Stable oxygen isotope variability in two contrasting glacier river catchments in Greenland

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Abstract. Analysis of stable oxygen isotope ($\delta^{18}O$) characteristics is a useful tool to investigate water provenance in glacier river systems. In order to attain knowledge on the diversity of $\delta^{18}O$ variations in Greenlandic rivers, we examined two contrasting glacierized catchments disconnected to the Greenland Ice Sheet (GrIS). At Mittivakkat Gletscher River, a small river draining a local temperate glacier in Southeast Greenland, diurnal oscillations in $\delta^{18}O$ occurred with a three-hour time lag to the diurnal oscillations in runoff. The mean annual $\delta^{18}O$ was -14.68 ± 0.18 ‰ during the peak flow period. A hydrograph separation analysis revealed that the ice melt component constituted 82 ± 5 % of the total runoff and dominated the observed variations during peak flow in August 2004. The snowmelt component peaked between 10:00 and 13:00 hours, reflecting the long travel time and an inefficient distributed subglacial drainage network in the upper part of the glacier. At Kuannersuit Glacier River on the island Qeqertarsuaq in West Greenland, the $\delta^{18}O$ characteristics were examined after the major 1995-1998 glacier surge event. The mean annual $\delta^{18}O$ was -19.47 ± 0.55 ‰. Despite large spatial variations in the $\delta^{18}O$ values of glacier ice on the newly formed glacier tongue, there were no diurnal oscillations in the bulk meltwater emanating from the glacier in the post-surge years. This is likely a consequence of a tortuous subglacial drainage system consisting of linked-cavities, which formed during the surge event. Overall, a comparison of the $\delta^{18}O$ compositions from glacial river water in Greenland shows distinct differences between water draining local glaciers and ice caps (between -23.0 ‰ and -13.7 ‰) and the GrIS (between -29.9 ‰ and -23.2 ‰). This study demonstrates that water isotope analyses can be used to obtain important information on water sources and the subglacial drainage system structure that are highly desired for understanding glacier hydrology.

1 Introduction

There is an urgent need for improving our understanding of the controls on water sources and flow paths in Greenland. As in other parts of the Arctic, glacierized catchments in Greenland are highly sensitive to climate change (Milner et al., 2009; Blaen et al., 2014). In recent decades freshwater runoff from the Greenland Ice Sheet (GrIS) to adjacent seas has increased significantly (Hanna et al., 2005, 2008; Bamber et al., 2012; Mernild and Liston, 2012), and the total ice mass loss from the GrIS contributes with 0.33 mm sea level equivalent yr^{-1} to global sea level rise (1993-2010; Vaughan et al. 2013). In addition, ice mass loss from local glaciers (i.e. glaciers and ice caps peripheral to the GrIS; Weidick and Morris, 1998) has resulted in a global sea level rise of 0.09 mm sea level equivalent yr^{-1} (1993-2010; Vaughan et al. 2013). The changes in runoff are coupled to recent warming in Greenland (Hanna et al., 2012, 2013; Mernild et al., 2014), an increasing trend in precipitation and changes in precipitation patterns (Bales et al., 2009; Mernild et al., 2015a), and a decline in albedo (Bøggild et al., 2010; Tedesco et al., 2011; Box et al., 2012; Yallop et al., 2012; Mernild et al., 2015b). Also, extreme surface melt events have occurred in recent years (Tedesco et al., 2008, 2011; van As et al., 2012) and in July 2012 more than 97% of the GrIS experienced surface melting (Nghiem et al., 2012; Keegan et al., 2014). In this climate change context, detailed catchment-scale studies on water source and water flow dynamics are urgently needed to advance our knowledge of the potential consequences of future hydrological changes in Greenlandic river catchments.

Analysis of stable oxygen isotopes is a very useful technique to investigate water provenance in glacial river systems. Stable oxygen isotopes are natural conservative tracers in low-temperature hydrological systems (e.g. Moser and Stichler, 1980; Gat and Gonfiantini, 1981; Haldorsen et al., 1997; Kendall et al., 2013). Consequently, oxygen isotopes can be applied to determine the timing and origin of changes in water sources and flow paths because different water sources often have isotopically different compositions due to their exposure to different isotopic fractionation processes. Since the 1970s, this technique has been widely used for hydrograph separation (Dinçer et al., 1970). Most often a conceptual two-component mixing model is applied, where an *old water* component (e.g. groundwater) is mixed with a *new water* component (e.g. rain or snowmelt), assuming that both components have spatial and temporal homogeneous compositions. The general mixing model is given by the equation

$$QC = Q_1C_1 + Q_2C_2 + \ldots , \qquad (1)$$

where the discharge Q and the isotopic value C are equal to the sum of their components. This simplified model has limitations when a specific precipitation event is analysed because the water isotope composition in precipitation (*new water*) may vary considerably during a single event (e.g. McDonnell et al., 1990) and changes in contributions from secondary *old water* reservoirs may occur (e.g. Hooper and Shoemaker, 1986). Nevertheless, water isotope mixing models still provide valuable information on spatial differences in hydrological processes on diurnal to annual timescales (Kendall et al., 2013).

In glacier-fed river systems, the principal water sources to bulk runoff derive from ice melt, snowmelt, rainfall and groundwater components. Depending on the objectives of the study and on the environmental setting, hydrograph separation of glacial rivers has been based on assumed end-member isotope-mixing between two or three prevailing components (Behrens et al., 1971, 1978; Fairchild et al., 1999; Mark and Seltzer, 2003; Theakstone, 2003; Yde and Knudsen, 2004; Mark and McKenzie, 2007; Yde et al., 2008; Bhatia et al., 2011; Kong and Pang, 2012; Ohlanders et al., 2013; Blaen et al., 2014; Dahlke et al., 2014; Hindshaw et al., 2014; Meng et al., 2014; Penna et al., 2014; Rodriguez et al., 2014; Zhou et al., 2014). As glacierized catchments vary in size, altitudinal range, hypsometry, degree of glaciation, and thermal and morphological glacier types, isotope hydrograph separation often requires that the primary local controls on runoff generation are identified in order to analyse the variability in isotope time-series. In detailed studies it may even be necessary to divide a main component, such as ice melt, into several ice facies sub-components (Yde and Knudsen, 2004). However, in highly glacierized catchments the variability in oxygen isotope composition is generally controlled by seasonal snowmelt and ice melt with episodic inputs of rainwater, whereas contributions from shallow groundwater flow may become important in catchments, where glaciers comprise a small proportion of the total area (e.g. Blaen et al., 2014).

In this study, we examine the stable oxygen isotope composition in two Greenlandic glacier river systems, namely Mittivakkat Gletscher River (13.6 km^2) which drains a local non-surging glacier in Southeast Greenland, and Kuannersuit Glacier River (258 km^2) which drains a local glacier on the island Qeqertarsuaq, West Greenland. The latter experienced a major glacier surge event in 1995-1998. Our aim is to gain insights into the variability and controls of the oxygen isotope composition in contrasting glacierized river catchments located peripheral to the GrIS (i.e. the river systems do not drain meltwater from the GrIS). Besides a study by Andreasen (1984) at the glacier Killersuaq in West Greenland, this is the first study of oxygen isotope dynamics in rivers draining glacierized catchments peripheral to the GrIS.

2 Study sites

2.1 Mittivakkat Gletscher River, Ammassalik Island, Southeast Greenland

Mittivakkat Gletscher (65°41' N, 37°50' W) is the largest glacier complex on Ammassalik Island, Southeast Greenland (Figure 1). The entire glacier covers an area of 26.2 km^2 in 2011 (Mernild et al., 2012) and has an altitudinal range between 160 and 880 m a.s.l. (Mernild et al., 2013a). Bulk meltwater from the glacier drains primarily westwards to the proglacial Mittivakkat Valley and flows into the Sermilik Fjord. The sampling site is located at a hydrometric station 1.3 km down-valley from the main subglacial meltwater portal. The hydrological catchment has an area of 13.6 km^2, of which 9.0 km^2 are glacierized (66%). The maritime climate is Low Arctic with annual precipitation ranging from 1400 to 1800 mm water equivalent (w.e.) yr^{-1} (1998-2006) and a mean annual air temperature (MAAT) at 515 m a.s.l. of -2.2 °C (1993-2011) (updated from Mernild et al., 2008a). There are no observations of contemporary permafrost in the area, and the proglacial vegetation cover is sparse.

The glacier has undergone continuous recession since the end of the Little Ice Age (Knudsen et al., 2008; Mernild et al., 2011). In recent decades the recession has accelerated and the glacier has lost approximately 29% of its volume between 1994 and 2012 (Yde et al., 2014), and surface mass balance measurements indicate a mean thinning rate of 1.01 m w.e. yr^{-1} between 1995/1996 and 2011/2012 (Mernild et al., 2013a). Similar to other local glaciers in the Ammassalik region, Mittivakkat Gletscher is severely out of contemporary climatic equilibrium (Mernild et al., 2012, 2013b) and serves as a representative location for studying the impact of climate change on glacierized river catchments in Southeast Greenland (e.g. Mernild et al., 2008b, 2015b; Bárcena et al., 2010, 2011; Kristiansen et al., 2013; Lutz et al., 2014).

