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

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1 **The hydrology of the proglacial zone of a High-Arctic glacier**
2 **(Finsterwalderbreen, Svalbard): atmospheric and surface water fluxes.**

3

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15

16 **Abstract**

17 Proglacial areas are expanding globally as a consequence of sustained glacier retreat, but
18 there are very few studies focusing on their hydrology. This paper examines the surface and
19 atmospheric water fluxes over a complete annual cycle in the proglacial area of the Svalbard
20 glacier Finsterwalderbreen (77° N), through a combination of field measurements, physical
21 modelling and statistical estimation. Precipitation in winter (226 mm) exceeded that in
22 summer (29 mm), and over the course of the annual cycle total precipitation exceeded total
23 evaporation (141 mm), although evaporative outputs from the proglacial area exceeded
24 precipitation inputs during the dry summer. Runoff was highly irregular in time, with much
25 of the total annual flow being concentrated into two relatively brief, early-to-mid summer
26 intervals, the greater of which was characterised by the release of subglacially-stored water.
27 Water fluxes were dominated by meltwater supply from the glacier: the total annual glacial
28 runoff ($7.38 \times 10^7 \text{ m}^3$) was an order-of-magnitude greater than the precipitation flux delivered
29 directly to the proglacial area, and two orders-of-magnitude greater than evaporative losses
30 from it. Outputs of meltwater from the proglacial area were not significantly different from
31 inputs over the duration of the melt season, so surface water storage does not appear to be
32 important in the studied catchment, despite episodes of flooding over shorter timescales. A
33 synthesised description of the seasonal hydrological cycle in Finsterwalderbreen's proglacial
34 area is presented, which can be viewed as a set of hydrological boundary conditions for
35 comparable high-latitude locations. Further study of these conditions is required, because the
36 challenging nature of hydrometry in the high-latitudes has the potential to limit progress in
37 understanding environmental change there.

38

39 **Key words** Arctic, Svalbard, proglacial, precipitation, runoff, evaporation.

40 **PACS codes** 92.40.Vq, 92.40.We, 92.40.Zg

41

42 **1. Introduction**

43 **1.1 Proglacial hydrology**

44 Proglacial areas, located immediately in front of glaciers and strongly influenced by
45 fluxes of water and sediment from them, are expanding globally as a consequence of
46 sustained glacier retreat. Studies of specifically proglacial hydrology are few in number,
47 reflecting the tendency for research in glacierized catchments to focus principally on glacial
48 processes (e.g. Willis, 2005). Hydrological data from the Arctic are sparse even today, and in
49 the Norwegian Arctic archipelago of Svalbard, for example, there are currently only 8
50 continuously-recording hydrometric stations, in 5 locations (Sund, 2008). Yet detailed
51 information on the quantity of water stored as snow, ice, groundwater and in lakes, and on the
52 exchange of water between these stores, from the atmosphere as precipitation and from
53 catchments as evaporation and runoff, is important both for the development of Arctic
54 communities and for scientific understanding of the Arctic hydrological cycle and its
55 response to enhanced atmospheric warming (Anisimov et al., 2007; Bates et al., 2008).

56 Results from studies of glacial hydrology do provide information concerning the
57 nature of meltwater inputs to the proglacial zone. Observations from Svalbard indicate that
58 the annual thaw typically commences in early June (Repp, 1988; Vatne *et al.*, 1995;
59 Hodgkins *et al.*, 1997), when air temperatures begin to rise consistently above zero (Hanssen-
60 Bauer *et al.*, 1990). However, the onset of runoff typically lags the increase in energy inputs,
61 since significant volumes of meltwater are temporarily stored in the snowpack and in various
62 glacial reservoirs (Repp, 1988; Vatne *et al.*, 1996; Hodgkins, 2001; Hodson et al., 2005).
63 Early season discharge may therefore be highly variable, with little indication of diurnal
64 cycling until later in the melt season, when such stores are depleted. Results from numerous
65 studies indicate that peak runoff typically occurs during July and August in Svalbard (Repp,
66 1988; Vatne *et al.*, 1995; Hodgkins *et al.*, 1997; Wadham *et al.*, 1997; Hodson et al., 1998;
67 Hodgkins, 2001), when air temperatures are at a maximum (Hanssen-Bauer *et al.*, 1990). The

68 sudden release of large volumes of stored meltwater to the proglacial zone from subglacial
69 reservoirs has been observed during the summer (Wadham *et al.*, 2001).

70 The presence of cold, impermeable surface ice on many glaciers in Svalbard often
71 results in a significant proportion of meltwater being directed to the margins and routed to the
72 proglacial zone in lateral, ice-marginal channels (Hodgkins, 1997; Hagen *et al.*, 2000).
73 However, in instances where runoff is able to access the glacier bed, meltwater may be routed
74 to the proglacial zone subglacially and may emerge under artesian pressure in front of the
75 glacier terminus (Vatne *et al.*, 1995; Wadham *et al.*, 1998; Hodson *et al.*, 2005). No direct
76 observations of the annual freeze-up in Svalbard have been reported to date, although it is
77 likely that the cessation of runoff occurs in early October, when air temperatures begin to fall
78 consistently below zero (Hanssen-Bauer *et al.*, 1990). In some instances, stored meltwater
79 may continue to issue from ice-marginal and subglacial reservoirs during the winter, forming
80 extensive icings (*aufeis* or *naledi*) in the proglacial zone (Vatne *et al.*, 1995; Hodgkins *et al.*,
81 1997; Wadham *et al.*, 2000; Hodgkins *et al.*, 2004; Hodson *et al.*, 2005).

82

83 **1.2 Aims**

84 The purpose of this paper is to quantify and analyse the hydrology of the proglacial
85 area of a high-arctic glacier, focusing on surface and atmospheric water fluxes. A
86 combination of field measurements and modelling will be used to determine temporal
87 variation in precipitation, evaporation and runoff over the course of an annual cycle, and
88 daily and total cumulative fluxes of water from each of these sources will be quantified. This
89 will lead to a synthesis of the annual proglacial surface hydrological regime. The subject of
90 this study is the proglacial area of the Svalbard glacier Finsterwalderbeen: this glacier has
91 already been the subject of a variety of glaciological and hydrological research over the past
92 decade, including studies of its thermal regime (Ødegård *et al.*, 1997), surface and sub-
93 surface hydrochemistry (Wadham *et al.*, 1998, 2000, 2001; Cooper *et al.*, 2002), fluvial

94 sediment transfer (Hodson and Ferguson, 1999; Hodgkins et al., 2003), the spatial and
95 temporal variation of winter accumulation (Hodgkins et al., 2005) and glacier dynamics
96 (Nuttall et al., 1997; Nuttall and Hodgkins, 2005; Hodgkins et al., 2007). A companion paper
97 will focus on sub-surface water fluxes in the proglacial area, and on the complete annual
98 water budget.

99

100 **2. Study site description**

101 The proglacial zone of Finsterwalderbreen is located at 77° 31' N, 15° 19' E in
102 Svalbard, Norway (Figure 1), and is part of a 65.7 km² (43.5 km² glacierized) catchment. The
103 catchment is mostly devoid of vegetation, except above the most recent glacial trimline and
104 on terminal moraines delimiting the proglacial zone, where a sparse Arctic flora survives.
105 The bedrock geology is diverse, comprising Precambrian basement and Carboniferous
106 through Cretaceous sedimentary units (Dallmann *et al.*, 1990). The mean annual air
107 temperature at 35 m a.s.l. is -3.9 °C, and mean monthly air temperatures are only positive
108 during the summer, although even then they remain <6.0 °C (Hanssen-Bauer *et al.*, 1990).
109 Annual precipitation totals lie in the range 180-440 mm w.e., with the bulk being delivered as
110 snow during the winter months (Hansen-Bauer *et al.*, 1990).

111 The proglacial zone itself consists of a sandur and a moraine complex situated
112 between the glacier terminus and the coastline of Van Keulenfjorden (Figure 1); it has a total
113 area of 4.3 km², most of which has only relatively recently been exposed by the retreat of the
114 glacier from the limits of a surge event which occurred between 1898 and 1918 (Nuttall *et al.*,
115 1997). The sediments comprising the proglacial zone contain material from all elements of
116 the catchment lithology. The proglacial zone is underlain by permafrost, the upper layers of
117 which thaw during the summer months to form a shallow active layer. The proglacial zone is
118 constrained to the east, north and west by a series of compounded terminal moraines, which
119 mark the limits of previous glacial advances. The remainder of the moraine complex

120 comprises kames and kettles interspersed with relict outwash terraces and hummocky
121 moraines. Many of the kettles and other depressions in this zone contain perennial lakes and
122 pools, the largest of which has a surface area of 0.03 km². Many of these lakes and pools are
123 connected by a network of small channels that convey surface runoff, which comprises
124 varying proportions of snowmelt, rainfall and meltwater derived from in-situ thawing of the
125 active layer, along a topographic gradient from the moraine complex to the sandur.

