Modeling the response of subglacial drainage at Paakitsoq, West Greenland, to 21st century climate change

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Abstract

Although the Greenland Ice Sheet (GrIS) is losing mass at an accelerating rate, much uncertainty remains about how surface runoff interacts with the subglacial drainage system and affects water pressures and ice velocities, both currently, and into the future. Here, we apply a physically-based, subglacial hydrological model to the Paakitsoq region, west Greenland, and run it into the future to calculate patterns of daily subglacial water pressure fluctuations in response to climatic warming. The model is driven with moulin input hydrographs calculated by a surface routing model, forced with distributed runoff. Surface runoff and routing are simulated for a baseline year (2000), before the model is forced with future climate scenarios for the years 2025, 2050 and 2095, based on the IPCC's Representative Concentration Pathways (RCPs). Our results show that as runoff increases throughout the 21st century, and/or as RCP scenarios become more extreme, the subglacial drainage system makes an earlier transition from a less efficient network operating at high water pressures, to a more efficient network with lower pressures. This will likely cause an overall decrease in ice velocities for marginal areas of the GrIS. However, short-term variations in runoff, and therefore subglacial pressure, can still cause localized speedups, even after the system has become more efficient. If these short-term pressure fluctuations become more pronounced as future runoff increases, the associated late-season speedups may help to compensate for the drop in overall summer velocities, associated with earlier transitioning from a high to a low pressure system.

1. Introduction

Recent studies suggest that the Greenland Ice Sheet (GrIS) is losing mass at an accelerating rate [*Rignot and Kanagaratnam*, 2006; *Joughin et al.*, 2008; *Pritchard et al.*, 2009], with a doubling of mass loss in the first decade of the 21st century [*Khan et al.*, 2010]. This is partly due to changes in surface mass balance (SMB), where increased accumulation of snowfall is more than offset by increased surface ablation [*Fettweis et al.*, 2011; *Sasgen et al.*, 2012; *Box*, 2013], and partly due to dynamic thinning and acceleration of ocean-terminating outlet glaciers in response to ocean warming and increased calving [*Howat et al.*, 2007, 2011; *Khan et al.*, 2010; *Seale et al.*, 2011]. Projections of the ice sheet's contribution to 21st century sea level rise (2081 – 2100 relative to 1986 – 2000) suggest that for the IPCC's Representative Concentration Pathway (RCP) 8.5, SMB changes may contribute 0.07 (0.03 - 0.16) m, whereas dynamic changes may contribute 0.05 (0.02 – 0.07) m [*Church et al.*, 2013]. In comparison, SMB changes and dynamic changes are estimated to be contributing relatively equal rates of mass loss from the GrIS currently [*van den Broeke et al.*, 2009].

Some of the uncertainty in the future dynamic contribution comes from limited process understanding of how basal conditions beneath land- and ocean-terminating outlet glaciers will respond to changes in surface meltwater production and penetration through the ice sheet. Several recent studies have shown that marginal areas of the GrIS respond dynamically at a range of time scales (hourly to annually) to variations in the rate of surface meltwater production [*Zwally et al.*, 2002; *Joughin et al.*, 2008; *Shepherd et al.*, 2009; *Bartholomew et al.*, 2010, 2011; *Colgan et al.*, 2011; *Hoffman et al.*, 2011; *Andrews et al.*, 2014]. It has been suggested that variation in the rate of meltwater production alters the rate at which water reaches the bed via crevasses and moulins, which affects subglacial water pressures and therefore rates of sliding [*Schoof*, 2010; *Hewitt*, 2013; *Werder et al.*, 2013]. In this respect, the ablation areas of the GrIS are similar to temperate valley glaciers [*Iken and Bindschadler*, 1986; *Mair et al.*, 2002; *Bingham et al.*, 2003].

Recent evidence from the GrIS shows that ice velocities increase in the short-term (hours to days) in response to increased meltwater delivery to the bed, either by hydrofracture initiating lake drainage [*Das et al.*, 2008; *Doyle et al.*, 2014; *Joughin et al.*, 2013; *Tedesco et al.*, 2013], or by increased melt production and more rapid routing of water to existing moulins [*Shepherd et al.*, 2009; *Banwell et al.*, 2013]. There is less certainty surrounding seasonal and annual velocity changes and how these are affected by variations in melt delivery at similar timescales [e.g., *Moon et al.*, 2014]. Recent evidence suggests that in the ablation areas of GrIS outlet glaciers, warmer (cooler) summers with higher (lower) overall melt rates are associated with reduced (enhanced) summer velocities [*van de Wal et al.*, 2008; *Hoffman et* *al*., 2011; *Sundal et al.,* 2011, *Sole et al.,* 2013; *Tedstone et al.,* 2013]. A proposed explanation is that higher water delivery to the subglacial drainage system promotes the channelization of the drainage pathways at the expense of a more distributed system. This lowers steady-state water pressures, reduces transient peaks in water pressure, and reduces sliding. This mechanism has support from theoretical idealised modeling studies [*Schoof,* 2010; *Hewitt,* 2013; *Werder et al.,* 2013]. Conversely, near the equilibrium line and in the lower accumulation areas of the GrIS where ice is thicker, it has been suggested that summer and annual velocities may increase in higher runoff years. The rationale is that a distributed system is more likely to survive in these areas, and thus the delivery of more water to the bed in high runoff years will increase steady-state water pressures, increase transient peaks in water pressure, and promote sliding. This theory is supported by recent evidence collected over 5 years (2008 – 2013) from the accumulation area of Russell Glacier [*Doyle et al*, 2014].

Although higher air temperatures in the future will likely cause increased meltwater production [*Graversen et al.,* 2011], much uncertainty still remains regarding the sensitivity and response of the ice sheet's dynamics to a warmer climate [*Church et al*, 2013; *Vaughan et al.,* 2013]. There is a need, therefore, to develop coupled process-based models of glacier surface mass balance and of surface and subglacial hydrology, in order to examine how current climate controls water delivery to the ice-sheet bed and affects subglacial water pressures, and how these processes will change in response to realistic scenarios of future climate change.

