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Data evaluation and numerical modeling of hydrological interactions between active layer, lake and talik in a permafrost catchment, Western Greenland

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SUMMARY

This study investigates annual water balance conditions and their spatiotemporal variability under a wide variety of atmospheric driving conditions in the periglacial permafrost catchment of Two Boat Lake in Western Greenland. The study uses and combines a comprehensive hydrological multi-parameter dataset measured at the site with site conceptualization and numerical model development, application and testing. The model result reproduces measured lake and groundwater levels, as well as observations made by time-lapse cameras. The results highlights the importance of numerical modeling that takes into account and combines evapotranspiration with other surface and subsurface hydrological processes at various depths, in order to quantitatively understand and represent the dynamics and complexity of the interactions between meteorology, active layer hydrology, lakes, and unfrozen groundwater below permafrost in periglacial catchments. Regarding these interactions, the water flow between the studied lake and a through talik within and beneath it is found to be small compared to other water balance components. The modeling results show that recharge and discharge conditions in the talik can shift in time, while the lake and active layer conditions in the studied catchment are independent of catchment-external landscape features, such as the unfrozen groundwater system below the permafrost and the nearby continental-scale ice sheet.

1. Introduction

Periglacial hydrology has during the last decade been a research topic of growing interest, especially in the climate change context. Permafrost soils and bedrock, which may contain significant fractions of ice, are highly sensitive to climate change. Climate-driven responses in landscapes, ecosystems and water systems may then be both heterogeneous and complex. A general lack of data and process understanding related to freezing and thawing processes in periglacial environments has been identified (Vaughan et al., 2013).

The hydrological cycle and water balance conditions of periglacial regions are influenced by ice, snow and frozen soil, with large seasonal fluctuations in the surface energy balance (Woo et al., 2008). The presence of permafrost, and its seasonal freeze and thaw processes may control the distribution and routing of water across the landscape (Quinton and Carey, 2008). Permafrost and its variability are therefore important for hydrological and water balance responses to change in high latitude areas (Kane et al., 1989; Hinzman et al., 1991; Karlsson et al., 2012). Sublimation has also been shown to be significant in the water balance of cold region environments (Reba et al., 2011; MacDonald et al., 2010) where the major part of the annual runoff is generated by snowmelt (McCann and Cogley, 1972). In general, the hydrological impacts of snow and surface and subsurface freezing processes accentuate the importance of including winter and spring measurements in hydrological and water balance assessments of catchments (Hirashima et al., 2004; Verrot and Destouni, 2015).

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Thorough water balance studies have been carried out for the near-surface hydrological system in the active layer of permafrost catchments (Søgaard and Hasholt, 2001; Woo and Marsh, 2005; Helbig et al., 2013), wetlands in permafrost areas (Jessen et al., 2014) and thermokarst lakes (Fedorov et al., 2014; Karlsson et al., 2014). With regard to the hydrogeology of cold regions, Hinzman et al. (2013) summarized the current status of research progress including regional studies (Cheng and Jin, 2013), process studies (Sjöberg et al., 2013) and simulation studies (Frampton et al., 2013; Wellman et al., 2013; McKenzie and Voss, 2013). However, the linking of surface hydrological processes to subsurface hydrology below the permafrost is not as well investigated. Even though groundwater recharge and discharge are reduced by the presence of permafrost, water exchange of deep and shallow groundwater through taliks exists and needs to be taken into consideration in catchments with permafrost (Bense et al., 2009; Frampton et al., 2011, 2013; Ge et al., 2011; Grenier et al., 2013; Bosson et al., 2012, 2013a; Kane et al., 2013). In a changing climate, talik growth might increase (Romanovsky et al., 2010; Vaughan et al., 2013) with increased contact between deep and shallow groundwaters as a consequence. Therefore, there is a need for an integrated approach in hydrological modeling that can take both supra- and sub-permafrost groundwaters into account.

To our best knowledge, only a few studies have considered the links and interactions between both surface and subsurface water, and shallow and deep groundwater for a whole catchment. However, a modeling study by Bosson et al. (2013a) investigated the exchange of shallow and deep groundwaters in a far-future colder Sweden. The modeling was based on simulated climate time series, but comparison to measurements could not be done for the possible future cold state of the investigated Swedish catchment. The present study uses a comprehensive dataset, presented in Johansson et al. (2015) for a periglacial catchment in Western Greenland, to test some key aspects of the modeling approach and simulation results of Bosson et al. (2013a). The catchment, which is called Two Boat Lake, drains into a lake underlain by a through talik and is located in an area with continuous permafrost.

By use of the collected dataset and applied numerical modeling, this study aims to analyze the long-term water balance conditions and their spatial and intra-annual temporal variability, and to identify and quantify the main hydrological flow paths and processes in the permafrost catchment of Two Boat Lake. The study uses and combines meteorological and hydrological data on surface water and groundwater levels, along with conceptual site and process understanding, to construct and calibrate a numerical model that quantifies the catchment scale hydrology and hydrogeology.

Methodologically, the annual thawing and freezing dynamics of the active layer and its influence on hydrological flows are represented in the modeling by soil temperature and its influence on hydraulic properties; this uses the same modeling approach as Bosson et al. (2010, 2012) and tests it against the available dataset. In particular, the numerical modeling approach of Bosson et al. (2010, 2012) used time-varying hydraulic properties, instead of a thermal model component, in order to describe thawing and freezing processes; the present study tests this approach on site-specific data from the Two Boat Lake permafrost catchment.

2. Materials and methods

2.1. Methodology

The investigated catchment of Two Boat Lake (TBL) in the Kangerlussuaq area, Western Greenland, is characterized by a dry climate, where running surface water that is not related to glacial melt is uncommon. Site investigations in the catchment were initiated in 2010 and have resulted in a wealth of information on the catchment hydrology and local meteorological conditions constituting the basis for the conceptual and numerical modeling described in the present paper. Discharge measurements of surface water are often a keystone in water balance studies for calibrating hydrological models, but such measurements could not be performed in the dry TBL catchment. The measurement installations, sampling and resulting data for the TBL catchment (until December 31, 2013) are described in detail in Johansson et al. (2015). The methodology employed in the present study is illustrated in Fig. 1 and includes: (i) constructing a conceptual and numerical hydrological model of the TBL area, (ii) calculating the water balance of the TBL catchment for different time periods, and (iii) studying the intra-annual hydrological interactions of the active layer, lake and talik.

All available data from the site are used in the analysis and conceptualization (A and B in Fig. 1). The methodology for modeling permafrost dynamics in MIKE SHE (Graham and Butts, 2005), which is the numerical modeling tool used in the present study, is described in Bosson et al. (2012). This methodology together with the conceptual model and site data are used to build up the numerical model (C in Fig. 1), which is calibrated and evaluated on an annual basis using data from the investigated period 2010–2013 (D in Fig. 1). The calibration targets of this first modeling stage are to reproduce the temporal pattern in the measured lake level variation and to achieve a water balance of the TBL catchment for the simulated period in accordance with observed values. A sensitivity analysis of (i) hydraulic properties in the active layer and bedrock and (ii) temperature-based time-varying processes in the active layer and on the ground surface are performed in order to reach these calibration targets.

In order to also relate the relatively short period of local data to long-term water balance conditions, 35 years of meteorological data from Kangerlussuaq (25 km from TBL) are used to define wet, dry and normal years with respect to precipitation (P). The three selected years are simulated using the calibrated model from stage one with the aim to analyze the sensitivity of the water balance components to weather conditions. In a last step, simulation results are also evaluated with respect to spatial and intra-annual temporal hydrological and hydrogeological variations within the catchment. In the following sections, the steps corresponding to each box B–F in Fig. 1 are described.

2.2. Site description and data overview

The TBL study catchment is situated approximately 25 km Northeast of the settlement Kangerlussuaq, in West Greenland (Fig. 2). This is the most extensive ice-free part of Greenland and includes a broad range of Arctic environments, from the coast, through plains and a hilly terrain, to the ice sheet margin. The

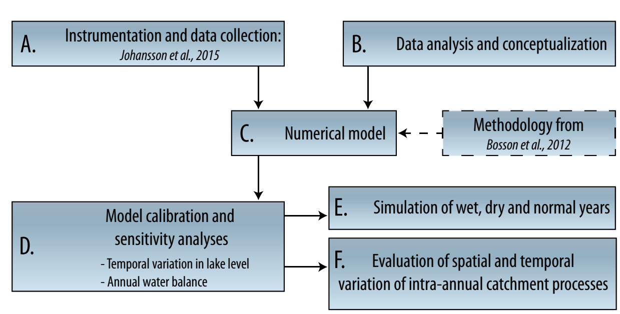

Fig. 1. Overall workflow from data collection to construction and evaluation of conceptual and numerical models of the hydrology and hydrogeology in the Two Boat Lake catchment, Western Greenland.