2.2 Kuannersuit Glacier River, Qeqertarsuaq, West Greenland

Kuannersuit Glacier (69°46' N, 53°15'W) is located in central Qeqertarsuaq (formerly Disko Island), West Greenland (Figure 1). It is an outlet glacier descending from the Sermersuaq ice cap and belongs to the Qeqertarsuaq-Nuussuaq surge cluster (Yde and Knudsen, 2007). In 1995, the glacier started to surge down the Kuannersuit Valley with a frontal velocity up to 70 m per day (Larsen et al., 2010). By the end of 1998 or beginning of 1999, the surging phase terminated and the glacier went into its quiescent phase, which is presumed to last more than hundred years (Yde and Knudsen, 2005a). The 1995-1998 surge of Kuannersuit Glacier is one of the largest land-terminating surge events ever recorded; the glacier advanced 10.5 km down-valley and approximately 3 km^3 of ice were moved to form a new glacier tongue (Larsen et al., 2010).

Kuannersuit Glacier River originates from a portal at the western side of the glacier terminus and the sampling site is located 200 m down-stream (Yde et al., 2005a). The catchment area has an altitude range of 100-1650 m a.s.l. and covers 258 km^2 of which Kuannersuit Glacier constitutes 103 km^2 of the total glacierized area of 168 km^2 (Yde and Knudsen, 2005a). The valley floor consists of unvegetated outwash sediment, dead-ice deposits and ice-cored, vegetated terraces. The proglacial area of the catchment is situated in the continuous permafrost zone (Yde and Knudsen, 2005b), and the climate is polar continental (Humlum, 1999). There are no meteorological observations from the area, but at the coastal town of Qeqertarsuaq (formerly Godhavn) located 50 km to the southwest the MAAT were -2.7 °C and -1.7 °C in 2011 and 2012, respectively (Cappelen, 2013).

3 Methods

3.1 Sampling protocol and isotope analyses

In total, 287 oxygen isotope samples were collected from Mittivakkat Gletscher River during the years 2003-2009 (Table 1). Most of the sampling campaigns were conducted in August at the end of the peak flow period (i.e. the summer period with relatively high runoff). The most intensively sampled period was from 8 August to 22 August 2004, where sampling was conducted with a 4-hour frequency supplemented by short periods of higher frequency sampling. In the years 2005 and 2008, meltwater was also collected during the early melt season (i.e. the period before the subglacial drainage system is well-established) to evaluate the seasonal variability in the $\delta^{18}O$ signal. An additional 40 river samples were collected for multi-sampling tests.

During five field seasons in July 2000, 2001, 2002, 2003 and 2005, a total of 180 oxygen isotope samples were collected from Kuannersuit Glacier River (Table 2) and another 44 river samples were collected for multi-sampling tests. In addition, 13 ice samples were obtained along a longitudinal transect at the centreline of the newly formed glacier tongue with 500 m sampling increments in July 2001, and 23 ice samples were collected along a transverse transect with 50 m sampling increments in July 2003. The transverse transect crossed the longitudinal transect at a distance of 3250 m from the glacier front. Seven samples of rainwater were collected in a Hellmann rain gauge located in the vicinity of the glacier terminus in July 2002.

All water samples were collected manually in 20 ml vials. Ice samples were collected in 250 ml polypropylene bottles or plastic bags before being slowly melted and decanted to 20 ml vials. The vials were stored in cold (~5 °C) and dark conditions to avoid fractionation related to biological activity.

The relative deviations (δ) of water isotope compositions ($^{18}O/^{16}O$) were expressed in per mil (‰) relative to Vienna Standard Mean Ocean Water (0 ‰) (Coplen, 1996). The stable oxygen isotope analyses were performed at the Niels Bohr Institute, University of Copenhagen, Denmark, using mass spectrometry with an instrumental precision of ± 0.1 ‰ in the oxygen isotope ratio ($\delta^{18}O$) value.

The oxygen isotope data from this study is available in the supplement (Tables S1-S6).

3.2 Multi-sample tests

In Mittivakkat Gletscher River, we conducted three multi-sample tests at 14:00 hours on 9, 15 and 21 August 2004 to determine the combined uncertainty related to sampling and analytical error. During the multi-sample tests samples were collected simultaneously (within three minutes). The tests show standard deviations of 0.08 ‰ (n = 25), 0.06 ‰ (n = 5) and 0.04 ‰ (n = 10), respectively, which are lower than the instrumental precision (± 0.1 ‰).

In Kuannersuit Glacier River, multi-sample tests were conducted in 2001, 2002 and 2003, showing a standard deviation of ± 0.16 ‰ (n = 5), ± 0.13 ‰ (n = 17) and ± 0.44 ‰ (n = 22), respectively. The multi-sample test in 2003 showed a standard deviation significantly larger than the instrumental precision (± 0.1 ‰). This deviation cannot be explained by the presence of a few high $\delta^{18}O$ values. The most plausible explanation is that the glacier runoff was not well-mixed in 2003, possibly because different parts of the drainage system merged close to the glacier portal.

3.3 Runoff measurements

Stage-discharge relationships were used to determine runoff at each study site. The accuracy of individual runoff measurements is within ± 7 % (e.g. Herschy, 1999). For details on runoff measurements we refer to Hasholt and Mernild (2006) for Mittivakkat Gletscher River and Yde et al. (2005a) for Kuannersuit Glacier River. In short, at Mittivakkat Gletscher River the runoff measurements were conducted at a hydrometric monitoring station located after the braided river system had changed into a single river channel about 500 m from the river outlet. The station was installed in August 2004 and recorded water stage every 10 minutes during the peak flow period. At Kuannersuit Glacier River the runoff measurements were obtained at a hydrometric monitoring station installed in July 2001 at a location where the river merges to a single channel. Water stage was recorded every hour during the peak flow period. The station was destroyed during the spring river break-up in 2002.

4 Results

4.1 $\delta^{18}O$ characteristics

At Mittivakkat Gletscher River, the early melt season is characterised by an increasing trend in $\delta^{18}O$. In 2005 the $\delta^{18}O$ values in the early melt season were coincident with the $\delta^{18}O$ values during the peak flow period (Figure 2a; Table 1). This indicates that the onset of ice melt commenced before the early melt season sampling campaign. In contrast, the 2008 onset of ice melt was delayed and snowmelt totally dominated the bulk composition of the river water, except on 30 May 2008 when a rainfall event (19 mm in the nearby town of Tasiilaq located 10 km to the southeast of the Mittivakkat Gletscher River catchment; Cappelen, 2013) caused a positive peak in $\delta^{18}O$ of ~1 ‰ (Figure 2b). This difference between the early ablation seasons in 2005 and 2008 is consistent with the meteorological record from Tasiilaq, which shows that the region received a large amount of precipitation in May 2008 (140 mm) compared to a dry May 2005 (17 mm; Cappelen, 2013). Episodic effects on $\delta^{18}O$ by precipitation seem common throughout the ablation season. For instance, another short-term change occurred on 14 – 15 August 2005 (Figure 2a), where a negative peak in $\delta^{18}O$ of ~2 ‰ coincided with a snowfall event (14 mm in Tasiilaq; Cappelen, 2013) and subsequent elevated contribution from snowmelt.

During the peak flow periods, the mean annual $\delta^{18}O$ was -14.68 ± 0.18 ‰ (Table 1). We use the 2004 time-series to assess oxygen isotope dynamics in the Mittivakkat Gletscher River during the peak flow period when the subglacial drainage system is assumed to be well-established, transporting the majority of meltwater in a channelized network (Mernild, 2006). In Figure 3, the 2004 $\delta^{18}O$ time-series is shown together with runoff (at the hydrometric station), air temperature (at a nunatak at 515 m a.s.l.) and electrical conductivity (at the hydrometric station; corrected to 25 °C). There was no precipitation during the entire sampling period, except for some drizzle on 8 August prior to the collection of the first sample. The time-series shows characteristic diurnal variations in $\delta^{18}O$ composition, e.g. on 9-10 and 16-18 August 2004. However, the diurnal pattern was severely disturbed at around 03:00 hours on 11 August 2004. The hydrograph shows that during the falling limb the diurnal trend in runoff was interrupted, coinciding with an air temperature increase and a change in $\delta^{18}O$ from decreasing to slightly increasing values. The runoff stayed almost constant until a rapid 39 % increase in runoff occurred at 13:00 hours on 12 August 2004, accompanied by an increase in $\delta^{18}O$ and decrease in electrical conductivity. Thereafter, runoff remained at an elevated level for more than two days before returning to a diurnal oscillation of runoff. Hydrograph separation of water sources is a helpful tool to elucidate the details of this event (see section 4.3).