In this study, we apply an existing physically-based, subglacial hydrological model [*Banwell et al.,* 2013] to the Paakitsoq region, west Greenland. The model is fed with moulin input hydrographs calculated using a surface routing model [*Banwell et al.,* 2012b, 2013; *Arnold et al.,* 2014], which is driven with distributed runoff calculated by a positive degree-day (PDD) model. The PDD model is first validated against the output from a more physically-based SMB scheme for the 2005 melt season [*Banwell et al.,* 2012a, 2013; *Arnold et al.,* 2014]. We then use the PDD model to generate a suite of surface runoff grids into the 21st century in line with future climate scenarios, based on the IPCC's framework of RCPs. These are then used to force our surface and subglacial hydrological models. The model outputs of spatially and temporally varying subglacial water pressure distributions will help to inform the on-going debate surrounding the links between surface melt, basal sliding and surface velocity patterns on the GrIS, and will allow a better forecast of the response of marginal areas of the ice sheet to climate change over the 21st century.

2. Study site and available data

The Paakitsoq region (~2,300 km^2; Figure 1) is located on the western margin of GrIS [*Banwell et al.*, 2012a], northeast of Jakobshavn Isbrae. The region was chosen because of the availability of various data sets including (i) hourly meteorological data measured at three GC-Net stations, JAR 1, JAR 2, and Swiss Camp, used to drive the melt models [*Steffen and Box,* 2001]; (ii) coastal precipitation and temperature data for 1985 to 2004 from the Asiaq Greenland Survey Station 437 (190 m a.s.l., 4 km west of the ice margin), also used to drive the melt models; (iii) a 750 m resolution bed digital elevation model (DEM) [*Plummer et al.,* 2008] for the subglacial routing model (resampled to 100 m using bilinear interpolation), and a 30 m resolution surface DEM taken from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) global DEM for the surface melt and routing models (smoothed using a 6 cell medium filter and then resampled to 100 m using bilinear interpolation); and (iv) proglacial stream discharge data measured at the Asiaq Station for validation of the complete hydrological model through comparison of modeled and measured proglacial discharge [*Banwell et al.*, 2013]. In this study, we focus on a ~200 km^2 subglacial catchment (and its corresponding supraglacial catchment), which is entirely within the ablation area of the ice sheet, and extends ~25 km inland from the margin and feeds the proglacial Asiaq Station (Figure 1).

The major development presented here, compared to previous studies undertaken in the region [e.g. *Banwell et al.,* 2012a, 2012b, 2013; *Arnold et al.,* 2014], is that the melt/hydrological model is run into the future. For this, we use meteorological data from the Meteorological Research Institute's CGCM-3 model (version 20110831; ensemble r1i1p1; PCDMI), run as part of the 5th Climate Model Intercomparison Project (CMIP5) (see Section 3.3.2 for more details). Monthly precipitation and temperature values from 2006 to 2100 were retrieved for the grid cell incorporating Paakitsoq (68^o – 70^o N, 309^o – 311^o E) for three RCP scenarios; 2.6, 4.5 and 8.5 [*Church et al.,* 2013; *Vaughan et al.,* 2013]. The RCPs are defined by their total radiative forcing pathway (cumulative measure of human emissions of greenhouse gases from all sources expressed in W m^{-2}) and level by 2100, based on an internally consistent set of socioeconomic assumptions [*van Vuuren et al.*, 2011]. RCP 2.6 assumes a peak in radiative forcing at 2.6 W m^{-2} (~490 ppm CO_2 equivalent) before 2100 followed by a decline; RCP 4.5 describes a stabilization without overshoot pathway to 4.5 W m^{-2} (~650ppm CO2 equivalent) by 2100; and RCP 8.5 describes a rising radiative forcing pathway leading to 8.5 W m^{-2} (~1370 ppm CO_2 equivalent) by 2100.

3. Methods

3.1. The subglacial routing model

The subglacial routing model is derived from the Extended Transport (EXTRAN) block of the US Environmental Protection Agency's Storm Water Management Model (SWMM), which was originally designed to simulate sewage pipe systems [*Roesner et al.,* 1988]. *Arnold et al.* [1998] adapted the original EXTRAN code to model subglacial drainage through ice-walled conduits by including equations to simulate the dual processes of conduit enlargement due to the release of frictional heat in the flowing water, and conduit closure in response to ice deformation [*Spring and Hutter*, 1981]. The subglacial model has previously been applied successfully to both Haut Glacier d'Arolla, a valley glacier in Switzerland [*Arnold et al.,* 1998], and to the Paakitsoq region of the GrIS [*Banwell et al.*, 2013]. The model is only briefly described below as *Banwell et al.* [2013] provide a detailed description and performance analysis of the model in the same region we are modeling here.

There are key differences between the current study and that of *Banwell et al.* [2013]. By focusing solely on the 2005 melt season, *Banwell et al.* [2013] ran the model at a higher temporal resolution (1 h), and forced it with distributed runoff calculated by a SMB model (as opposed to a PDD model as we do here). This enabled *Banwell et al.* [2013] to investigate changes in subglacial water pressures patterns on intra-seasonal, daily, and hourly timescales. In contrast, the present study focuses primarily on inter-decadal changes in surface runoff and subglacial water pressure through the 21st century.

In its present form, our subglacial hydrology model routes water flow through a series of circular conduits that join at vertical 'junctions', with wider junctions (representing moulins) routing meltwater from the surface to the subglacial system. It is assumed that most of the surface water entering moulins flows quickly to the base of the ice sheet [cf. *Björnsson*, 1982]. The model for the study area is formulated such that the subglacial system is predominantly channelized (as opposed to distributed), which is a realistic assumption because ice within a few kilometres of the ice-sheet margin is relatively thin and therefore conducive to rapid development of channelized flow in the early part of the melt season [*Pimentel and Flowers*, 2010; *Banwell et al.*, 2013; *Sole et al*., 2013]. However, as the temporary storage and release of water in a distributed subglacial drainage system is not explicitly accounted for by the model, subglacial water routing may sometimes occur too rapidly in our model, notably in the early part of the melt season.

