The hydrology of the proglacial zone of a high-Arctic glacier (Finsterwalderbreen, Svalbard): Sub-surface water fluxes and complete water budget.

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Abstract

Proglacial areas receive fluxes of glacial meltwater in addition to their own hydrological inputs and outputs, while in high latitudes the seasonal development of the active layer also affects their hydrology. This paper supplements a previous study of the surface and atmospheric water fluxes in the proglacial area of the Svalbard glacier Finsterwalderbreen (77° N), by focusing on the sub-surface water fluxes of the active layer, and bringing together all the components of the proglacial water balance over a complete annual cycle. Particular attention is given to the transitional zone between the moraine complex and the flat sandur. Sub-surface water in the moraine complex (sourced mainly from snowmelt, lake drainage and active-layer thawing), is exchanged with sub-surface water from the sandur (sourced mainly from glacier-derived snow- and icemelt), across a largely distinct boundary. Hydraulic head and specific discharge were monitored in a transect of wells spanning this boundary. A hydraulic gradient from the moraine complex to the sandur is maintained throughout the melt season, although this is reversed first briefly when glacial runoff floods the sandur, and then diurnally from mid-melt-season, as peak daily flow in the proglacial channel network drives sub-surface water in the sandur towards the moraine complex. It is estimated that the active layer does not freeze up until mid-December at this location, so that sub-surface water flow may be maintained for months after the cessation of surface runoff. However, the magnitude of sub-surface flow is very small: the total, annual flux from the moraine complex to the sandur is 11 mm, compared with 1073 mm of total, annual runoff from the whole catchment (glacier included). Furthermore, when considering the water balance of the entire proglacial area, there are unlikely to be significant, seasonal storage changes in the active layer.

Key words proglacial, active layer, hydraulic conductivity, water balance, water budget, Svalbard.

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1. Introduction

Proglacial areas are expanding globally as a consequence of sustained glacier retreat (Zemp et al., 2008), and can be characterized as highly dynamic fluvial environments (Warburton, 1999). Given the intractability of most proglacial areas, and the complex experimental design necessary for monitoring multiple hydrological fluxes over sustained periods in such dynamic environments, it is unsurprising that still very few comprehensive water balance studies are available for glacierized catchments as a whole, or for proglacial areas in particular. Water balances for glaciers themselves can be derived from mass-balance data assuming that inter-annual storage is insignificant (e.g. Hagen et al., 2003), but these shed little light on the hydrological functioning of catchments, and on the interaction and relative contributions of different drainage pathways and potential stores. As water and sediment fluxes from glaciers globally are likely to increase over the coming decades (ACIA, 2004; Meehl et al., 2007) an enhanced understanding of the hydrological functioning of proglacial areas would be beneficial: the purpose of this paper is to contribute to this understanding – building on a previous paper which dealt with surface and atmospheric water fluxes in a proglacial area in the Norwegian high-Arctic archipelago of Svalbard (Hodgkins et al., 2009) – by analyzing sub-surface (active-layer) water fluxes and bringing together the complete water balance over an annual cycle.

Studies of sub-surface hydrology in Svalbard have tended to focus on sub-permafrost groundwater (e.g. Haldorsen et al., 1996; Booji et al., 1998; Haldorsen and Heim, 1999), with relatively little attention paid to water flow within the active layer. However, results from several hydrochemical studies suggest that the annual formation of the active layer is hydrologically significant: observations indicate that the annual formation of the active layer in Svalbard typically commences following snowpack recession in early June (Herz & Andreas, 1966; Stäblein, 1971), when mean air temperatures begin to rise consistently above zero (Hanssen-Bauer et al., 1990). Downward-thawing velocities are initially high, although variations in microtopography and the persistence of patchy snow cover may result in the development of an irregular permafrost table with thawed troughs and frozen ridges, though this irregularity tends to even out as the melt season progresses. The potential for sub-surface water storage and flow in the active layer increases in line with the gradual increase in the depth of the permafrost table, which constitutes the lower boundary layer for water movement (Pecher, 1994). Sub-surface flow in the active layer may increasingly contribute to throughputs of runoff in the proglacial zone as the melt season progresses (Pecher, 1994; Hodson et al., 1998); this effect may be enhanced following precipitation events, due to the displacement of sub-surface water by infiltrating precipitation.

Available studies indicate that Arctic catchments often exhibit a pattern in which runoff appears significantly to exceed precipitation (Killingtveit et al., 2003). This can be attributed to a combination of measurement errors, non-representative locations of precipitation stations, and net glacial ablation. Førland et al. (1997) considered that precipitation underestimation for upland areas by coastally-located gauges may fully account for the discrepancy between precipitation measured at Ny-Ålesund and runoff from the nearby Bayelva catchment. Groundwater storage has often been regarded as insignificant in glacierized catchments in Svalbard (e.g. Hagen et al., 2003), usually owing to the presence of permafrost, although there is little evidence available and groundwater springs are not unusual (Haldorsen and Heim, 1999). With regards to the active layer itself, the observation that total evaporation at elevations <50 m above sea level (a.s.l.) in Svalbard may exceed precipitation by up to about 160% during the summer indicates that water storage there is often sufficient to maintain the rate of evaporation during dry periods (Harding and Lloyd, 1997).

1.1. Aims

The purpose of this paper is to quantify and analyze the sub-surface hydrology of the proglacial area of a high-Arctic glacier, focusing in general on water fluxes in the active layer, and in particular on the transitional zone between the moraine complex and the sandur (see Section 2, ‘Study site description’). Time series of both active layer development and of hydraulic head in the active layer were acquired by in-situ monitoring over the course of the 1999 melt season, in order to elucidate sub-surface hydraulic gradients and flow paths in the transitional zone. The saturated hydraulic conductivity of the sediments comprising the active layer was assessed, in order to enable the time series of hydraulic head to be used to determine specific discharge at the boundary between the moraine complex and the sandur, and thus facilitate the calculation of total sub-surface water fluxes to and from the moraine complex. These fluxes will then be combined with previously-determined surface and atmospheric fluxes (Hodgkins et al., 2009) to present a comprehensive, annual, proglacial water balance.

2. Study site description

The proglacial zone of Finsterwalderbreen is located at 77º 31′ N, 15º 19′ E in the Norwegian High Arctic archipelago of Svalbard (Fig. 1). It is part of a catchment situated on the southern side of Van Keulenfjorden which drains northwards to the sea from a maximum elevation of 1065 m a.s.l. The catchment is constrained to the east, south and west by high mountain ridges, and has a total area of 65.7 km^2, of which 43.5 km^2 is currently glacierized. The non-glacierized part of the catchment comprises steep, scree-covered mountain slopes, with the exception of the proglacial zone itself, which consists of a flat sandur (mostly between 10–20 m a.s.l.) surrounded by a moraine complex (mostly between 20–50 m a.s.l.), situated between the glacier terminus and the coastline of Van Keulenfjorden (Fig. 1). The characteristics of the proglacial zone have been described in detail in Hodgkins et al. (2009).

Of particular interest for this paper is the transitional zone between the moraine complex and the sandur: the former consists of a series of compounded ridges (marking the limits of previous advances) enclosing a hummocky terrain of kames and kettles (many of which contain small lakes) composed largely of glacial diamicton, interspersed with relict outwash terraces; the latter is a relatively uniform, low-gradient surface, composed largely of fluvial sediments, across which glacier meltwater streams braid extensively (Fig. 1). Sub-surface waters from the moraine complex, sourced mainly from snowmelt, lake drainage and active-layer thawing, are exchanged with those from the sandur, sourced mainly from glacier-derived snow- and icemelt, across the largely distinct boundary between the two; this exchange was the focus of field measurements, described in Section 3.

3. Methods: Determination of sub-surface water fluxes between the moraine complex and the sandur

3.1. Hydraulic head monitoring

Hydraulic head was monitored for a total of 36 days, from 19:00 on day 192 (11 July) to 11:00 on day 227 (15 August). At the start of the monitoring period, five PVC-tube monitoring wells (Fig. 2) were sited along a gently-sloping transect spanning the transitional zone, approximately 1 km downstream from the glacier terminus (Fig. 1). The characteristics of these wells have been described in detail by Cooper et al. (2002), in a study of the hydrochemistry of waters in the active layer. Following a period of equilibration, Druck PDCR1830 pressure transducers were used to sample pressure head in each well at 20-s intervals, and record hourly means (potential error ±0.1%). Pressure-head values were calibrated with a measurement of elevation head derived from field surveying, to give the record of hydraulic head. No significant net change in the elevation of the sandur due to aggradation or degradation was detected at the wells transect (with one exception, noted in Section 4.1). The potential error range for hydraulic head is estimated to be ±5%. Active layer depth during the monitoring period was measured every 2–6 days by driving a steel stake into the ground at each well, until the resistance of the uppermost surface of the permafrost was encountered (estimated error range ±5%).