In the Kuannersuit Glacier River, the sample-weighted mean annual $\delta^{18}O$ was -19.47 ± 0.55 ‰ during the peak flow period (a sample-weighted value is applied because the number of samples per year deviated between 2 and 109). In Figure 4, the variations in $\delta^{18}O$ are presented together with runoff for the period 14 – 31 July 2001. The 2001 runoff measurements showed diurnal oscillations with minimums around 10:00 – 12:00 hours and maximums at 19:00 – 20:00 hours, correlating with reversed oscillations in solutes (Yde et al., 2005a) and poorly with suspended sediment concentrations (Knudsen et al., 2007). However, the variability of $\delta^{18}O$ did not correlate with runoff or any of these variables. While some of the episodic damming and meltwater release events appear as peaks on the runoff time-series, the peaks in the $\delta^{18}O$ time-series coincided with rainfall events (e.g. on the nights of 21 July and 29 July 2001). Besides these episodic peaks, a lack of diurnal fluctuations in $\delta^{18}O$ characterised the $\delta^{18}O$ time-series.

Figure 5 shows the diurnal $\delta^{18}O$ variations during four July days without rainfall in the years 2000-2003. There were no diurnal oscillations in 2000, 2001 and 2002. In 2003, the fluctuations were much larger than in the preceding years, but the highest $\delta^{18}O$ (-19.03 ‰) was measured at 21:00 hours and low $\delta^{18}O$ prevailed during the night (~-21.0 ‰). This diurnal variability was also reflected in the standard deviations of the measurements taken over the 24-hour periods, which increased from ± 0.07 ‰ in 2000 to ± 0.11 ‰, ± 0.23 ‰ and ± 0.70 ‰ in 2001, 2002 and 2003, respectively. The corresponding diurnal amplitudes for 2000-2003 were 0.28 ‰, 0.42 ‰, 0.64 ‰ and 2.85 ‰, respectively. Although these measurements from a single day each year are insufficient to represent the conditions for the entire peak flow period, they may indicate post-surge changes in the structure of subglacial hydrological system which are worth addressing in detail in future studies of the hydrological system of surging glaciers.

4.2 $\delta^{18}O$ end-member components

On Mittivakkat Gletscher, three snow pits (0.1 m sampling increments) were excavated at different altitudes in May 1999, showing a mean $\delta^{18}O$ composition of -16.5 ± 0.6 ‰ (hereafter the uncertainty of $\delta^{18}O$ is given by the standard deviation) in winter snow (Dissing, 2000). The range of individual samples in each snow pit varied between -14.5 ‰ and -19.5 ‰ (269 m a.s.l.; mean $\delta^{18}O$ = -16.24 ± 1.35; *n* = 36), -13.8 ‰ and -21.2 ‰ (502 m a.s.l.; mean $\delta^{18}O$ = -17.11 ± 2.13; *n* = 21) and -11.9 ‰ and -21.6 ‰ (675 m a.s.l.; mean $\delta^{18}O$ = -16.18 ± 2.70; *n* = 26) (Dissing, 2000). Also, two ice-surface $\delta^{18}O$ records of 2.84 km and 1.05 km in length (10 m sampling increments) were obtained from the glacier terminus towards the equilibrium line (Boye, 1999). The glacier ice $\delta^{18}O$ ranged between -15.0 ‰ and -13.3 ‰ with a mean $\delta^{18}O$ of -14.1 ‰ (Boye, 1999), and the theoretical altitudinal effect (Dansgaard, 1964) of higher $\delta^{18}O$ towards the equilibrium line altitude (ELA) was not observed. The reasons for an absence of a $\delta^{18}O$ lapse rate are most likely due to the limited size and altitudinal range (160-880 m a.s.l.) of Mittivakkat Gletscher, but ice dynamics, ice age and meteorological conditions, such as frequent inversion (Mernild and Liston, 2010), may also have an impact. The $\delta^{18}O$ of summer rain has not been determined in this region, but at the coastal village of Ittoqqortoormiit, located ~840 km to the north of Mittivakkat Gletscher, observations show monthly mean $\delta^{18}O$ in rainwater of -12.8 ‰, -9.1 ‰ and -8.8 ‰ in June, July and August, respectively (data available from the International Atomic Energy Agency database WISER). Based on these observations it is evident that end-member snowmelt has a relatively low $\delta^{18}O$ compared to end-member ice melt and that these two water source components can be separated. Contributions from rainwater will likely result in episodic increase in the $\delta^{18}O$ of bulk meltwater.

In the Kuannersuit Glacier River system, the glaciological setting differed from the Mittivakkat Gletscher River system. During the surge event of Kuannersuit Glacier, the glacier front advanced from ~500 m a.s.l. down to 100 m a.s.l., while a significant part of the glacier surface in the accumulation area was lowered by more than 100 m to altitudes below the ELA (~1100-1300 m a.s.l.). A helicopter survey in July 2002 revealed that the post-surge accumulation area ratio was less than 20 % (Yde et al., 2005a). Hence, we assume that the primary post-surge water source during the peak flow period is ice melt, particularly from ablation of the new glacier tongue. The mean $\delta^{18}O$ value of glacier ice collected along the longitudinal and transverse transects was -20.5 ± 1.0 ‰ (n = 36). This is consistent with $\delta^{18}O$ values of glacier ice located near the glacier front, showing mean $\delta^{18}O$ of -19.4 ± 0.9 ‰ (n = 20) in a section with debris layers formed by thrusting and -19.8 ± 1.1 ‰ (n = 37) in a section without debris layers (Larsen et al., 2010). In contrast to the setting at Mittivakkat Gletscher River, it was likely that other ice melt component in bulk runoff from Kuannersuit Glacier comprised water from several ice facies sub-component sources with various $\delta^{18}O$ values and spatial variability. During the surge event, a thick debris-rich basal ice sequence was formed beneath the glacier and exposed along the glacier margins and at the glacier terminus (Yde et al., 2005b; Roberts et al., 2009; Larsen et al., 2010). The basal ice consisted of various genetic ice facies, where different isotopic fractionation processes during the basal ice formation resulted in variations in the $\delta^{18}O$ composition. The $\delta^{18}O$ in massive stratified ice was -16.6 ± 1.9 ‰ (n = 10); in laminated stratified ice it was -19.6 ± 0.7 ‰ (n = 9) and in dispersed ice it was -18.8 ± 0.6 ‰ (n = 41) (Larsen et al., 2010). Also, during the termination of the surge event in the winter 1998/1999 proglacial naled was stacked into ~3 m thick sections of thrust-block naled at the glacier front, as the glacier advanced into the naled (Yde and Knudsen, 2005b; Yde et al., 2005b; Roberts et al., 2009). Naled is an extrusive ice assemblage formed in front of the glacier by rapid freezing of winter runoff and/or proglacial upwelling water mixed with snow. A profile in a thrust-block naled section showed a $\delta^{18}O$ of -20.1 ± 0.5 ‰ (n = 60; excluding an outlier polluted by rainwater; Yde and Knudsen, 2005b). With regards to the end-member compositions of snowmelt and rainwater at Kuannersuit Glacier River, it was not possible to access snow on the upper part of the glacier, so no $\delta^{18}O$ values on snowmelt were measured. Rainwater was collected during rainfall events in July 2002, showing a wide range in $\delta^{18}O$ between -18.78 ‰ and -6.57 ‰ and a median $\delta^{18}O$ of -10.32 ± 4.49 ‰ (n = 7; Table S6).

4.3 Hydrograph separation

The conditions for conducting hydrograph separation during the peak flow period were different for the two study catchments. At Mittivakkat Gletscher River it was possible to distinguish between the $\delta^{18}O$ values of end-member ice melt and snowmelt components, and there were diurnal oscillations in $\delta^{18}O$. In contrast, the available data from Kuannersuit Glacier River did not allow hydrograph separation in the years following the surge event. Here, there were no diurnal oscillations in $\delta^{18}O$, and the composition and importance of the snowmelt component were unknown. Hence, we will continue by using the 2004 time-series to construct a two-component hydrograph separation (equation 1) during a period without precipitation for Mittivakkat Gletscher River.

First, we apply time-series cubic spline interpolation to estimate $\delta^{18}O$ at one-hour time-step increments, matching the temporal resolution of the runoff observations. This approach allows a better assessment of the diurnal $\delta^{18}O$ signal. For instance, a best-fit analysis shows that overall the $\delta^{18}O$ signal lags three hours behind runoff (r^2 = 0.66; linear correlation without lag shows r^2 = 0.58), indicating the combined effect of the two primary components, snowmelt and ice melt, on the $\delta^{18}O$ variations. The diurnal amplitude in $\delta^{18}O$ ranged between 0.11 ‰ (11 August 2004) and 0.49 ‰ (16 August 2004). However, there was no statistical relation between diurnal $\delta^{18}O$ amplitude and daily air temperature amplitude (r^2 = 0.28), indicating that other forcings than variability in surface melting may have a more dominant effect on the responding variability in $\delta^{18}O$.

Based on the assumption that snowmelt and ice melt reflect their end-member $\delta^{18}O$ compositions (-16.5 ‰ and -14.1 ‰, respectively), a hydrograph showing contributions from snowmelt and ice melt is constructed for the 2004 sampling period (Figure 6). The ice melt component constituted 82 ± 5 % (where ± indicates the standard deviation of the hourly estimates) of the total runoff and dominated the observed variations in total runoff (r^2 = 0.99). This is expected late in the peak flow period, where the subglacial drainage mainly occurs in a channelized network in the lower part of the glacier (Mernild, 2006). The slightly decreasing trend in the daily snowmelt component was likely a consequence of the diminishing snow cover on the upper part of the glacier. The snowmelt component peaked around 10:00-13:00 hours each day, reflecting the long distance from the melting snowpack to the proglacial sampling site and the possible existence of an inefficient distributed subglacial drainage network in the upper part of the glacier.