126 The sandur, which extends north-eastwards from the glacier terminus, is 1.5 km long
127 and has a total area of 0.9 km², with an altitudinal range of 10–50 m a.s.l. The morphology of
128 the sandur changes with increasing distance downstream, forming three distinct zones similar
129 to those described by Krigstrom (1962) for Icelandic sandar. In the proximal zone, which
130 extends approximately 0.5 km north-eastwards from the glacier terminus, runoff is conveyed
131 mainly in a network of deep, narrow channels incised into coarse, gravelly sediments. In the
132 intermediate zone, which extends approximately 0.5 km north-eastwards from the proximal
133 zone, flow is conveyed in a network of braided channels incised into fine, sandy sediments.
134 Many of these channels shift position frequently and some only convey flow during periods
135 of high discharge. In the distal zone, which extends approximately 0.5 km north-eastwards
136 from the intermediate zone, flow is conveyed in a network of very wide and ill-defined
137 channels incised into fine, silty sediments. Many of these channels overflow during periods
138 of high discharge, producing extensive areas of shallow flooding on the sandur surface. The
139 channels in the distal zone converge where the sandur abuts the moraine complex to form a
140 single well-defined channel, hereafter referred to as the Outlet (Figure 1), which breaches the
141 terminal moraines before issuing into Van Keulenfjorden 1.5 km further downstream.

142

143 **3. Methods**

144 Meteorological time series were acquired over an 11-month period that included the
145 1999 melt season, in order to facilitate the calculation of total atmospheric water fluxes to and

146 from the proglacial zone. Data concerning over-winter snow accumulation were acquired
147 prior to the onset of the thaw associated with the 2000 melt season, in order to validate the
148 use of regression relationships between altitude and snow depth derived from end-of-winter
149 snow surveying conducted on the glacier in 1999, and thus facilitate the estimation of the
150 total snowmelt water flux to the proglacial zone at the start of the 1999 melt season. Time
151 series of discharge at points of input to and output from the proglacial channel network were
152 acquired in order to facilitate the calculation of total surface water fluxes. Points of input
153 comprised both the eastern and western ice-marginal channels at the glacier terminus
154 (hereafter Terminus East and Terminus West), while all outputs were accounted for in the
155 Outlet, where it breaches the terminal moraines (Figure 1).

156

157 **3.1 Meteorological monitoring**

158 An Automatic Weather Station (AWS) was sited in the moraine complex,
159 approximately 0.75 km north of the glacier terminus (Figure 1), and operated for a total of
160 338 days, from 13:00 on day 113 (23 April) in 1999 to 20:00 on day 85 (25 March) in 2000.
161 Air temperature and vapour pressure deficit were measured at a height of 2.0 m above the
162 ground surface with a Campbell Scientific HMP45C temperature and relative humidity probe,
163 housed within a Campbell Scientific URS1 un aspirated radiation shield. The potential error
164 range for these measurements is $\pm 2.0\%$, as specified by the instrument manufacturer. Global
165 radiation (direct and diffuse), was also measured at a height of 2.0 m with a Kipp & Zonen
166 SP-LITE pyranometer (potential error range $\pm 2.5\%$). Wind speed was measured at a height of
167 2.2 m with an RM Young 05103 Wind Monitor (potential error range $\pm 10.0\%$). Rainfall was
168 measured from 17:00 on day 175 (24 June) to 12:00 on day 229 (17 August) during the 1999
169 melt season, using a Campbell Scientific ARG100 tipping bucket rain gauge (sensitivity 0.2
170 mm of water per tip). All variables were initially measured at 20-second intervals and
171 compiled as hourly and daily means, with the exception of global radiation and rainfall,

172 which were compiled as hourly and daily totals. From 16:00 on day 229 (17 August), the
173 measuring interval was increased to 5 minutes and the compilation interval to four hours, in
174 order to minimise power consumption and data storage in advance of the winter months.

175

176 **3.2 Evaporation modelling**

177 Actual evaporation is usually estimated as a fixed percentage of potential evaporation,
178 or as a function of potential evaporation and soil moisture conditions. It may also be
179 estimated as a residual term in a water balance calculation. However, evaporation
180 measurements from Svalbard are “almost non-existent” (Killingtveit et al., 2003). For this
181 study, daily evaporation fluxes from the surface of the proglacial zone were determined by
182 summing hourly evaporation fluxes calculated using a modified version of the general
183 combination model developed by Granger and Gray (1989) for non-saturated surfaces. To
184 account for the departure from saturated conditions, this model makes use of the concept of
185 relative evaporation (the ratio of actual evaporation to potential evaporation, the latter being
186 defined as that which would occur under saturated conditions) and its relation to the relative
187 drying power of the air (the ratio of the drying power of the air to the sum of the drying
188 power of the air and total available energy from net radiation):

$$189 \quad E = (\Delta E_r G_n / \Delta E_r + \gamma) + (\gamma E_r E_a / \Delta E_r + \gamma) \quad (1)$$

190 where E is actual evaporation (mm h^{-1}), Δ is the slope of the saturation vapour pressure curve
191 ($\text{mb } ^\circ\text{C}^{-1}$) (which defines the relationship between saturation vapour pressure and air
192 temperature), γ is the psychrometric constant ($\text{mb } ^\circ\text{C}^{-1}$), E_a is the drying power of the air (mm
193 h^{-1}), G_n is the total available energy from net radiation (mm h^{-1}) and E_r is relative
194 evaporation. Full details of the method can be found in Granger and Gray (1989).

195

196 **3.3 Unmonitored atmospheric water fluxes**

197 Data concerning over-winter snow accumulation in the proglacial zone were not

198 acquired prior to the onset of the thaw associated with the 1999 melt season. However,
199 because the bulk of annual precipitation in this location is delivered as snow during the
200 winter months, it is important that this component of the total atmospheric flux is estimated.
201 Use was therefore made of snow depth and density data collected in end-of-winter surveys
202 conducted on the glacier in 1999, and on the glacier and in the proglacial zone itself in 2000.
203 Full details of how these data were collected are given in Hodgkins et al. (2005).

204 A frequently-encountered problem is that, while precipitation usually increases with
205 elevation, most precipitation gauges are located in the lowlands (Killingtveit et al., 2003).
206 Hanssen-Bauer et al. (1996) found that the ratio between true and measured precipitation at
207 various sites in Svalbard varied between 1.26 for the summer and 1.70 for the winter.
208 However, the proglacial zone considered here is a small area with a restricted elevation range
209 (about 10–50 m a.s.l.): in summer, the precipitation gauge was located within the zone
210 (Figure 1), while in winter, precipitation was measured directly by probing across the zone.
211 Therefore the problematic extrapolation of precipitation data from a distant gauge is not
212 necessary here.

213 Relationships between elevation and winter accumulation were assessed using linear
214 regression models, using input data from 106 and 75 glacier-wide locations surveyed at the
215 end of winter in 1999 and 2000 respectively, and used to estimate mean accumulation in the
216 proglacial zone. The relationship is rather stable (1999 slope = 0.002, 2000 slope = 0.003;
217 Table 1, Hodgkins et al., 2005), yielding similar accumulation estimates of 0.23 m w.e.
218 (1999) and 0.24 m w.e. (2000). The mean end-of-winter snow depth in the proglacial zone in
219 2000, based on 39 probed locations over the interval from days 103–107 (12–16 April), was
220 0.25 m w.e., which is very close to the estimated depth and gives confidence that the
221 regression model is valid over the limited elevation range of the proglacial zone. The
222 potential error range for the snow depth measurements is estimated to be $\pm 10.7\%$, as
223 determined by averaging the standard errors from multiple measurements at all 39 locations.

224 The 1999 winter accumulation regression estimate is therefore regarded as reliable.

225

226 **3.4 Discharge monitoring**

227 Channel discharge was monitored in stable reaches for a total of 55 days, from 17:00
228 on day 175 (24 June) to 12:00 on day 229 (17 August). Stage was measured at each gauge at
229 20-second intervals using a Druck PDCR1830 pressure transducer (potential error $\pm 0.1\%$)
230 and compiled as hourly means. The stage records were converted into discharge time series
231 using a rating curve (Table 1) derived from discrete velocity-area measurements using a
232 Valeport BFM001 flow meter (potential error for velocity measurements is $\pm 2.2\%$; the
233 potential error range for the distance and depth measurements is estimated to be $\pm 10\%$, due
234 mainly to turbulent flow conditions).