3.2. Saturated hydraulic conductivity testing

The saturated hydraulic conductivity (K_{sat}) of the sediments comprising the active layer in the transitional zone is required to determine specific discharge, and hence quantify sub-surface water fluxes. K_{sat} (m s^{-1}) was assessed using falling-head slug tests in the monitoring wells (Bouwer and Rice, 1976; Bouwer, 1989) and determined from

$$K_{sat} = \frac{R_i^2 \ln(L_r/R_e)}{2L_i}\frac{1}{t}\ln\left(\frac{\psi_0}{\psi_t}\right) \tag{1}$$

where R_i and R_e are the internal and external radii of the well tubing respectively (m), L_r is the effective radius over which the increase in pressure head is dissipated (m), L_i is the length of the screened intake through which water can enter (m), t is the time since $\psi = \psi_0$ (s), ψ is pressure head in the well (m), ψ_0 is the maximum displacement in pressure head at time $t = 0$ (m) and ψ_t is the displacement in pressure head at $t = t$ (m) (Bouwer and Rice, 1976). As L_r is unknown, the dimensionless ratio $\ln(L_r/R_e)$ was estimated from

$$\ln\frac{L_r}{R_e} = \left\{\frac{1.1}{\ln(L_w/R_e)} + \frac{a + b\ln\left[(L_b - L_w)/R_e\right]}{L_i/R_e}\right\}^{-1} \tag{2}$$

where L_w is the distance from the bottom of the well to the water table (m), a and b are dimensionless functions of L_i/R_e and L_b is the distance from the water table to the upper surface of the permafrost (m)(Bouwer, 1989). K_{sat} values obtained in this way ranged from 6.01×10^{-5} m s^{-1} for sandur sediments to 4.08×10^{-4} m s^{-1} for moraine complex sediments.

3.3. Specific discharge and sub-surface water flux calculation

Sub-surface water fluxes, Q (m^3 s^{-1}) (Fig. 3A), were determined as the product of mean hourly values of specific discharge (m s^{-1}) at the well situated closest to the boundary between the moraine complex and the sandur (Well 4) and the cross-sectional area of that boundary, A (m^2), using

$$Q = \left(K_{sat} \frac{dH}{dL} \right) A \tag{3}$$

where dH/dL is the hydraulic gradient between Wells 4 and 5, determined by dividing the difference in mean hourly values of hydraulic head by the distance between the wells, and A is determined by multiplying the saturated layer depth L_b (m) by the moraine complex-sandur boundary length (7000 m): given the flat nature of the sandur, it is probably reasonable to assume that all of the boundary is active concurrently. Hourly values of L_b were determined from

$$L_b = \psi + \left(L_a - L_s \right) \tag{4}$$

where L_a is the depth of the active layer (m) (the distance from the ground surface to the upper surface of the permafrost) and L_s is the distance from the bottom of the well to the ground surface (m).

3.4. Unmonitored sub-surface water fluxes

As with the surface and atmospheric water fluxes at Finsterwalderbreen discussed in Hodgkins et al. (2009), monitoring of hydraulic head and active layer development commenced some time after the onset of the thaw associated with the 1999 melt season, and ceased some time before the annual freeze-up. Sub-surface water fluxes during these missed intervals may have been significant, particularly during the latter period, since the relationship between decreasing air temperature and refreezing is subject to the zero-curtain effect, whereby the release of latent heat stabilizes the temperature of the active layer at 0 °C for a prolonged period, delaying the progression of the freezing front (Boike et al., 1998). However, the development of a robust annual hydrological budget requires these missed fluxes to be quantified.

The first step was to estimate variation in the thickness of the active layer during the pre-monitoring interval. Monitoring-interval data exhibited an almost perfect linear relationship between cumulative, positive, hourly air temperature and active layer depth at all five of the monitoring wells (R^2 values >0.99 in all cases), reflecting the dominance of conductive heat transfer during the melt season. Active layer development at Well 4 was therefore predicted using a linear regression model, constructed from all available input terms for the interval during which active layer depths were monitored, i.e. days 192–227 (11 July–15 August). The second step was to model the freeze-back of the active layer following the cessation of monitoring. Freeze-back at Well 4 was again predicted using regression models, based on significant linear relationships between cumulative, negative, hourly air temperature and the progression of the downward- and upward-moving freezing fronts in the active layer at a comparable site in Svalbard during this interval (Roth and Boike, 2001). Daily, unmonitored, sub-surface water fluxes were then estimated by multiplying values of estimated L_b (a linear function of active layer depth, again using input terms from the monitoring interval – see Equation 4) by the mean value of specific discharge determined during the period of monitoring. A summary of the regression models is provided in Table 1.

4. Results: Sub-surface water fluxes between the moraine complex and the sandur

4.1. Temporal variation in hydraulic head and specific discharge

The temporal pattern of hydraulic head during the period of monitoring (Fig. 3A) was characterised by three periods of markedly different behaviour:

(1) From day 192–196 (11–15 July), hydraulic-head values in those wells sited in the moraine complex (Wells 4 and 5) were high and relatively invariable, while values on the sandur (Well 3) were lower and more variable (Fig. 3A; note that Wells 1 and 2 are excluded from the figures and discussion, as their behaviour was almost identical to that of Well 3, so they add no additional insight); a hydraulic gradient was maintained from the moraine complex to the sandur throughout this period.

(2) From day 197–209 (16–28 July), water levels in all of the wells were somewhat higher and more variable than previously. Peak seasonal values of hydraulic head in all of the wells were recorded in the interval from day 199–202 (18 July–21 July), when the surface of the sandur became flooded in response to peak seasonal flow in the proglacial channel network (Wadham et al., 2001; Hodgkins et al., 2009). During this interval, hydraulic-head values in Wells 3 and 4 periodically exceeded those in Well 5, reversing the hydraulic gradient from the moraine complex to the sandur. As the floodwaters subsided, it became apparent that the surface of the sandur had been eroded between Wells 3 and 4, forming a depression into which channel waters were able to flow.

(3) From day 210–227 (29 July–15 August), water levels in Well 3 were elevated in comparison to interval 1, reflecting the routing of a greater proportion of flow down the western margin of the sandur following the peak seasonal, proglacial flow. A greater degree of diurnal variability was recorded in Wells 3 and 4 during interval 3, along with significant temporal variation in peak daily values of hydraulic head. While peak daily hydraulic-head values in Well 4 closely tracked the diurnal pattern of flow in the proglacial channel network with a 1–2 hour delay, those in Well 3 typically exhibited a 10–14 hour delay. Since similar values of hydraulic head were maintained in Wells 3 and 4 throughout this period, the temporal variation in peak daily values resulted in the reversal of the hydraulic gradient on the sandur on a daily basis. However, consistently high values of hydraulic head in Well 5

maintained an overall hydraulic gradient from the moraine complex to the sandur throughout this interval.

Time series of specific discharge at Well 4 are presented in Fig. 3B. The pattern of specific discharge during the period of monitoring was characterised by a trend of fairly constant discharge (ranging from 2.10×10^{-7} m s^{-1} to 1.64×10^{-6} m s^{-1}), punctuated by a short period of recharge (peak value -1.14×10^{-6} m s^{-1}), which occurred in response to peak seasonal flow in the proglacial channel network and reflects the temporary reversal of the hydraulic gradient. The greater degree of diurnal variability in discharge following the period of recharge reflects the daily inflow and outflow of channel waters to and from the depression formed between Wells 3 and 4. High values of peak daily discharge of ~1.40×10^{-6} m s^{-1} on days 221 (9 August) and 226 (14 August) reflect elevated water levels in the moraine complex following heavy rainfall (Hodgkins et al., 2009: Fig. 2).