The most likely reason for an abrupt change in glacial runoff, such as the one observed during the early morning of 11 August 2004 followed by the sudden release of water 34 hours later, is a roof collapse causing ice-block damming of a major subglacial channel. The hydrograph separation (Figure 6) shows that the proportion between ice melt and snowmelt remained almost constant after the event commenced, indicating that the bulk water derived from a well-mixed part of the drainage system, which was unaffected by the large diurnal variation in ice melt generation. This suggests that the functioning drainage network transported meltwater from the upper part of the glacier with limited connection to the drainage network on the lower part. Meanwhile, ice melt was stored in a dammed section of the subglacial network located in the lower part of the glacier, and suddenly released when the dam broke at 13:00 hours on 12 August (Figure 6). In the following hours ice melt comprised up to 94 % of the total runoff. On 13 August the snowmelt component peaked at noon but then dropped markedly and in the evening it only constituted 4 % of the total runoff. On 14 August there were still some minor disturbances in the lower drainage network, but from 15 August the drainage system had stabilized and the characteristic diurnal glacionival oscillations had taken over (Figures 3 and 6).

4.4 Uncertainties in $\delta^{18}O$ hydrograph separation models

The accuracy of end-member hydrograph separation models is limited by the uncertainties of the estimated values of each end-member component, the uncertainty of the cubic spline interpolation at each data point and the uncertainty of $\delta^{18}O$ in the river. While the uncertainty of $\delta^{18}O$ in the river is likely to be relatively small, the uncertainties of each end-member component must be kept in mind (e.g., Cable et al., 2011; Arendt et al., 2015). The assumption of discrete values of each end-member component is unlikely to reflect the spatial and temporal changes in bulk $\delta^{18}O$ of snowmelt, ice melt and rainwater. For instance, Raben and Theakstone (1998) found a seasonal increase in mean $\delta^{18}O$ in snow pits on Austre Okstindbreen, Norway, and episodic events such as passages of storms (e.g., McDonnell et al., 1990; Theakstone, 2008) or melting of fresh snow in the late ablation season may cause temporal changes in one component. Also, snowpacks have a non-uniform layered structure with heterogeneous $\delta^{18}O$ composition and isotopic fractionation is likely to occur as melting progresses and the snowpack is mixed with rainwater (e.g., Raben and Theakstone, 1998; Lee et al., 2010). It is also difficult to assess how representative snow pits and ice transects are for the bulk $\delta^{18}O$ value of each component. Spatial differences in $\delta^{18}O$ may exist within and between snow pits but the overall effect on the isotopic composition of the water leaving the melting snowpack at a given time is unknown.

4.5 Longitudinal and transverse $\delta^{18}O$ transects

Glacier ice samples were collected on the surface of Kuannersuit Glacier to gain insights into the spatial variability of $\delta^{18}O$ on the newly formed glacier tongue. Both the longitudinal and transverse transects showed large spatial fluctuations in $\delta^{18}O$ (Figure 7). The longitudinal transect was sampled along the centreline but showed unsystematic fluctuations on a 500 m sampling increment scale. In contrast, the transverse transect, which was sampled 3250 m up-glacier with 50 m increments, showed a more systematic trend where relatively high $\delta^{18}O$ values were observed along both lateral margins. From the centre towards the western margin an increasing trend of 0.46 ‰ per 100 m prevailed, whereas the eastern central part showed large fluctuations in $\delta^{18}O$ between -22.69 ‰ and -20.08 ‰. The total range of measured $\delta^{18}O$ in glacier ice along the transverse transect was 4.14 ‰. A possible explanation of this marked spatial variability may be that the ice forming the new tongue derived from different pre-surge reservoirs on the upper part of the glacier. If so, it is very likely that the marginal glacier ice was formed at relatively low elevations (high $\delta^{18}O$ signal), whereas the glacier ice in the western central part mainly derived from high elevation areas of Sermersuaq ice cap (low $\delta^{18}O$ signal). At present, there are only few comparable studies on transverse variations in $\delta^{18}O$ across glacier tongues. Epstein and Sharp (1959) found a decrease in $\delta^{18}O$ towards the margins of Saskatchewan Glacier, Canada. Hambrey (1974) measured a similar decrease in $\delta^{18}O$ towards the margins of Charles Rabots Bre, Norway, in an upper transect, whereas a lower transect showed wide unsystematic variations in $\delta^{18}O$. Hambrey (1974) concluded that in the upper transect the marginal ice derived from higher altitudes than ice in the centre, whereas in the lower transect the wide variations were related to structural complexity of the glacier. However, both of these studies are based on few samples. Hence, it therefore remains unknown whether a high spatial variability in $\delta^{18}O$ is a common phenomenon or related to specific circumstances such as surge activity or presence of tributary glaciers.

5 Discussion

5.1 Differences in $\delta^{18}O$ between Mittivakkat Gletscher River and Kuannersuit Glacier River

A significant difference between the $\delta^{18}O$ dynamics in Mittivakkat Gletscher River and Kuannersuit Glacier River is the marked diurnal oscillations in the former and the lack of a diurnal signal in the latter during the peak flow period. At Mittivakkat Gletscher River, the 2004 hydrograph separation analysis showed a three-hour lag of $\delta^{18}O$ to runoff caused by the difference in travel time for ice melt and snowmelt. Meltwater in the early melt season was dominated by snowmelt with relatively high $\delta^{18}O$ and weak diurnal oscillations; whereas diurnal oscillations with amplitudes between 0.11 ‰ and 0.49 ‰ existed during the peak flow period due to mixing of a dominant ice melt component and a secondary snowmelt component. Diurnal oscillations in $\delta^{18}O$ are common in meltwater from small, glacierized catchments; for instance, at Austre Okstindbreen, Norway, the average diurnal amplitude is approximately 0.2 ‰ (Theakstone, 1988; Theakstone and Knudsen, 1989; 1996a,b; Theakstone, 2003). The largest diurnal amplitudes in $\delta^{18}O$ (up to 4.3 ‰) have been observed in small-scale GrIS catchments, such as at Imersuaq and “N Glacier”, where large differences in $\delta^{18}O$ exist between various ice facies and snowmelt (Yde and Knudsen, 2004; Bhatia et al., 2011).

The lack of strong diurnal oscillations as observed in the post-surge years at Kuannersuit Glacier River indicates either a mono-source system, a well-mixed drainage network, or a multi-source system, where the primary components have similar $\delta^{18}O$ compositions. The expected primary component, glacier ice melt, has lower $\delta^{18}O$ than bulk runoff and there must be additional contributions from basal ice melt (similar $\delta^{18}O$ composition as runoff), snowmelt (unknown $\delta^{18}O$ composition) or rainwater (higher $\delta^{18}O$ composition than runoff). We therefore hypothesize that the presence of a well-mixed drainage network is the most likely reason for the observed $\delta^{18}O$ signal in the bulk runoff from Kuannersuit Glacier. During the surge event the glacier surface became heavily crevassed and the pre-existing drainage system collapsed (Yde and Knudsen, 2005a). It is a generally accepted theory that the drainage system of surging glaciers transforms into a distributed network where meltwater is routed via a system of linked cavities (Kamb et al., 1985; Kamb, 1987), but little is known about how subglacial drainage systems evolve into discrete flow systems in the years following a surge event. In the initial quiescent phase at Kuannersuit Glacier, frequent loud noises interpreted as drainage system roof collapses were observed, in addition to episodic export of ice blocks from the portal, suggesting ongoing changes to the englacial and subglacial drainage system. A consequence of these processes is also visible on the glacier surface, where circular collapse chasms formed above marginal parts of the subglacial drainage system (Yde and Knudsen, 2005a).

Lack of diurnal oscillations in $\delta^{18}O$ has previously been related to other causes at non-surging glaciers. At Glacier de Tsanfleuron, Switzerland, sampling in the late melt season (23-27 August 1994) showed no diurnal variations in $\delta^{18}O$, which was interpreted by Fairchild et al. (1999) as a consequence of limited altitudinal range (less than 500 m) of the glacier. An alternative explanation may be that snowmelt only constituted so small a proportion of the total runoff in the late melt season that discrimination between snowmelt and ice melt was impossible. At the glacier Killersuaq, an outlet glacier from the ice cap Amitsulooq in West Greenland, Andreasen (1984) found that diurnal oscillations in $\delta^{18}O$ were prominent during the relatively warm summer of 1982, whereas no diurnal $\delta^{18}O$ oscillations were observed in 1983 because the glacier was entirely snow-covered throughout the ablation season, due to low summer surface mass balance caused by the 1982 El Chichón eruption (Ahlstrøm et al., 2007).