235 At the Terminus gauges, damage inflicted to the instrumentation by rafted ice blocks
236 and bedload, and recurrent channel migration, meant that stage records were only partly
237 rated. Although intervals of missing data are typical in discharge time series from unstable,
238 glacially-fed systems, a continuous record is required here to facilitate flux calculations.
239 Short intervals of missing data (< 12 hours) were interpolated geometrically (Synergy
240 Software, 1997). Longer intervals of missing data at the Terminus East gauge were predicted
241 from the continuous Outlet discharge record, using a linear regression model constructed
242 using hourly input terms for the period from days 187–193 (6–12 July), as this excluded non-
243 linear behaviour caused by stored water release (Wadham et al., 2001): Table 1. Longer
244 intervals of missing data at the Terminus West gauge (> 12 hours) were predicted from mass
245 conservation (Table 1): this approach assumes that the sandur runoff budget is in steady-state
246 and that inputs from water sources other than the ice-marginal channels are negligible. While
247 results presented below indicate that the latter assumption is valid, observations of
248 widespread flooding on the sandur during periods of high flow indicate that the former
249 assumption is likely to be invalid over shorter timescales (Cooper, 2003; Cooper et al., 2002).

250 3.5 Unmonitored surface water fluxes

251 Monitoring of channel discharge and glacial ablation commenced some time after the
252 onset of the thaw associated with the 1999 melt season and ceased some time before the
253 annual freeze-up. While surface water fluxes during these missed periods are likely to have
254 been relatively small, the aims of assessing the full annual hydrological cycle and of fully
255 quantifying all fluxes require that they be estimated.

256 Ablation on the glacier terminus was monitored for a total of 51 days, from days 178–
257 228 (27 June–16 August), by measuring surface lowering at an aluminium stake every 2–4
258 days. The potential error range for ablation measurements is estimated to be $\pm 10\%$, due
259 mainly to uncertainty caused by surface roughness. Nevertheless, there are not enough data to
260 model melt throughout the year at the scale of the entire glacier, so the approach taken was to
261 find a relationship between melt on the terminus and runoff at the Outlet. A temperature-
262 index model of melt on the glacier terminus was therefore developed, in order to estimate
263 melt outside the monitoring period. Temperature-index melt models, which are based on
264 empirical relationships between air temperature and ablation, have been widely applied in
265 glacial environments and have proven to be powerful tools, despite their relative simplicity
266 (Hock, 2003). In this instance, we used the model form

$$267 \quad Abl = \begin{cases} f(T_a - T_0), & T_a > T_0 \\ 0, & T_a \leq T_0 \end{cases} \quad (2)$$

269 where Abl is specific melt (mm w.e.), f is a derived melt factor (the mean of individual
270 ablation measurements divided by mean air temperature since the last measurement, mm w.e.
271 $^{\circ}\text{C}^{-1}$), T_a is mean air temperature ($^{\circ}\text{C}$) and T_0 is a threshold temperature beyond which melt is
272 assumed to occur (in this case, 0°C). An advantage of using this model form was the ease
273 with which it enabled irregular time intervals in the measured ablation data to be simulated.
274 The mean derived melt factor, $6.8 \text{ mm w.e. } ^{\circ}\text{C d}^{-1}$ (range $2.5\text{--}11 \text{ mm w.e. } ^{\circ}\text{C d}^{-1}$) is
275 consistent with other reported degree-day factors for ice, which are in the range $5.4\text{--}20 \text{ mm}$

276 w.e. °C d⁻¹ (Hock, 2003). The total modelled melt for the period of monitoring, 1722 mm,
277 compares with the observed value of 1684 mm: a 3% difference. Linear regression of
278 modelled melt on observed melt yields a highly significant ($p<0.001$) relationship with an R^2
279 value of 0.66. Model error, based on the Root-Mean-Squared Error (RMSE) expressed as a
280 percentage of the mean observed melt, is $\pm 21.5\%$. The performance of the melt model is
281 therefore satisfactory, and daily values of ablation for the entire 1999 melt season were
282 modelled using mean daily air temperature as the input series.

283 Daily discharge fluxes at the Outlet were then estimated from daily values of
284 modelled ablation, using a regression relationship for the whole of the period during which
285 ablation was monitored. The delay between ablation on the glacier terminus and flow
286 response in the Outlet was accounted for by lagging the discharge series by 1 day. Linear
287 regression of Outlet discharge on modelled melt yields a highly significant ($p<0.001$)
288 relationship with an R^2 value of 0.74. The total modelled Outlet flux for the period of
289 monitoring, $4.94 \times 10^7 \text{ m}^3$, compares with the observed value of $4.79 \times 10^7 \text{ m}^3$: a 3% difference.
290 Model error (RMSE as a percentage of mean observed melt) is $\pm 36.2\%$. The performance of
291 the discharge regression model is therefore satisfactory, and it has been used to estimate
292 unmonitored surface water fluxes.

293

294 **4. Results: atmospheric and surface water fluxes**

295 **4.1 Temporal variation in meteorology**

296 Time series of mean daily air temperature, vapour pressure deficit, wind speed and
297 total daily global radiation and rainfall are presented in Figure 2. Mean daily air temperatures
298 were consecutively positive (mean 4.7° C, range 0.5–10.4° C) during the period from days
299 156–262 (5 June–19 September), giving a duration for the 1999 melt season of 107 days.
300 Outside this period, mean daily air temperatures were significantly lower (mean –5.7° C) and
301 more variable (range –23.0 to 3.8° C). Mean daily vapour pressure deficits and mean daily

302 wind speeds were high and variable during the melt season, especially on days when global
303 radiation totals and mean air temperatures were high. Outside this period, both mean daily
304 vapour pressure deficits and mean daily wind speeds were lower and less variable, reflecting
305 more stable atmospheric conditions resulting from reduced daily global radiation totals and
306 lower mean daily air temperatures. Daily global radiation totals were high and relatively
307 invariable at the start of the period of monitoring, reflecting frequent clear-sky conditions and
308 the receipt of significant quantities of diffuse radiation reflected by the snowpack. Daily
309 global radiation totals gradually declined to zero with the onset of the polar night on day 296
310 (23 October). Rainfall was recorded on 15 days during the period from days 176–228 (25
311 June–16 August), although significant amounts only fell on three of these, with a maximum
312 daily total of 8.0 mm (on day 226, 14 August).

313 Killingtonveit et al. (2003) state that in Arctic catchments the summer potential
314 evaporation may be very significant, due to high net radiation during days with 24 hours'
315 sunlight, although actual evaporation may still be small as there is little vegetation, little
316 rainfall and soils tend to dry up easily after snowmelt. Time series of total daily available
317 energy and evaporation from the surface of the proglacial zone are presented in Figure 3.
318 Daily evaporation totals were low at the start of the period of monitoring, reflecting the high
319 albedo of the snowpack and the limited available energy. Following snowpack recession,
320 daily evaporation totals became higher and more variable and remained so for the duration of
321 the melt season. The maximum daily evaporation total was 4.4 mm on day 190 (9 July).
322 Variability was largely driven by global radiation during this time interval, although rainfall
323 and low mean daily vapour pressure deficits associated with the passage of maritime air
324 masses from the south-west were intermittently significant. The mean daily total evaporation
325 in the period from the start of June to the end of August was 1.4 mm. In the period between
326 the end of the melt season and the onset of the polar night, daily evaporation totals declined
327 rapidly to zero ahead of the gradual decline in daily global radiation totals, reflecting

328 increasing long-wave radiative losses and the progressive cooling of the land surface during
329 the annual freeze-up. Evaporation effectively stopped (daily totals ≤ 0.1 mm) on day 249 (6
330 September).

331

332 **4.2 Daily and cumulative atmospheric water fluxes**

333 Daily total rainfall and evaporation are presented in Figures 2(E) and 3(B),
334 respectively. Since rainfall was only monitored in the 53-day period from day 176–228 (25
335 June–16 August), it is possible that a significant proportion of total annual rainfall was
336 missed. Since there is no way of realistically predicting missed rainfall totals, the cumulative
337 total recorded is considered to be a minimum estimate of total rainfall in 1999: the
338 cumulative total of 29.4 mm equates to a total cumulative atmospheric water flux of 1.26×10^5
339 m^3 (assuming spatial representativeness). The highest daily rainfall flux (3.44×10^4 m^3 on day
340 226 (14 August]) accounted for about 27% of the total cumulative atmospheric water flux.