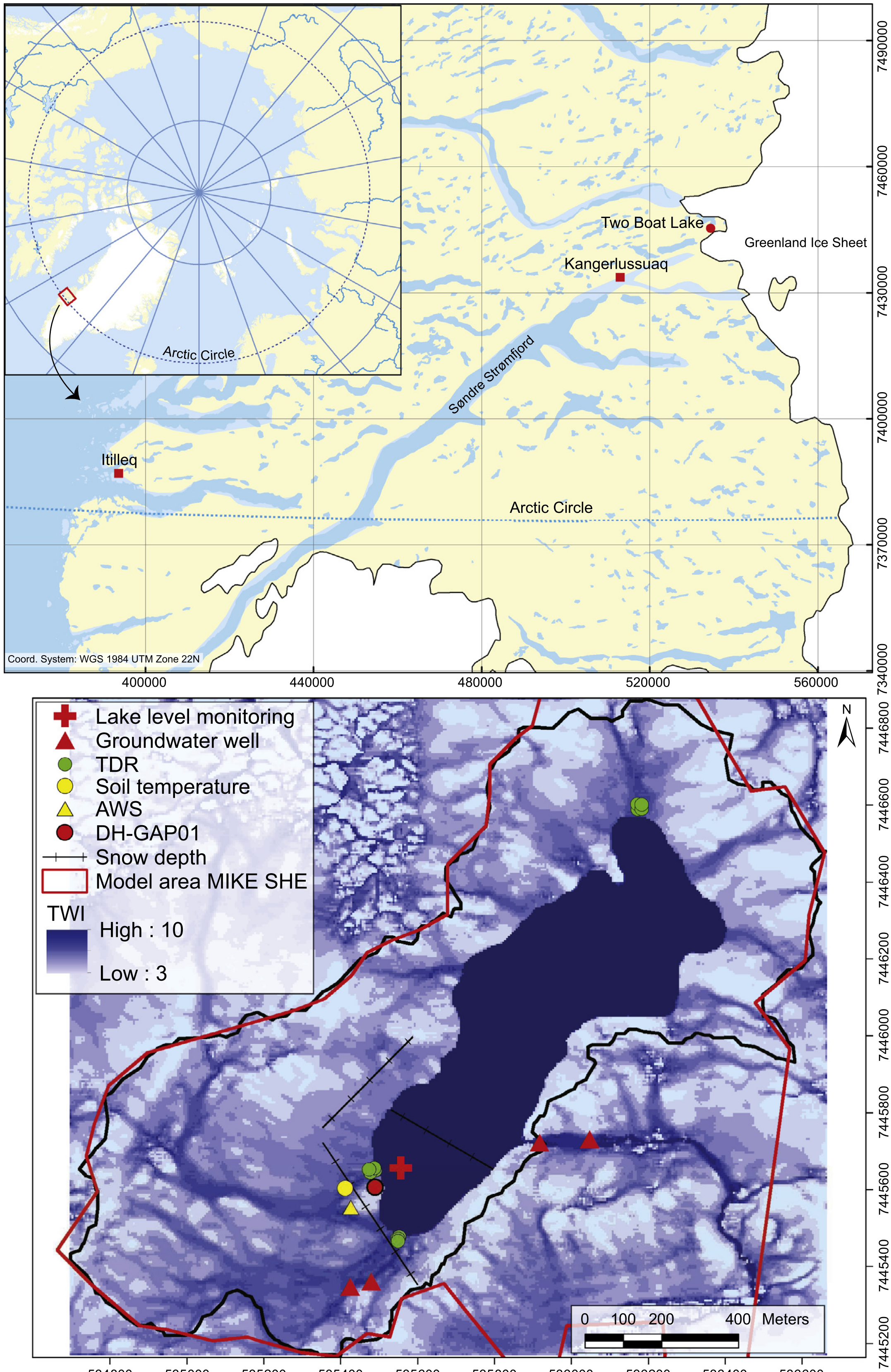

Fig. 2. The regional area of Kangerlussuaq (upper) and the catchment area of Two Boat Lake (lower). In the lower figure the Topographical Wetness Index (TWI) within the catchment is illustrated together with the MIKE SHE model area and observation points of interest within the catchment.

Table 1
Data from Johansson et al. (2015) used in the conceptual and numerical modeling of the Two Boat Lake catchment. Data are divided into time-series, measurement campaigns, soil sampling and geometries.

Time series	Measurement campaigns	Soil sample analyses	Geometries
Lake water level	Evaporation	Retention curves	Topography (DEM)
Groundwater level in the active layer	Sublimation	Hydraulic conductivity	Lake bathymetry
Hydraulic head in bedrock below lake[a]	Snow depth	Grain size distributions	Lake threshold
Soil water content	Active layer depth	Porosity	
Soil temperature	Infiltration capacity		
Precipitation (liquid and solid)			
Wind speed			
Wind direction			
Radiation (long and short wave)			
Barometric pressure			
Relative humidity			
Air temperature			
Time lapse photos			

[a] Data described in Harper et al. (2011).

regional area is characterized by continuous permafrost, interrupted by taliks. The active layer in the area between Kangerlussuaq and the ice sheet margin has a thickness of 0.15–5 m (van Tatenhove and Olesen, 1994). The study site includes a lake, with a verified through talik beneath it (Harper et al., 2011), and its surrounding catchment area (Fig. 2 and Table 1).

Despite its close vicinity to the ice sheet, ~500 m, the lake is a precipitation-driven lake and no surface inflow of glacial melt water occurs (Johansson et al., 2015). Except from the Watson River, extending from the ice sheet toward the fjord in Kangerlussuaq, no rivers are present in the area. A digital elevation model (DEM) for the TBL catchment was developed (Clarhäll, 2011), which describes both the catchment topography and the lake bathymetry. The surface water divide is well defined around the catchment according to the DEM and was also verified in the field. The groundwater divide in the active layer is assumed to coincide with the surface water divide. The catchment elevation ranges from ~335 to 500 m above sea level. The small valleys in the northern and southern parts of the catchment are characterized by high topographical wetness index values (TWI, Beven and Kirkby, 1979) whereas the hill slopes and local topographical highs have low TWI (Fig. 2); high and low TWI correspond to wet and dry soil conditions, respectively.

The soils in the TBL catchment are dominated by till and glaciofluvial deposits, which in a large parts are overlain by a layer of aeolian silt to fine sand. In some topographically low areas close to the lake, the aeolian silt/fine sand is overlain by a layer of peaty silt. The total depth of the aeolian sediments and glacial deposits ranges from 7 to 12 m in the valleys to zero along the hill sides. In the lake, a sediment thickness of up to 1.5 m has been observed (Petrone, 2013). In general, the hydraulic conductivity (K) and porosity (n) decrease with depth. The assumption that the observed depth of glacial deposits in the valleys of the catchment is also applicable for the depth of glacial deposits present below the lake implies that the lake sediments are underlain by approximately 10 m of till.

Vegetation in the TBL catchment is dominated by dwarf-shrub heath. There are no trees, and bushes rarely exceed a height of 0.5 m. The dataset from Johansson et al. (2015) used in the conceptualization and numerical modeling of TBL is summarized in Table 1 and the conceptual interpretation of the data is described in Section 2.3–2.5.

2.3. Permafrost, active layer and talik

Temperature profiles from a nearby borehole, DH-GAP03 (Harper et al., 2011), situated between the ice sheet and TBL, show that the permafrost thickness is ~335 m in the area. Another borehole, DH-GAP01 (Fig. 2), was drilled in a 60° inclination under the TBL, reaching a vertical depth of ~190 m. Based on borehole temperature data, it is shown that the upper 20 m of the DH-GAP01 borehole is located in permafrost. The transition from frozen to unfrozen ground coincides with the lake shoreline, i.e., temperature data shows that a talik exists under the lake. The mean temperature in the bedrock under the lake is 1.2 °C. There are no indications of decreasing temperatures in the deepest part of the borehole, which would have been the case if the talik had been a closed talik. Thus, the available data indicate that there is a through talik beneath the lake.

Soil temperature monitoring (Fig. 2) indicates that the deepest active layer during the period 2010–2013 was 0.9 m, with the variations between the monitored years being relatively small (Johansson et al., 2015). The different periods of frozen and active conditions in the TBL catchment (Table 2 and Fig. 3C), used in the numerical modeling, are defined based on soil temperature data and soil moisture data from the TBL catchment. The thawing and freezing periods in Table 2 are characterized by temperatures fluctuating just above and below 0 °C. In order to identify when water on the ground surface is in liquid state or frozen, air temperature data is used. In general, the thawing of the active layer is initiated in the beginning of May and the freezing starts in mid-September.

2.4. Meteorological data

The regional climate is dominated by relatively stable high-pressure cells over the ice sheet. The mean annual corrected P at the station in Kangerlussuaq (Figs. 2 and 3B) is 173 mm (measured 1977–2013) (Cappelen ed., 2014; Johansson et al., 2015). An automatic weather station (AWS) was installed at TBL in 2011 (Fig. 2) in order to study the local weather conditions prevailing within the catchment. The annual corrected P at TBL in 2012 and 2013 was 365 mm and 269 mm, respectively (Johansson et al., 2015), approximately 1.8 times greater than the P measured in Kangerlussuaq in the same years. In these study years, the major part of the precipitation fell in April–May and August–October, and approximately 40% of the annual precipitation fell as snow. Mean annual air temperature (MAAT) in Kangerlussuaq is −4.8 °C (Cappelen ed., 2014). The MAAT observed at TBL during the time period 2011–2013 was −4.3 °C. During the same period, non-freezing temperatures persisted at TBL throughout the summer months. The winter temperatures were typically between −10 and −20 °C. However, above-freezing temperatures occurred several times each winter, implying that there has been more than one snowmelt event each year. This has also been confirmed by time-lapse photos.

Table 2
Definition of active and frozen periods for the period 2011–2013. Data are given for the soil depths 10, 25, 50 and 75 cm.[a] Ground surface data are based on air temperature and used in the numerical modeling to define when overland water is mobile or immobile. The different periods presented in the table are also illustrated in Fig. 3C.

	Ground surface	10 cm	25 cm	50 cm	75 cm
Thawing period 2011	3 May–2 June	No data	8 June–13 June	30 June–5 July	28 August–3 November
Active period 2011	2 June–17 September	No data	13 June–19 Sep	5 July–21 September	No active period
Freezing period 2011	17 September–3 October	No data	19 September–5 October	21 September–29 October	No active period
Frozen period 2011–2012	3 October–1 May	4 November–9 May	5 October–17 May	29 October–9 June	3 November–16 July
Thawing period 2012	1 May–5 May	9 May–23 May	17 May–26 June	9 June–13 June	16 July–31 July
Active period 2012	5 May–30 September	23 May–18 October	26 June–6 October	13 June–8 October	31 July–1 October
Freezing period 2012	30 September–6 October	18 October–13 November	6 October –30 October	8 October–23 November	1 October–23 December
Frozen period 2012–2013	6 October–3 May	13 November–25 May	30 October–1 June	23 November–21 June	23 December–1 August
Thawing period 2013	3 May–3 June	25 May–1 June	1 June–6 June	21 June–25 June	1 August–14 August
Active period 2013	3 June–	1 June–	6 June–	25 June–	14 August–

[a] Time series data are presented in Johansson et al. (2015).