5.2 $\delta^{18}O$ compositions in glacier rivers

It is clear from the studies at Mittivakkat Gletscher and Kuannersuit Glacier that glacier rivers have different $\delta^{18}O$ compositions. The bulk meltwater from Mittivakkat Gletscher has a $\delta^{18}O$ composition similar to the water draining the nearby local glacier Hobbs Gletscher and to waters from studied valley and outlet glaciers in Scandinavia, Svalbard, European Alps, Andes and Asia (Table 3). The $\delta^{18}O$ composition of Kuannersuit Glacier is lower and similar to the $\delta^{18}O$ composition of the glacier Killersuaq (Table 3). Currently, the lowest $\delta^{18}O$ compositions are found in bulk meltwater draining the GrIS in West Greenland (Table 3), but there is a lack of $\delta^{18}O$ data from Antarctic rivers. Estimations of $\delta^{18}O$ based on δD measurements suggest $\delta^{18}O$ values of -32.1 ‰, -34.4 ‰ and -41.9 ‰ in waters draining Wilson Piedmont Glacier, Rhone Glacier and Taylor Glacier, respectively (Henry et al., 1977).

The differences in $\delta^{18}O$ in glacial rivers are due to a combination of geographical effects related to altitude, continentality and latitude (Dansgaard et al., 1973) and temporal effects that work on various time-scales and in specific environments. These temporal effects include a seasonal effect (Dansgaard, 1964), a monsoonal effect (Tian et al., 2001; Kang et al., 2002), a precipitation amount effect (Holdsworth et al., 1991) and a palaeoclimatic effect (Reeh et al., 2002). For instance, the altitude and continentality effects cause low $\delta^{18}O$ in rivers draining the GrIS compared to rivers draining valley glaciers at similar latitudes (Table 3). More data on the $\delta^{18}O$ composition and dynamics in glacial rivers is needed to improve the understanding of how the relative influence of geographical and temporal effects varies on local and regional scales.

6 Conclusions

In this study, we have examined the oxygen isotope hydrology in two of the most studied glacierized river catchments in Greenland to improve our understanding of the prevailing differences between contrasting glacial environments. This study has provided insights into the variability and composition of $\delta^{18}O$ in river water draining glaciers and ice caps adjacent to the GrIS.

The following results were found:

- The Mittivakkat Gletscher River on Ammassalik Island, Southeast Greenland, has a mean annual $\delta^{18}O$ of -14.68 ± 0.18 ‰ during the peak flow period, which is similar to the $\delta^{18}O$ composition in glacier rivers in Scandinavia, Svalbard, European Alps, Andes and Asia. The Kuannersuit Glacier River on Disko Island, West Greenland, has a lower mean annual $\delta^{18}O$ of -19.47 ± 0.55 ‰, which is similar to the $\delta^{18}O$ composition in bulk meltwater draining an outlet glacier from the ice cap Amitsulooq but higher than the $\delta^{18}O$ composition in bulk meltwater draining the GrIS.
- In Mittivakkat Gletscher River the diurnal oscillations in $\delta^{18}O$ were conspicuous. This was due to the presence of an efficient subglacial drainage system and diurnal variations in the ablation rates of snow and ice that had distinguishable oxygen isotope compositions. The diurnal oscillations in $\delta^{18}O$ lagged the diurnal oscillations in runoff by approximately three hours. A hydrograph separation analysis revealed that the ice melt component constituted 82 ± 5 % of the total runoff and dominated the observed variations in total runoff during the peak flow period in 2004. The snowmelt component peaked between 10:00 and 13:00 hours, reflecting the long travel time and a possible inefficient distributed subglacial drainage network in the upper part of the glacier.
- In contrast to Mittivakkat Gletscher River, Kuannersuit Glacier River showed no diurnal oscillations in $\delta^{18}O$. This is likely a consequence of glacier surging. In the years following a major surge event, where Kuannersuit Glacier advanced 10.5 km, meltwater was routed through a tortuous subglacial conduit network of linked cavities, mixing the contributions from glacier ice, basal ice, snow and rainwater.

This study has showed that environmental and physical contrasts in glacier river catchments influence the spatio-temporal variability of the $\delta^{18}O$ compositions. In Greenlandic glacier rivers, the variability in $\delta^{18}O$ composition is much higher than previously known ranging from relatively high $\delta^{18}O$ values in small-scale coastal glacierized catchments to relatively low $\delta^{18}O$ values in GrIS catchments. This study demonstrates that water isotope analyses can be used to obtain important information on water sources and subglacial drainage system structure that are highly desired for understanding glacier hydrology.

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Table 1. Summary of $\delta^{18}O$ mean and range in bulk water samples at Mittivakkat Gletscher River.

Year	Campaign period	n	$\delta^{18}O_{mean}$	$\delta^{18}O_{max}$	$\delta^{18}O_{min}$
2003	11 – 13 Aug	4	-14.42	-14.30	-14.65
2004	8 – 22 Aug	103	-14.55	-14.19	-14.91
2005	30 May – 12 Jun	29	-14.71	-14.35	-15.16
	23 – 26 Jul	19	-14.10	-13.74	-14.41
	11 – 19 Aug	44	-14.73	-14.13	-16.43
2006	11 – 16 Aug	11	-14.85	-14.26	-15.42
2007	2 – 10 Aug	17	-14.69	-14.07	-15.11
2008	29 May – 11 Jun*	28	-16.92	-15.92	-17.35
	10 – 16 Aug	15	-14.84	-14.47	-15.20
2009	8 – 16 Aug	17	-14.88	-14.56	-15.13

* collected at a sampling site c. 500 m closer to the glacier front

Table 2. Summary of $\delta^{18}O$ mean and range in bulk water samples at Kuannersuit Glacier River.

Year	Campaign period	n	$\delta^{18}O_{mean}$	$\delta^{18}O_{max}$	$\delta^{18}O_{min}$
2000	24 – 27 Jul	21	-19.80	-19.47	-19.97
2001	14 – 31 Jul	109	-19.25	-17.82	-19.55
2002	14 – 15 Jul	21	-19.01	-18.75	-19.39
2003	18 – 26 Jul	27	-20.43	-19.03	-21.88
2005	19 – 24 Jul	2	-19.42	-19.32	-19.51

Table 3. Maximum and minimum $\delta^{18}O$ in glacier rivers.

Site	Sampling period	Latitude	Longitude	Maximum (‰)	Minimum (‰)	Reference
Greenland						
Mittivakkat Gletscher *(local glacier)*	2003-09	65º41'N	37º50'W	-13.7	-17.4	This paper
Kuannersuit Glacier *(ice cap outlet)*	2000-05	69º46'N	53º15'W	-17.8	-21.9	This paper
Hobbs Gletscher *(local glacier)*	2004	65º46'N	38º11'W	-14.7	-15.1	Yde, unpublished data
Imersuaq *(GrIS outlet)*	2000	66º07'N	49º54'W	-24.3	-29.9	Yde and Knudsen (2004)
Killersuaq *(ice cap outlet)*	1982-83	66º07'N	50º10'W	-19.5	-23.0	Andreasen (1984)
Leverett Glacier *(GrIS outlet)*	2009	67º04'N	50º10'W	-23.2	-24.2	Hindshaw et al. (2014)
Isunnguata Sermia *(GrIS outlet)*	2008	67º11'N	50º20'W	-26.2 [a]		Yde, unpublished data
'N' Glacier *(GrIS outlet)*	2008	68º03'N	50º16'W	~ -23.3	~ -28.3	Bhatia et al. (2011)
Scandinavia and Svalbard						
Austre Okstindbreen, Norway	1980-95	66º00'N	14º10'E	-11.8	-14.4	Theakstone (2003)
Storglaciären, Sweden	2004 & 2011	67º54'N	18º38'E	-10.9	-15.9	Dahlke et al. (2014)
Austre Grønfjordbreen, Svalbard	2009	77º56'N	14º19'E	-11.2 [a]		Yde et al. (2012)
Dryadbreen, Svalbard	2012	78º09'N	15º27'E	-13.0	-15.5	Hindshaw et al. (2016)
Longyearbreen, Svalbard	2004	78º11'N	15º30'E	-12.3	-16.7	Yde et al. (2008)
European Alps						
Glacier de Tsanfleuron, Switzerland	1994	46º20'N	07º15'E	~ -7.8	-12.2	Fairchild et al. (1999)
Dammagletscher, Switzerland	2008	46º38'N	08º27'E	-13.3	-17.3	Hindshaw et al. (2011)
Hintereisferner, Austria	1969-70	46º49'N	10º48'E	~ -13.8	~ -19.4	Behrens et al. (1971)
Kesselwandferner, Austria	1969-70	46º50'N	10º48'E	~ -14.8	~ -18.1	Behrens et al. (1971)
Andes						
Cordillera Blanca catchments, Peru	2004-06	9º-10ºS	77º-78ºW	-13.3	-15.3	Mark and McKenzie (2007)
Juncal River, Chile	2011-12	32º52'S	70º10'W	~ -16.4	~ -18.0	Ohlanders et al. (2013)
Asia						
Hailuogou Glacier River, China	2008-09	29º34'N	101º59'E	-13.7	-17.6	Meng et al. (2014)
Kumalak Glacier No. 72, China	2009	41º49'N	79º51'E	-9.8 [a]		Kong and Pang (2012)
Urumqi Glacier No. 1, China	2009	43º07'N	86º48'E	-8.7 [a]		Kong and Pang (2012)

[a] Single sample

Figure captions

Figure 1. Location map (A) of the study areas at (B) Mittivakkat Gletscher River, Southeast Greenland (image from Landsat 8 OLI on 3 September 2013); and at (C) Kuannersuit Glacier River, West Greenland (image from Landsat 8 OLI on 8 July 2014).

