated by snowmelt and rainfall) and from glacier areas (generated by snow and ice melt) minus

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lake net evaporation (lake evaporation minus on-lake precipitation). For simplicity, the terms land runoff, glacier runoff and net evaporation are used to refer to several water-balance components. Assuming a simplified depiction of lake geometry, the modeled mean annual lake-water storage changes were divided by the corresponding lake area in order to provide mean annual lake-level change estimates for the four studied lakes.

The conceptual model presented herein 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 closed lake basins on the TP. A schematic illustration of the model structure including several model contexts and components is presented in Fig. S1 in the Supplement.

For the interpolation of the HAR10 raster points (centroid of the raster cell) to each HRU unit, the regionalization procedure implemented in J2000g was used. This combines Inverse Distance Weighting (IDW) with an optional elevation correction. All data sets, except air temperature, were regionalized using only IDW. Net radiation and evapotranspiration were calculated following the Food and Agriculture Organization of the United Nations (FAO) proposed use of the Penman–Monteith model (Allen et al., 1998). The long-wave radiation part of the FAO56 calculation was modified according to the recommendations of Yin et al. (2008). For this study, the commonly used approach for calculating net long-wave radiation over water surface (e.g., Jensen, 2010) was implemented in JAMS. For the estimation of open-water evaporation rates from large lakes, the Penman equation modified with the addition of an empirical estimation of the lake heat storage (Jensen et al., 2005) was used. As suggested by Valiantzas (2006), the

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reduced wind function proposed by Linacre (1993) was applied for the estimation of evaporation from large open-water body surfaces. The J2000 snow module that combines empirical or conceptual approaches with more physically-based routines (Nepal et al., 2014) was used in this study. Glacier ice-melt rate was calculated according to an extended temperature-index approach (Hock, 1999). Soil-water budget and runoff processes were simulated using a simple water storage approach implemented in the J2000g soil module (Krause and Hanisch, 2009). A more detailed description of the theoretical and methodological background of the model components and enhancements are given in the Supplement.

3.2 Delineation of spatial model entities

In order to provide spatially distributed information of landscape characteristics for the hydrological modeling, the Hydrological Response Units (HRUs) approach (Flügel, 1995) was applied. Using ArcGIS software, HRUs with similar hydrological behaviour were delineated by overlaying topographic-related and land-cover information. Land-cover data were reclassified in five hydrologically relevant classes: water, wetland, grassland, barren land and glacier. Soil and hydro-geology information were not included in the overlay analysis due to a lack of detailed data. In the absence of detailed spatially-distributed information of landscape parameters, as is the case for the TP, the spatial scale of model entities usually becomes coarser; whereby, the spatial homogeneity of a given HRU decreases. Nevertheless, 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 in the lower elevations (smaller number of large polygons). The total number of HRUs varied 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 spatial and temporal static calibration parameters. Land-surface parameters were derived from field studies or literature values. Due to the limited availability of soil information for the TP, observed 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 representation of the (often limited) calibration data, which is referred to as the equifinality problem (e.g., Beven, 2001; Beven and Freer, 2001). Due to a lack of calibration data, default settings or parameter values given in the literature were used in this study (see Table S1 in the Supplement).

Following Huintjes (2014) and Mölg et al. (2014), a precipitation-scaling factor was implemented as additional model parameter to account for (i) HAR10 precipitation overestimation related to atmospheric model errors and/or (ii) sublimation of blowing or drifting snow which was neglected in the model. Mölg et al. (2014) proposed a parameter range between 0.5 and 0.8 for the precipitation-scaling factor to apply to the HAR10 data in the Zhadang glacier area in the Nam Co basin. Due to the high uncertainty of the range of the precipitation-scaling factor in various regions on the TP (Huintjes, 2014), model runs with precipitation-scaling factors varying between 0.3 and 1 (0.05 was used as factor increment) were performed. 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. Simulated mean annual lake-volume changes of each model run were compared 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

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lake-volume change was defined as 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 following precipitation-scaling factors: 0.80 (Nam Co), 0.75 (Tangra Yumco), 0.85 (Paiku Co) and 0.50 (Mapam Yumco). The possible reasons for the lower parameter value for the Mapam Yumco basin compared to the other basins are discussed in Sect. 5.2.

Similar to the calibration process, data scarcity limited the establishment of rigorous and systematic validation tests. For an independent assessment of the snow model capabilities, modeled snow water equivalent (SWE) simulations of the Nam Co basin were compared 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 as snow-covered at days when the amount of SWE was larger than 1 mm. Then, SCAF within the Nam Co basin was calculated for each month and hydrological years (September through August). To evaluate further the plausibility of the model results, simulated glacial melt quantities in the Nam Co basin were compared with the results of SEB/MB models applied for the Zhadang glacier in the Nam Co basin, using the HAR10 data (Huintjes, 2014; Mölg et al., 2014). Further plausibility analysis of individual water-balance components have been conducted by comparing model outputs with results from other studies published in the literature (Sect. 5.1).

4 Results

Section 4.1 presents the modeling results of several hydrological components in the Nam Co basin, focusing on seasonal and year-to-year variations. The modeling results of all basins are compared in Sect. 4.2, assessing spatiotemporal variations of

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water-balance components and similarities and differences in the water balance from a regional perspective for the 2001–2010 period.