341 Daily evaporation totals were determined for the entire period of monitoring, which
342 included almost all of the period in which evaporation could occur. The total determined
343 during the period of monitoring (141 mm) is therefore considered to equate to total annual
344 evaporation in 1999, which in turn equates to a total cumulative atmospheric water flux from
345 the proglacial zone of 6.08×10^5 m^3 (assuming spatial representativeness). The highest daily
346 evaporation total (4.4 mm on day 190 [9 July]) corresponds to a total daily atmospheric water
347 flux of 1.89×10^4 m^3 (about 3% of the total flux). During the 53-day period in which rainfall
348 was monitored, a cumulative total of 98.7 mm of evaporation was recorded, which exceeds
349 the cumulative rainfall total by about 240%. The estimated water-equivalent snow
350 accumulation at the end of the 1999 winter season (226 mm) equates to a total atmospheric
351 water flux to the proglacial zone of 9.70×10^5 m^3 . Total annual precipitation (255 mm)
352 therefore exceeded total annual evaporation (141 mm) by about 81% in 1999.

353 Various sources of potential error have been identified concerning the determination
354 of atmospheric water fluxes to and from the proglacial zone, including those associated with
355 the use of instrumentation, field techniques and empirical equations. With regard to the
356 rainfall water flux, the main sources of potential error relate to the catch efficiency of the rain
357 gauge and the fact that rainfall was only monitored for about 50% of the melt season: in the
358 former case, it is reasonable to assume that the catch efficiency of the rain gauge was >95%,
359 given that the mean wind speed on days when rain fell during the period of monitoring was
360 low (2.71 m s^{-1}) (Bruce & Clark, 1990); in the latter case, it is conceivable that the annual
361 rainfall total may have been about 100% greater than the monitored total, but given that the
362 bulk of total annual precipitation in Svalbard comprises snowfall (Hanssen-Bauer *et al.*,
363 1990), even a doubling of the rainfall total in this instance only generates a potential error
364 range of $\pm 20.2\%$ for total annual precipitation. Since these various sources of potential error
365 are multidirectional and therefore non-additive, realistic estimates of error may be determined
366 by combining all of the potential errors probabilistically as the root of the sum of the squares
367 of individual error sources (Topping, 1972). This approach gives a probable error for the
368 rainfall water flux in the period of monitoring of $\pm 20.8\%$.

369 With regard to the snowpack water-equivalent flux, the main sources of potential
370 error relate to the high standard error associated with the snow-depth measurements and the
371 fact that the snow depth was estimated from a regression on elevation. However, repeated
372 measurements in 1999 and 2000 showed that it was the spatial variation of accumulation
373 which contributed by far the most to overall error, being greater, for instance, than inter-
374 annual variability (Hodgkins *et al.*, 2005). Killingtveit *et al.* (2003) make the same point in
375 suggesting that residual errors in water balance calculations are probably related mainly to
376 problems of precipitation correction. The probable error range for the snowpack flux may
377 therefore be estimated by combining the standard errors associated with the snow depth and
378 density measurements probabilistically, giving $\pm 43.7\%$ (the greatest proportional uncertainty

379 of all the fluxes). With regard to the evaporation flux, the main potential sources of error
380 relate to the use of instrumentation and empirical equations, the latter of which include
381 potential errors associated with assumed values and approximations, in addition to an overall
382 standard error. Probable errors for the evaporation flux are summarised in Table 2.

383

384 **4.3 Temporal variation in runoff**

385 Time series of discharge at points of input to and output from the proglacial channel
386 network are presented in Figure 4. The seasonal pattern of discharge was characterised by
387 two periods of high and variable flow interspersed with periods of low and relatively
388 invariable flow. The first period of high flow occurred during the first 11 days of monitoring,
389 over days 175–186 (24 June–5 July). During the first 4 days of this period, peak daily
390 discharge at the Outlet rose rapidly from $<5 \text{ m}^3 \text{ s}^{-1}$ to $>33 \text{ m}^3 \text{ s}^{-1}$ and diurnal cycling became
391 evident. Weather conditions during this period were warm and windy (Figure 2) and the
392 snowline was observed to retreat rapidly up the lower reaches of the glacier. Localised
393 flooding was observed on the sandur in the time interval from days 179–180 (28–29 June).
394 The second period of high flow occurred during the middle of the melt season, over days
395 195–207 (14–26 July). During the first 4 days of this period, weather conditions were again
396 warm and windy and peak daily discharge at the Outlet rose dramatically from $<11 \text{ m}^3 \text{ s}^{-1}$ to
397 its seasonal maximum of $>60 \text{ m}^3 \text{ s}^{-1}$ at 20:00 on day 199 (18 July): massive bank erosion was
398 observed at the Terminus West gauge on this day, along with large numbers of rafted ice
399 blocks and enhanced turbidity (Hodgkins et al., 2003). Wadham et al. (2001) have described
400 the occurrence of seasonal outburst floods from Finsterwalderbreen, with the hydrochemical
401 signature of released waters suggesting a subglacial origin. Widespread flooding was
402 observed on the sandur during the time interval from days 199–202 (18–21 July). In contrast
403 to these high-magnitude events, flow in the intervening periods was relatively low and
404 invariable, save for increasingly well-defined diurnal cycling.

405 **4.4 Daily and cumulative surface water fluxes**

406 Total daily glacier ablation (measured and modelled) and total daily surface water
407 flux from the proglacial zone (measured and estimated) are presented in Figure 5. A total
408 cumulative water flux of $4.79 \times 10^7 \text{ m}^3$ was discharged at the Outlet gauge during the 53-day
409 period from days 176–228 (25 June–16 August). During the same time interval, a combined
410 total cumulative water flux of $4.88 \times 10^7 \text{ m}^3$ was discharged at the Terminus West and
411 Terminus East gauges (about 64% at the West and 36% at the East). The difference between
412 the total cumulative water fluxes at the Terminus and Outlet gauges of $0.09 \times 10^7 \text{ m}^3$ (about
413 2% of the total cumulative flux at the Outlet) falls well within the probable error range for
414 discharge determined at all three gauges (Table 3), so is not regarded as significant. During
415 the 51-day period from days 178–228 (27 June–16 August), a cumulative total of 1.68 m w.e.
416 ablation was recorded on the glacier terminus. Some 0.6 km^2 of the terminus was drained by
417 supraglacial channels flowing directly to the proglacial channel network, i.e. water which was
418 not accounted for at the Terminus West or East gauges. This gives a total cumulative
419 supraglacial water flux of $1.01 \times 10^6 \text{ m}^3$: only 2% of the combined flux at the Terminus gauges
420 during the same time interval. During the first period of high flow, a total water flux of
421 $1.15 \times 10^7 \text{ m}^3$ (about 24% of the total cumulative flux) was discharged at the Outlet. During
422 the second period of high flow, a total water flux of $2.16 \times 10^7 \text{ m}^3$ (about 45% of the total
423 cumulative flux in less than 23% of the monitoring period) was discharged at the Outlet. Of
424 this total, $3.91 \times 10^6 \text{ m}^3$ (about 8% of the total cumulative flux) was discharged in one day
425 (day 199, 18 July, <2% of the monitoring period), when discharge at the Outlet attained its
426 seasonal maximum.

427 A total estimated water flux of $2.42 \times 10^7 \text{ m}^3$ was discharged at the Outlet outside the
428 period of monitoring (Figure 5). Summing the combined unmonitored water fluxes and the
429 cumulative water flux at the Outlet during the period of monitoring gives a total annual flux

430 of $7.21 \times 10^7 \text{ m}^3$. Of this total, about 9% was discharged prior to the onset of monitoring, about
431 66% was discharged during the period of monitoring and about 25% was discharged after its
432 cessation. It is likely that predicted peaks and troughs in runoff outside the period of
433 monitoring are over- and under-estimates respectively, as they cannot take into account the
434 variable, modulating effect of meltwater storage, and therefore that flow in the missed
435 periods was relatively low and invariable.