Figure 2. $\delta^{18}O$ time-series of meltwater draining Mittivakkat Gletscher in (a) 2005 and (b) 2008.

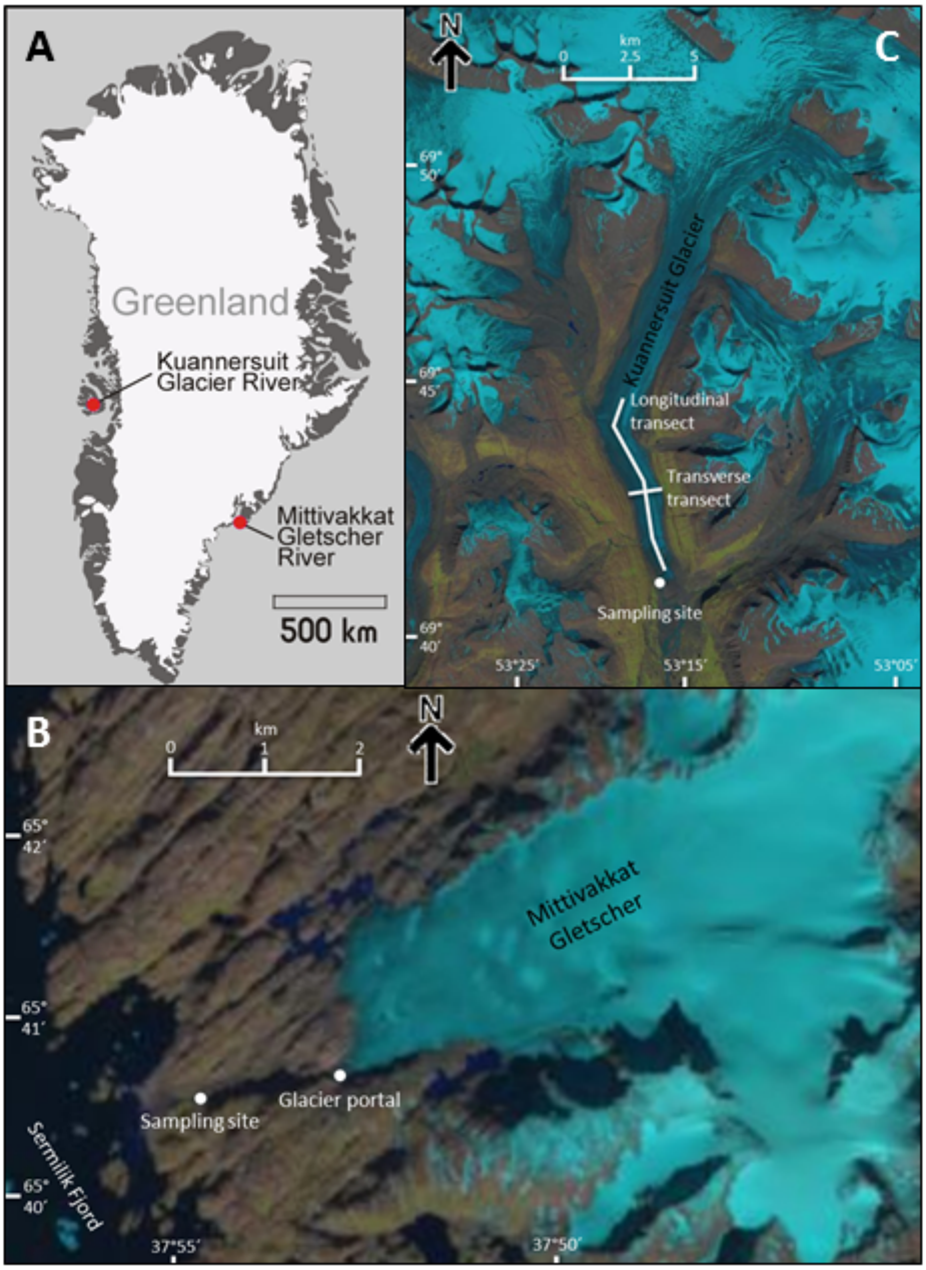
Figure 3. Time-series of $\delta^{18}O$, discharge, air temperature and electric conductivity in meltwater draining Mittivakkat Gletscher in 8-21 August 2004.

Figure 4. Time-series of $\delta^{18}O$ (red curve) and discharge (black curve) in Kuannersuit Glacier River during the period 14-31 July 2001.

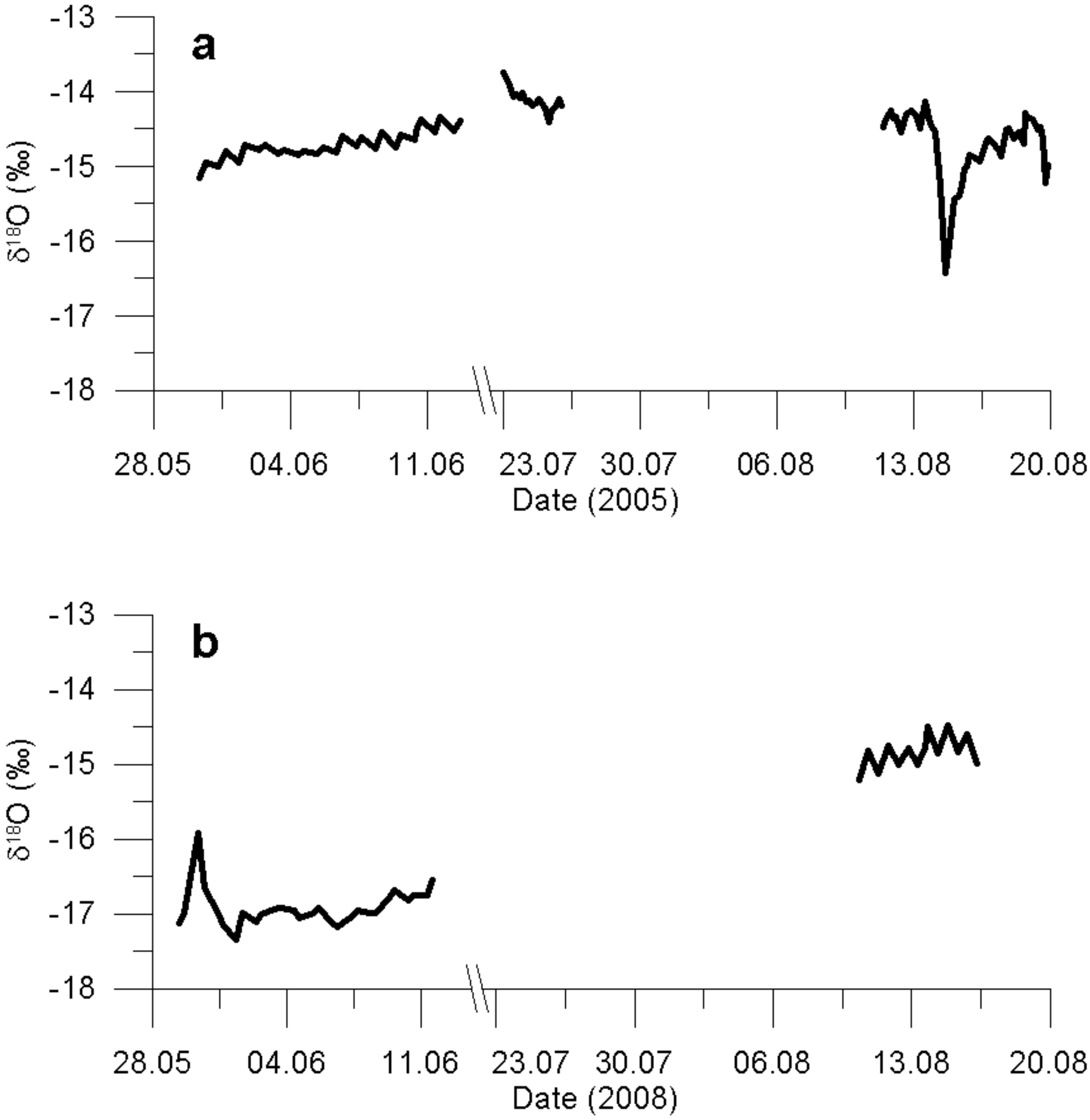
Figure 5. Diurnal $\delta^{18}O$ variations in Kuannersuit Glacier River on studied days in July in the post-surge years 2000-2003. Multi-sample tests conducted in 2001, 2002 and 2003 showed standard deviations of ± 0.16 ‰, ± 0.13 ‰ and ± 0.44 ‰, respectively.