3.2. Subglacial drainage system analysis

Here, we briefly describe the key boundary conditions of the model: the subglacial catchment feeding the Asiaq station; the overall structure of the subglacial drainage system within this catchment; and the surface catchment feeding the moulins that feed the subglacial drainage system. As these boundary conditions are very similar to those employed by *Banwell et al.* [2013], we refer the reader to that study for a fuller description.

3.2.1. Subglacial catchment and drainage network delineation

To define the subglacial catchment area and the structure of the subglacial drainage network (i.e. the locations and connectivity of the drainage conduits), we assume that water flows along the steepest subglacial hydraulic potential gradient (following *Shreve* [1972]). The total subglacial hydraulic potential (Φ) (Pa) is the sum of the elevations, and pressure potentials and can be defined as:

$$\Phi = \rho_w \, g \, Z_b + k \, \rho_i \, g \, (Z_s - Z_b) \quad (1)$$

where ρ_w is water density (1000 kg m^{-3}), ρ_i is ice density (917 kg m^{-3}), g is acceleration due to gravity (9.81 m s^{-1}), Z_b is the bed elevation (m), Z_s is surface elevation (m), and k is a spatially uniform flotation fraction, defined as the ratio of water pressure to ice overburden pressure (P_w/P_i), with k = 1 representing water at the ice overburden pressure, and k = 0 representing atmospheric pressure (adapted from *Shreve* [1972] and following *Banwell et al.* [2013]).

First, using the 100 m resolution surface and bed DEMs, we calculate hydraulic potential surfaces for a range of realistic k values from 0.5 to 1.0 [*Thomsen and Olesen*, 1991; *Thomsen et al.*, 1991] (note that throughout the rest of this paper, the term ‘*k* value’ refers to the k in Equation 1). Although the k value is likely to be spatially and temporally variable in reality, it must be fixed for the purpose of defining the subglacial catchment area and drainage network and is considered to be a long-term average for steady-state conditions [*Hagen et al*., 2000; *Willis et al.*, 2012].

Second, to calculate the patterns of flow accumulation (i.e. upstream area) and thereby delineate the subglacial drainage network and catchment area for each k value, we run the lake and catchment identification algorithm (LCIA) [*Arnold*, 2010; later used by *Banwell et al.*, 2012b, 2013, *Arnold et al.*, 2014]. This algorithm allows us to delineate the subglacial drainage network and subglacial catchment area for each k value [*Banwell et al.*, 2012b, 2013]. See *Arnold* [2010] for a full description of the LCIA and *Banwell et al.* [2012b; 2013] and *Arnold et al.* [2014] for full details of its application to the Paakitsoq region.

3.2.2. Supraglacial catchment delineation

To calculate the surface meltwater input locations to the subglacial routing model, we assume that all depressions in the surface DEM contain an 'open' moulin in its lowest cell, implying that all lakes have already drained by hydrofracture to leave an open moulin. The assumption that a moulin has the potential to form in the lowest grid cell of every depression gives a moulin density of 0.25 km^{-2} [*Banwell et al*., 2013], which is similar to those mapped from satellite imagery by *Colgan and Steffen* [2009] (0–0.89 km^{-2}) and *Zwally et al.* [2002] (0.2 km^{-2}) for the Paakitsoq region.

This study differs to the study by *Banwell et al.* [2013], in which lakes drain only if they reach a threshold volume of water during the melt season (and in which lakes that do not reach a given threshold volume, overflow into downstream catchments). The 'open' moulin assumption reduces our model's ability to capture short-term fluctuations in water pressures (and associated inferred short-term fluctuations in ice velocity) resulting from lake drainage events. However, *Banwell et al.* [2013] concluded that longer-term periods of sustained water pressures (and associated longer-term fluctuations in ice velocity) are not a direct result of lake drainage events; instead, lake drainage events probably play a key role in opening up moulins, which can subsequently transport large quantities of water rapidly from the surface to the ice-bed interface for the remainder of the melt season. This is supported by evidence that lake drainage events have mainly short (< 1 - 2 days) effects on ice dynamics [*Das et al*. 2008; *Hoffman et al.*, 2011; *Doyle et al*., 2013; *Tedesco et al*., 2013]. Thus, our assumption that moulins are always 'open' is unlikely to have a significant effect on longer-term water pressures. Instead, our assumption that moulins are 'open' can be seen as an end-member that allows a maximum volume of surface meltwater to reach the subglacial drainage system through the melt season.

If we assume that moulins are vertical shafts routing water directly from the surface to the bed [*Björnsson,* 1982; *Catania et al.,* 2008], with each moulin having its own supraglacial catchment supplying it with runoff, the size and shape of the entire Paakitsoq supraglacial catchment is highly dependent on the shape of the subglacial catchment feeding the Asiaq station [*Banwell et al.,* 2013]. The LCIA is run for the surface DEM in order to identify which lake locations (assumed to all contain 'open' moulins) supply melt to the subglacial catchment for each specified *k* value.

3.2.3. *k* value selection

As explained more fully in *Banwell et al.* [2013], subglacial catchments defined for different *k* values will be associated with different volumes of surface meltwater due to the varying extents of the supraglacial catchments that supply water to the bed. In order to choose a suitable *k* value to define the subglacial drainage system structure and catchment area, *Banwell et al*. [2013] compared the total volume of the measured proglacial discharge at the Asiaq station to the total volume of modeled net runoff (calculated by their SMB model) within supraglacial catchments which supply melt to subglacial catchments delineated for *k* values ranging from 0.5 to 1 for the melt season of 2005. Although *Banwell et al*. [2013] found that a value of k = 0.925 produced the best agreement between modeled and observed runoff, a value of k = 0.95 produced the largest surface catchment feeding the Asiaq station. We adopt this value in this study as it enables us to investigate the impacts of introducing the largest volume of surface meltwater to the subglacial system that is physically plausible.