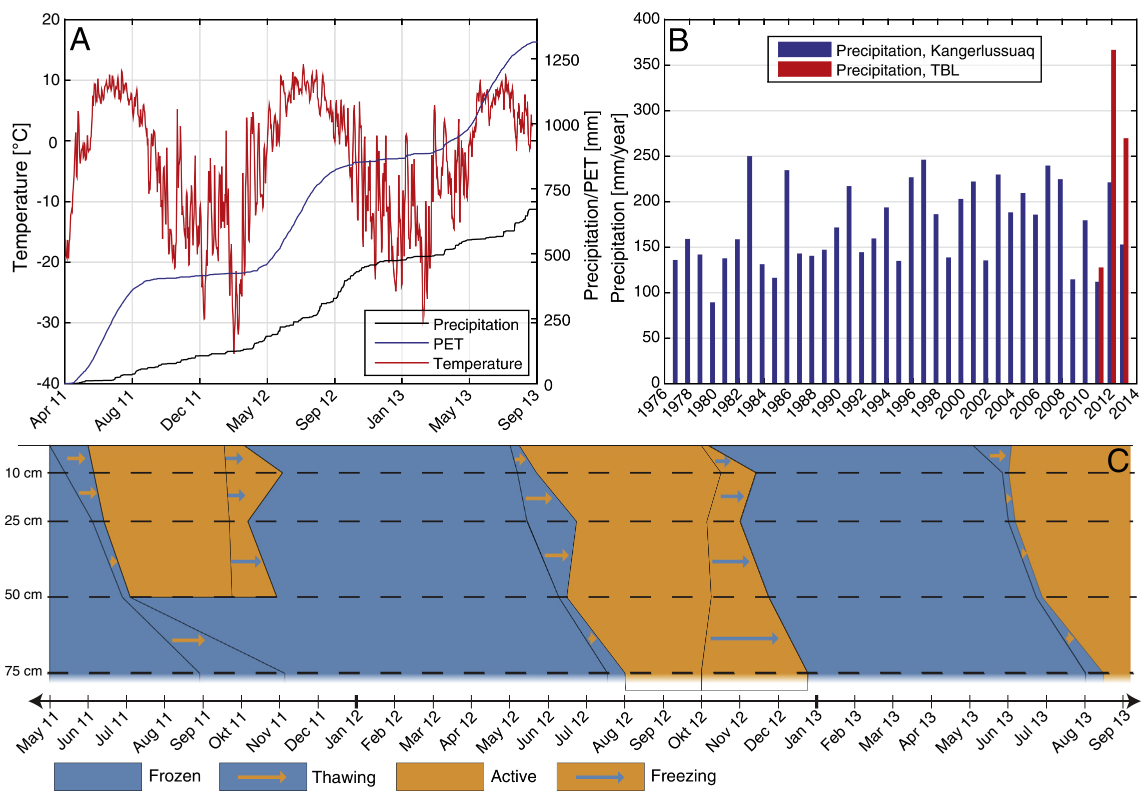

Fig. 3. (A) Precipitation (*P*), potential evapotranspiration (PET), and temperature (*T*) from the local automatic weather station (AWS) at TBL for the period 2011–2013. (B) Annual precipitation measured in Kangerlussuaq 1977–2013. For the period 2011–2013 the annual *P* measured at TBL is also marked. (C) A schematic figure of the different periods of frozen, freezing, thawing and active conditions based on the time periods presented in Table 3.

In general, the wind speed has been low, with typical values below 5 m/s. The strongest winds, above 10 m/s, occurred in winter. Snow depth measurements performed in April 2011 indicate an irregular snow cover, with snow accumulation in the valleys, and snow blowing away from wind exposed areas on hillsides and ridges (Johansson et al., 2015). In April 2013, sublimation measurements were performed at three sites in the catchment resulting in a mean sublimation rate of 0.63 mm/day for the three-day observation period (Johansson et al., 2015).

The MIKE SHE model requires climate data on air temperature (*T*), potential evapotranspiration (PET) and precipitation (*P*). Local data from the AWS for the period April 2011–September 2013 is used in the present calibration and sensitivity analysis, Fig. 3A. PET is calculated with the Penman formula (Penman, 1948) based on local data from the AWS (Johansson et al., 2015).

Precipitation measured in Kangerlussuaq during the period 1977–2013 is analyzed with regard to total annual *P*. The years falling in the range of the lower 20th percentile and the upper 80th percentile are defined as dry and wet years, respectively. An annual *P* below 134 mm is accordingly defined as dry, and annual *P* above 220 mm is defined as a wet year. The median value for the time period 1977–2013 is 159 mm. In the time series from Kangerlussuaq, the year 2011 is in the lower 20th percentile, 2012 is in the upper 80th percentile, and 2013 is close to the median value. Accordingly, 2011 is selected as representative of a dry year, 2012 as a wet year and 2013 as a normal year. Hence, by analyzing time series from Kangerlussuaq it could be concluded that the short measurement period at TBL covers both a wet year (2012), a dry (2011) year and a normal year (2013), and the MIKE SHE model representing these types of years can be run by use of only locally weather data. Time series of *T*, *P* and PET from TBL are used as driving data for each year in order to produce long-term wet, dry and normal hydrological conditions of the catchment, with the aim to analyze the hydrological and hydrogeological responses to various meteorological driving conditions.

2.5. Hydrological data

The lake level has never exceeded its threshold level for surface water outflow since the measurements started in September 2010; the last time an outflow was observed was in 2009. Isotopic data of groundwater, surface water and precipitation indicate that surface water outflow and/or near-surface outflow via the active layer occurs on a regular basis. Groundwater wells (Fig. 2) situated downstream of TBL have an ^{18}O-signature similar to that in the lake water, whereas water from groundwater wells upstream of the lake has a precipitation-like signature (Fig. 4A). In addition, electrical conductivity values from the TBL catchment span the interval 120–145 µS/cm; lakes with long-term negative water balance in the area exhibit conductivity values as high as 3000 µS/cm (Aebly and Fritz, 2009).

In general, there is a total annual variation in lake level of ~30 cm (Johansson et al., 2015). The strongest transients occur in summer due to evapotranspiration processes and in the initial phase of the active period in the spring due to water generated by snowmelt (Fig. 4B and C). Monitoring data show a time-lag between increased snowmelt and lake level, because the main part of the melt water is trapped as depression storage and forms small ponds in the catchment, Fig. 4C. The ponds re-freeze as the temperature drops during certain periods in March and April. As the uppermost soil layer thaws, the retaining effect of the ponds disappears and the water can flow to the lake.

Hydraulic head in DH-GAP01 is measured at a vertical depth of 128 m relative to the pressure transducer in TBL. The mean difference in hydraulic head between the lake and the observation point in the bedrock during the period 2010–2013 is calculated to be 2.15 m with a consistently greater hydraulic head in the lake (Fig. 4D); this implies a downward mean hydraulic gradient of 0.017, i.e., recharge from the lake to the talik. A hydraulic test was performed in DH-GAP01 in the summer of 2010 and the hydraulic conductivity of the bedrock below the lake is calculated to be 2–5 $*$ 10^{-8} m/s with the interval representing the different methods used to evaluate the test (Harper et al., 2011), with most of the hydraulic capacity attributed to the fractures in the rock.

Soil moisture is monitored using the Time Domain Reflectory (TDR) technique in 43 points distributed over the whole depth of the active layer in a set of vertical profiles organized as three clusters placed in relatively close vicinity to the lake (Fig. 2, Johansson et al., 2015). The data have been used both to define the frozen and active periods, and to describe storage changes in the active layer. The uppermost soil layer (5–10 cm) has a very high porosity and accordingly a high saturated water content. The soil moisture response to precipitation is fast in the uppermost layers whereas a delay has been noticed as depth increases (Johansson et al., 2015).

3. Conceptual and numerical models of the TBL hydrology

3.1. Conceptual model and water balance calculation

During the active period, three runoff components are defined (Fig. 5A): surface water flow to and from the lake (R_s), and groundwater exchange between the lake and the active layer (R_{al}) and via the talik (R_{gw}). Precipitation (P) that falls as rain is assumed to be uniformly distributed in the catchment during the active period and equal to that measured at the AWS (after correction for wind losses. The lake evaporation (E) is assumed to be equal to the calculated PET and uniformly distributed over the lake, whereas the terrestrial evapotranspiration (ET) is dependent on soil moisture, vegetation and soil type. Two storage components are identified by measurements: storage change in the lake (ΔH) and in the