4.2. Daily and cumulative sub-surface water fluxes

Active layer depth at each well increased linearly throughout the period of monitoring, at a rate of ~0.01 m d^{-1} (Fig. 4A). Significant spatial variation in active layer depth was observed on the sandur, reflecting local variations in channel proximity and thermal erosion. Despite the deepening of the active layer, water levels in the wells sited in the moraine complex remained relatively constant, resulting in a progressive increase in the thickness of the saturated layer as the season progressed.

Total daily sub-surface water fluxes are presented in Fig. 4B. A total cumulative sub-surface water flux of 9.24×10^3 m^3 was discharged from the moraine complex to the sandur during the 34-day period from days 193–226 (12 July–14 August). Total daily sub-surface water fluxes were positive throughout this time interval, except for on day 199 (18 July), when 2.04×10^2 m^3 was recharged to the moraine complex from the sandur. Very low positive total daily sub-surface water fluxes during the following 2 days reflect shorter, subsequent periods of recharge. The highest total daily sub-surface water flux was recorded on day 226 (14 August), when 4.69×10^2 m^3 (about 5% of the total cumulative sub-surface water flux) was discharged from the moraine complex to the sandur following heavy rainfall.

Total sub-surface water fluxes outside the monitoring period were estimated as the product of the number of missed days of monitoring and the mean daily sub-surface water flux measured during the period of monitoring (2.72×10^2 m^3). The number of missed days was estimated by subtracting the number of days in the period of monitoring (34) from the number of days during which mean daily air temperatures were consecutively positive: 107 days from 5 June–19 September (Hodgkins et al., 2009: Fig. 2). A cumulative total sub-surface water flux of 1.99×10^4 m^3 is therefore estimated to have been discharged from the moraine complex to the sandur outside the period of monitoring. Adding this missed total to the monitored total gives a total annual sub-surface water flux of 2.91×10^4 m^3, of which about 32 % was monitored.

4.3. Sub-surface water flux uncertainties

Various sources of potential error have been identified concerning the calculation of sub-surface water fluxes, including those associated with the use of instrumentation, field techniques and extrapolation in both space and time. Of these potential sources of error, some are quantifiable and thus susceptible to probabilistic analysis, while others are systematic and more difficult to constrain. For example, the disturbance associated with digging and then back-filling holes for the installation of the monitoring wells into the coarse-grained sediments of the moraine complex probably affected the saturated hydraulic conductivity of the surrounding sediments, but to what extent is unknown. Furthermore, with regard to upscaling from specific-discharge estimates at set points on the boundary between the moraine complex and the sandur to sub-surface water flux estimates for the boundary as a whole, it is acknowledged that values of saturated hydraulic conductivity probably vary significantly across the moraine complex, and that the depth of the active layer and thickness of the saturated layer also probably exhibit significant spatial variability.

In view of the above, the sub-surface water fluxes must be viewed as first-order estimates and treated with an appropriate degree of caution, given that it is not possible to determine realistic error estimates with the data available. However, in order to assess the robustness of the flux estimates, a sensitivity analysis was conducted, based on Equation 3, and presented in Table 2. For this analysis, values of the three parameters K_{sat}, dH/dL and L_b were varied between –50% to +50% of the measured/modelled values used to determine the fluxes presented in Section 4.2. The modified values were then used to re-calculate the total annual sub-surface flux (because of the linear form of the equation, varying any of the three parameters by the same proportion has an identical numerical outcome). In addition, fluxes were re-calculated assuming that the length of the hydrologically-active boundary between the moraine complex and the sandur was either constant (at 7000 m, as assumed for the fluxes presented in Section 4.2) or varied linearly, between zero and the seasonal maximum (7000 m) from the start of drainage to 1 July, and from 1 October to end of drainage. From the results given in Table 2, it seems unlikely that the calculated sub-surface flux is in error by an order of magnitude, based on errors in measured/modelled saturated hydraulic conductivity, hydraulic gradient and saturated layer depth. Furthermore, varying the length of the boundary that is hydrologically active has only a minor effect on the calculated fluxes, as its impact is greatest when the rate of sub-surface discharge is smallest: early or late in the melt season, or during the long period of recession flow after the cessation of surface melt.

5. Discussion:

5.1. The annual, proglacial, sub-surface hydrological regime at Finsterwalderbreen

The results presented in Section 4 provide quantitative insights into hydrological pathways in proglacial areas underlain by permafrost. The annual cycle in proglacial atmospheric and surface water fluxes at Finsterwalderbreen was described in detail in Hodgkins et al. (2009). We are able here to develop that description with detail of the variation in sub-surface water fluxes, which are rarely measured in glacierized environments. This description can be used as a context for understanding both the hydrological functioning of high-latitude, glacierized catchments and material fluxes from such catchments (e.g. Wadham et al., 2000; Cooper et al., 2002; Hodgkins et al., 2003).

The annual formation of the active layer commences following the recession of the snowpack, although persistent snow patches may initially delay thawing in some areas. Active layer formation in the Finsterwalderbeen proglacial area is estimated to have commenced around 14 June in 1999. As the melt season proceeds, lake levels in the moraine complex fall in response to the gradual deepening of the active layer and resultant water loss. Consequently, an increasing proportion of runoff is routed from the moraine complex to the sandur via sub-surface flow paths, resulting in the gradual disappearance of many ephemeral surface channels, and a dominant, sub-surface hydraulic gradient from moraine complex to sandur becomes established.

Peak seasonal flow in the proglacial, surface channel network tends to occur in mid-to-late July in this location, in response to high rates of ablation on the lower reaches of the main glacier; it may be accompanied by subglacial outburst floods, submerging the sandur for several days at a time (Wadham et al., 2001). The impact of such events on sub-surface flow is significant, since the dominant hydraulic gradient from the moraine complex to the sandur is temporarily reversed, allowing floodwaters from the proglacial channel network to recharge sub-surface water levels in the active layer at the boundary zone of the moraine complex. During preceding and succeeding periods of lower flow, the hydraulic gradient is maintained from the moraine complex to the sandur, although the gradient beneath the sandur itself is reversed diurnally, in response to diurnal flow variations in the proglacial channel network. The rate of sub-surface discharge from the moraine complex increases as the melt season proceeds, reflecting the progressive deepening of the active layer and the supply of water from the interior of the moraine complex. A degree of flushing occurs following periods of rainfall, as infiltrating precipitation displaces water stored in the active layer.

The refreezing of the active layer commences in early-to-mid October, as two freezing waves begin to advance: one from the ground surface and one from the permafrost table (Marlin et al., 1993). However, complete refreezing may take 6–8 weeks, since the release of latent heat upon freezing offsets the initial drop in temperature (French, 2007). This temporal pattern appears to be typical for high-latitude, permafrost-influenced catchments. For instance, Humlum (1998) found that seasonal maximum thaw depth in the active layer was reached in late September at Qeqertarsuaq, Greenland (69˚ 15’ N, mean annual air temperature –5.1˚ C) and that the closure of the active layer occurred between late December and late January. In the somewhat more northerly Finsterwalderbeen proglacial area, closure is estimated to have occurred around 11 December in 1999 (Fig. 4A).

5.2. The complete (atmospheric, surface and sub-surface) annual proglacial hydrological budget at Finsterwalderbreen

The results presented in this paper, in combination with those presented in Hodgkins et al. (2009), also enable the complete, annual hydrological budget of the proglacial zone to be determined. This allows the relative importance of the various hydrological pathways in the proglacial zone to be identified. The annual, steady-state hydrological budget of the proglacial zone may be represented by the simple water-balance model

$$W_{PZ} = W_P + W_R - W_E - W_{SSS} - W_{SR} \pm \Delta W_S \qquad (5)$$

where W_{PZ} is the net proglacial water flux, W_P is the precipitation water flux, W_R is the channel recharge water flux (active-layer discharge from the sandur to the moraine complex),

W_E is the evaporation water flux, W_{SSS} is the sub-surface seepage water flux (active-layer discharge from the moraine complex to the sandur), W_{SR} is the surface runoff water flux (mainly snowmelt and lake drainage from the moraine complex) and ΔW_S is the change in water storage. A schematic of this model is presented in Fig. 5, with the addition of both glacial runoff, which effectively constitutes a proglacial throughput, and bulk runoff, which is the sum of glacial runoff and the net proglacial water flux. Specific values of the water balance terms are also given in Table 3. In both cases, the water fluxes given are annual totals, derived from the data presented in this paper and in Hodgkins et al. (2009), with the exception of the value for surface runoff from the moraine complex, which was determined by balance, assuming zero change in water storage.