A value of k = 0.95, equivalent to an average subglacial water pressure that is 95% of ice overburden, may seem high, but as suggested by *Banwell et al.* [2013], it is likely that conduit paths become established early in the summer when water pressures are very high due to lower discharge. Once established, conduits are likely to remain fixed in those locations, since they are unlikely to migrate laterally to areas of the bed with a lower hydraulic potential. We also note that specific conduit locations are relatively insensitive to the range of *k* values we test [*Banwell,* unpublished PhD thesis, 2012], so predicted pressure fluctuations are unlikely to be a strong function of the *k* value used to determine the catchment size. Therefore we use k = 0.95 to determine: i) the size and shape of the subglacial catchment feeding the Asiaq station; ii) predict subglacial conduit paths; iii) specify the number and locations of moulins, and therefore the size and shape of the supraglacial catchment.

3.2.4. Subglacial network configuration

Figure 2 shows the inferred locations of individual conduits, moulins, junctions and outflow points overlaid onto the subglacial flow accumulation map for the subglacial catchment for $k = 0.95$. Junctions are placed along conduits, such that no conduit segment is longer than 1000 m (conduit segments longer than this reduce model stability [*Roesner et al*., 1988]). As also found by *Banwell et al*. [2013], all moulin locations fall almost exactly on the paths of subglacial conduits (within 100 m), giving us confidence that the modeled conduit locations follow realistic paths. For each model time step, water reaching all of the marginal outflow points is cumulated and compared with measured proglacial discharge for that time period.

All parameter values for conduits and moulins must be set at the beginning of the model run. Following *Banwell et al.* [2013], we assume conduits to have an initial cross-sectional area (CSA) of 3.14 m^2 (equivalent to a diameter of 2 m) and roughness of 0.05 $m^{-1/3}$ s^{-1}, moulins to have a fixed CSA of 2 m^2, and junctions which are not moulins to have a fixed CSA of 0.1 m^2. We also assume that all conduits are empty at the beginning of the model run. To prevent conduits from experiencing high creep closure rates at this time, we apply an initial 24 h spin-up period where no wall melt or creep closure occurs (i.e. the *Spring and Hutter* [1981] equations are turned 'off'), and a subsequent 24 h spin-up period where the *Spring and Hutter* [1981] equations gradually become effective in a linear way with time [*Banwell et al.*, 2013]. During this initial spin-up period (total time = 48 h), the total discharge in the subglacial system is very low (e.g. for 2005, the maximum discharge in hour 48 on June 2 is 8 m^3 s^{-1}, compared to a maximum of 206 m^3 s^{-1} on July 18).

3.3. Input hydrographs

The subglacial model is driven with moulin input hydrographs that are generated using the melt output from the PDD model, which is then routed in each sub-catchment to its appropriate moulin using a surface routing model. Full details of this surface routing model are given in *Banwell et al.* [2012b, 2013] and *Arnold et al.* [2014]. As previously mentioned, we assume that depressions in our surface DEM do not fill to form lakes; there is an 'open' moulin in the lowest cell of every depression.

In the following sections, we first describe the PDD model, and explain how it is used to generate the surface runoff for the 2005 mass balance year (1 September 2004 – 31 August 2005), in order to validate the PDD-modelled runoff against SMB-modelled runoff for the time period 1 June to 31 August 2005. Second, we outline our strategy for future climate forcing and explain how the PDD model is used to model surface runoff over the 21st century.

3.3.1. Positive degree-day model

Like all PDD models, our model is forced entirely using temperature data. Although the concept involves a simplification of complex processes that are more accurately described by the surface energy-balance equations, the approach is justified because of the high correlation between temperature and various components of the energy-balance equation [*Braithwaite*, 1981; *Ohmura*, 2001; *Hock*, 2005]. Longwave incoming radiation and the turbulent heat fluxes depend strongly on temperature, and temperature in turn is affected by shortwave radiation [*Ohmura*, 2001; *Hock*, 2005]. The PDD approach is therefore still used extensively for modeling GrIS surface melt [*Braithwaite*, 1995; *Abdalati et al.*, 2001; *Mote*, 2003; *Hanna et al.*, 2006; *Rignot and Kanagaratnam*, 2006], and has been shown to provide estimates of melt that are comparable to more complex EB modeling [*van de Wal*, 1996].

We use a degree-day factor (DDF) of 8.9 mm per PDD for ice [*Braithwaite and Olesen*, 1989; *Braithwaite*, 1995], and a DDF of 3.6 mm per PDD for snow [*McMillan et al.*, 2007]. Following *Arendt et al.* [2009], the PDD model calculates the total melt, M (mm water equivalent (w.e.)), produced in a surface grid cell at each time interval (Δt, equal to 24 hours), using the following equations:

$$M = -T(z)\ \delta[T(z)]\ \mathrm{DDF}_{\mathrm{snow/ice}}\ \Delta t + P(z)\ \delta[-T(z)] \qquad (2)$$

$$T(z) = T_{\mathrm{aws}} + (z - z_{\mathrm{aws}})\ \Gamma_T \qquad (3)$$

$$P(z) = P_{\mathrm{aws}}k + (z - z_{\mathrm{aws}})\ \Gamma_P\ P_{\mathrm{aws}} \qquad (4)$$

where T is the daily average air temperature (^{o}C), P is the daily total precipitation (rain and snow, mm w.e.), z is elevation (m), $\mathrm{DDF}_{\mathrm{snow/ice}}$ is the degree-day factor for snow/ice (mm $^{o}\mathrm{C}^{-1}\ \mathrm{d}^{-1}$), and the subscript 'aws' refers to values measured at the automatic weather station at JAR1. Values of T_{aws} and P_{aws} are adjusted for elevation using constant temperature and precipitation lapse rates Γ_T (^{o}C m^{-1}), Γ_P (% m^{-1}). δ determines the threshold between positive temperatures for melt and negative temperatures for accumulation of solid precipitation, such that $\delta[T] = 1$ when $T > 0$, and $\delta[T] = 0$ when $T \leq 0$.