436 Numerous sources of potential error have been identified concerning the acquisition
437 of time series of discharge in both melt seasons, including those associated with the use of
438 instrumentation, field techniques and statistical procedures. Probable errors for the surface
439 water flux are summarised in Table 3. There is unquantifiable error associated with the melt
440 model (aside from the error determined by comparison with observed melt) from the use of a
441 melt factor derived from ablation measurements from an ice surface: this likely over-predicts
442 melt early in the summer when the glacier is still snow-covered. On the other hand, only 9%
443 of total runoff is estimated to occur in the period prior to the commencement of monitoring,
444 so there is unlikely to be a significant impact on the magnitude of the calculated flux. The
445 melt factor should be appropriate for the period after the cessation of monitoring, when
446 glacier ice is exposed.

447

448 **5. Synthesis: the annual proglacial hydrological regime at Finsterwalderbreen**

449 The data presented above provide insights into the nature of the annual proglacial
450 hydrological regime at Finsterwalderbreen, particularly in terms of the significance of diverse
451 hydrological pathways in both space and time. The following discussion synthesises these
452 data with field observations and results from previous studies to produce a qualitative
453 framework outlining the principal variations in surface hydrology in the proglacial zone of
454 Finsterwalderbreen over the course of an annual cycle. This framework can be used as a
455 context for understanding fluvial material fluxes from this and similar catchments, and as a

456 basis for assessing the effects of climate change on local and regional hydrological regimes.

457 **5.1 Winter (December–March)**

458 Winter is essentially a dormant phase in the annual proglacial hydrological cycle.
459 During the winter months, when mean air temperatures are at an annual minimum and
460 continuous darkness prevails, the ground surface in the proglacial zone remains entirely
461 frozen and blanketed by snow cover down to sea level. The only significant hydrological
462 events are the development of an icing in the area where the western ice-marginal channel
463 issues into the proglacial channel network, and the receipt of significant quantities of
464 snowfall, typically in mid-to-late winter (February–March). The development of the icing
465 occurs as a result of the freezing of subglacial drainage, which issues throughout the winter
466 from an artesian upwelling situated at the glacier terminus (Wadham *et al.*, 2000). Some
467 additional water is supplied to the icing by snowmelt on the glacier terminus during warmer
468 periods associated with the passage of maritime air masses (Wadham *et al.*, 2000). Ice layers
469 form beneath and within the proglacial snowpack during such periods, as snowmelt
470 percolates downwards and then refreezes. Sublimation may occur, but to what extent is
471 unknown: some estimates suggest that up to about 20 % of winter snowfall may be lost by
472 sublimation in situations where air humidity is very low (French, 2007). The return of
473 daylight in mid-February following the cessation of the polar night has little initial impact on
474 hydrological activity in the proglacial zone, since the progressive increase in the receipt of
475 energy inputs from solar radiation is very gradual and largely offset by the high albedo of the
476 snowpack.

477

478 **5.2 Spring (April–May)**

479 Spring is a phase of increasing activity in the annual proglacial hydrological cycle.
480 Although mean air temperatures remain low, the onset of continuous daylight in mid-April
481 results in a further gradual increase in the receipt of energy inputs from solar radiation as the

482 sun climbs progressively higher in the sky. While the high albedo of the snowpack continues
483 to offset much of the increase in energy inputs from solar radiation, the potential for
484 snowmelt gradually increases, especially during warmer periods associated with the passage
485 of maritime air masses. Ice layers continue to form beneath and within the proglacial
486 snowpack during such periods. As the sun climbs progressively higher and mean daily air
487 temperatures begin periodically to rise above zero in late May, snowmelt becomes more
488 sustained. However, the onset of runoff is delayed as significant volumes of percolating
489 snowmelt are temporarily stored in the snowpack. The presence of basal ice layers beneath
490 the snowpack prevents percolating snowmelt from accessing the underlying frozen ground
491 surface and may increase lateral flow velocities within the snowpack (Fountain, 1996).

492

493 **5.3 Summer (June–September)**

494 Summer is the most active phase in the annual proglacial hydrological cycle and
495 encompasses the annual melt season (duration about 100 days), during which time mean air
496 temperatures attain an annual maximum and >99% of total annual runoff occurs. During the
497 annual thaw, snowmelt is almost entirely radiation-driven (e.g. Hodgkins, 2001; Hodson et
498 al., 2005), since the controlling influence of the temperature of the snow-covered land surface
499 restricts the sensible heat flux into the snowpack (Nakabayashi *et al.*, 1996; Harding &
500 Lloyd, 1997). Melt rates are thus largely similar from year-to-year and the date of the final
501 disappearance of the snowpack and the length of the snow-free period are determined largely
502 by the depth of over-winter snow accumulation at the onset of the melt season (Nakabayashi
503 *et al.*, 1996; Harding & Lloyd, 1997).

504 Runoff usually commences in early June, when the onset of consistently positive
505 mean daily air temperatures triggers rapid and sustained melting of the snowpack. Snowmelt
506 is initially routed laterally within the basal layers of the snowpack to either frozen lake
507 surfaces situated in topographic depressions in the moraine complex or to the surface of the

508 sandur. Increasing inputs of snowmelt from the lower reaches of the main valley glacier
509 initiate the onset of continuous flow in the proglacial channel network by mid-June. By this
510 time, the lake network in the moraine complex is largely thawed and lake levels are high,
511 having been recharged by the receipt of significant volumes of snowmelt. Excess snowmelt is
512 subsequently conveyed as surface runoff in small, ephemeral channels that drain along a
513 topographic gradient from the interior of the moraine complex to the sandur.

514 The rate of evaporation begins to assume significance following the recession of the
515 snowpack, reflecting the increase in air temperatures and the abundance of surface water
516 available for evaporation. However, as the melt season proceeds, the rate of evaporation
517 gradually declines, reflecting the progressive drying out of the ground surface. Nevertheless,
518 evaporation may exceed precipitation by as much as 235% over the course of the melt season,
519 indicating that water storage in the active layer is sufficient to maintain evaporation during
520 dry periods. Harding & Lloyd (1997) similarly found that total evaporation at elevations <50
521 m a.s.l. in Svalbard may exceed precipitation by about 160% during the summer.

522 The first period of significant flow in the proglacial channel network occurs in either
523 late June or early July, in response to sustained melting of the snowpack on the lower reaches
524 of the glacier. By this time, the annual melting of the icing is also usually well under way.
525 However, if prevailing weather conditions are cold and cloudy, the onset of sustained melting
526 of the snowpack and the icing may be delayed until mid-to-late July. Peak flow in the
527 proglacial channel network typically occurs during periods of good weather in mid-to-late
528 July, in response to high rates of ablation on the lower reaches of the glacier. The oversupply
529 of significant volumes of meltwater to the subglacial drainage system during such periods has
530 the potential to trigger subglacial outburst floods, which issue from the glacier terminus and
531 cause the sandur to become flooded for several days (Wadham *et al.*, 2001; Cooper, 2003).

532 Killingtveit *et al.* (2003) note that runoff is dominated by snowmelt in June and July
533 in Svalbard, while in August and September it is mainly derived from rainfall and glacial

534 melt: catchments with higher proportion of icemelt-supplying glacier cover tend to have
535 relatively higher runoff in August and September than non-glacierized catchments. The return
536 of nights in late August as the sun descends progressively lower in the sky results in a gradual
537 decline in energy inputs from solar radiation throughout the remainder of the summer and a
538 gradual decline in hydrological activity in the proglacial zone. Flow recession occurs in the
539 proglacial channel network, reflecting reduced meltwater inputs from ablation on the glacier,
540 and the rate of evaporation falls sharply, reflecting an increasing excess of outgoing long-
541 wave radiation over incoming short-wave radiation and a progressive cooling of the ground
542 surface prior to the annual freeze-up.

543

544 **5.4 Autumn (October–November)**

545 Autumn is a phase of decreasing activity in the annual proglacial hydrological cycle.
546 The cessation of continuous flow in the proglacial channel network probably occurs in early-
547 to-mid October, when air temperatures begin to fall consistently below zero and the nights
548 become progressively longer. The annual formation of the icing in the area where the western
549 ice-marginal channel issues into the proglacial channel network is initiated by the freezing of
550 subglacial drainage, which continues throughout the autumn months from the artesian
551 upwelling. By the start of November, permanent darkness again prevails and hydrological
552 activity in the proglacial zone essentially becomes dormant, save for inputs of snowfall and
553 the development of the icing.