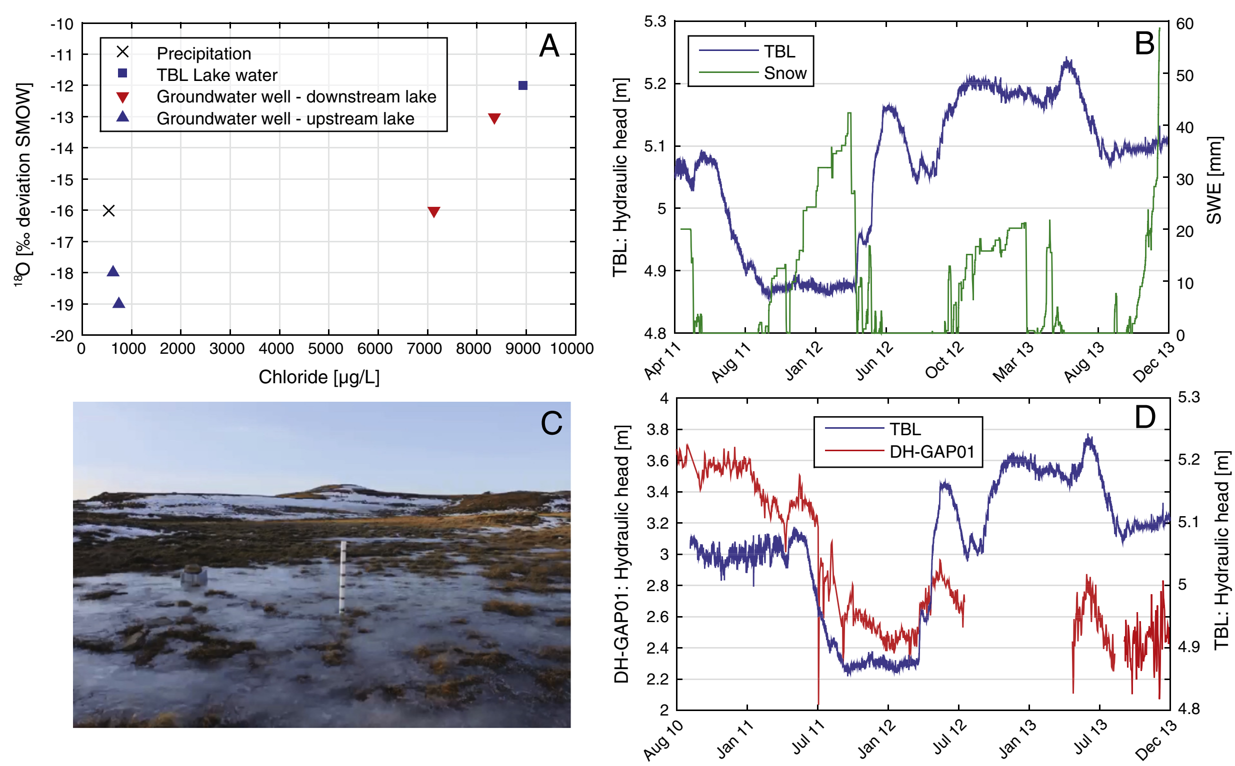

Fig. 4. ^{18}O – data from precipitation, lake water and groundwater wells upstream and downstream the lake plotted against chloride concentration (A). The response in lake level to snowmelt is illustrated in (B), and (C) shows a photo of ponded water in low topographical points after snowmelt. Hydraulic head in TBL and in the talik beneath the lake measured in borehole DH-GAP01 (D).

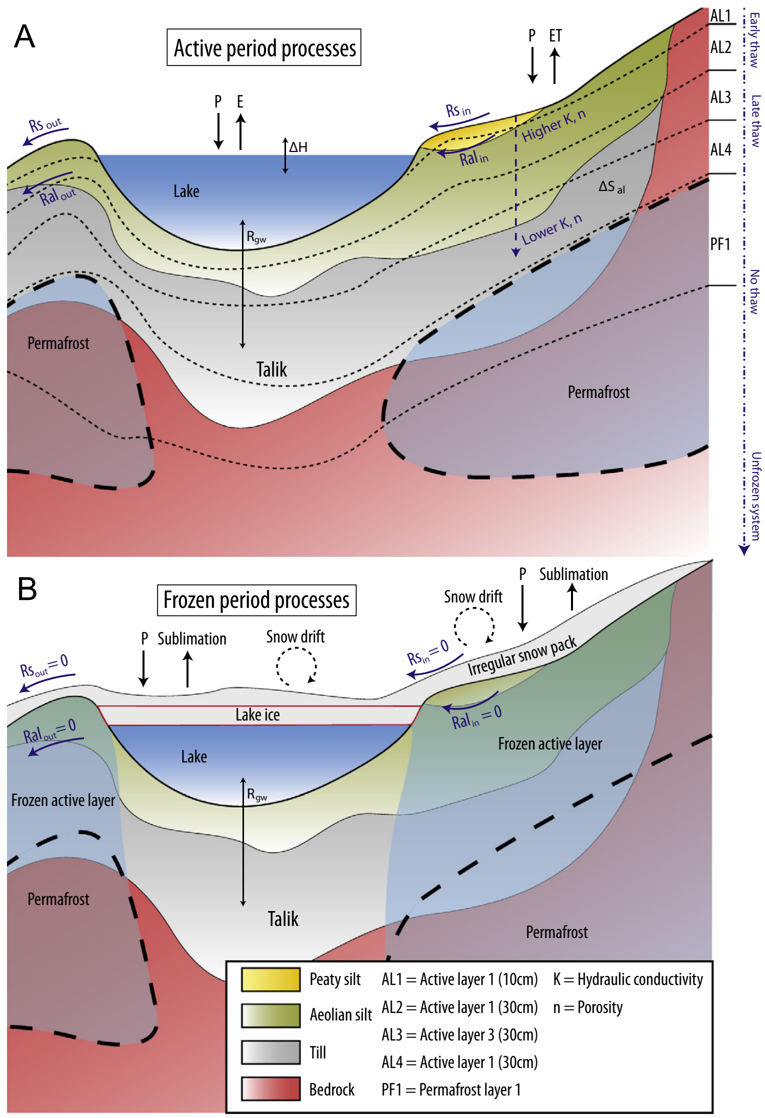

Fig. 5. Conceptual model of the hydrological flows during the active (A) and frozen (B) period in the Two Boat Lake catchment. The conditions for precipitation (P), evaporation (E) and evapotranspiration (ET) as well as runoff from active layer (R_{al}), water exchange with the talik (R_{gw}), and surface water runoff (R_s) are illustrated. The arrows illustrate the main hydrological flows for each period. The figure is not to scale; in reality, the lake is much deeper than the active layer and the lake ice reaches a depth well below the maximum depth of the active layer. A schematic illustration of the upper 5 calculation layers of the numerical model (AL1-4, PF1) is added to the conceptual model in Figure A.

active layer (ΔS_{al}). For the relatively short period in time when local data is available, the storage change in the permafrost is assumed to be zero (Fig. 5A).

In the frozen period, R_s and R_{al} are zero (Fig. 5B). Due to the presence of the through talik, R_{gw} is non-zero also in the frozen period. Snowdrift, resulting in snow re-distribution, and mass loss via snow sublimation influence the lake water balance (Fig. 5B). Losses and gains of water due to snow drift over the catchment boundary are assumed to cancel out.

The fluxes defined in Fig. 5, together with data on P, PET, ΔS_{al}, ΔH and R_{gw} (Johansson et al., 2015) allow for a simple calculation of the annual water balance by applying Eqs. (1) and (2):

$$\text{Lake area}: \; P - E + (Rs_{in} + Ral_{in}) - R_{gw} - (Rs_{out} + Ral_{out}) = \Delta H \tag{1}$$

$$\text{Terrestrial catchment area}: \; P - \text{ET} - (Ral_{in} + Rs_{in}) = \Delta S_{al} \tag{2}$$

First, the water balance for the lake is calculated (Eq. (1)) and in the second step the water balance for the terrestrial areas is calculated (Eq. (2)). The residual in (Eq. (1)), i.e. the inflow to the lake (Rs_{in} + Ral_{in}), links the two equations. The calculated inflow from (Eq. (1)) is used in (Eq. (2)) to calculate the residual of the water balance of the terrestrial catchment area, interpreted as the total evapotranspiration, ET, required to close the catchment water balance. In Eq. (2) data from the TDR is used to make an approximation of the water storage change of the active layer. The results of the calculations, Table 3, show which processes dominate the water balance.

3.2. Numerical model

In order to evaluate the long-term water balance and to analyze spatial and intra-annual variations of hydrological fluxes in the TBL catchment, a numerical model was established based on the conceptual model of frozen and active periods. The numerical model also takes the transients of thawing and freezing into account (Fig. 3C).

The present hydrological simulations were carried out with the physical, distributed three-dimensional coupled model MIKE SHE-MIKE 11 (Graham and Butts, 2005; Refsgaard et al., 2010).

Table 3
Annual water balance based on data from Johansson et al. (2015) and the conceptual model presented in Fig. 5. The water balance components are given in m^3 and in mm normalized over the area for which the water balance was calculated (i.e. the lake or the terrestrial area).

	mm	m^3
Lake area		
P	281	106,000
E (=PET)	446	169,000
(Rs_{out} + Ral_{out})	0	0
R_{gw}	15	6000
ΔH	94	36,000
Residual: (Rs_{in} + Ral_{in})	274	105,000
Terrestrial area		
P	281	332,000
Ral_{in} + Rs_{in} (from Eq. (1))	88	104,000
ΔS_{al}	14	17,000
Residual: ET	179	211,000

The model domain (Fig. 2), with a horizontal resolution of 10 × 10 m, covers the whole catchment area as well as downstream areas in order to avoid boundary effects. The active layer (Fig. 5A) and the permafrost are represented by four layers each, with the bottom boundary in the lowest layer set at 200 m depth. A head boundary condition is applied in the lowest calculation layer (150–200 m depth) using the hydraulic head measured in DH-GAP01 (Fig. 4D). The upper boundary conditions are applied using P and PET data, and a no-flow boundary condition is applied at the lateral boundaries. Depressions formed by ice wedges, which are reflected as areas with a high TWI are the major flow paths of surface water from the catchment to the lake (Fig. 2). The small streams are represented as a river network in the model.

Since MIKE SHE does not support thermal modeling, dynamics in the active layer and the permafrost are simulated using the modeling approach of Bosson et al. (2010, 2012). In previous studies with MIKE SHE, the use of time-varying hydraulic properties was not implemented in the code. Bosson et al. (2010, 2012) divided the year into frozen and active periods, and each period was calculated separately. However, in the present study the hydraulic properties are allowed to change on a daily basis and annual cycles are simulated in the same model. Time-varying properties, with daily resolution, of hydraulic conductivities, surface roughness and infiltration capacity are used to mimic frozen and thawed conditions in the active layer. The permafrost is implemented as layers with very low hydraulic conductivity. The area under the lake, i.e., the talik, is continuously unfrozen, enabling a contact between the lake and the unfrozen groundwater system below the permafrost throughout the year (Fig. 5A). The methodology for permafrost modeling in MIKE SHE is further described in Appendix A.

K-values are applied to each type of soil in the different layers (Table 4) and the changes in hydraulic properties follow the scheme of active and frozen periods presented in Fig. 3C and Table 2. Based on hydraulic tests in the borehole below the lake

Table 5
Maximum LAI-values for each vegetation class. The LAI values are time varying and reduced to zero during winter.