The PDD model requires an initial grid of snow distribution across the entire supraglacial catchment. However, since the spatially distributed snow depth as calculated by the SMB model [*Banwell et al.*, 2013] on 31 August 2004 after it had been run for a full mass balance year (1 September 2003 to 31 August 2004) is zero, we initialise the PDD model with a zero snow depth. The model is then able to accumulate snow over the winter and into the following summer of the 2005 mass balance year.

Although refreezing in the snowpack is of no importance over the entire summer in the ablation zone of the GrIS, it is still an important factor to account for in the short-term as it can reduce the net amount of meltwater that becomes 'runoff' (i.e. the portion of water which does not refreeze in the snowpack) [*Lefebre et al.*, 2002; *van Pelt et al*., 2012]. Following *Radic and Hock* [2011], annual refreezing R (cm) is related to annual mean air temperature T_a (^oC) by

$$R = -0.69\, T_a + 0.0096 \qquad (5)$$

where the lower boundary of R is zero across the entire catchment glacier, and the upper boundary of R is assumed equal to accumulated snow in the ablation area. Daily melt refreezes until the accumulated melt in one day (i.e. 24 hours) exceeds the potential refreezing, at which point it is treated by the PDD model as runoff. For example, we calculate that between June 1 and August 31 2005, only ~0.5% of the total melt across the surface catchment refreezes early in the summer and does not become runoff immediately.

To validate the PDD model, we compared the daily runoff values calculated by the PDD model across the supraglacial catchment to the daily runoff values calculated by the SMB model used in the study by *Banwell et al*. [2013] for the time period June 1 to August 31 2005. The Pearson's correlation coefficient between these two data sets is 0.84 (significant at $p < 0.00001$). Furthermore, the total runoff calculated during the melt season by the PDD model (5.6×10^8 m^3) underestimates by only 8% the total runoff calculated by the SMB model (6.1×10^8 m^3).

We are using a form of static mass balance modelling, in which the surface elevation remains the same; a technique frequently used by other studies [e.g. *Bougamont et al.*, 2005; *de Woul* *and Hock*, 2005; *Radic and Hock*, 2006]. This is appropriate, as future changes in surface elevation will have a much smaller effect on surface runoff (as a result of lapse-rate driven air temperature changes) compared to the effects of RCP-driven air temperature changes. Moreover, the effect of surface mass loss (ice/snow) on the ice overburden pressure at the bed (and therefore conduit opening/closure rates) will have a much smaller effect on subglacial water pressures than will the future increase in meltwater entering the subglacial system due to increased surface melt. Finally, while future changes in surface topography might lead to minor changes in the size and shape of surface catchments, studies suggest that the locations of surface depressions, and therefore moulins, are unlikely to vary greatly over the next century due to the overriding control of bedrock topography [*Echelmeyer et al*., 1991; *Sergienko*, 2013].

3.3.2. Future climate forcing

We applied a statistical downscaling method, referred to as 'local scaling' [*Salathé*, 2005], to the CGCM-3 output (originally at a resolution of 125 km) in order to better represent local subgrid-scale features and dynamics [*Giorgi et al.,* 2001; *Radic and Hock*, 2011]. This method effectively corrects for the lapse rate by accounting for the elevation difference of the local grid point relative to the climate model grid, and has been shown to produce estimations of local temperature and precipitation that are comparable to empirical observations [e.g. *Radic and Hock,* 2006]. We subsequently bias-corrected the monthly climate model output series using the average difference over a period of 20 years (1 January 1985 – 31 December 2004) between the climate model data and monthly and precipitation temperature data measured at the Asiaq station. We calculated the future temperature time series (T_i) as:

$$T_i\,(t) = T_{i,\,\mathrm{GCMf}}\,(t) + [T_{i,\,\mathrm{measured}} - T_{i,\,\mathrm{GCMh}}] \qquad i = 1, \ldots 12 \qquad (6)$$

where $T_{i,\mathrm{GCMf}}$ is the mean monthly temperature (^{o}C) for month i from the future run of the GCM for the years (t) 2006 to 2100; $T_{i,\,\mathrm{measured}}$ is the mean measured temperature (^{o}C) for month i over the bias-correction period 1985 to 2004; $T_{i,\,\mathrm{GCMh}}$ is the mean temperature (^{o}C) of the historical run of GCM for month i over the bias-correction period 1985 to 2004.

To calculate future precipitation rates, the local scaling method simply multiplies the large-scale simulated precipitation at each local grid point by a seasonal scale factor; precipitation is scaled equally throughout the year. The future precipitation time series (P_i) is

$$P_i(t) = P_{i,\,\mathrm{GCMf}}(\mathrm{t}) * [P_{i,\,\mathrm{measured}} / P_{i,\,\mathrm{GCMh}}] \qquad i = 1, \ldots, 12 \qquad (7)$$

where $P_{i,\,\mathrm{GCMf}}$ is the monthly precipitation sum for month *i* from the future run of the GCM for the period 2006 to 2100; $P_{i,\mathrm{measured}}$ is the mean measured precipitation, for month *i*, over the bias-correction period 1985 to 2004; $P_{i,\mathrm{GCMh}}$ is the mean precipitation of historical run of GCM for month *i* over the bias-correction period 1985 to 2004.

Using the output from the three RCP scenarios (2.6, 4.5 and 8.5), and with the initial assumption of zero snow depth on 31 August, we ran the PDD model for three chosen mass balance 'years' over the next century (2025, 2050 and 2095). We used the mean of the climate data from the decade around each of the three chosen years to improve reliability (e.g. the decade of 2020 to 2030 was used to represent the year 2025). For each of the three years, we used the calculated surface runoff for the time period 1 June to 31 August to drive the surface routing model, which produced moulin hydrographs to drive the subglacial routing model.