4.1 Seasonal and inter-annual variations of hydro-climatological components in the Nam Co basin

4.1.1 Air temperature and precipitation

The basin-average mean annual air temperature (MAAT) in the Nam Co basin was −3.5 °C for the 2001–2010 period. The years in the second half of the decade were warmer compared to the first half (Fig. 3a, right panel), which is mostly related to higher winter temperatures. In terms of seasonal variation of air temperature the warmest month was July with a spatially averaged MAAT of 7 °C and the coldest month was January with an average MAAT of −14 °C (Fig. 3a, left panel). The monthly MAAT was below 0 °C for a seven-month period (October through April) (Fig. 3a, left panel). The modeled annual precipitation varied between 270 mm (2005, 2006) and 550 mm (2008) during the 10 year period (Fig. 3a, right panel) with a mean annual value of ∼ 400 mm. The annual precipitation totals showed a high variability compared to the 10 year annual mean. In particular, 2005 and 2006 were dry years and 2008 was an extreme wet year.

4.1.2 Snow dynamics

As simulated by the model, snow began accumulating in mid-September, reaching a first smaller peak in November and the maximum peak between April and May, followed by rapid snow decay between May and June and a slower decrease until August (Fig. 4). A relatively small amount of modeled snow (between 10–15 %) was lost through sublimation during the winter compared to the amount of snow released through snowmelt during the spring and summer (Fig. 3b, left panel).

The modeled snow-covered area in the Nam Co basin was on average 25 % greater than MODIS snow cover (SCAF 28 vs. 21 %, Fig. 4). During the winter months November through April the modeled SCAF overestimated the MODIS SCAF by a factor of 2. During the months May through October, however, the modeled snow-cover extent was approximately 40 % lower compared to MODIS. Yearly values of modeled SCAF and MODIS SCAF are correlated ($r = 0.86$), indicating that the model captures inter-annual variability quite well.

4.1.3 Glacier melt

Glacier melt (snow and ice melt) was concentrated between June and September (Fig. 3c, left panel). The modeled glacier runoff ranged from 900 to 1700 $mm\,yr^{-1}$ (Fig. 3c, right panel) with a 10 year average of $\sim$ 1300 mm. Air temperature and the amount, timing, and form of precipitation were the principal controlling factors of the year-to-year variations. For example, the lower glacier melt amounts in the years 2001 and 2008 were caused by higher snow accumulation rates between May and June which led to a shortening of the ice-melt period. In contrast, lower snowfall rates at the beginning and during the ablation season in specific years (e.g., 2006) resulted in excessive ice melt, because the glacier ice was snow free for a longer time. As a consequence of higher temperatures during the last years of the study-period's decade, the modeled percentage of snowfall declined, causing higher ice-melt rates in relatively wet years (e.g., 2010).

4.1.4 Evapotranspiration and runoff

About 80 % of simulated annual actual evapotranspiration (AET) occurred during the growing season (May through October) (Fig. 3d, left panel). Modeled AET had its maximum in July when the availability of soil-water and energy was highest. About 70 % of annual precipitation was lost to the atmosphere by AET and did not enter the runoff. The modeled mean annual AET amounted to 290 mm, varying between 240 mm (2006) and 320 mm (2008, 2009) during the study period (Fig. 3d, right panel). The inter-annual variations of modeled AET were determined principally by precipitation ($r = 0.76$). In-

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deed, modeled AET in the year 2009 was as high as in the wettest year 2008. Analyzing the temporal variations of further climate factors controlling AET (air temperature, wind, humidity, radiation) it was noted that the year 2009 stood out due to relatively high wind speeds during this year, thus intensifying the evapotranspiration process. The simulated runoff from non-glacierized land surface varied between 50 mm in dry years (e.g., 2006) and 200 mm in wet years (e.g., 2008). The modeled year-to-year variability of runoff on non-glaciated land surfaces was strongly related to inter-annual variations of precipitation ($r = 0.94$).

The combination of various influencing variables such as local climate, topography, land cover, soil and hydro-geological properties leads to a spatially distributed pattern of runoff generation within the catchment. Figure 5 illustrates the variability of simulated mean annual basin-wide precipitation total and runoff from non-glacierized land surface and glacierized areas related to altitude. The area–altitude relation (hypsometry) for non-glacierized land surface and glacier areas was derived based on mean elevations of single model entities. Larger precipitation amounts in the high mountainous and hilly headwater areas resulted in higher land runoff estimates compared to lower elevation areas (Fig. 5). Indeed, the increase of the modeled mean annual land runoff rates with altitude was higher than the elevation-dependent increase of mean annual precipitation. The non-glacierized high-elevation areas characterized by sparse vegetation, poorly developed soils, steep topography and lower air temperatures indicated smaller soil-water contents and lower AET rates compared to lower elevation bands, resulting in higher runoff rates. However, the modeled runoff from low-altitude glacier areas significantly exceeded the land runoff in the same elevation band, 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 decreased with altitude.