Figure 6. Hydrograph showing the separation of the discharge in Mittivakkat Gletscher River (black curve) into an ice melt component (red curve) and a snowmelt component (blue curve) during the period 8-21 August 2004. The error of the ice melt and snowmelt components depends on the constant end-member estimates and the cubic spline interpolation. The arrow indicates the onset of the abrupt change in discharge.

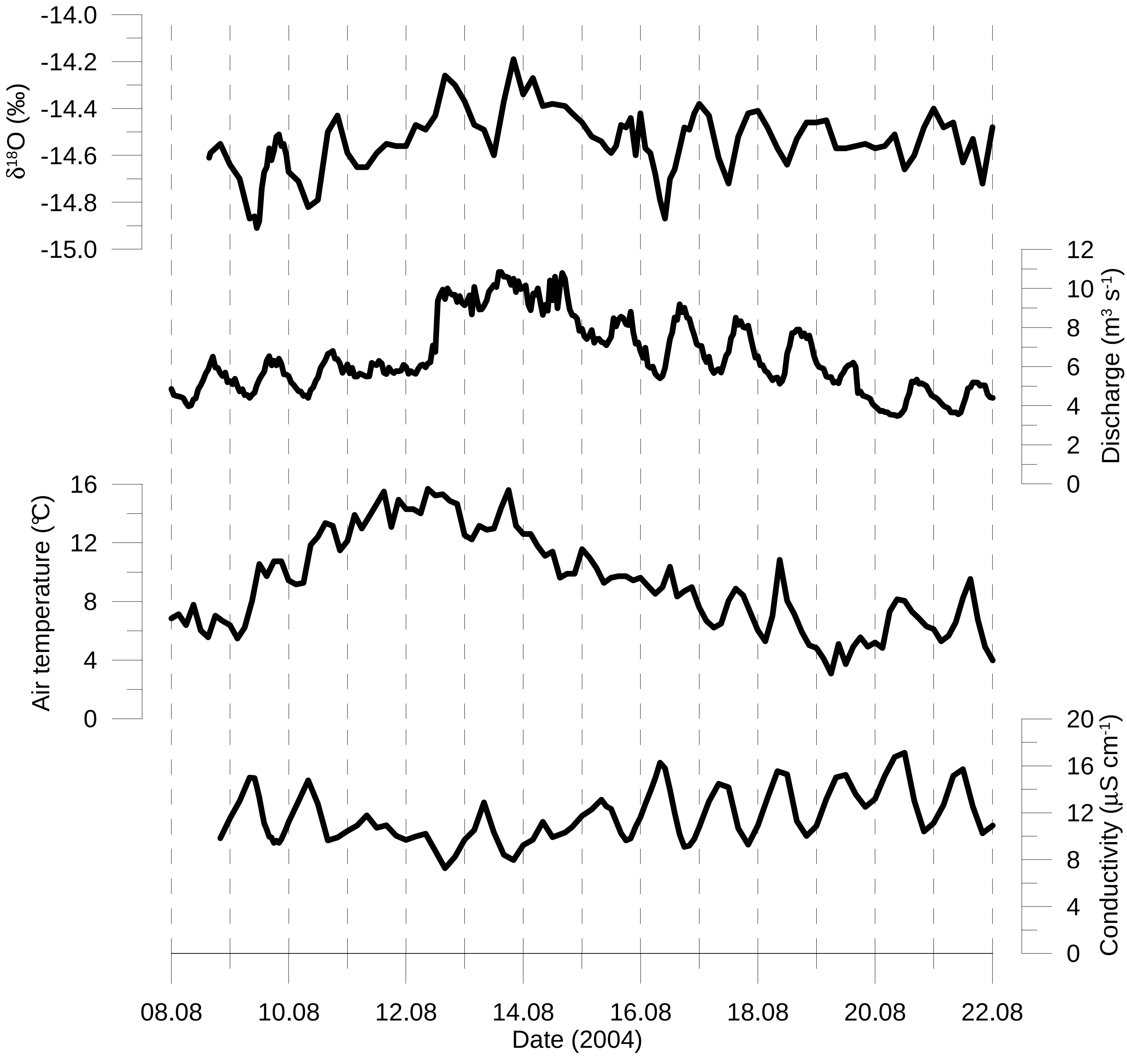
Figure 7. Variations in $\delta^{18}O$ of glacier ice along a longitudinal transect and a transverse transect on Kuannersuit Glacier. The transverse transect crosses the longitudinal transect at a distance of 3,250 m from the glacier terminus.

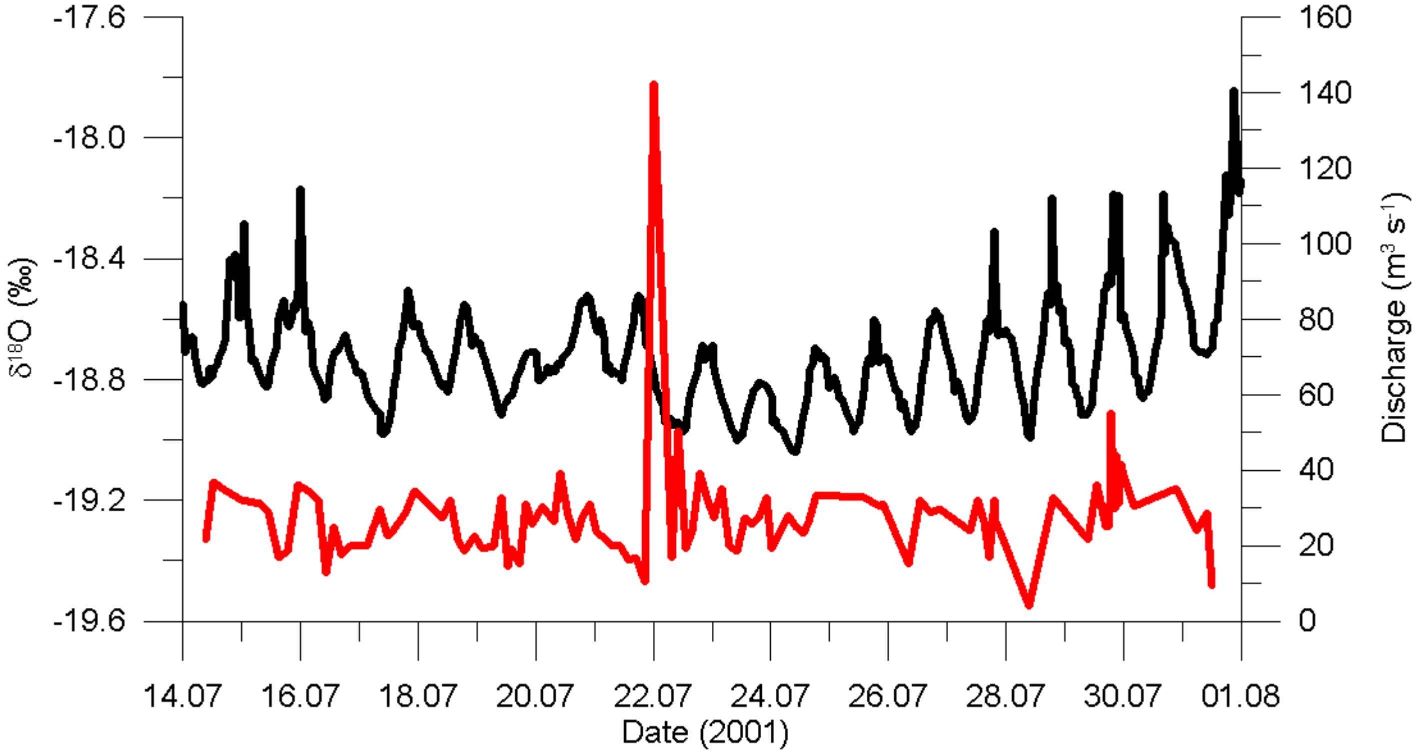


A
Greenland
Kuannersuit
Glacier River
Mittivakkat
Gletscher
River
500 km
C
N
km
0
2.5
5
69°
50′
69°
45′
69°
40′
Kuannersuit Glacier
Longitudinal
transect
Transverse
transect
Sampling site
53°25′
53°15′
53°05′
B
N
km
0
1
2
65°
42′
65°
41′
65°
40′
Mittivakkat
Gletscher
Glacier portal
Sampling site
Sermilik Fjord
37°55′
37°50′