As the PDD model requires daily meteorological data, and the future climate data is only monthly, we calculated the average temperature and precipitation per day using a ten-year (1995 – 2004) baseline period of temperature and precipitation data measured at the Asiaq station. A 'baseline year', which we call the year 2000 hereafter, was aggregated from the baseline period in a similar way to how the RCP runs were averaged over the decade around one year. To calculate daily temperature, the mean monthly temperature average for 2000 was linearly interpolated across consecutive months, from the 15$^{\mathrm{th}}$ day of one month to the 15$^{\mathrm{th}}$ day of the next. We calculated an additive factor to relate the 2000 daily temperature to the mean monthly average, thus enabling us to estimate future daily temperatures. To calculate daily precipitation, we calculated the mean number of 'precipitation days' (defined as days where precipitation > 0 mm) per month in 2000, and divided equally the total modeled monthly precipitation by the number of precipitation days, to give the average precipitation per day (on the days on which precipitation occurred).

4. Results

4.1. Surface runoff through the 21$^{\mathrm{st}}$ century

Figure 3 displays the total modeled summer runoff volumes over the Paakitsoq catchment for the three target years for each of the three RCPs, and for the baseline year (2000). Under RCP

2.6, runoff is predicted to increase from 3.46 x 10^8 m^3 in 2025 to a maximum of 5.30 x 10^8 m^3 by 2050, and then drop back down to 4.15 x 10^8 m^3 by 2095. Thus, under RCP 2.6, runoff remains comparable to 2000 (4.01 x 10^8 m^3), although it is somewhat greater than that during the middle part of the century. Under RCP 4.5, runoff is higher than in 2000 and steadily increases over the century, from 4.63 x 10^8 m^3 in 2025 to 6.47 x 10^8 m^3 in 2095. RCP 8.5 runs show a marked increase in runoff in the latter half of the century, with runoff volume almost doubling from 6.86 x 10^8 m^3 in 2050 to 13.3 x 10^8 m^3 by 2095.

The calculated daily runoff series for 2025, 2050 and 2095 under the three RCP scenarios and for 2000 are shown in Figure 4. Although runoff series for all three RCP scenarios are similar in pattern with each other and with the series for 2000, the changes in runoff magnitude compared to 2000 do not vary evenly throughout the melt season – and this general finding becomes even more apparent as the 21st century progresses. In 2025, runoff under RCP 2.6 is always lower in magnitude than in 2000; under RCP 4.5, runoff is comparable in magnitude with 2000 in June and August, but substantially greater in July; under RCP 8.5, runoff magnitude is comparable with that in 2000 in June, greater in July (though not as high as under RCP 4.5), and lower than 2000 and both the other RCP scenarios in August. By 2050, runoff under RCP 2.6 is comparable in magnitude with that in 2000 throughout the melt season; under RCP 4.5, runoff magnitude is comparable with 2000 in June, but increasingly rises above it in July and August; under RCP 8.5, runoff magnitude is approximately twice that for 2000 from early June to late July, but decreases in August. Finally, by 2095, runoff magnitude under RCP 2.5 is slightly lower than that in 2000 in June, and slightly above in August; under RCP 4.5, runoff is substantially above that in 2000 throughout the summer; and under RCP 8.5, runoff is four times greater than in 2000, with most of July and August experiencing runoff volumes > 15 x 10^6 m^3 d^{-1}.

As the century progresses, and for the more intense RCP scenarios, our model suggests an increase in the areal extent of high surface runoff. Figure 5 shows that by 2095 under RCP 4.5 and RCP 8.5, the total surface area experiencing > 1000 mm w.e. runoff extends further inland compared to 2000. However, under RCP 2.6, there are few observable differences in the extent and magnitude of surface runoff by the end of the century compared to 2000. The most noticeable increase in the extent and magnitude of surface runoff occurs under RCP 8.5, where surface runoff for 2095 along the margin of the ice sheet (~8500 mm w.e.) is about twice what it is in 2000, and surface runoff production at the furthest inland part of the surface catchment (~4000 mm w.e.) is four times what it is in 2000. Notably, the surface runoff production at the most inland part of the surface catchment in 2095 is comparable to what it is nearest to the margin in 2000 under RCP 8.5.

4.2. Subglacial water pressure through the 21st century

For each model run, daily subglacial water pressure is calculated in all 47 moulins and 95 junctions shown in Figure 2. Here we analyse the water pressures variations for each future year for each RCP scenario and for 2000. We do this for the sample of 11 moulins and 6 junctions labelled in white text on Figure 2, which are representative of the hydrological conditions beneath different parts of the entire catchment. We carry out two stages of analysis. First, to highlight the difference between the lower and upper ablation areas, we group the moulins/junctions into: i) those < 10 km of the ice margin; and ii) those > 10 km from the margin, and analyse the average value of P_w/P_i through the melt season for the different model runs. Second, to quantify the amount of time during the melt season that basal sliding is likely to be high, we analyse the percentage of time throughout the summer that each moulin/junction is at or above ice overburden pressure (i.e. $P_w/P_i \geq 1$) for each model run. This threshold is based upon previous observational [*Iken and Bindschadler*, 1986; *Kamb*, 2001; *Andrews et al.*, 2014] and modeling [*Schoof*, 2010; *Hewitt*, 2013] studies that suggest that enhanced basal sliding is likely to occur when subglacial water pressures approach or exceed ice overburden pressures.

Figure 6 shows the results of the first stage of analysis. For each future year for each RCP scenario, and for 2000, the variation in average P_w/P_i is shown for a) moulins/junctions < 10 km of the ice margin; and b) moulins/junctions > 10 km of the ice margin. In general, within each group of moulins/junctions, the overall patterns in P_w/P_i appear to follow a similar trend for all future years and for all RCP scenarios. The time series show an early-season peak in P_w/P_i (higher and more pronounced for moulins > 10 km from the margin), followed by a period of elevated water pressure (again, generally longer and with higher P_w/P_i, for moulins > 10 km from the margin), and then a decrease to a lower, fluctuating, mid- to late-season value.