4.1.5 Lake evaporation

The seasonal cycle of the modeled lake evaporation and the water-surface temperature is illustrated in Fig. 3e (left panel). Based on the ARC-Lake water-surface temperature, the Nam Co froze up in mid-January, or as late as early February during the 2001–2010 period. The frozen period ended between early-April and late-April. The late freeze-up dates are related to the large water volume of the Nam Co. The seasonality of the water-surface temperature with increasing temperatures in spring and decreasing temperatures in autumn is an indicator for the seasonal variation of the heat storage of the Nam Co (Haginoya et al., 2009). The released heat in autumn acts as energy source for evaporation. Thus, the modeled evaporation was higher in autumn than in spring (Fig. 3e, left panel).

The model-simulated mean annual lake evaporation for Nam Co ranged from 710 to 860 $mm\,yr^{-1}$ during the last decade (Fig. 3e, right panel) with an annual mean of 770 mm. The year 2001 exhibited the lowest evaporation rate; whereas, 2009 indicated the highest evaporation amount. Inter-annual variations in modeled lake evaporation were related to specific combinations of air temperature, wind speed, radiation, water-surface temperature, and humidity in individual years (Fig. 6). The higher air temperatures toward the end of the study period's decade as provided by HAR10 are mostly related to an increase in winter temperatures. This either does not affect lake-surface water temperatures (the lake is frozen in winter), or is an artifact in HAR10. In 2001, relatively low air temperatures, wind speeds, net radiation and water-surface temperatures, and relatively high humidity related to the other years led to less evaporation. In contrast, in 2009, net radiation, wind speeds, air temperatures, and water-surface temperatures were comparatively high and humidity relatively low, enhancing the evaporation process.

4.2 Regional comparative analysis of multiple 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-to-September precipitation is approximately 80 % of the annual total in the Nam Co and

Tangra Yumco basins and only around 60 % in the Paiku Co and Mapam Yumco basins (Fig. 7). This indicates a higher influence of the Westerlies in the Paiku Co and Mapam Yumco basins. The seasonal dynamics of simulated snow and ice melt, land runoff, terrestrial evapotranspiration, and lake evaporation in the Tangra Yumco, Paiku Co and Mapam Yumco basins are similar to the seasonal variations described above for the Nam Co basin (Sect. 4.1).

Table 4 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 varied between 700 and 900 mm yr^{-1} in the four basins for the study period. In all four basins, the modeled on-lake precipitation was three to four times lower than the water loss by lake evaporation. Because of unlimited water availability, the modeled mean annual lake evaporation was approximately three to four times higher than the land AET (see Table 4).

Due to higher precipitation amounts in the eastern part of the study region, the simulated mean annual AET was higher in the east (~ 290 mm in the Nam Co basin) than in the west (~ 170 mm in the Mapam Yumco basin) (Table 4). According to the decreasing precipitation gradient from east to west, the simulated mean annual land runoff in the Nam Co basin (~ 130 mm) was also around twice as high as in the Mapam Yumco basin (~ 60 mm) during the study period. The percentage of the annual precipitation which was lost by terrestrial AET increases with decreasing precipitation totals from 69 % (Nam Co basin) to 74 % (Mapam Yumco basin) (Table 4). Within all the catchments, the AET/precipitation ratio decreases with altitude as described for the Nam Co basin (Sect. 4.1.4).

On average, the modeled mean annual glacier runoff rates in the Nam Co and Tangra Yumco basins (~ 1300 mm) were approximately two to four times higher than in the Paiku Co (~ 300 mm) and Mapam Yumco (~ 600 mm) basins, which was caused principally by lower air temperatures (~ 2 °C less) in the glacierized areas of the Paiku Co and Mapam Yumco basins.

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4.2.2 Regional differences in the water balance

The water volume of the four lakes included in this study generally increases during the months of June through September. This is due to the following factors: (i) runoff from the non-glacierized land surface generated by snowmelt and direct rainfall, and (ii) water from glacial melting. Then the lake's water-volume decreases until May of the following year, primarily as a result of lake evaporation or sublimation. The relation between water loss through net evaporation and water gain through runoff from non-glacierized and glacier areas in essence sets each basin's water balance. 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 period for the four basins are summarized in Table 4. Comparative values for the mean annual lake-level changes derived from remote-sensing data also are given in Table 4.

The total water inflow in the Tangra Yumco and Nam Co basins exceeded the water loss by a factor of 1.4 or 1.5, respectively. Because of the small portion of glacier areas (1–2 % of the total basin area, Table 1), the mean annual contribution of modeled glacier-melt water to total basin runoff volume was only 14 % in the Tangra Yumco and 19 % in the Nam Co basin. In contrast to Nam Co and Tangra Yumco, the water loss term for Paiku Co exceeded the water gain terms by 10 %; whereas, in the Mapam Yumco basin the water gain and loss terms tended to balance each other out. The glacierization in the Paiku Co basin is about three to six times larger than in the other three basins, but the glacier melt contribution to the total basin runoff volume was only 30 % during the study period. The differences in the water balance among the four basins over the study period were primarily caused by relatively lower land runoff contributions in the Paiku Co and Mapam Yumco basins compared to the Nam Co and Tangra Yumco basins. This is related to lower precipitation totals in the Paiku Co and Mapam Yumco basins compared to the other two basins.