In the year studied, precipitation exceeded evaporation by a little over 80%, though during the summer season, the evaporation rate was almost five times that of precipitation. Runoff was simply determined here as precipitation minus evaporation, as monitoring the extensive network of small, surface streams draining the moraine complex was not a feasible task. The value of runoff (104 mm a^{-1}) is small compared to other Svalbard values given by Killingtveit et al. (2003), though this can be explained by the inclusion of glacial runoff in those other values. However, assuming no significant storage changes, the water budget balances without any obvious difficulties or anomalies, so there is no indication of an apparent precipitation deficit, as identified at some other Arctic catchments: again, this may be partly attributable to the separation of glacial water fluxes from specifically proglacial ones in this study. Killingtveit et al. (2003) considered the main uncertainties in high-latitude water balances to be: (1) the distribution of precipitation, and (2) the rate of evaporation. Regarding (1), the Finsterwalderbreen proglacial area has a limited elevation range (10–50 m a.s.l., the maximum being moraine crests of limited spatial extent), so minimal extrapolation is required, but the hummocky topography of the moraine complex contributes to the relatively large uncertainty in winter precipitation, in particular; this is reflected in the large, proportional error term for the precipitation flux in Fig. 5. Regarding (2), the modelled rate of evaporation from the proglacial area of Finsterwalderbreen (141 mm a^{-1}) compares favourably with other estimates from non-glacierized (and the non-glacierized parts of glacierized) Svalbard catchments, which are in the range 51–200 mm a^{-1} (Jania and Pulina, 1994; Killingtveit et al., 1994; Bruland, 2001; Mercier, 2001). Killingtveit et al. (2003) determined average annual evaporation (as a function of air temperature, based on evaporation pan measurements at Ny-Ålesund) for glacier-free areas at three locations in Svalbard to be about 80 mm a^{-1}.

The notable exclusion from the water-balance model is the contribution to total, annual glacial runoff by over-winter subglacial drainage. However, a reliable estimate for this value may be derived by calculating the water-equivalent volume of the proglacial icing, which typically accumulates over an area of about 0.3 km^2. Previous coring investigations have revealed that the icing typically has a mean thickness of 1.5 m (Wadham et al., 2000). Assuming an ice density of 900 kg m^{-3}, a water-equivalent volume of about 4.05×10^5 m^3 of winter subglacial drainage is implied. This estimate equates to <1% of average annual glacial runoff and is therefore unlikely to be a significant source of error through its contribution to storage changes. Neither is there any indication of significant changes in active-layer water storage; the magnitude of the sub-surface water flux is an order of magnitude smaller than the atmospheric and surface fluxes, and fully two orders of magnitude smaller than the bulk catchment runoff (Table 3).

6. Conclusions

Understanding the water balance of glacierized catchments is important both for furthering scientific understanding of the hydrological functioning of snow- and ice-fed systems at a time of rapid environmental change, and for the management of snow- and ice-derived water resources of the world's major mountain chains, likewise in the context of change (Barnett et al., 2005; Bates et al., 2008). Even in high latitudes, remote from centres of population, changes in the storage and release of freshwater may have important implications for the functioning of aquatic ecosystems, ocean currents and ice-sheet stability (Das et al., 2008; Hanna et al., 2008; Mernild et al., 2008; Milner et al., 2009; Schofield et al., 2010).

This contribution has demonstrated that active-layer depth, hydraulic head and specific discharge may be successfully monitored as part of a water balance study in permafrost-influenced catchments. There are a range of practical limitations necessarily associated with monitoring sub-surface processes in remote and relatively intractable areas such as the Finsterwalderbreen catchment in Svalbard, but some aspects of the environment compensate for these: for instance, the air temperature-active-layer depth relationship is very linear, allowing early-season thawing and late-season freezing to be modelled quite straightforwardly. The active layer itself responds quite sensitively to forcing from proglacial surface hydrology, with diurnal reversals of the hydraulic gradient between the moraine complex and the sandur taking place from the mid-melt-season onwards, and clear flux peaks related to rainfall.

The results obtained are consistent with previous water balance studies from Svalbard, though this contribution is distinctive in quantifying active-layer fluxes, although the total annual flux from the moraine complex to the sandur, at 11 mm, is very small compared to the total annual catchment runoff, at 1073 mm. While the total water balance was determined assuming no significant changes in any plausible stores, the consistency of the various measured or estimated values for the balance components suggests there are unlikely to be significant gains from or losses to storage in the active layer. Uncertainties in the sub-surface variables are difficult to quantify, although given the small magnitude of the annual totals, are unlikely to have an important effect on the water balance calculation. Probably the principal source of uncertainty is the representativeness of the location used for well monitoring: this is essentially a matter of judgement. In any case, it is clear that throughputs from the adjacent glacier – 1697 mm of runoff in the season studied (Hodgkins et al., 2009) – dominate the proglacial area hydrologically, underlining the important role of glacially-derived water fluxes in this high-latitude region.

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Figure captions

Fig. 1. Location of the study site within the Svalbard archipelago (inset) and configuration of the Finsterwalderbreen proglacial area (main). The limits of the moraine complex are shown with a solid white line; the part of the moraine complex that drains to the sandur is delimited by the long-dashed line; the boundary between the moraine complex and sandur is shown by the short-dashed line. WT marks the position of the wells transect, along which five wells are located (Fig. 2A). Aerial photograph acquired by UK Natural Environment Research Council Airborne Research and Survey Facility in 2003.

Fig. 2. (A) Detail map of the wells transect, located in Fig. 1. Wells 1 and 2 behave so similarly to Well 3 that their data are excluded from subsequent figures, for clarity. (B) Example of a monitoring well, comprising a pair of rigid, plastic tubes, the bottoms of which are sealed; the buried length of each tube features a screened intake into which sub-surface waters can flow. One tube was used for water sampling for hydrochemical studies (Cooper et al., 2002), while the monitoring instruments described in the text are secured at the bottom of the other tube: further details are given in Cooper et al. (2002).

Fig. 3. Temporal variation in (A) hydraulic head and (B) specific discharge during the period of monitoring. Well 4 is in the transitional zone between the moraine complex and the sandur; Well 5 is representative of the former, Well 3 of the latter. Note that positive values indicate discharge from the moraine complex to the sandur and negative values indicate recharge from the sandur to the moraine complex.

Fig. 4. Proglacial active layer depth/thickness (A) and daily sub-surface water flux between the moraine complex and sandur (B). Note that positive values indicate discharge from the

moraine complex to the sandur and negative values indicate recharge in the opposite direction.

Fig. 5. Schematic representation of the complete, annual water budget for the Finsterwalderbreen proglacial area. Blue arrows represent inputs, red arrows represent outputs, and other arrows represent internal transfers; broken lines represent minor multi-directional, stores/exchanges that cannot be quantified from the data available. All of the water fluxes presented in the figure are given in m^3, with estimates of probable error, except for channel recharge, active-layer discharge and surface runoff, for which it is not possible to determine realistic error estimates with the data available: these fluxes must therefore be viewed as first-order estimates. The determination of errors in all other water fluxes are described in detail in Hodgkins et al. (2009), with additional estimates as follows: (1) the error for the precipitation water flux is an average of probable errors previously determined for the rainfall and snowpack water-equivalent fluxes, weighted by proportion of total precipitation; (2) the probable error for the glacial runoff water flux is an average, weighted in proportion to overall contribution, of probable errors from the Terminus East and West gauging stations (average-weighted by time, as these errors varied temporally: Hodgkins et al., 2009) and the supraglacial water flux determined from ablation measurements.