Vegetation class	Max LAI
Betula	1.06
Wetland	0.76
Barren	0
Salix	0.99
Heath	0.55
Grass Exposed	0.55

(Section 2.5), the K of the bedrock in the talik is set to $2 * 10^{-8}$ m/s. This value, which is the only available data on K for the bedrock, is influenced by a fracture zone below the lake. Therefore, the hydraulic conductivity of the bedrock on the terrestrial parts of the catchment is set to a lower value, $1 * 10^{-9}$ m/s (Table 4). Leaf area index (LAI) and root depth are required in the evapotranspiration calculations in MIKE SHE. Time varying LAI is assigned to the different prevailing vegetation types (Table 5) in accordance with Shaver et al. (2007). The root depth is set to 30 cm for all vegetation classes based on from vegetation investigations and excavations of the active layer in the TBL area.

First, the model is calibrated for the period with available site data: 2011–2013. Model sensitivity analysis is carried out with regard to hydraulic properties of the talik and the active layer and to seasonal dynamics in the active layer and on the ground surface. Based on the calibrated model, the normal, dry and wet years (Section 2.4) are simulated. Each of the dry, wet and normal years are then cycled for 10–15 years, until inter-annual steady-state conditions are reached. In addition, a 37-year simulation is performed based on climate data from Kangerlussuaq. The simulations representing dry, wet and normal years are driven by local data from TBL only, whereas the 37-year simulation is driven by data from the nearby station in Kangerlussuaq and contains a natural sequence of different meteorological conditions for a longer period.

4. Results

4.1. Modeled water balance and comparison with observations

4.1.1. Simulation of the period with available field data

The MIKE SHE model was run for the period March 1st 2011–September 1st 2013, with a sensitivity analysis carried out for the K-values, the bottom boundary condition, and the thawing and freezing dynamics, with the aim to get as good a match as possible between measured and calculated lake levels. Based on the results of the sensitivity analysis with different K-values in the soil and the talik bedrock, the applied K-values presented in Table 4 were not changed in the model (Fig. 6), and the measured groundwater head from DH-GAP01 was applied as the bottom boundary condition in the final calibrated model. To reproduce the right lake response to snow melt and thawing of the active layer, the soil

Table 4
Horizontal (K_h) and vertical (K_v) hydraulic conductivity of different types of soil based on the soil map presented in Clarhäll (2011).

	10[a] cm	25[b] cm		50[b] cm		75[b] cm	
	K_v = K_h (m/s)	K_h (m/s)	K_v (m/s)	K_h (m/s)	K_v (m/s)	K_h (m/s)	K_v (m/s)
Bedrock, terrestrial parts	1.00E−09	1.00E−09	1.00E−09	1.00E−09	1.00E−09	1.00E−09	1.00E−09
Aeolian silt-fine sand	2.60E−04	2.20E−05	2.20E−06	2.20E−05	2.20E−06	2.20E−05	2.20E−06
Till	3.60E−04	4.76E−04	4.76E−05	4.76E−04	4.76E−05	4.76E−04	4.76E−05
Peaty silt	2.20E−04	0.00022[a]	2.20E−05	–	–	–	–

[a] Data from infiltration capacity measurements Johansson et al. (2015).
[b] Data from grain size distributions Johansson et al. (2015).

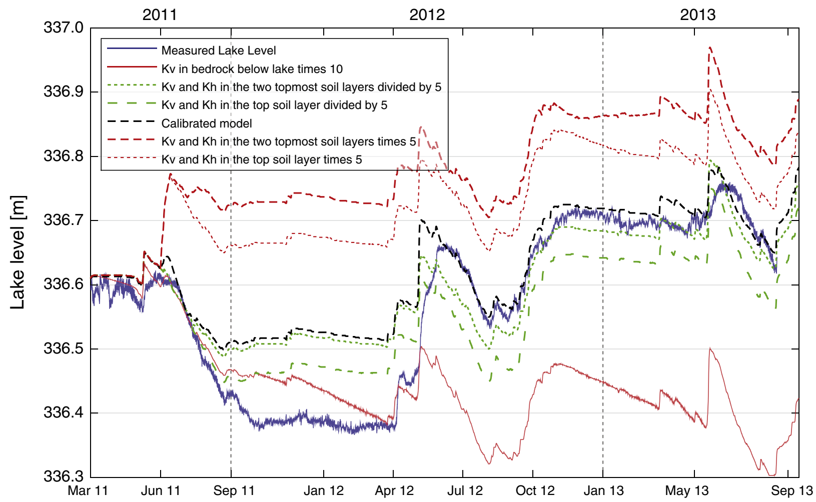

Fig. 6. Measured and calculated lake level of the TBL for the period March 2011–September 2013. Results from the sensitivity analysis of *K*-values in the bedrock below the lake and in the active layer are also illustrated.

Table 6
Annual water balance from the MIKE SHE model of TBL. The annual mean values are based on modeling results for the period September 1st 2011–August 31st 2013. The water balance is calculated for the total catchment area, the lake area and the terrestrial part of the catchment. All values are given in mm and are normalized over the respective areas (catchment, lake, land).

	mm/year
Whole catchment area (P-ET-R_{out} = ΔS)[a]	
P	281
ET	250
ΔS	27
R	4
R_{gw}	4
R_{out}	0
Lake area (Eq. (1))	
P	281
E	442
(Rs_{in} + Ral_{in})	276
(Rs_{out} + Ral_{out})	0
R_{gw}	16
ΔH	99
Terrestrial area (Eq. (2))	
P	281
ET	192
Ral_{in} + Rs_{in}	79
ΔS_{al}	10

[a] R_{out} = Rs_{out} + Ral_{out}, ΔS = ΔH + ΔS_{al}.

temperature at a depth of 5 cm, rather than air temperature, was set to control the surface roughness (Manning number). No changes were made to the parameters controlling infiltration capacity.

A mean annual water balance for the whole catchment calculated for the two hydrological years September 1st 2011–August 31st 2013 (Table 6) implies a mean annual P of 281 mm per year and a mean ET of 250 mm per year. The total annual storage change in the lake and in the active layer, ΔS, is 27 mm per year, and the recharge to the talik, R_{gw}, is 4 mm per year. Considering only the terrestrial part of the catchment, ET is 192 mm per year, the discharge to the lake (Rs_{in} + Ral_{in}) is 79 mm per year and ΔS_{al} is 10 mm per year (notation explained in Fig. 5). The residual components of the water balance, obtained by analyzing the time series data (Johansson et al., 2015) and presented in Table 3, are then of the same order of magnitude as the simulated values (Table 6).

The transient behavior of the lake level is mostly well captured by the calibrated model (Fig. 6). For the period April 2012–September 2013, lake-level increases and decreases are primarily explained by the modeled snow melting, thawing of the active layer and summer evapotranspiration, respectively. The mean error between modeled and observed lake level for this period is less than 0.05 m. However, more significant discrepancy between the calculated and the measured drawdown of lake level is seen for the summer of 2011, leading to lake-level errors in the range of 0.10–0.15 m for the fall and winter of 2011.

In contrast to the summer periods 2012 and 2013, the measured lake drawdown in summer 2011 is thus not fully explained by the corresponding modeled losses of water by evaporation and groundwater recharge to the talik. For the possibility that the latter, R_{gw}, might be underestimated in the model, a sensitivity analysis carried out of the hydraulic conductivity of the bedrock below the lake shows that increased R_{gw} does not resolve the lake-level problem. Specifically, increasing R_{gw} by increasing the vertical conductivity in the bedrock causes unrealistic lake drawdown, compared to data, in all winter periods (Fig. 6).

However, sensitivity analysis with regard to *K*-values in the active layer does show a particularly large effect for the first year of simulation, 2011 (Fig. 6). After May 2012, temporal differences are relatively small among the tested cases in this sensitivity analysis. A main reason for these results is that the snowmelt period was less intensive and the summer was much dryer in 2011 than in 2012 and 2013. This makes the modeling of 2011 more sensitive to applied hydraulic properties and initial conditions in the active layer. This particular sensitivity implies that this first year of simulation should be considered as an initialization phase for the modeling, for which somewhat less consistency may be acceptable between measured and calculated lake levels than in the following years. Since the base set-up of hydraulic properties in the model (Table 4) provides the best match between measured and calculated lake levels for the period May 2012–September 2013, model *K*-values were not changed in order to improve model fit to the 2011 data.

With regard to possible model-fit improvement, it is noted that the observed lake level in September 2011 can be well reproduced in the MIKE SHE modeling by applying a low initial groundwater table (no results from this scenario simulation are shown in Fig. 6). However, by doing this, the high lake level in April 2012 is then not as well reproduced by the model. With more than 100 observed snow-drift events occurring during the winter 2011–2012 (Bosson et al., 2013b), such temporary model discrepancy for lake level may be due to temporary snow addition to the catchment precipitation, which is not accounted for in the model. In addition, measurement error contributions to lake-level results are also possible, in particular due to undercatch of winter precipitation under windy conditions (i.e., too small correction of measured data) and/or shorter periods of potential drifting of the pressure transducer (the transducer was maintained and calibrated twice a year). Over time, however, precipitation gains and losses of snow due to wind drift tend to balance each other out and the pressure transducer was regularly calibrated, leading to the overall good model results for the lake level.

4.1.2. Long term water balance based on data from Kangerlussuaq

To address and identify the main uncertainties in the calculated water balance conditions for the relatively short calibration period, water balances for long-term normal, wet and dry periods were studied based on meteorological data from Kangerlussuaq (Section 2.4). It is considered unlikely that the TBL catchment should be exposed to extreme weather conditions during several consecutive years. However, the three tested normal, wet and dry year cases do represent a variety of water balance components under a relatively wide range of meteorological driving conditions (Fig. 7).