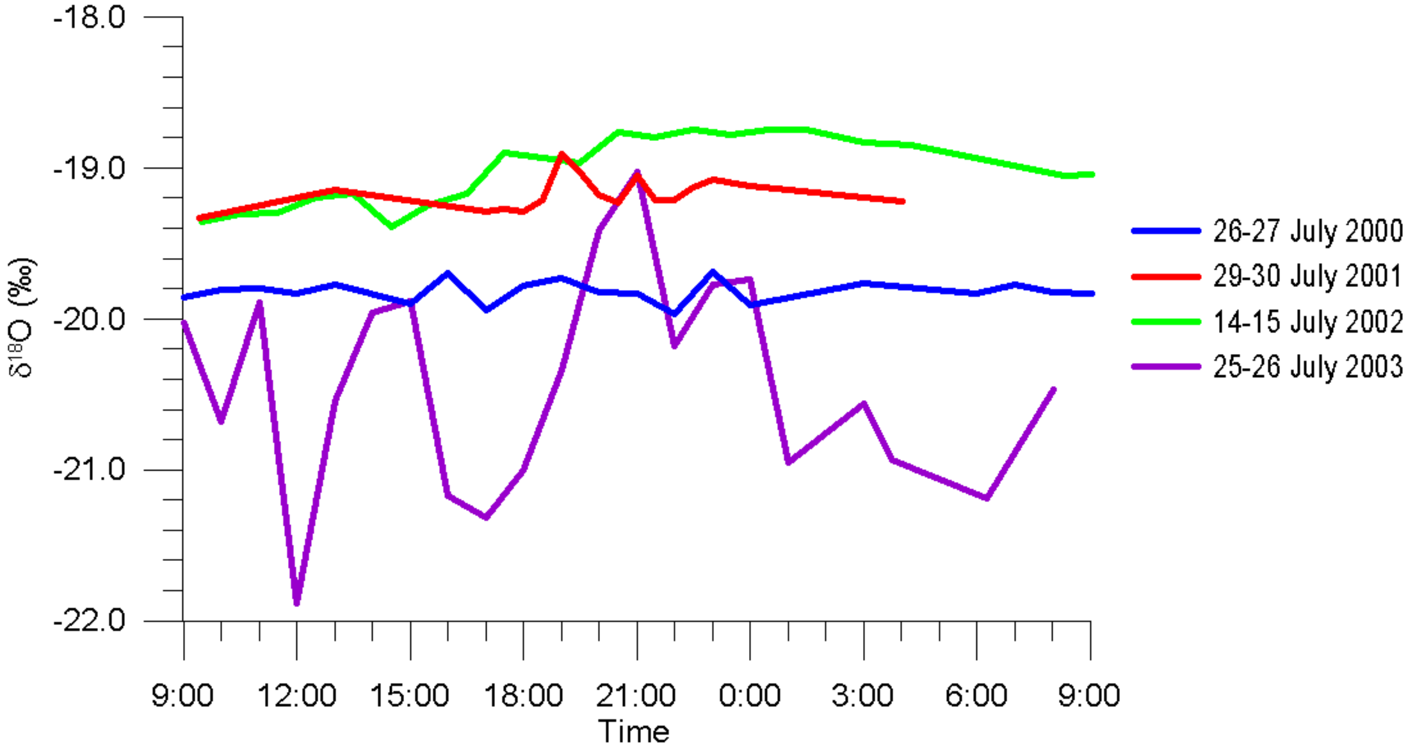


a
-13
-14
-15
-16
-17
-18
δ18O (‰)
28.05
04.06
11.06
23.07
30.07
06.08
13.08
20.08
Date (2005)
b
-13
-14
-15
-16
-17
-18
δ18O (‰)
28.05
04.06
11.06
23.07
30.07
06.08
13.08
20.08
Date (2008)

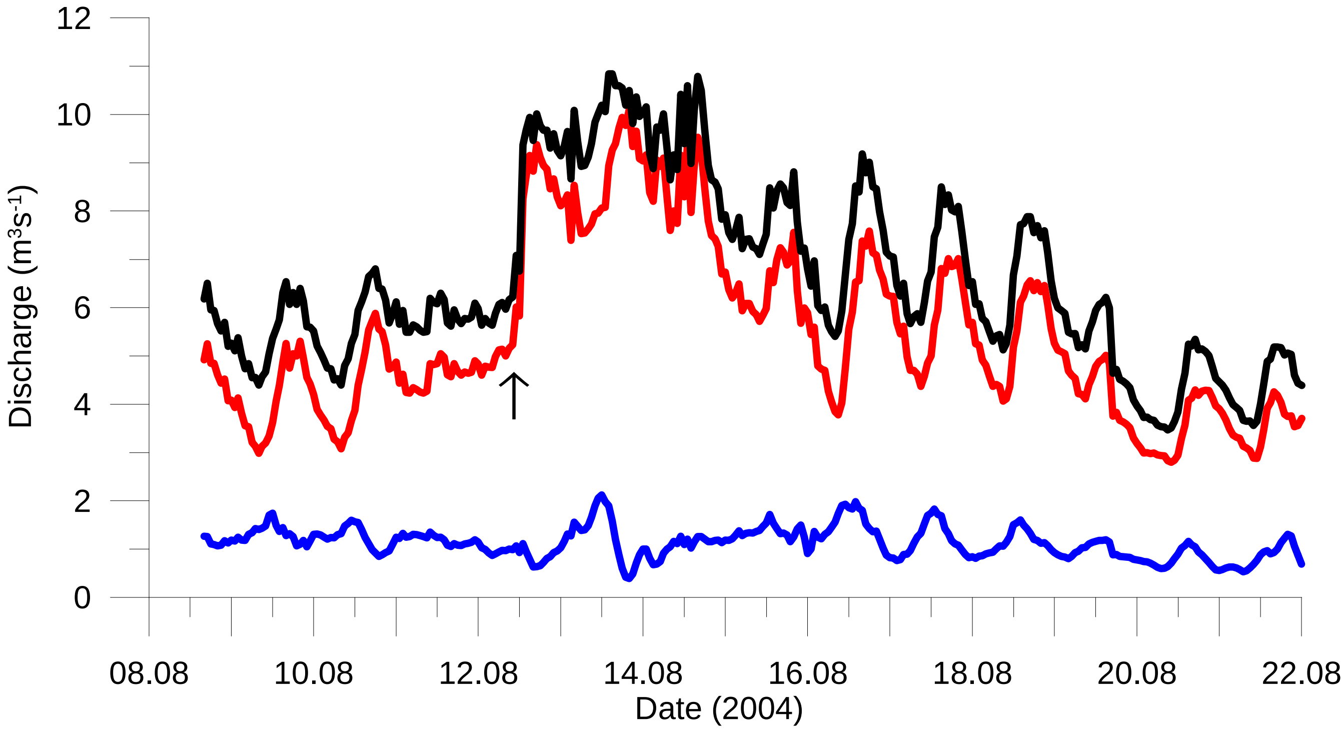

-14.0
-14.2
-14.4
-14.6
-14.8
-15.0
δ18O (‰)
12
10
8
6
4
2
0
Discharge (m3 s-1)
16
12
8
4
0
Air temperature (°C)
20
16
12
8
4
0
Conductivity (µS cm-1)
08.08
10.08
12.08
14.08
16.08
18.08
20.08
22.08
Date (2004)



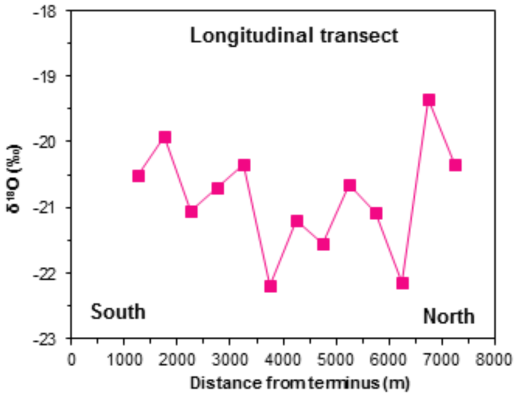
-17.6
-18.0
-18.4
-18.8
-19.2
-19.6
δ18O (‰)
160
140
120
100
80
60
40
20
0
Discharge (m3 s-1)
14.07
16.07
18.07
20.07
22.07
24.07
26.07
28.07
30.07
01.08
Date (2001)

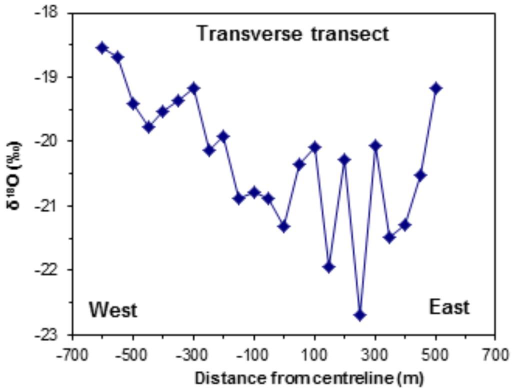


-18.0
-19.0
-20.0
-21.0
-22.0
δ18O (‰)
9:00
12:00
15:00
18:00
21:00
0:00
3:00
6:00
9:00
Time
26-27 July 2000
29-30 July 2001
14-15 July 2002
25-26 July 2003



12
10
8
6
4
2
0
Discharge (m^3s^{-1})
08.08
10.08
12.08
14.08
16.08
18.08
20.08
22.08
Date (2004)


Longitudinal transect
δ18O (‰)
-18
-19
-20
-21
-22
-23
South
North
0
1000
2000
3000
4000
5000
6000
7000
8000
Distance from terminus (m)


Transverse transect
δ18O (‰)
-18
-19
-20
-21
-22
-23
West
East
-700
-500
-300
-100
100
300
500
700
Distance from centreline (m)