Figure 1 (black/white)
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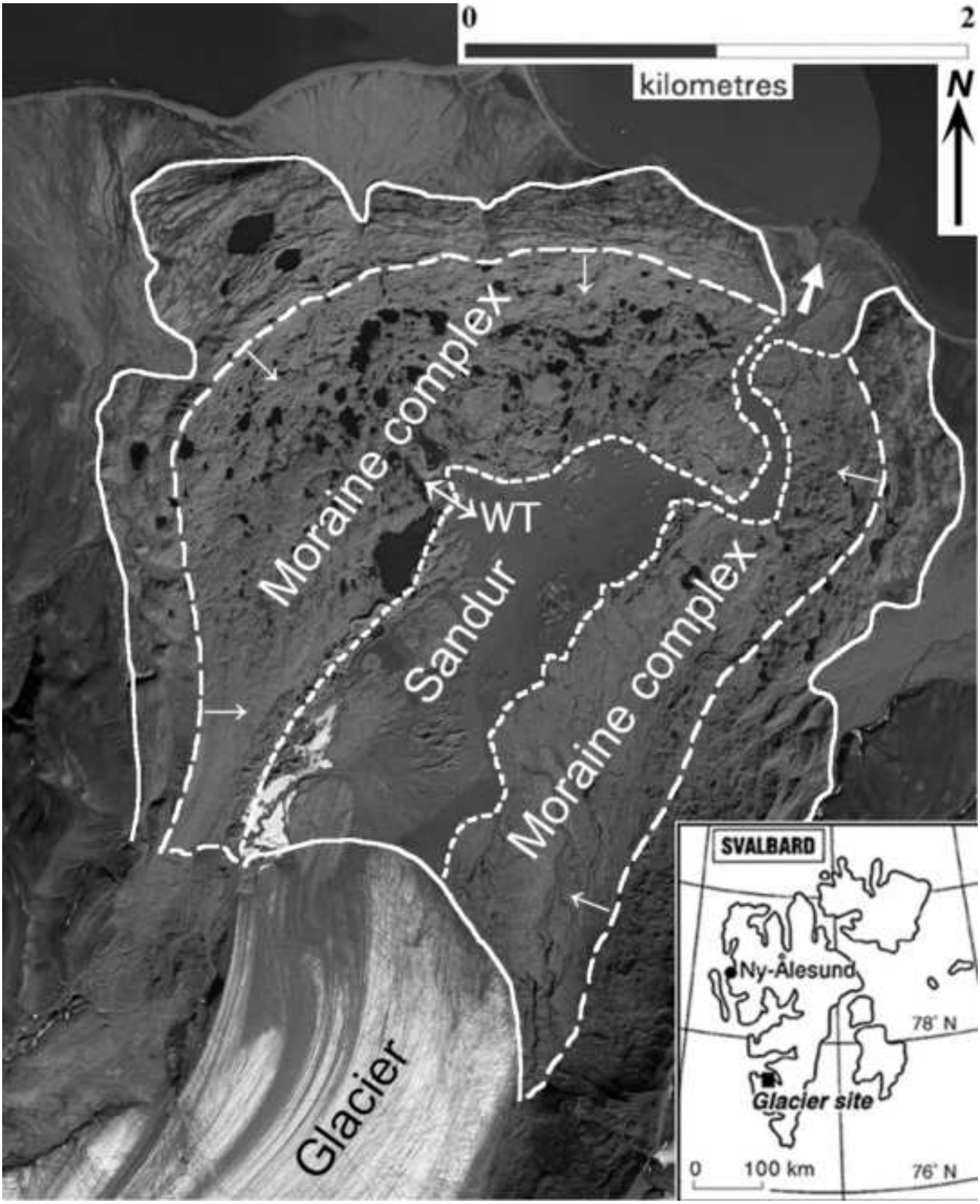

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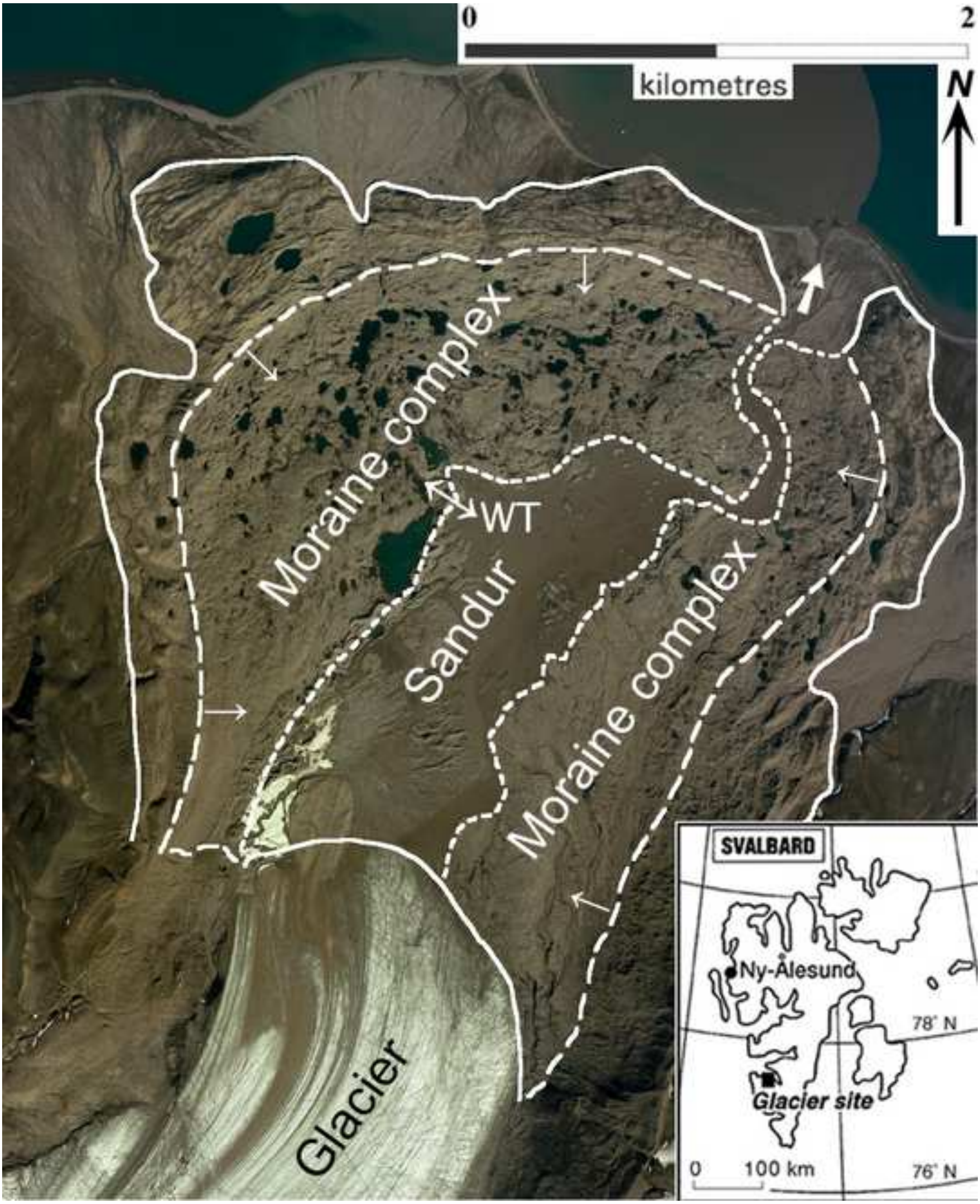