Figure 8a–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 per-

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centage deviations from the 10 year average of several hydrological system components are presented in Fig. 8a–d (lower panels). The modeled annual lake-volume changes of all studies lakes are highly correlated to inter-annual variations of land runoff ($r \approx 0.99$). No correlation was found between annual glacier melt amounts and lake-volume changes in the four basins.

Over the study period, annual relative lake-volume changes in the four basins indicated similar patterns. A relatively high correlation was found between the Nam Co and Tangra Yumco basin ($r = 0.82$) which are the basins with a higher proportion of June-to-September precipitation compared to the Paiku Co and Mapam Yumco basins. Although the modeled annual glacier runoff was above the 10 year average in the year 2006 in all basins, lower precipitation amounts led to less land runoff, causing a lake-volume decrease in this year in all basins. In contrast, the year 2008 might be considered 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. To explore the mechanism controlling spatial patterns of precipitation in single years was beyond the scope of this study.

5 Discussion

5.1 Comparison with other studies

Due to the scarcity of field measurements in the TP region, the research on spatiotemporal variations of water-balance components and on their environmental controls is limited. Evaporation over lake surfaces has been estimated for only 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 on the TP vary between 700 and 1200 mm. Our results are within this range (Table 4). There are only few stud-

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ies for the TP for assessing the actual evapotranspiration over alpine grassland, based on measurements and model estimations (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 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 south-central TP, the simulated mean annual AET ranged from 100 to 300 mm, with generally higher values in the east and lower values in the drier regions in the west. Our model-simulated AET for the four basins varied 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 of 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 ($\sim 150\,\text{mm yr}^{-1}$) to the west ($\sim 50\,\text{mm yr}^{-1}$) 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).

In order to evaluate the plausibility of glacier-runoff simulations, the modeled glacial-melt quantities in the Nam Co basin were compared with the results of SEB/MB models applied for the Zhadang glacier in the Nam Co basin, using the HAR10 data (Huintjes, 2014; Mölg et al., 2014). The comparison revealed that the modeled basin-wide mean annual glacier runoff of ∼ 1300 mm is in the same order of magnitude as simulated by the SEB/MB models for the Zhadang glacier. The modeled year-to-year variability with lower melt rates in the years 2001–2004 and 2008, and higher melt rates in the years 2005–2007, 2009 and 2010 agreed well with the findings of Mölg et al. (2014). Nevertheless, the range between individual years was more pronounced using the energy-balance approach. Regarding the low contribution of glacier melt water to total runoff,

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it was also found by Li, B. et al. (2014) that glacier runoff plays a minor role compared to snowmelt and rainfall runoff components from non-glacierized areas. For the period 2006–2011, they estimated a glacier runoff contribution of 15 % to the total runoff in a sub basin of the Nam Co basin, the Qugaqie basin (8.4 % glacierized coverage area), using an energy-balance based glacier-melt model and the "Gridded Subsurface Hydrologic Analysis (GSSHA) model" (Downer and Ogden, 2004).

5.2 Limitations and uncertainties

Hydrological modeling studies in ungauged or poorly gauged basins are generally fraught with uncertainty most related to the fact that reliable input data are not available and process dynamics are often unknown and/or unobservable. Hydrological predictions in data-scarce regions are severely hindered by systematic or random model-input errors, model-parameter uncertainty and model-structure inadequacies (Sivapalan, 2003). Model-input uncertainty stems from the fact that HAR10 grid points surrounding the lakes were used as input for the modeling of lake evaporation and estimation of on-lake precipitation. As stated in many studies (e.g., Knoche et al., 2014), precipitation input is the primary source of uncertainty for hydrological modeling studies in data-scarce regions. The HAR10 precipitation output was compared to rain-gauge data and to 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. The comparison of HAR10 precipitation with observations primarily located near population centers in lower elevations in the eastern part of the Plateau revealed a non-systematic error pattern (Maussion et al., 2014). In particular, precipitation totals in the summer of years 2007–2010 indicate a positive bias. The reasons for these discrepancies for some years are unclear. The differences after 2007 may be associated with changes in the global data assimilation system (Maussion et al., 2014). However, the uncertainty of HAR10 precipitation in complex terrain such as in the case study basins, where no measurements are available, is unknown. Fol-

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lowing Huintjes (2014) and Mölg et al. (2014), a precipitation-scaling factor was applied in all basins. The precipitation-scaling factors were kept constant for the entire 10 year period. Indeed, this may have had an impact on inter-annual variations of modeling results if there is a non-systematic error pattern in the HAR10 precipitation data.

Uncertainty arises also from the fact that the precipitation-scaling factor can compensate not only input data errors, but also model-structure inadequacies, in particular wind-induced sublimation of suspended snow above the snow pack which 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. Blowing-snow sublimation was neglected in our modeling approach, due to the complexity of this process in complex terrain (Vionnet et al., 2014). However, this process is judged to be important in the study area, due to the relatively dry near-surface conditions and higher wind speeds during the winter months. 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 the 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 a.m.) rather than at several times during the day. That means that MODIS indicates snow cover at 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).