554

555 **6. Conclusions**

556 Research in high-latitude hydrology remains challenging. Hydrological research
557 infrastructure in high-latitude catchments remains very limited, and the extreme seasonality
558 reduces the utility of many standard techniques, e.g. even where weir structures have been
559 built, they typically fail to capture early-season runoff adequately because of snow- and ice-

560 blocking of channels (e.g. Sund, 2008). Significant challenges persist in measuring
561 precipitation reliably and representatively; this not only hinders process analysis and water
562 resources management, but also makes climate change detection difficult (e.g. Førland and
563 Hanssen-Bauer, 2003). Measuring and monitoring the discharge of even moderately-sized,
564 glacially-fed rivers is a demanding task because of the temporal and spatial instability of their
565 flow regimes, particularly if continuous, complete time series are required, as exemplified by
566 the work presented here.

567 Research into the surface and atmospheric water fluxes of the proglacial zone of
568 Finsterwalderbreen in 1999 has indicated that winter dominates the delivery of precipitation
569 (226 mm a^{-1}), while summer is rather dry (29 mm a^{-1}). Other parts of Svalbard, such as the
570 north-west, are wetter in summer, to the extent that summer precipitation may occasionally
571 exceed the winter total (Hodson et al., 2005). Measurements and even estimates of
572 evaporation are uncommon in the high latitudes, but a modified version of the general
573 combination model for non-saturated surfaces (Granger & Gray, 1989) worked well in this
574 case, given the availability of a range of meteorological variables, including vapour pressure
575 deficit, wind speed and global radiation, as well as air temperature. Total annual precipitation
576 (255 mm) exceeded total annual evaporation (141 mm) by about 81%, although evaporation
577 can be appreciably in excess of precipitation during the dry summer, indicating short-term
578 water storage, probably in the active layer.

579 Monitoring of proglacial discharge for 53 days between late-June and mid-August is
580 estimated to have captured 66% of total annual runoff: fluxes outside the monitoring period
581 could be simulated by combining a temperature-index melt model with a runoff regression
582 relationship, an approach which performed well for the instrumental period and was
583 parsimonious in terms of data requirements. 47% of total annual runoff occurred in 23 days
584 encompassing two high-flow periods in early- to mid-summer, so even within the *c.* 100-day
585 melt season, runoff is temporally-concentrated. Despite episodes of flooding, likely leading to

586 short-term water storage, the total annual input and output of runoff to and from the
587 proglacial zone were not significantly different at the timescale of the melt season, indicating
588 that storage is unlikely to have been important overall, or from season to season.

589 A range of uncertainties are associated with the fluxes derived by measurement,
590 modelling and estimation in this paper, but these can be rigorously quantified
591 probabilistically. The greatest flux uncertainties are associated with runoff, because it is at
592 least an order-of-magnitude higher than any other flux, e.g. total glacial runoff delivered to
593 the sandur is $7.38 \times 10^7 \text{ m}^3 \text{ a}^{-1}$, whereas the total precipitation flux to the sandur is $2.29 \times 10^5 \text{ m}^3$
594 a^{-1} . However, the proportional uncertainty associated with precipitation is two-and-a-half
595 times that associated with runoff, mainly as a result of the considerable spatial heterogeneity
596 in winter accumulation.

597 The synthesised description of the seasonal hydrological cycle presented above can be
598 viewed as a set of hydrological boundary conditions, serving as a context for understanding
599 fluvial material fluxes from this and similar catchments, and as a basis for assessing the
600 effects of climate change on local and regional hydrological regimes. A companion paper to
601 this one will look at the sub-surface fluxes in the Finsterwalderbreen catchment, and at the
602 complete annual water budget.

603

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609

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715 Finsterwalderbreen, southern Spitsbergen, Svalbard. *Annals of Glaciology*, 24, 262–
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717

718 **Figure captions**

719 Fig. 1. (Clockwise from top left) Location of Finsterwalderbreen within the Svalbard
720 archipelago (inset). Topographic map of the glacier terminus and proglacial area, elevations
721 in m a.s.l. (Fox, 1995). 1995 aerial photograph of the glacier terminus and proglacial area
722 (subset of aerial photograph S95 1113© Norwegian Polar Institute): discharge monitoring
723 locations are indicated (note that many stream courses apparent on the map and photograph,
724 e.g. X, are not currently active, and that all of the runoff from the catchment is channelled
725 through the outlet); the locations of the Automatic Weather Station (AWS) and a Wells
726 Transect (WT), used to monitor sub-surface water fluxes, are also indicated. Upstream views
727 of the Outlet on 24 June (discharge ca. $5 \text{ m}^3 \text{ s}^{-1}$) and 21 July (discharge ca. $25 \text{ m}^3 \text{ s}^{-1}$) 1999;
728 the lighter colour of the runoff on 21 July is a result of the angle of the sun, rather than lower
729 turbidity. High-elevation view of the Finsterwalderbreen proglacial area looking northeast,
730 showing discharge monitoring locations (although the East gauge is just off the right of the
731 image).

732

733 Figure 2. Meteorological time-series from the AWS located in the proglacial zone. (A) Air
734 temperature ($^{\circ} \text{C}$). (B) Vapour pressure deficit (kPa). (C) Wind Speed (m s^{-1}). (D) Global
735 radiation (W m^{-2}). (E). Precipitation as rainfall (mm). Days of year 110–330 correspond to 20
736 April–26 November.

737

738 Figure 3. Time series of daily (A) radiation flux (MJ m^{-2} ; NSW, NLWR and NAWR are
739 Net Short-Wave, Net Long-Wave and Net All-Wave Radiation Fluxes, respectively) and (B)
740 evaporation from the surface of the proglacial zone (mm). Days of year 110–330 correspond
741 to 20 April–26 November.

742

743

744 Figure 4. Time series of (A) stage (m) and (B) discharge ($\text{m}^3 \text{s}^{-1}$) at each of the gauges.
745 Terminus West and East constitute inputs to the proglacial channel network, and Outlet
746 constitutes output. Days of year 175–230 correspond to 24 June–18 August.

747

748 Figure 5. Time series of daily (A) measured and modelled glacier ablation (mm w.e.) and (B)
749 measured and estimated (from the regression of Outlet discharge on modelled melt) total
750 surface water flux from the proglacial zone (m^3). Days of year 110–330 correspond to 20
751 April–26 November.

752

753 **Tables**

Gauge	Compilation Period	Equation	<i>n</i>	R^2	<i>s.e.</i> (%)
Outlet	17:00 175–12:00 229	$Q_O = 1.43e^{3.23S}$	13	0.99	4.4
Terminus West	17:00 183–12:00 187	$Q_W = 19.5S + 1.28$	2	-	-
Terminus West	16:00 191–12:00 195	$Q_W = 5.76S + 2.57$	2	-	-
Terminus West (interpolation) ^a	17:00 175–16:00 183 13:00 187–15:00 191 13:00 195–10:00 229	$Q_W = Q_O - Q_E$	-	-	-
Terminus East	16:00 183–15:00 195	$Q_E = 16.4S + 1.16$	9	0.98	12.5
Terminus East (interpolation) ^a	17:00 175–15:00 183; 16:00 195–10:00 229	$Q_E = 0.30Q_O + 0.67$	-	0.71	9.4

754

755 Table 1. Summary of rating curves compiled at each gauge during the 1999 discharge
756 monitoring period, together with interpolations used when adequate ratings could not be
757 determined. Q is discharge ($\text{m}^3 \text{s}^{-1}$) and subscripts O , W and E denote discharge at the Outlet,
758 Terminus West and Terminus East gauges, respectively. S is stage (m), n is the number of
759 discharge measurements, R^2 is the coefficient of determination and *s.e.* is the standard error
760 of the regression. ^a Q_O has been lagged by 1 hour to account for the mean transit time between
761 the glacier terminus and the Outlet.

762

763

Period	E_T	E_{WS}	E_{VPD}	E_G	E_{AV}	E_{EM}	E_P
13:00 113 1999–20:00 85 2000	2.0 ⁽³⁾	10.0	2.0	2.5	10.0 ⁽⁶⁾	7.2	27.8

764

765 Table 2. Summary of probable errors in the calculation of evaporation water flux from the
766 proglacial zone. Potential errors ($\pm\%$) are associated with the measurement of air temperature
767 (E_T), wind speed (E_{WS}), vapour pressure deficit (E_{VPD}) and global radiation (E_G); other
768 potential errors are associated with the use of assumed values (E_{AV}) and the standard error
769 from empirical models (E_{EM}). E_P is the probable error range of calculated evaporation values
770 for the specified time period. Figures in brackets denote the number of times each potential
771 source of error arose during calculations.