The annual precipitation in the TBL catchment is 163–366 mm per year and ET is 202–263 mm per year. The groundwater recharge R_{gw} ranges from 4 mm per year during wet and normal conditions to – 2 mm per year (discharge) under dry conditions; this indicates that the gradient between lake and bedrock changes direction when the catchment is exposed to "extremely" dry climate conditions. The water exchange with the underlying talik is around 4% of the net outflow from the lake catchment under wet conditions and 57% under normal climate conditions. However, R_{gw} is low compared to other fluxes in the catchment. The evapotranspiration ET is the dominating flux in the water balance in all cases, and higher than the annual P for dry climate conditions lead to a negative water balance and a decreased lake level (Fig. 7A and B).

The catchment has a positive water balance during both wet and normal years, with a net outflow of water over the catchment boundary. The ET from the terrestrial parts ranges over 138–199 mm per year and the inflow to the lake is in the range 25–159 mm per year (normalized over the terrestrial catchment area). Hence, an inflow of surface water and groundwater via the active layer to the lake occurs in all studied climate cases. The calculated lake level for wet and normal periods exceeds the

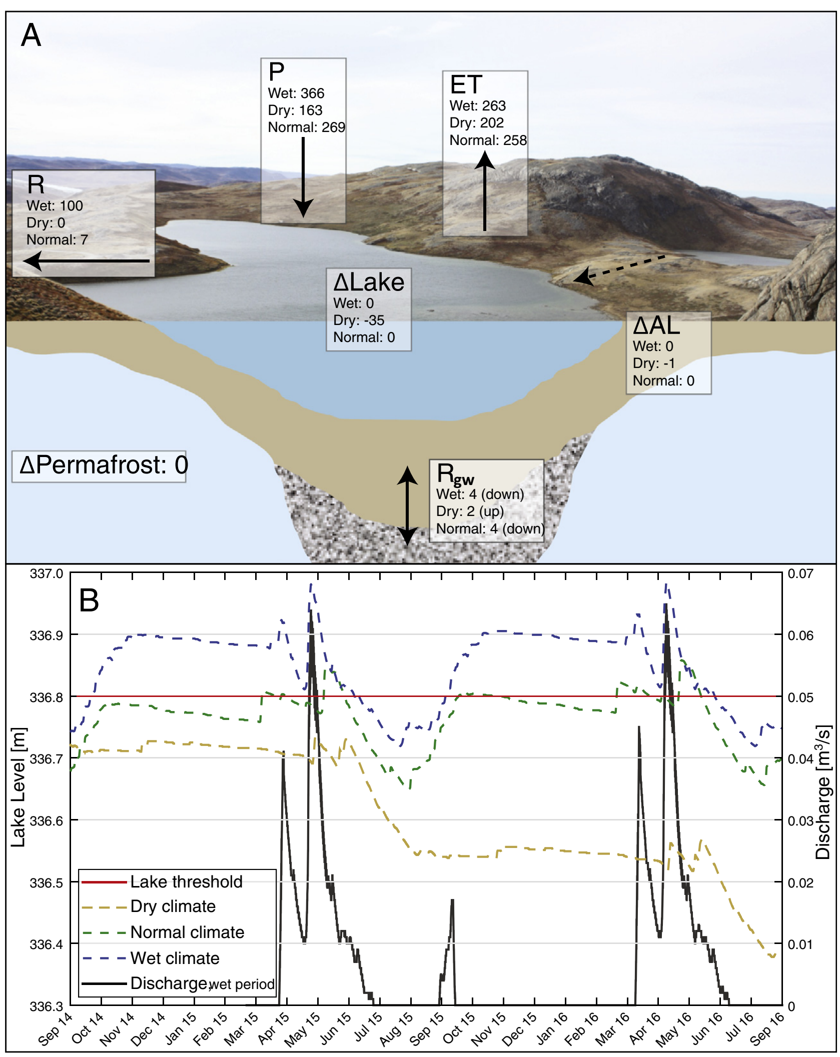

Fig. 7. Water balance for the simulated wet, dry and normal years (A). The water balance components total runoff (R), precipitation (P), total evapotranspiration (ET), water exchange with the bedrock in the talik (R_{gw}) and the storage components in the active layer (ΔAL), the lake (ΔLake) and in the permafrost (ΔPermafrost) are marked in the figure. Lake level variation during an annual cycle for dry, wet and normal weather conditions plotted together with the surface water outflow from the lake during the wet year (B).

lake threshold each year in the beginning of the active period as well as during the autumn rains (Fig. 7B); only the surface water outflow under wet conditions is illustrated in Fig. 7B.

With the aim to study if and when a surface water outflow may occur from TBL, a 37-year simulation was carried out based on climate data from Kangerlussuaq. The time series on P, T and PET, with PET calculated from T with the Thornthwaite method (Thornthwaite, 1948), are then used to drive the MIKE SHE model in two ways. One simulation uses driving weather data from the original time series from Kangerlussuaq, and one simulation uses corrected precipitation data from Kangerlussuaq in order to represent the local P at TBL. That is, the precipitation time series from Kangerlussuaq was in the latter case multiplied by a factor of 1.8 (Section 2.4). The lake level never exceeds its threshold for surface water outflow in the simulation with the original P, whereas a surface water outflow occurs from the lake for approximately 70% of the simulated years with the corrected (higher) P, see figure in Appendix A. The latest observed surface outflow in 2009 is reproduced in the model. The model results confirm field observations of no surface water outflow for the period September 2010–December 2013. The observed lake shore line on 336.4 m.a.s.l during LiDAR measurements in September 2011, on which the DEM is based, is also reproduced in the long term simulation.

4.2. Spatial and temporal variations of water balance components

4.2.1. Spatial variations

The spatial variations of hydrological flows within the catchment are evaluated by use of the calibrated model. The areas where surface water is concentrated and further transported to the lake are found in small valleys within the catchment. The spatial distribution of ponded water on the ground surface immediately after snowmelt (Fig. 8A) shows that areas with high TWI (Fig. 2) coincide with areas of large depth of ponded water and

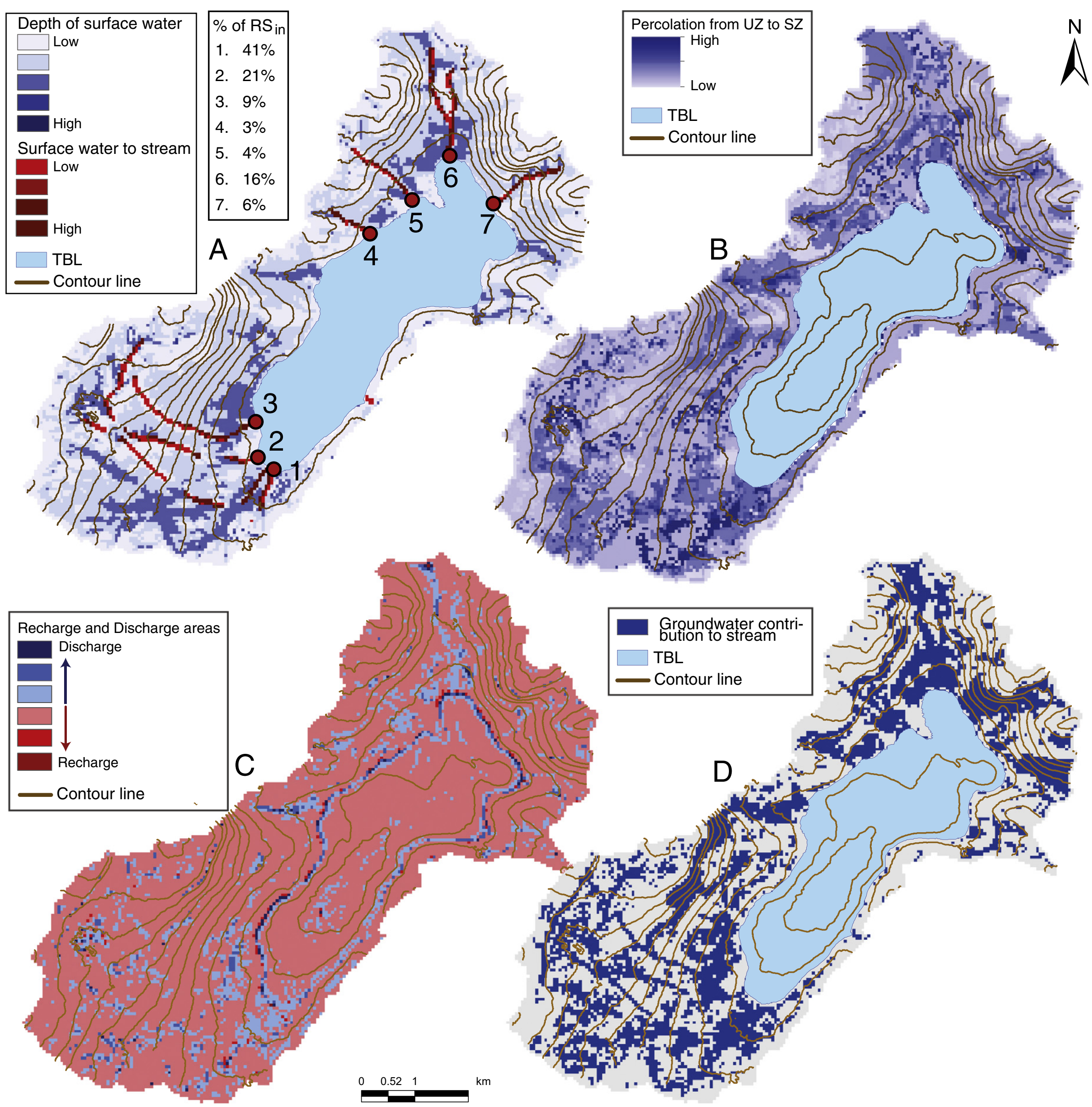

Fig. 8. The depth of ponded water after snowmelt and the areas where a surface water exchange with the streams occurs (A) and the simultaneous percolation from the unsaturated zone (UZ) to the saturated zone (SZ) (B). The contribution to total Rs_{in} from the largest sub-catchments 1–7 are listed in A. The mean annual groundwater recharge and discharge fluxes in the upper part of the active layer are shown in (C). In (D) the areas where groundwater transport to the streams occurs are illustrated. Note that these in general coincide with the groundwater discharge areas shown in (C).