As described in Sect. 3.3, multiple model runs were conducted using precipitation-scaling factors between 0.3 and 1, seeking a precipitation-scaling factor that best simulates satellite-derived lake-volume changes. Figure 2 indicates how sensitive the model results are to the precipitation-scaling factor. The lower precipitation-scaling factor of 0.50 for the Mapam Yumco basin compared to the other three basins, where scaling factors between 0.75 and 0.85 resulted in the "best" match between simulated and satellite-based water-volume changes, might be an indication either that HAR10 precipitation is significantly overestimated in this specific basin or that drifting-snow sublimation plays a greater role than in the other basins. Huintjes (2014) also found that the use of a precipitation-scaling factor lower than the factor applied to the Zhadang glacier in the Nam Co basin lead to more reliable glacier mass balance results for the Naimona'nyi glacier (Gurla Mandhata, south western TP) which is located close to the Mapam Yumco basin. A further source of uncertainty is that there may be errors in the satellite-derived water-volume data, which were used for the setting of the precipitation-scaling factor. Thus, errors in satellite-based lake-volume changes might have affected the setting of the precipitation-scaling factor and thereby the accuracy of model results. Apart from the precipitation-scaling factor, default settings of the J2000g model were applied or parameter values were taken from the literature due to missing calibration data. Therefore, model outputs might also be influenced by assumptions on certain key model parameters other than the precipitation-scaling factor.

In respect to the given limited data availability, further assumptions about or simplifications of the system were required. The currently implemented glacier-melt model component according to Hock (1999) is a simple, robust and easy to use methodology. The less complex model structure might be a major factor for a lower year-to-year variability compared to the SEB/MB model results (Huintjes, 2014; Mölg et al., 2014). Furthermore, effects of lake-groundwater interactions were neglected in the model applications to basins in the TP region, because the quantification of flow between aquifer systems and a deep lake is difficult (Rosenberry et al., 2014). However, it is unclear if and to what extent intermittent (at irregular time intervals) exfiltration (flow from groundwater to a lake) and infiltration (flow from a lake to groundwater) processes might occur, thereby impacting water-level changes. The stated values of lake-groundwater exchange rates do strongly vary within literature by more than five orders of magnitude. The relative contribution of exfiltration to input terms in lake-water budgets ranges from near 0 to 94 %, and infiltration contribution to loss terms ranges from near 0 to 91 % (Rosenberry et al., 2014).

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5.3 Factors influencing long-term lake-level changes

Under constant climatic conditions, closed lakes should eventually tend to attain a stable equilibrium, where several water-balance terms tend to balance each other out. Although the Paiku Co and Mapam Yumco were at a state near to the hydrologic equilibrium during the study period, the Nam Co and Tangra Yumco with a lake-level increase between 0.2 and 0.3 $m\,yr^{-1}$ indicate a non-equilibrium state. During recent decades, the central and eastern part of the TP exhibited an increasing trend in precipitation (Yin et al., 2013) which might be a primary reason for the positive water balances noted for the Nam Co and Tangra Yumco during the last decade. For a closed lake, where the lake evaporation is higher than the on-lake precipitation, the lake area and thus lake evaporation will increase over time as consequence of a positive water balance. Thus, this has the consequence that the lake-volume rate of change will be reduced over time. The time lag in the response of the area of a closed lake to climate fluctuations or, in other words, the time required to reach an equilibrium depends upon the climate over each lake and associated catchment area and the geomorphological characteristics of the lake. The more arid the climate conditions and the higher the rate of change of a given lake's area with volume, the faster the lake can adjust its area (and thus its net evaporation rate) to compensate for a given increment in net inflow (Mason, 1994). Thus, different lake response times need to be considered when analyzing long-term water-balance changes.

6 Conclusions and outlook

Lake-level changes as an indicator for differences in water-balance changes on the TP have become a focal point for research in recent years. This research highlights the possible benefit for conducting quantitative hydrological-modeling studies to explore the causes of lake-level changes. In this study, the hydrological model J2000g was extended and applied to endorheic lake basins on the TP aiming to provide a more

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quantitative understanding of the key factors controlling the water-balance on the TP. The model results indicated that the relative contribution of glacier runoff to total water inflow (between 15–30 %) plays a relatively minor role, compared to precipitation and snowmelt runoff components from non-glacierized areas. Indeed, the glacier-derived runoff contribution increases during dry years. The small glacier contribution to the basin water balance is related to the low percentages of glaciers in the selected basins (ranging between 1–7 %) during the study's period (2001–2010). It is concluded that the positive water balance in the Nam Co and Tangra Yumco basins was caused by higher precipitation totals and thus higher land runoff rates compared to the Mapam Yumco and Paiku Co basins with relative stable or slight negative water balances, respectively.

Data scarcity on the TP complicates model parameter setting and limits the implementation of rigorous and systematic validation testing. Therefore, model applications in such a data-scarce region have inherent uncertainty which 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).

In general, 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. Moreover, observations of blowing-snow sublimation and lake-groundwater exchanges would be very helpful to clarify their role in the water balance. Overall, future research should focus on model-independent data describing hydrological system components which 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 (continuously data available since 2012, planed 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. Moreover, atmospheric data at higher than 10 km × 10 km resolution would allow more reliable estimations of hydrological components in future modeling studies.