772

Gauge	Period	E_S	E_V	E_D	E_{RC}	E_{RM}	E_P
Outlet	17:00 175–10:00 229	0.1	2.2	10.0	4.4	-	11.1
Terminus West	17:00 175–16:00 183	-	-	-	-	-	18.4 ^a
Terminus West	17:00 183–12:00 187	0.1	2.2	10.0	10.0 ^b	-	14.3
Terminus West	13:00 187–15:00 191	-	-	-	-	-	16.4 ^a
Terminus West	16:00 191–12:00 195	0.1	2.2	10.0	10.0 ^b	-	14.3
Terminus West	13:00 195–10:00 229	-	-	-	-	-	18.4 ^a
Terminus East	17:00 175–15:00 183	-	-	-	-	9.4	14.6 ^c
Terminus East	16:00 183–15:00 195	0.1	2.2	10.0	6.4	-	12.1
Terminus East	16:00 195–10:00 229	-	-	-	-	9.4	14.6 ³

774

775 Table 3. Summary of probable errors in the time series of discharge at each gauge during the
776 1999 monitoring period. Potential errors ($\pm\%$) are associated with the measurement of stage
777 (E_S), the measurement of flow velocity (E_V), the measurement of channel depth (E_D), the
778 relevant rating curve (E_{RC}), the regression used for interpolation (Table 1)(E_{RM}); E_P is the
779 probable error range of discharge data for the specified time period. ^aindicates value
780 determined by probabilistically combining relevant values of E_P at the Outlet and Terminus
781 East, ^bindicates value is a maximum estimate and ^cindicates value determined by
782 probabilistically combining value of E_{RM} with relevant value of E_P at the Outlet.

Figure 1
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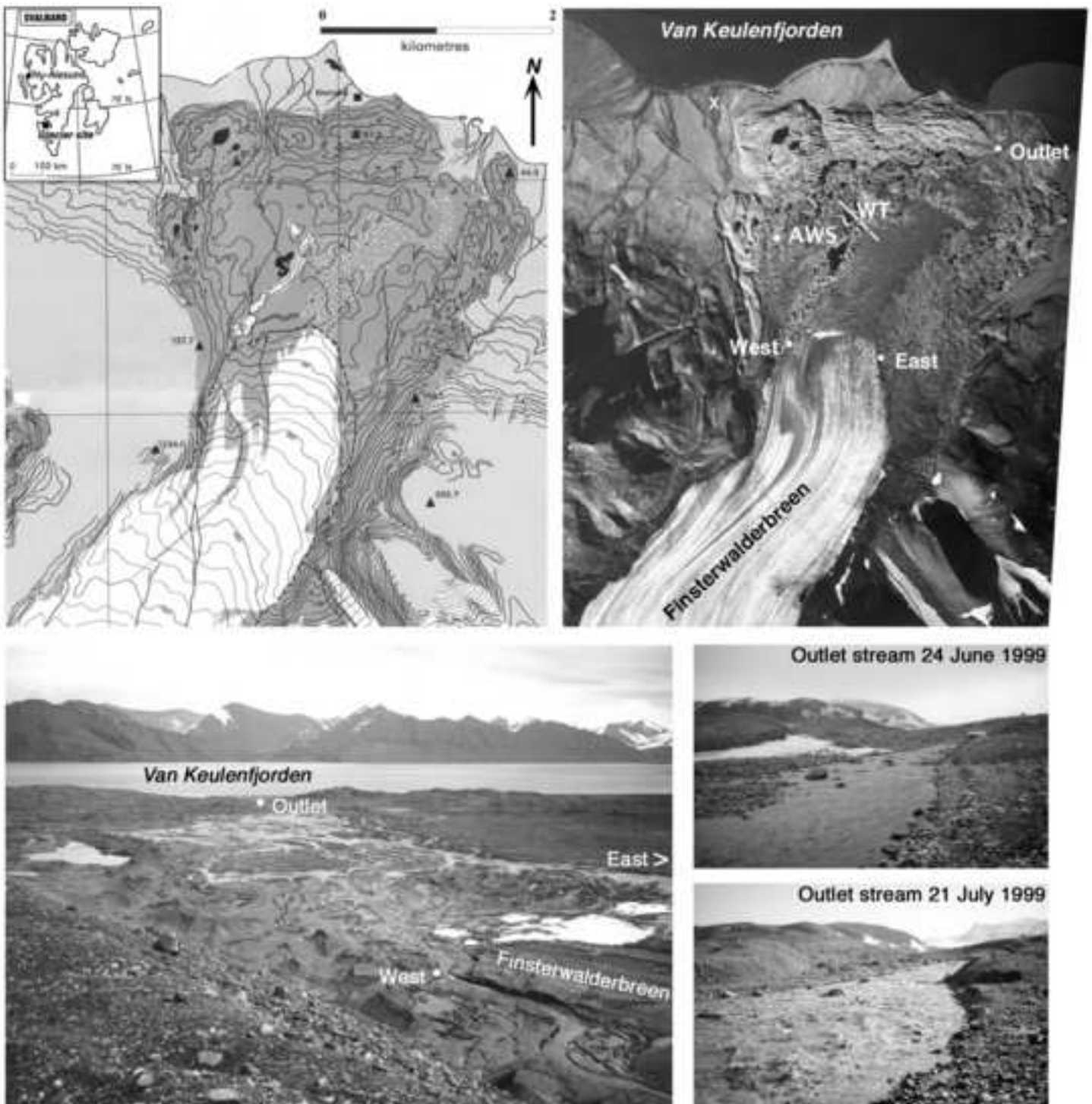


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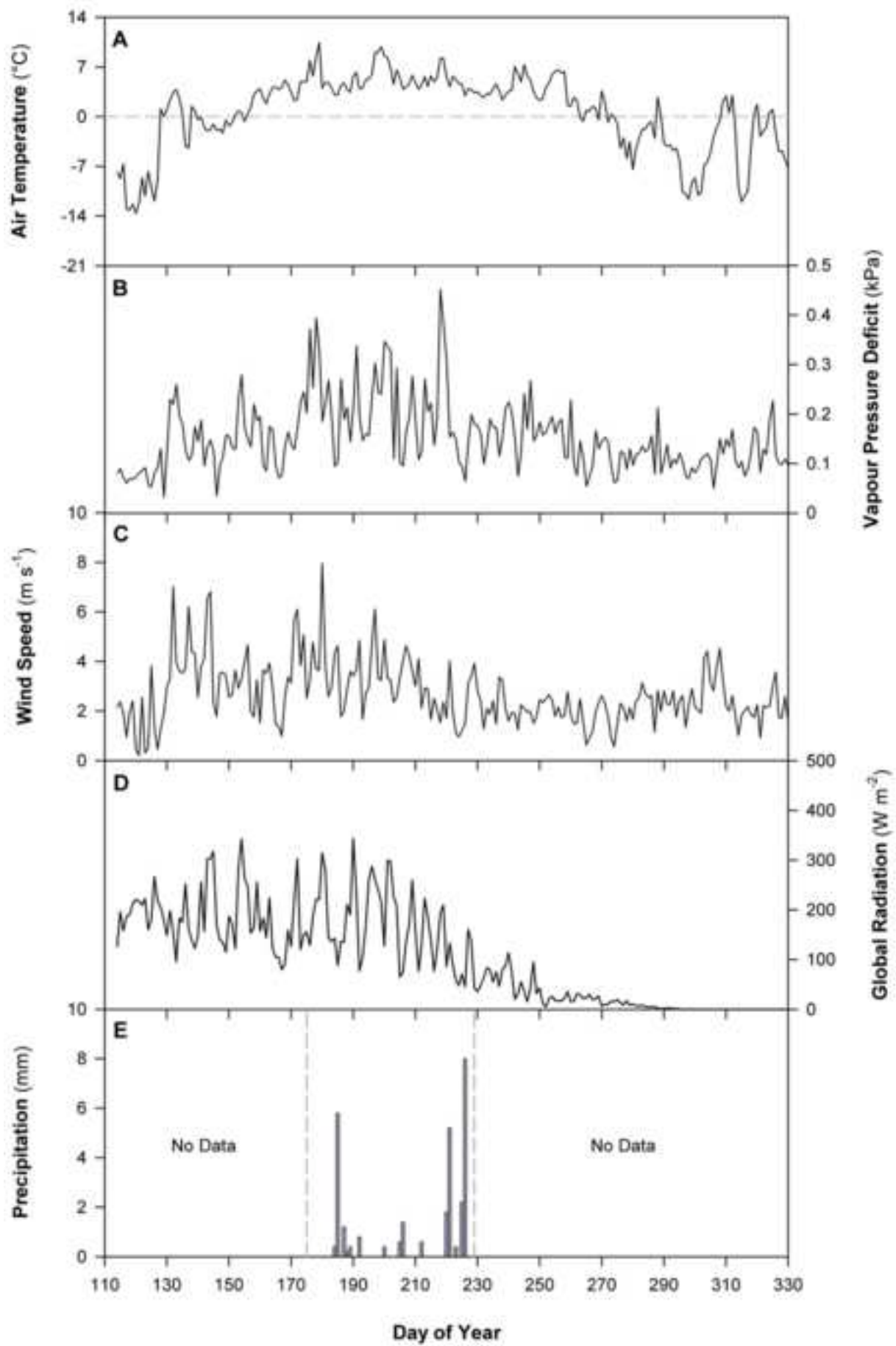