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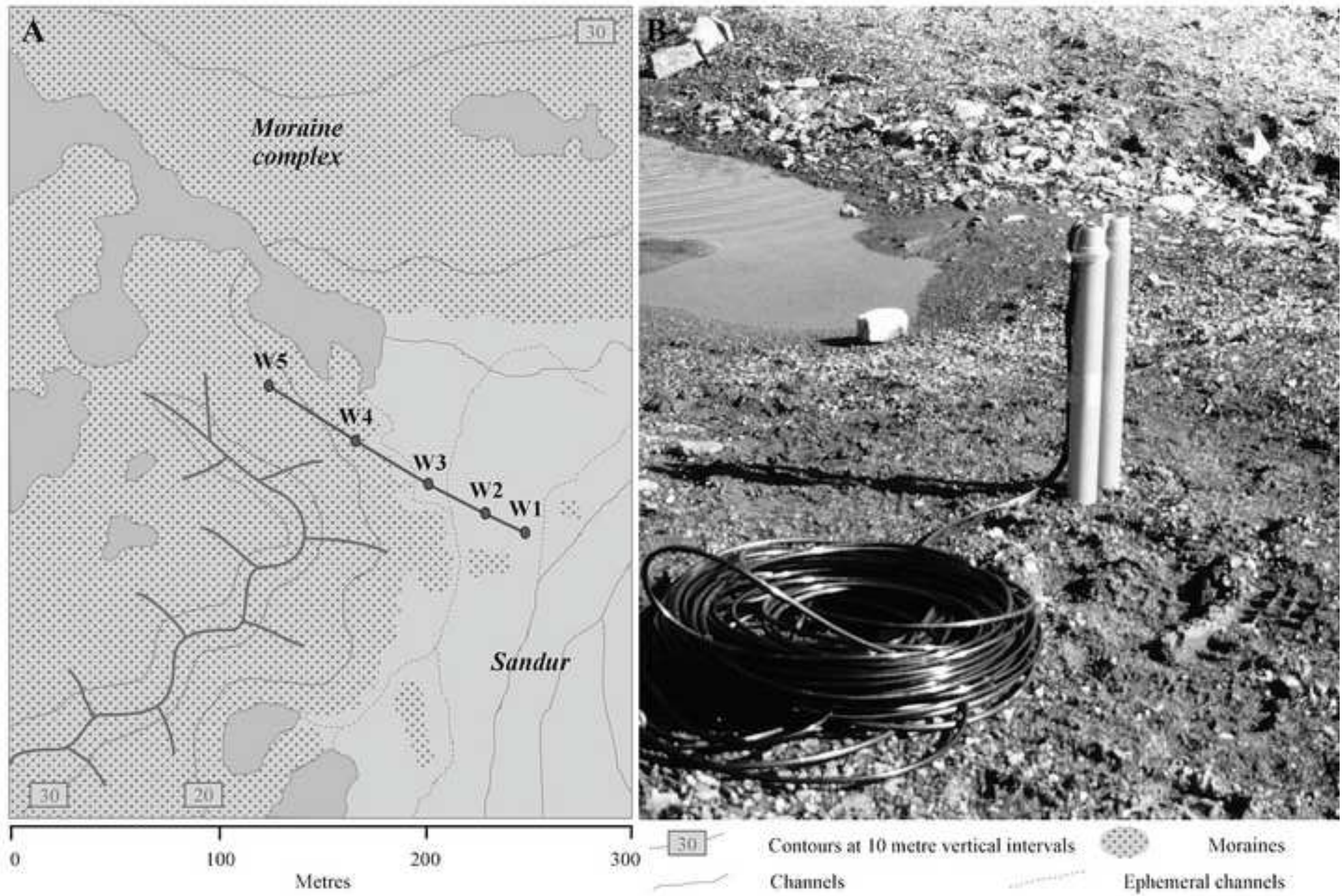

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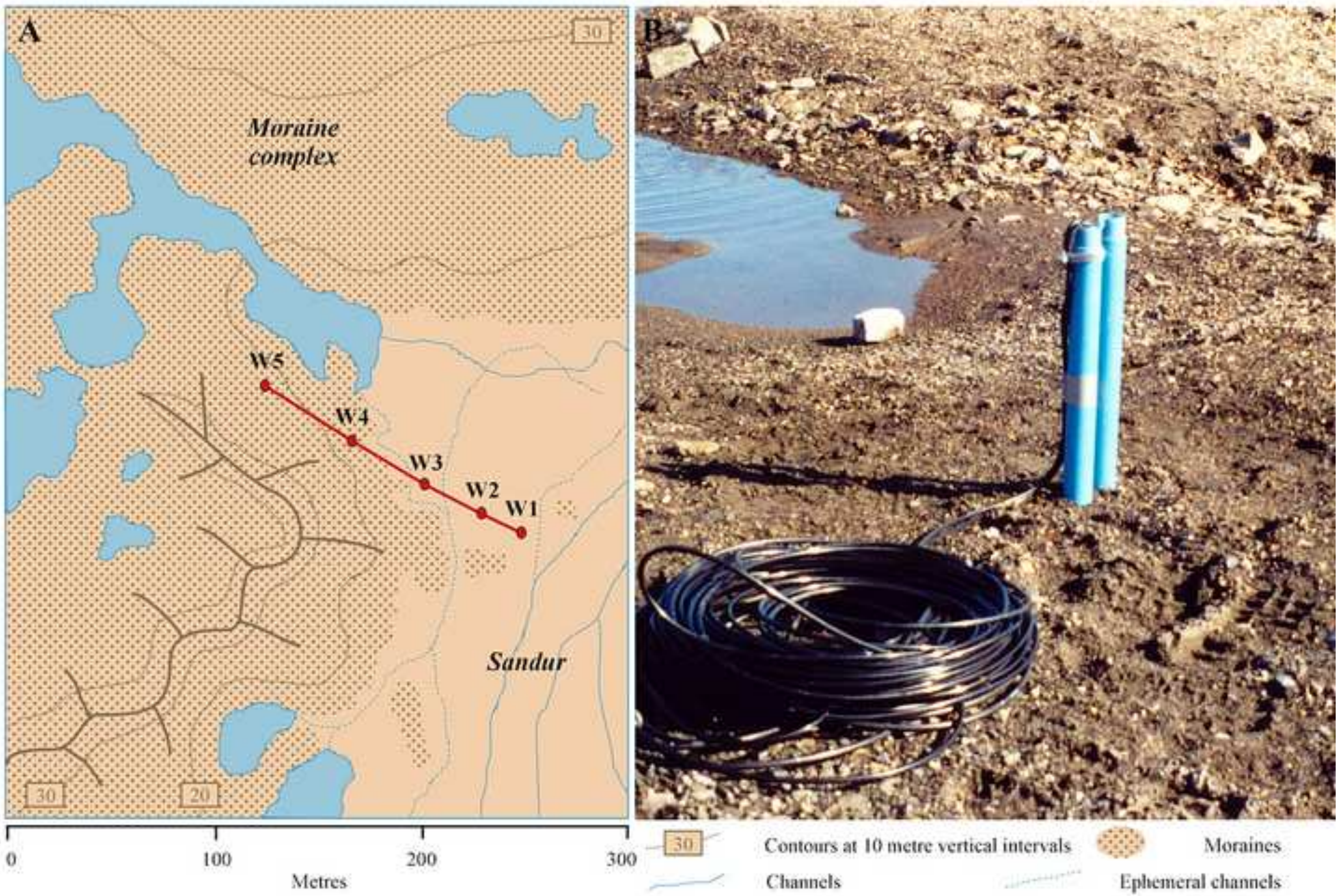

Figure 3 revised

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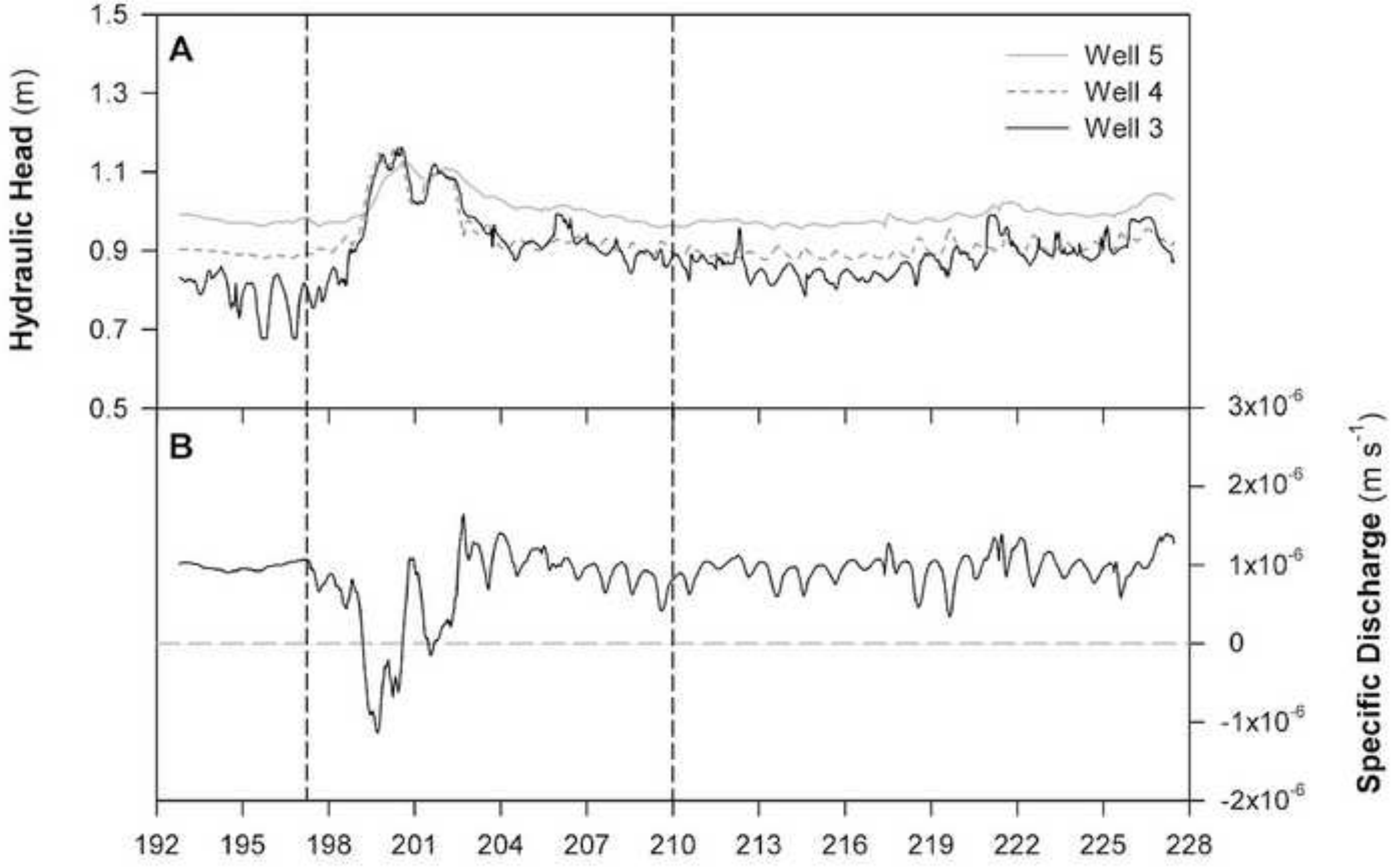