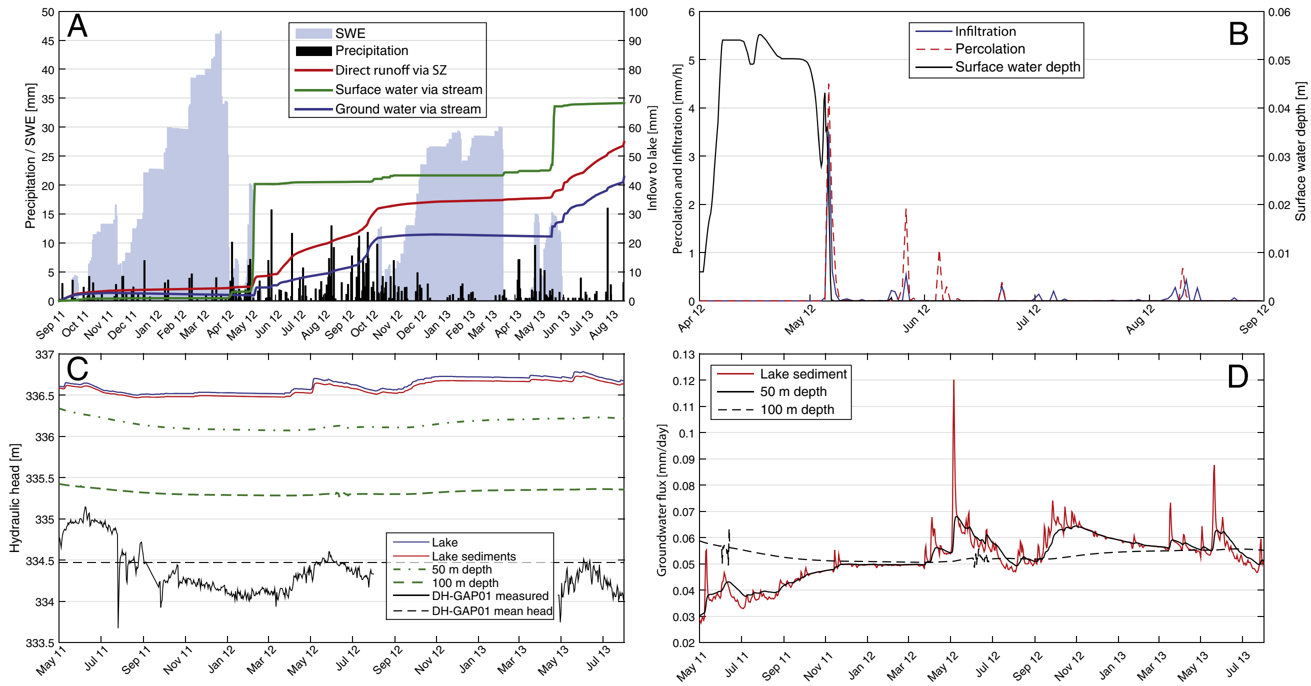

Fig. 9. Calculated snow water equivalents (SWE) co-plotted with measured precipitation (*P*) and calculated inflow to the lake divided into the three components surface water via streams, groundwater via streams and direct runoff from the active layer (A). The ponding of water due to snowmelt and the subsequent infiltration and recharge processes are illustrated in (B), the calculated hydraulic head in (C), and the vertical groundwater flux in the lake, the lake sediments and at different depths in the bedrock below the lake in (D).

areas where surface water flows into the streams. The greatest contributions of surface water inflow to the lake occur in the northern and southern valleys of the catchment: points 1, 2 and 6 in Fig. 8A, which are also characterized by high TWI (Fig. 2). The spatial pattern of infiltration to the unsaturated zone and percolation of water from the unsaturated to the saturated zone (Fig. 8B) follows the depth of ponded water, i.e., infiltration and percolation are high in areas with large surface water depth.

The annual mean distribution of groundwater recharge and discharge areas in the active layer shows the main discharge areas to be close to the lake and in the main valleys of the catchment (Fig. 8C). The shoreline is also a strong groundwater discharge area, i.e., there is seepage of groundwater from the active layer to the lake during the active period. Groundwater discharge areas, Fig. 8C, prevail in most of the areas where there is a transport of groundwater from the active layer toward the streams, as indicated by the dark blue areas in Fig. 8D. Runoff on ground surface is limited in the catchment, ponded water infiltrates the active layer (Fig. 8A and B) and a groundwater flow in the active layer toward the streams occurs (Fig. 8D). The main part of the lake acts as a recharge area in the upper part of the lake sediments (Fig. 8C), with only small areas in the deepest part of the lake representing the annual mean discharge. At greater depth, in the bedrock under the lake sediments, the whole lake acts as a recharge area, which means that the lake recharges the talik beneath it.

4.2.2. Intra-annual temporal variations

The ground surface in the TBL catchment area is generally quite dry. There were several snowmelt events every season (Fig. 9A), each followed by a period of ponded water, increased infiltration and recharge. However, a time lag is noticed between ponding and subsequent infiltration and groundwater recharge. An example is given in Fig. 9B where the interaction between ponding, infiltration and recharge close to outlet number 2 (Fig. 8A) is shown. Water is ponding on the ground surface directly after snowmelt when the ground is still frozen and no water can infiltrate. Once the infiltration starts, due to thawing of the upper part of the active layer, the ground is quickly saturated and water is transported from the unsaturated to the saturated zone and, accordingly, a quick decrease of ponded water occurs. As the thawing proceeds deeper into the active layer in June (Fig. 9B), more water can recharge the groundwater. The relatively small summer rains in July initiate an infiltration process, but the infiltrated amount of water is not large enough to percolate to the groundwater (Fig. 9B).

The main inflow of water to the lake occurs in the beginning of the active period. The first snowmelt events each year (Fig. 9A) do not cause inflows to the lake. Instead, the melt water is trapped in small pits in the frozen ground and the water refreezes as the temperature drops below zero in early spring. This process is also verified by the analysis of time lapse photos and by the measured lake level (Fig. 4). The last snowmelts in the beginning of May 2012 and mid May 2013 coincide in time with the thawing of the ground surface and the upper part of the active layer, which is when flow of water to the lake is being induced.

Approximately 50% of the annual runoff to the lake occurs during the initial phase of the active period, irrespective of studied year, and surface water is then the dominating contribution. The groundwater contribution to streams and the direct groundwater inflow to the lake from the active layer increases as the thawing of the active layer proceeds (Fig. 9A). The main part of the groundwater flowing into the lake is water from a thawing active layer, whereas the additional input of water due to rain is relatively small.

The interaction between the lake and the groundwater in the talik is continuously dominated by recharge from the lake to the talik (Fig. 9D). The hydraulic head in the lake sediments reflects the surface water level of the lake, which in turn is influenced by surface and near-surface processes, such as inflow to the lake and evapotranspiration. Deeper into the bedrock the transient behavior is smaller. The calculated hydraulic heads in the lake sediments and in the bedrock at 50 and 100 m depth are not sensitive to the transients in the bottom boundary condition. The same results are obtained for the time-varying measured head in DH-GAP01 as for the mean value of the same time series assigned as bottom boundary condition (Fig. 9C). The recharge from the lake to the talik is highest during the beginning of the active period (Fig. 9D), i.e. when the main part of the annual inflow to the lake occurs.

5. Discussion

5.1. Water balance of the TBL catchment

The MIKE SHE model together with long-term weather data from Kangerlussuaq made it possible to study typical water balance characteristics for a range of weather conditions and for the relatively short period of available local data. The calibrated MIKE SHE model shows good agreement with the water balance presented in Table 3 and with measured lake and groundwater levels, which was the basis for using the model to represent catchment hydrology under different weather conditions. A positive water balance is maintained in the catchment under both wet and normal conditions, as also indicated by electrical conductivity data (Section 2.5). The results from the 37-year simulation based on meteorological data from Kangerlussuaq indicate that surface water outflow occurs on a regular basis, which is also supported by isotopic data from the catchment (Section 2.5). During the period 2000–2008 the mean annual *P* in Kangerlussuaq was 205 mm, soundly above the long-term mean (Section 2.4). This indicates favorable conditions for surface water outflow in 2009, which indeed was the last time a surface water outflow was observed. The period 2009–2013 is characterized by dry years, which explains the absence of surface water outflow since the TBL site investigation started in 2010.

The results from the 37-year simulation also accentuate the importance of local weather data. The simulation with *P* from Kangerlussuaq, which is situated 30 km from TBL, yields a continuously negative water balance for the TBL catchment, whereas *P* corrected for local TBL conditions yields surface water outflow in 70% of the simulated years.

5.2. Limitations of the MIKE SHE model

The absence of a thermal component in the MIKE SHE modeling system is solved by applying time-varying properties of all major hydrogeological elements influenced by freezing/thawing (Bosson et al., 2010, 2012). By using data for soil temperature and by temporally varying the hydraulic properties, the transients in lake level and terrestrial inflows to the lake, as well as freeze and thaw dynamics in the active layer, are possible to mimic. The simulated lake level is generally somewhat higher than that measured during the frozen period. This might be due to the assumption of uniformly distributed snow cover over the catchment area and/or an underestimated PET during the frozen period, caused by underestimated sublimation.

Snow sublimation has been identified as an important process in cold region hydrology, especially during snow-blowing events when the snow is more exposed to sublimation forces (Pomeroy and Gray, 1995; Pomeroy et al., 1997). The analysis of time-lapse photos from the TBL area confirms both an irregular snow distribution and several snow-blowing events each winter (Johansson et al., 2015). Sublimation measurements in the TBL catchment in April 2013 yielded an average value of 0.63 mm per day (SWE). The conditions for sublimation were favorable during the observation period and this result should be interpreted as a maximum sublimation rate for the frozen period. However, the high mass deficit implied by available measured data emphasizes the importance of quantifying sublimation processes for the water balance of high Arctic catchments. Gauge undercatch and snow blowing events might contribute with a higher amount of snow to the catchment than what is measured by the AWS. Model account of snow additions to the catchment during such snow blowing events would require manual snow addition to the measured precipitation as input data. For the purposes of the present study, it was decided not to further manipulate the input data time series in order to reach a perfect fit between measured and observed lake levels.