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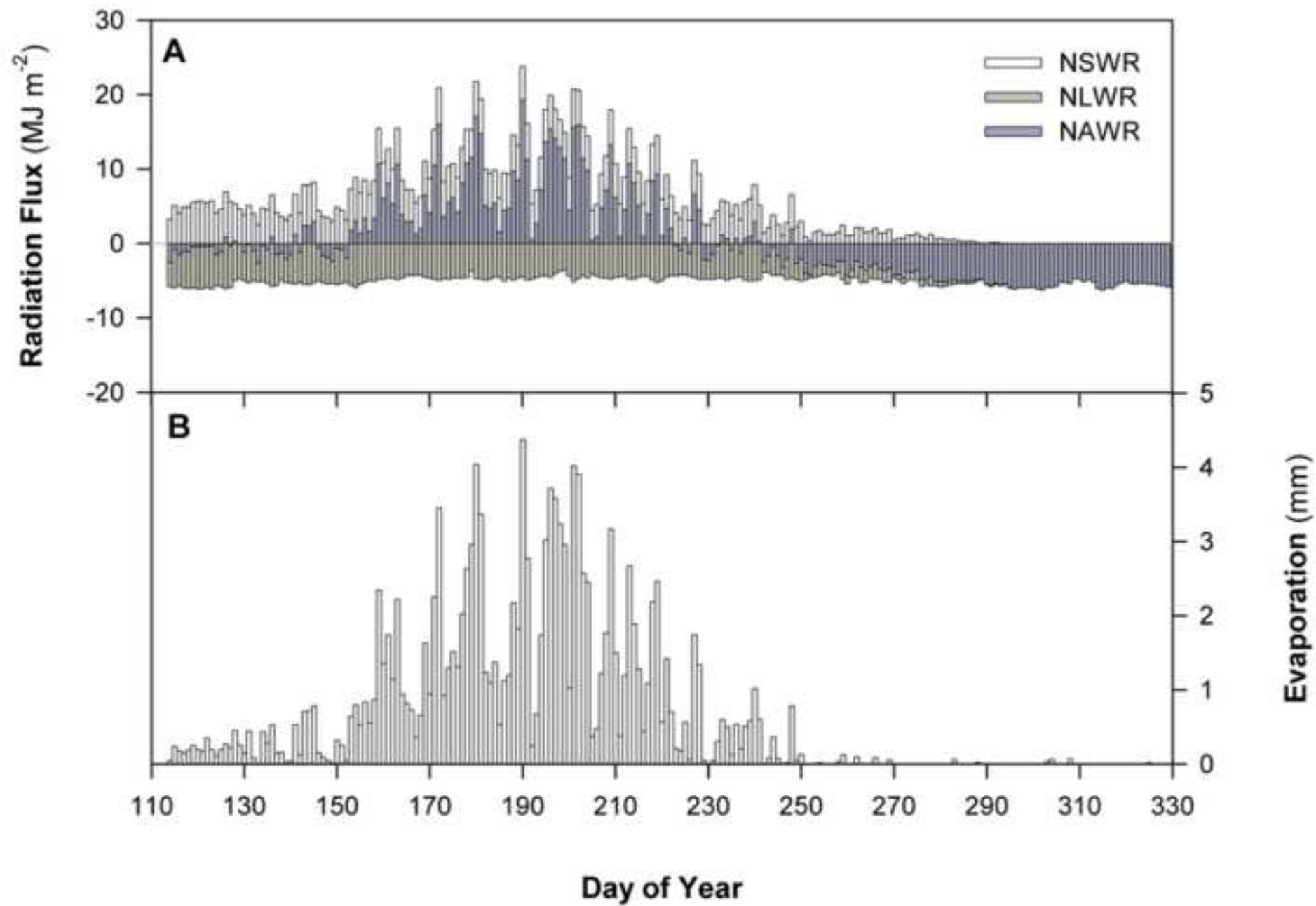


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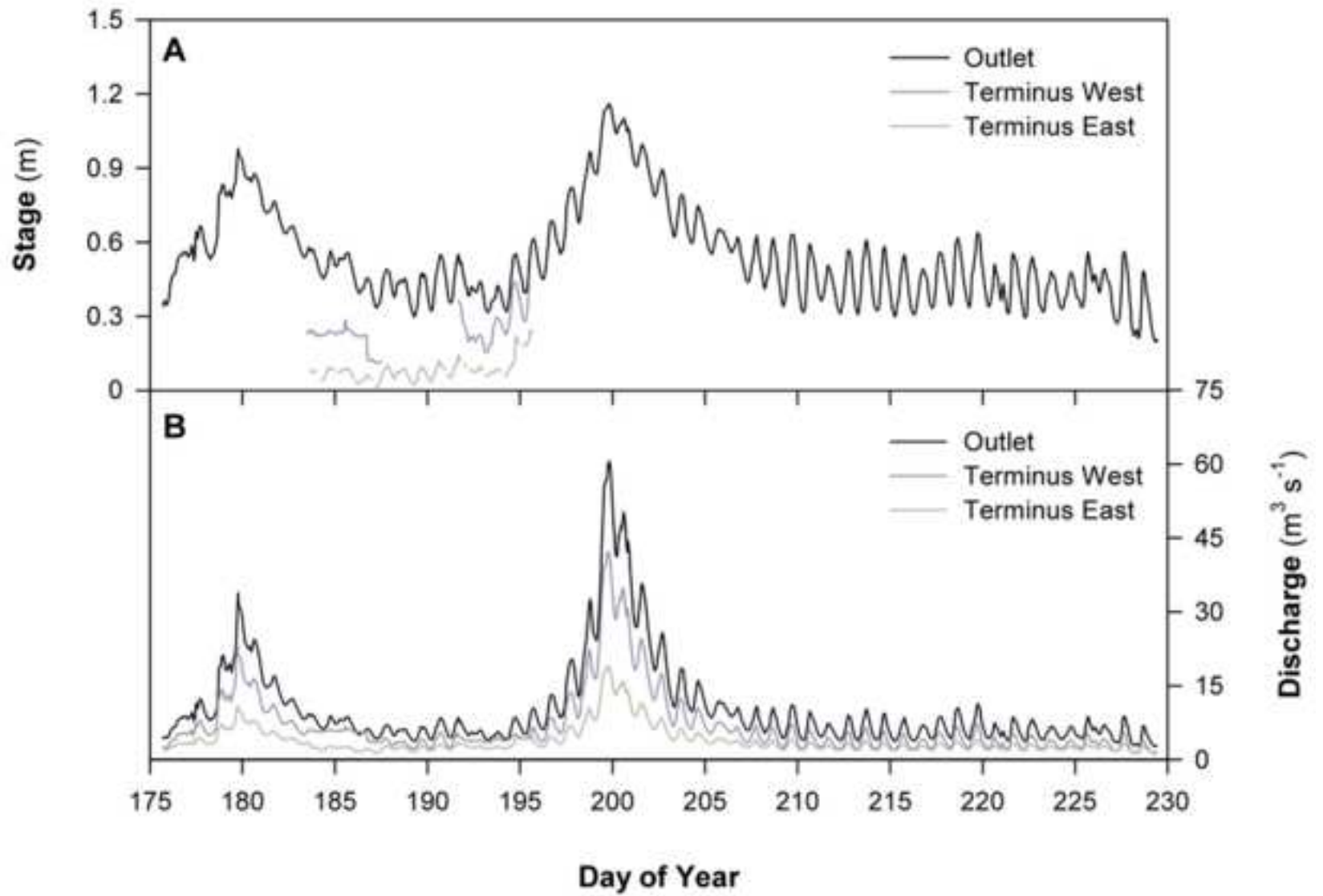


Figure 5
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