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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	30°42	81°28	2380	270	10	6.5	43	0.5	40
Mapam Yumco	4580	28°55	85°35	4440	420	10	1.5	64	2.5	22

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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

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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	–	–	–	–	–	–

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Table 4. Water-balance components, water-budget and lake-level changes for the four studied lake basins for the study period 2001–2010.

		Western basin ⟶ Eastern basin			
		Mapam Yumco	Paiku Co	Tangra Yumco	Nam Co
Water-balance components [mm yr^{-1}]					
Land	Precipitation	230	250	300	420
	AET (AET/precip.)	170 (0.74)	180 (0.72)	210 (0.70)	290 (0.69)
	Land runoff	60	70	90	130
Glacier	Precipitation	330	480	330	560
	Glacier runoff	600	320	1320	1320
Lake	On-lake precipitation	90	140	150	290
	Lake evaporation	710	910	840	770
	Net evaporation	620	770	690	480
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	0.23
LEGOS*		−0.05	−0.07	0.25	0.22

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

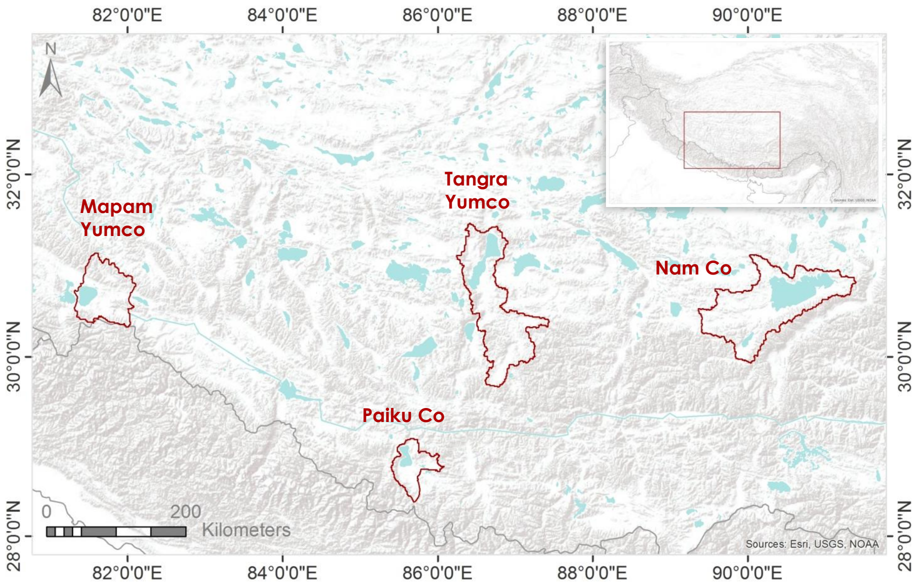

Figure 1. Location of the study region comprising four selected closed lake basins.

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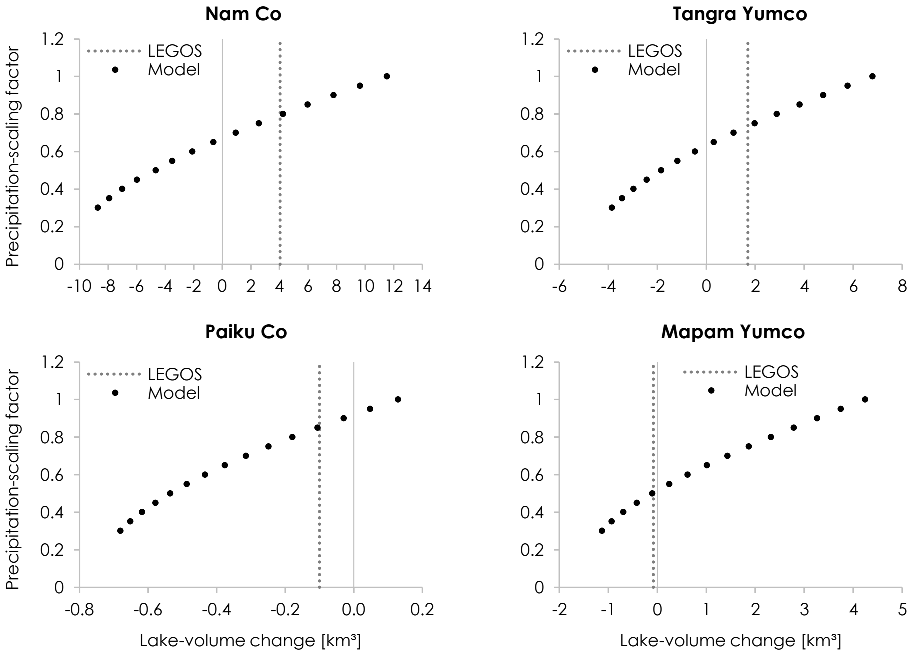

Figure 2. Model-simulated lake-volume changes for Nam Co, Tangra Yumo, Paiku Co and Mapam Yumco for the time periods given in Table 2 using precipitation-scaling factors varying between 0.3 and 1. Dotted line indicates lake-volume changes derived from remote-sensing data provided by LEGOS. At the point where model dots are closest to the dotted line is how the precipitation-scaling factor was set for each basin.