Figure 4

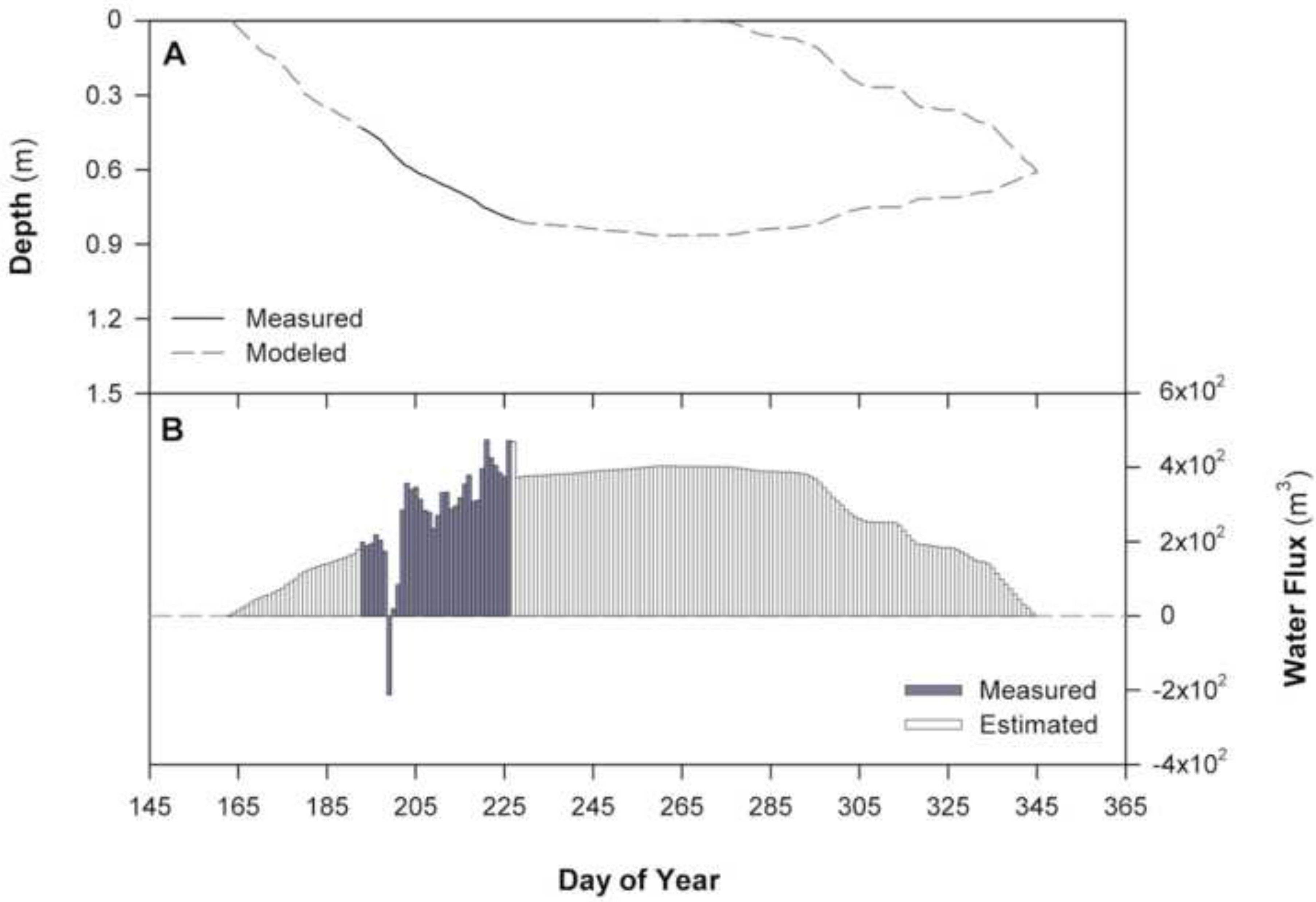
A
B
Depth (m)
0
0.3
0.6
0.9
1.2
1.5
Measured
Modeled
6x10^2
4x10^2
2x10^2
0
-2x10^2
-4x10^2
Water Flux (m^3)
Measured
Estimated
145
165
185
205
225
245
265
285
305
325
345
365
Day of Year

Figure 5 (black/white)
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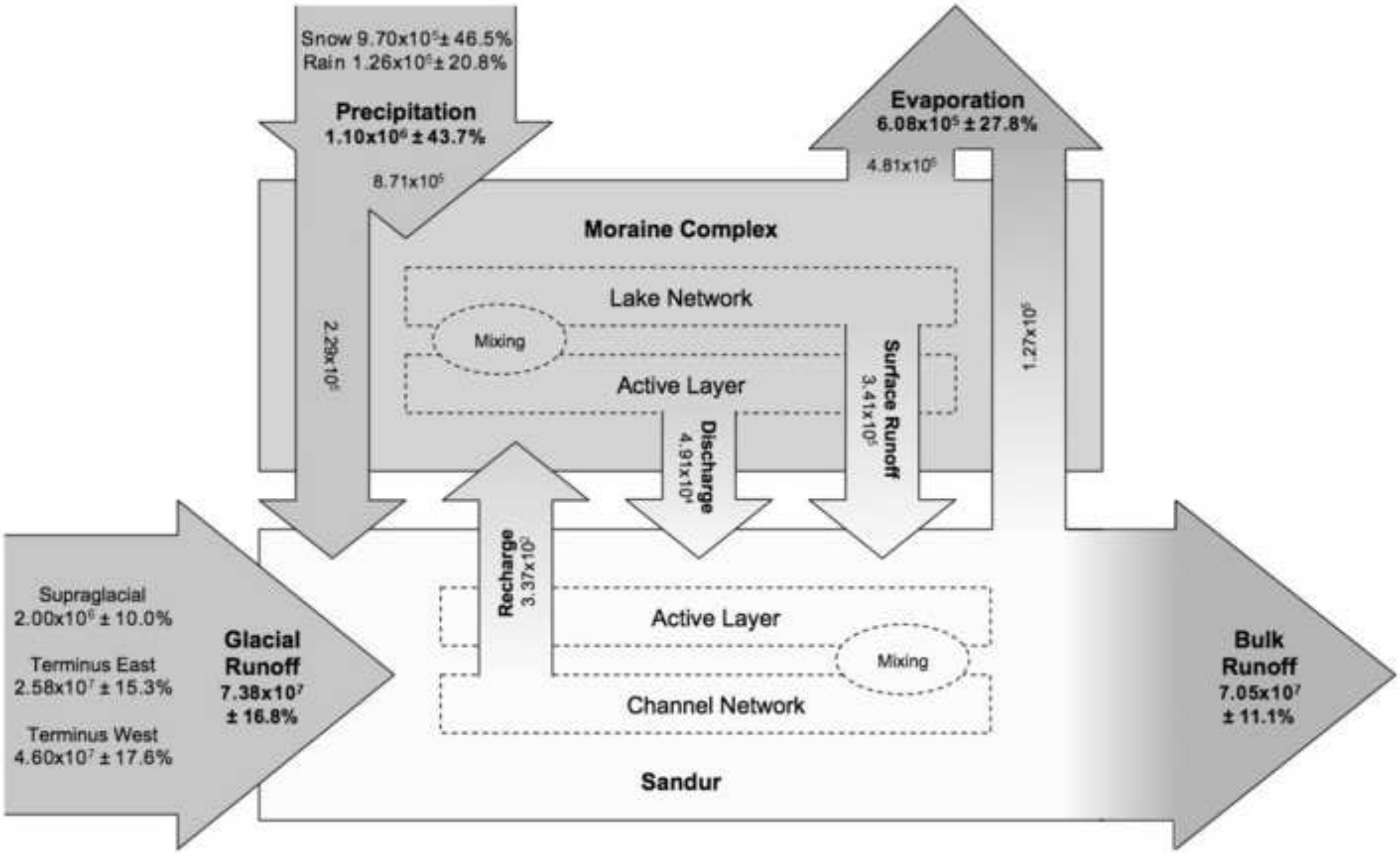