With the exception of RCP 2.6, where the maximum runoff occurs in 2050 instead of 2095, the highest peak in P_w/P_i occurs earlier in the melt season in 2095 compared to 2050, and earlier in 2050 compared to 2025 (Figure 6). For RCP 2.6, P_w/P_i peaks earliest in 2050, followed by 2095, then 2025. Again, with the exception of RCP 2.6, P_w/P_i, values also decrease to their lower mid- to late-season value earlier in the melt season in 2095 than in 2050, and earlier in 2050 compared to 2025 (i.e. compared to the mean value of P_w/P_i in the first few weeks of the melt season). For 2.6, P_w/P_i values decrease to their lower mid- to late-season mean earliest in 2050, followed by 2095, then 2025. Similarly, as the RCP scenarios get more extreme, the peaks in P_w/P_i also tend to occur earlier in the melt season, and then decrease to the mid- to late-season value earlier in the melt season than for less extreme RCP scenarios. However, when 2025 under RCP 2.6 is compared to 2000, we find that the transition to a lower mid- to late-season P_w/P_i value occurs even later than in 2000, and the peak in P_w/P_i also occurs even later than in 2000 (Figure 6). This is consistent with the result that the total modeled runoff for 2025 under RCP scenario 2.6 is less than the total modeled runoff for 2000 (Figure 3).

For moulins/junctions < 10 km of the margin (Figure 6a), subglacial water pressure fluctuates ultimately around a mid- to late-season mean $P_w/P_i \approx 0.45$, and this is reached by ~10 July for the majority of years and RCP scenarios. After the initial filling of the subglacial drainage system (i.e. from 1 to 10 June), and before the lower mid- to late-season P_w/P_i is reached, this group of moulins fluctuates around $P_w/P_i \approx 0.6$ (often peaking at a maximum of ~0.7 and decreasing to a minimum of ~0.45).

For moulins/junctions > 10 km of the margin (Figure 6b), subglacial water pressure fluctuates eventually around a mid- to late-season mean $P_w/P_i \approx 0.6$, which is reached by ~20 July for most years and RCP scenarios. After the initial subglacial drainage system filling, and before the mid- to late-season mean P_w/P_i is reached, this group of moulins fluctuates around $P_w/P_i \approx$ 0.8 (often peaking at a maximum of just over 1.0 and decreasing to the minimum of ~0.65).

For the purpose of the second stage of our analysis – identifying the percentage of time that moulins/junctions are at or above ice overburden pressure during the model run – three moulins and four junctions, those located < 5 km of the ice margin, are excluded from the analysis because the percentage of time that $P_w/P_i \geq 1$ is <1%. We analyse the pressures in the remaining 8 moulins (481, 494, 519, 532, 564, 582, 619 and 624) and 2 junctions (1014 and 10221) that are > 5 km from the margin (Figure 2).

Table 1 shows the average percentage of time that $P_w/P_i \geq 1$ for the 10 selected moulins/junctions, over the melt seasons of the three future years under each RCP scenario and for 2000. With the exception of RCP 2.6 for 2025, the average percentage of time that $P_w/P_i \geq 1$ decreases over the 21st century for each RCP scenario. Under RCP 2.6 and RCP 4.5, the largest decline in the percentage of time that $P_w/P_i \geq 1$ occurs between 2025 and 2050, whereas under RCP 8.5, the largest decline occurs between 2050 and 2095.

Under RCP 2.6, the average percentage of time that $P_w/P_i \geq 1$ increases slightly between 2000 and 2025 (by ~0.8%), decreases from 2025 until 2050 (by ~3%), and again decreases from 2050 until 2095 (by ~1.1%) (Table 1). The exception to this general trend is junction 1014 (Figure 2) (located in an area of relatively thick ice; ~530 m), where the percentage of time that $P_w/P_i \geq 1$ increases from 2000 until 2050 (by ~0.6%), before decreasing, like the other moulins/junctions, from 2050 until 2095 (by ~1.4%).

Under RCP 4.5, the percentage of time that $P_w/P_i \geq 1$ decreases slightly from 2000 until 2095. In the same way as under RCP 2.6, a larger decrease in subglacial water pressure occurs between 2025 and 2050 (~1.8%), than between 2000 and 2025 (~0.5%), and between 2050 and 2095 (~0.6%). However, between 2050 and 2095, four moulins/junctions (481, 494, 519 and 1014) experience a slight increase in pressure (~0.9%). These moulins/junctions are positioned under some of the thickest ice (mean = 565 m) and are also > 10 km from the ice margin.

Under RCP 8.5, the percentage of time that $P_w/P_i \geq 1$ decreases between 2000 and 2095, similar to RCP 4.5. But unlike under RCP 4.5, a larger decrease in pressure occurs between 2050 and 2095 (~2.2%) than occurs between both 2000 and 2025 (~1.4%) and 2025 and 2050 (~1.5%). However, four moulins/junctions (481, 494, 519 and 582) experience either a small increase or decrease (~0.5%) in pressure between 2000 and 2025, before experiencing a noticeable increase (~4.7%) in pressure between 2025 and 2050, and a noticeable decrease (~3.8%) from 2050 until 2095. These moulins/junctions are positioned under some of the thickest ice (> 500 m) and are also > 10 km from the ice margin.