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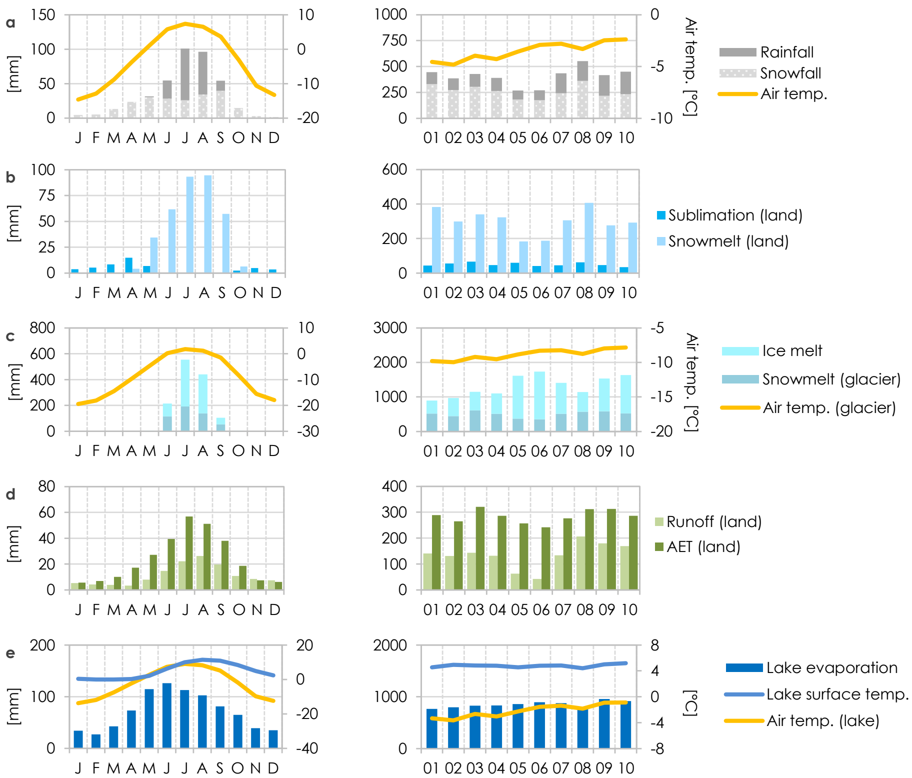

Figure 3. Mean monthly and inter-annual variations (2001–2010) of modeled hydro-climatological components in the Nam Co basin. **(a)** Basin-wide air temperature, rainfall, snowfall. **(b)** Snow sublimation and snowmelt for the non-glacierized land surface. **(c)** Glacier-wide air temperature and glacier melt (snow and ice melt). **(d)** Evapotranspiration and runoff from the non-glacierized land surface. **(e)** Air temperature over the lake, water-surface temperature and evaporation from the lake.

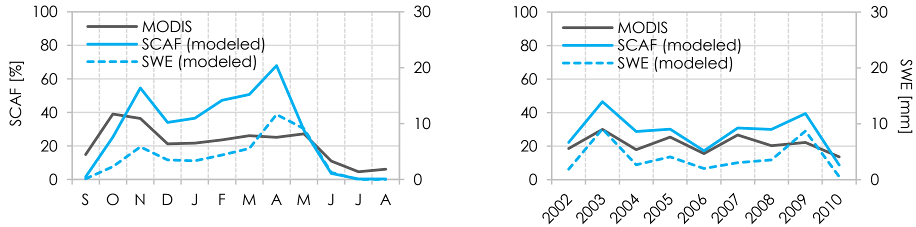

Figure 4. Mean monthly and inter-annual variations of modeled snow water equivalent (SWE) and snow-covered area fraction (SCAF) vs. SCAF derived from MODIS for the Nam Co basin.

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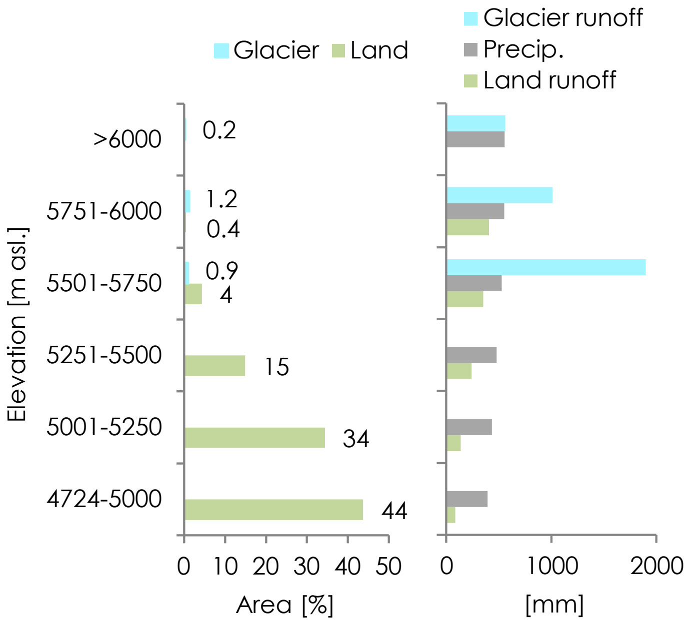

Figure 5. Hypsometry of non-glacierized land areas and glacier areas based on mean elevations of single model entities for the Nam Co basin (left panel). Variability of precipitation and runoff from non-glacierized land areas and glacier areas related to altitude (right panel).