It is possible to apply time-varying, spatially distributed PET and *P* in MIKE SHE, but this time variation has to be calculated outside the MIKE SHE tool. In future studies of periglacial areas with MIKE SHE, an external calculation of snow distribution, snow blowing events and sublimation could be motivated, such that these types of data become available and could be imported to the model as a time and space varying input if needed.

5.3. Hydraulic interactions between active layer, lake and underlying bedrock

The lake level dynamics, which are mainly governed by surface and near-surface processes, are well reproduced by the MIKE SHE model. The time-lag between snowmelt and inflow to the lake, observed by photos and lake level measurements, is also reflected in the modeling results. Irrespective of the studied year, runoff related to snowmelt is the main contribution to the annual inflow to the lake and is dominated by surface water flow, as was also concluded in previous studies of water balance in periglacial areas (McCann and Cogley, 1972). The groundwater contribution to the lake increases continuously during thawing of the active layer (as described by Woo and Marsh, 2005) and is sensitive to temperature and precipitation. A warm and wet summer period, such as that in 2012, increases the contribution to runoff from the thawing active layer.

Three different bottom boundary conditions have been tested in the MIKE SHE model: a no-flow boundary, a time varying head boundary condition with measured hydraulic head from DH-GAP01, and a constant head boundary representing the average value of the measured hydraulic head in DH-GAP01 for the simulated period. The lake level variation is almost independent of this applied bottom boundary condition, and reflects instead near-surface processes like evapotranspiration and freeze and thaw dynamics in the active layer. With a no-flow boundary condition, an upward gradient between the bedrock and lake is developed, contradicting field observations and the conceptual model. When applying a head boundary condition, based on measured data from DH-GAP01, a net recharge of water occurs from the lake to the talik. The magnitude of the groundwater flux and flow direction between lake and talik is in this case independent of time variation in the applied head boundary condition.

The fact that the same lake level dynamics is obtained irrespective of whether temporal variability in the bottom boundary condition is considered, in combination with the good agreement between calculated and measured lake levels, indicates that the lake is independent of catchment-external influences from the unfrozen system below the permafrost. Whether the co-variation of the hydraulic head in the lake and that in the borehole below the lake is indicative of a good hydraulic contact between the lake and the bedrock is a question requiring further investigation by larger-scale hydrogeological modeling that also includes ice sheet effects. Most likely, the groundwater dynamics in the talik are influenced both by the lake and by the large-scale groundwater system below the permafrost.

The recharge/discharge conditions between the lake and the underlying talik are likely changing in time, implying that water flow may switch between recharge and discharge at the lake-talik boundary. This is shown by the simulation results for the dry year conditions, with the lake level dropping so that the gradient between the lake and the talik shifts direction. It should be noted that groundwater pressure is only monitored in one point below the lake.

A previous model study by Bosson et al. (2010) has also shown that a talik can act both as a recharge and a discharge area. The TBL catchment is located at high altitude compared to other possible through taliks further downstream toward the fjord in Kangerlussuaq (Fig. 2), which further strengthens the possibility of the talik below TBL acting mostly as a recharging talik. Altitude differences between taliks are important for their respective recharging or discharging conditions, as also shown in previous simulation studies (Bosson et al., 2013a).

6. Conclusions

Annual water balance conditions and their spatiotemporal variability have here been quantified for the catchment of Two Boat Lake, Western Greenland. This has been achieved by site conceptualization and numerical analysis of the hydrological multi-parameter dataset presented in Johansson et al. (2015). Despite the absence of a thermal component, the developed MIKE SHE numerical model enabled identification and quantification of main hydrological flows, as well as mean water balance conditions and their possible long-term variations in the TBL catchment. The hydrological modeling approach used by Bosson et al. (2010, 2012), based on simulated climate time series for different scenarios of a colder far-future Sweden, has here been extended and tested with site-specific data for an Arctic catchment with present-day periglacial conditions. The applied model is able to reproduce measured lake and groundwater levels, as well as observations made by time-lapse cameras in the area.

The work highlights the importance of numerical modeling that takes into account and combines evapotranspiration and other surface and subsurface hydrological processes in order to quantitatively represent the dynamics and complexity of the interactions between meteorology, active layer hydrology, the lake and the unfrozen groundwater below permafrost. Simple hydrological models driven by topography can reflect the spatial distribution of surface waters and groundwater in the active layer, but advancement of the understanding of spatiotemporal variability and changes in hydrological processes within a permafrost catchment requires use of spatially distributed, transient and physically-based modeling.

Based on hydrological measurements that also reflect the links between the surface system and the subsurface water system below the permafrost, it is concludes here that water flow between the studied lake and its underlying talik is small relative to the other water balance components. The long-term simulations show that recharge and discharge conditions in the talik may shift in time. Temporal pressure changes in the talik reflect the measured and simulated local hydrological conditions at the lake surface, which could also be influenced by larger-scale regional processes in an ice-sheet dominated landscape. The present results indicate that the lake and the hydrology of the active layer in the TBL catchment are independent of catchment-external landscape features, such as the unfrozen groundwater system below the lake and the nearby ice sheet.

Acknowledgment

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Appendix A

Permafrost modeling in MIKE SHE

The approach to account for permafrost effects in MIKE SHE simulations has been described and applied in previous studies by Bosson et al. (2010, 2012, 2013a). Instead of a thermal component, the hydraulic properties are changed in the model in order to mimic the permafrost, the talik features and the active layer. To the part of the model domain containing permafrost, a continuously low hydraulic conductivity is applied, independently of the type of soil or bedrock. The hydraulic properties within the taliks are not changing in time and fully thawed conditions prevail the whole annual cycle. Bosson et al. (2010) identified three key model parameters to be changed in order to reach the right seasonal dynamics in the active layer: (i) the hydraulic conductivity, (ii) the leakage coefficient and (iii) the Manning number. These three parameters are changing throughout the year and are controlled by temperature data. In previous studies the use of time-varying properties could not be used in MIKE SHE. Bosson et al. (2010, 2012, 2013a) divided the year into active, freezing, frozen and thawing periods and each period had to be calculated separately. However, the modeling tool has been developed for the present study in order to allow time-varying hydraulic properties and several annual cycles are simulated in the same model.

The hydraulic conductivity values obtained by measurements (Table 4) are applied to each calculation layer for the active period (i.e. when the active layer is unfrozen) and they are reduced to 10^{-16} m/s during the frozen period to represent the low-permeable frozen soil and bedrock materials. The leakage coefficient in MIKE SHE controls the contact between the Overland Flow compartment and the Saturated Zone compartment, i.e. the infiltration capacity. This parameter is reduced to zero during the frozen period. For the active period, the values obtained by infiltration capacity measurements are applied (Table 4). The Manning number describes the surface roughness. The parameter is used in the permafrost simulations to control whether the surface water is mobile or immobile (frozen). The Manning number is set to a very low value in order to increase the surface resistance under frozen conditions. For numerical reasons the value cannot be set to zero; instead a value of 10^{-15} $m^{1/3}/s$ is applied. The Manning number during the active period is set to 5 in the terrestrial part of the catchment and to 0.5 in the lake.

Daily mean soil temperature data from different depths is used to define frozen and thawed conditions in the active layer, presented in Table 2, and air temperature data is used to determine whether surface water is frozen. Soil temperature data from the upper 10 cm of the soil profile controls the infiltration capacity.

Each calculation layer is controlled by soil temperature data from the corresponding depth (0.1 m; 0.25 m; 0.5 m and 0.75 m). The time step of the input data for hydraulic conductivity, infiltration capacity and surface roughness was set to 24 h in order to catch the periods of thawing and freezing. During periods of freezing (Fig. 3 and Table 2) the hydraulic properties are linearly decreased from values representing fully thawed to values

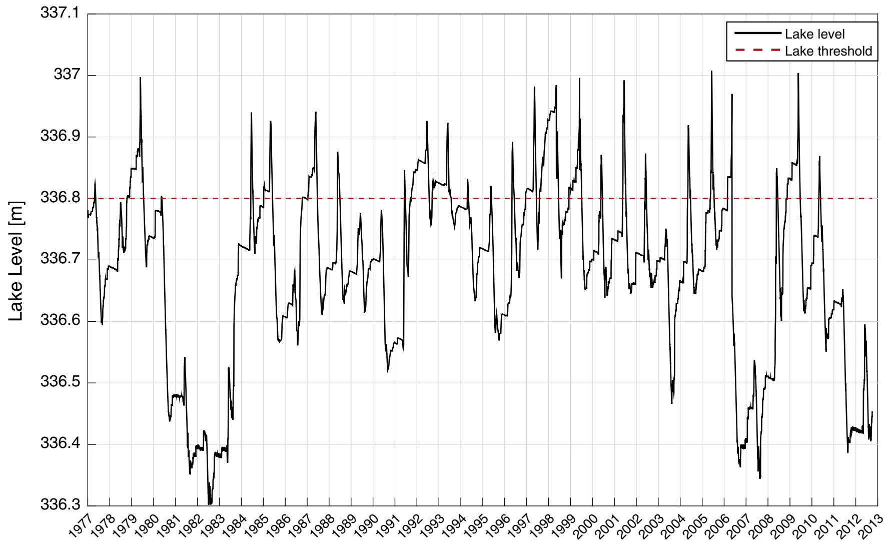

Fig. A1. Calculated lake level for the period 1977–2013. The lake threshold for surface water outflow is marked in the figure, a surface water outflow occurs in approximately 70% of the simulated years.

representing frozen conditions. The period of thawing (Fig. 3 and Table 2) is described accordingly; the properties are linearly increased from frozen to fully thawed conditions.

Results from 37-year simulation

See Fig. A1.

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