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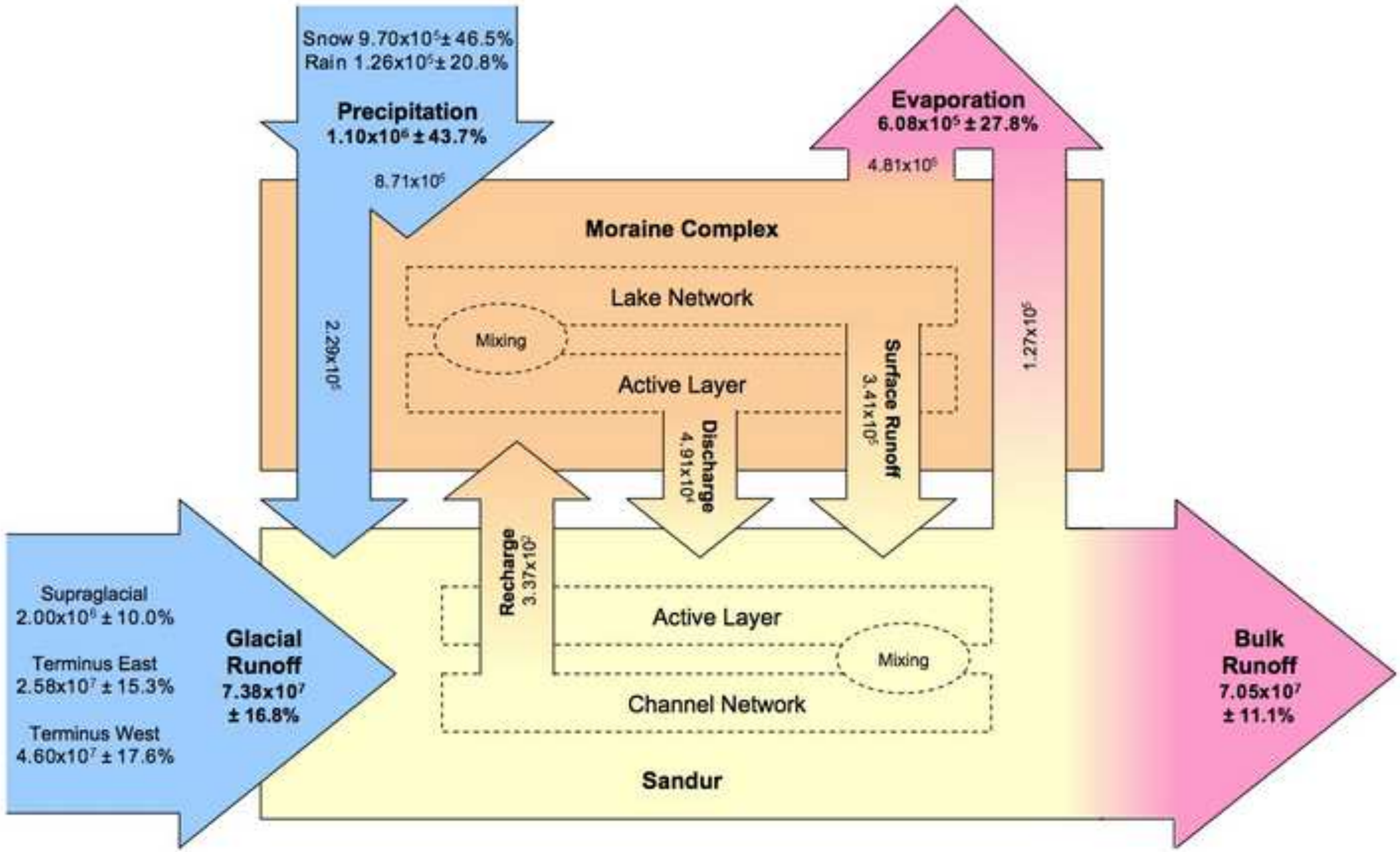

Tables

Table 1: Linear regression models used to predict active- and saturated-layer depths (L_a and L_b respectively) during unmonitored intervals at the start and end of the 1999 melt season. All coefficients are significant at $p<0.05$. T_a is hourly air temperature, positive or negative.

Independent variable	Dependent variable	Slope	Intercept	R^2	When applied
L_a	L_b	1.047	–0.134	0.981 ($n = 690$)	For duration of active layer formation
$\sum T_a^+$	L_a	0.012	–0.380	1.000 ($n = 2$)	During active-layer thawing
$\sum T_a^-$	L_a	-2.624×10^{-4}	0.000	0.837 ($n = 5$)	During active-layer freezing (surface-down)
$\sum T_a^-$	L_a	1.111×10^{-4}	1.179	0.635 ($n = 3$)	During active-layer freezing (bottom-up)

Table 2: Sensitivity analysis of sub-surface water flux using equation 3, for the entire period of sub-surface flow (monitored and modelled fluxes, from June to December).

	Constant boundary	Variable boundary
Results using parameter values given in text		
Total annual flux	4.91×10^{4} m^{3}	4.19×10^{4} m^{3}
Mean annual discharge	$3.12\pm1.56\times10^{-3}$ m^{3} s^{-1}	$2.66\pm1.77\times10^{-3}$ m^{3} s^{-1}
Corresponding results with any of K_{sat}, dH/dL or L_b modified by stated percentage		
+50%	7.36×10^{4} m^{3} $4.68\pm2.34\times10^{-3}$ m^{3} s^{-1}	6.29×10^{4} m^{3} $4.00\pm2.66\times10^{-3}$ m^{3} s^{-1}
+30%	6.38×10^{4} m^{3} $4.06\pm2.02\times10^{-3}$ m^{3} s^{-1}	5.45×10^{4} m^{3} $3.46\pm2.31\times10^{-3}$ m^{3} s^{-1}
+20%	5.89×10^{4} m^{3} $3.75\pm1.87\times10^{-3}$ m^{3} s^{-1}	5.03×10^{4} m^{3} $3.20\pm2.13\times10^{-3}$ m^{3} s^{-1}
–20%	3.93×10^{4} m^{3} $2.50\pm1.25\times10^{-3}$ m^{3} s^{-1}	3.35×10^{4} m^{3} $2.13\pm1.42\times10^{-3}$ m^{3} s^{-1}
–30%	3.44×10^{4} m^{3} $2.19\pm1.09\times10^{-3}$ m^{3} s^{-1}	2.93×10^{4} m^{3} $1.87\pm1.24\times10^{-3}$ m^{3} s^{-1}
–50%	2.45×10^{4} m^{3} $1.56\pm0.78\times10^{-3}$ m^{3} s^{-1}	2.10×10^{4} m^{3} $1.33\pm0.89\times10^{-3}$ m^{3} s^{-1}

Table 3: Finsterwalderbreen proglacial area annual water balance, 1999: summary.

Component	Value
Precipitation (winter)	226 mm
Precipitation (summer)	29 mm
Precipitation (total)	256 mm
Evaporation	141 mm
Precipitation - evaporation	115 mm
Surface runoff	104 mm (inferred)
Sub-surface discharge from moraine complex to sandur	11 mm
Sub-surface recharge from sandur to moraine complex	0.1 mm
Glacial runoff	1697 mm (43.5 km^2 glacierized area)
Bulk catchment runoff	1073 mm (65.7 km^2 total catchment area)