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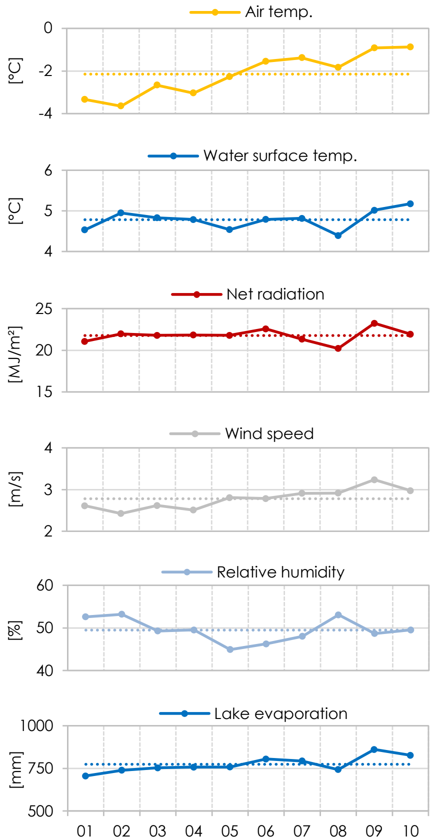

Figure 6. Inter-annual variations of modeled lake-surface water temperature, lake evaporation, and various climate factors (air temperature, net radiation, wind speed, relative humidity) over the lake Nam Co. Dotted lines represent the corresponding 10 year annual means.

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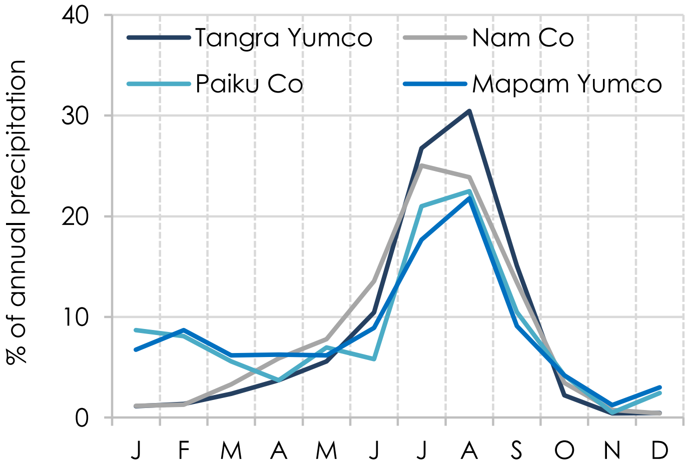

Figure 7. Monthly percentage of annual precipitation for the four studied basins.

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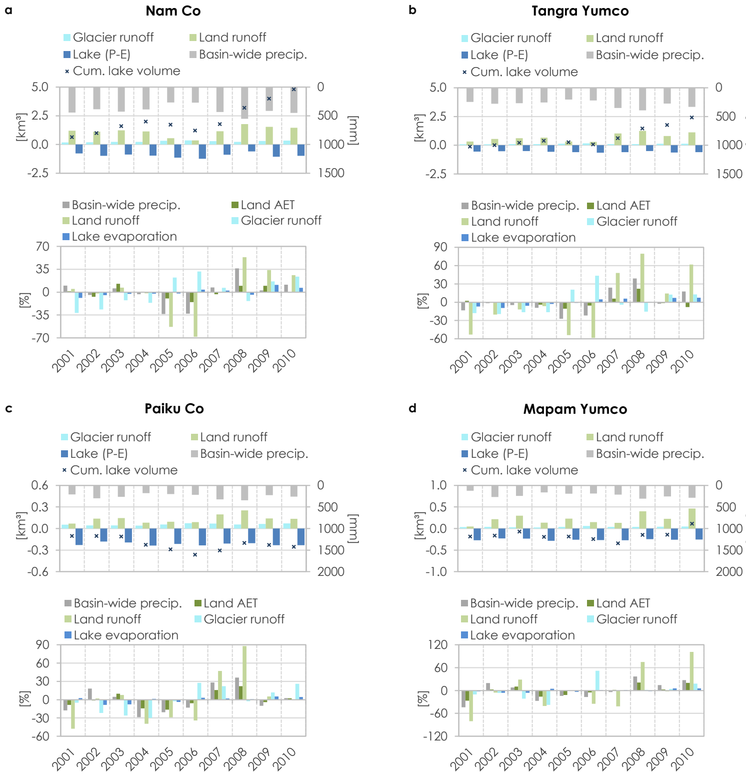

Figure 8. (**a–d**, upper panels) Cumulative lake-volume change (km^3), contribution of several water-balance components (km^3) to lake-volume change and annual basin-wide precipitation amounts ($mm\,yr^{-1}$) for the four studied basins. (**a–d**, lower panels) Annual percentage deviations from the 10 year average of several water-balance components for the four studied basins.