5. Discussion

5.1. Variations in runoff through the 21st century

The PDD model output for RCP scenarios 2.6, 4.5 and 8.5 suggests that summer surface runoff generally increases in magnitude throughout the 21st century at Paakitsoq. The exception is for RCP 2.6, where the total summer runoff for 2025 is slightly less than that for

2000, and where a small decrease in summer air temperatures from the middle to the end of the century results in lower summer runoff for 2095 than for 2050 (Figure 3). The general trend of increasing runoff is mainly due to an increase in meltwater production (due to increased air temperatures) as opposed to an increase in liquid precipitation. Given our focus on the ablation zone, this dominance increases with higher air temperatures, as snow is removed increasingly quickly to expose the lower-albedo ice surface below.

Although the summer average air temperatures increase under most RCP scenarios in the future, causing an increase in total summer runoff, the temperature increases do not occur evenly throughout the summer, and in some cases monthly temperatures decrease compared to 2000. The most obvious example of this is for the RCP 8.5 scenarios where August temperatures are relatively low compared with other scenarios. This results in lower runoff volumes in August than might be expected (Figure 4). This contrasts with the situation in June where RCP 8.5 temperature increases are relatively high compared to other scenarios, resulting in greater runoff volumes in June. The timing of future runoff increases (or decreases) during the summer might be expected to influence the evolution of the subglacial drainage system and patterns of steady-state and transient water pressure fluctuations over the summer.

Our results also suggest that as the century progresses, the areal extent of high surface runoff migrates inland and therefore enlarges (Figure 5). The upper region of the surface catchment experiences a four-fold increase in surface runoff from 2000 to 2095, whereas the marginal area of the surface catchment experiences only a doubling of runoff from 2000 to 2095 (Figure 5). This result is partly due to significant albedo feedback in the upper regions. A low albedo ice-surface predominates in the marginal regions for the majority of the melt season, even in 2000. However, a higher albedo snow-covered surface remains for the majority of the melt season in the upper regions in 2000, but is removed and replaced by a lower albedo ice-surface much earlier in the melt season by 2095. The fact that different parts of the catchment will experience different rates of runoff increase in the future might also be expected to affect the way in which the subglacial drainage system evolves over the summer, and patterns of steady-state and transient water pressure fluctuations might be expected to change more in some places than others.

5.2. Variations in subglacial water pressure through the 21st century

Although our model does not explicitly simulate the transition from a distributed system to a channelized system, the transition to a lower mean P_w/P_i during the melt season indicates that the season-long evolution of the conduits themselves increases the efficiency of the system from small, constricted conduits early in the summer, to larger, more efficient conduits later in the season (Figure 6). Additionally, the finding that the transition to a lower mean P_w/P_i occurs earlier for conduits nearer the margin (i.e. where ice is relatively thin and runoff rates are relatively high) (Figure 6a), than for those higher up in the catchment (i.e. where ice is thicker and runoff rates are lower) (Figure 6b), indicates an upglacier progression in the evolution of conduit efficiency throughout the summer. These findings are consistent with several previous studies undertaken in marginal areas of the GrIS [e.g. *Bartholomew et al*., 2010, 2011; *Colgan et al.*, 2012; *Banwell et al.*, 2013; Sole et al., 2013].

Given that: i) a low mean P_w/P_i value tends to be reached earlier in the summer for model runs with higher available surface runoff (i.e. runs under the more extreme RCP scenarios, and runs later in the century) (Figure 6); and ii) the percentage of time that $P_w/P_i \geq 1$ tends to decrease as surface runoff increases (Table 1), we infer that the transition from a relatively inefficient to a more efficient subglacial drainage system occurs earlier in the melt season as volumes of available surface runoff increase, in agreement with theory [e.g. *Rothlisberger*, 1972]. This is consistent with the result that the modeled runoff for 2025 under RCP 2.6 is less than for 2000 (Figure 3), and as a consequence the transition to a lower mid- to late-season pressure mean occurs earlier for 2000 than it does for 2025 under RCP 2.6 (Figure 6).

Uncertainty remains about whether future increases in surface runoff (and therefore increases in subglacial discharge) will increase or decrease basal sliding, and thus ice velocities, over short (days to weeks) and long (months to years) timescales. Given our finding that the subglacial drainage system generally transitions from an inefficient to a more efficient system earlier in the melt season as the century progresses and as RCP scenarios become more extreme, we suggest that future increases in surface runoff to the subglacial hydrological system will to lead to an overall reduction in ice velocities over monthly and yearly timescales. This conclusion is consistent with previous work undertaken in marginal areas of the GrIS, where the ice is sufficiently thin to enable an efficient subglacial system to become established during the melt season [e.g. *Bartholomew et al.*, 2010; *Schoof*, 2010; *Sundal et al.*, 2011; *Sole et al.*, 2013; *Tedstone et al.*, 2013].

Although the overall trend is a decrease in water pressure associated with an increase in subglacial system efficiency through the century and as RCP scenarios become more extreme, some moulins/junctions under particularly thick ice (> 500 m) exhibit slightly different behaviour. For example, under RCP 2.6, the percentage of time that $P_w/P_i \geq 1$ for junction 1014 (Figure 2) increases from 2025 until 2050, rather than decreasing like the other moulins/junctions. Similarly, under RCP 8.5, four moulins/junctions (481, 494, 519 and 582, Figure 2) experience a noticeable increase (~4.7%) in the percentage of time that $P_w/P_i \geq 1$ between 2025 and 2050, rather than a decrease like the rest of the moulins/junctions. This suggests that a certain runoff threshold is needed for the subglacial drainage system to experience a decrease, rather than an increase, in water pressure, and that this runoff threshold is higher for conduits beneath thicker ice than for those under thinner ice. Consequently, the transition from early-season high water pressure to mid- to late-season low pressure will occur latest for the thicker regions of the ice sheet. With this reasoning, we suggest that for inland ice that is above a certain thickness, the runoff threshold may not be reached under any of the RCP scenarios investigated in this study, meaning that water pressures could continue to increase, rather than decrease, in response to increases in runoff over the 21st century. This is supported by a recent study by *Doyle et al.* [2014], who presented observational data from > 100 km from the GrIS margin and demonstrated an average increase in ice velocities from mid- to late melt season.

However, over shorter timescales, our results suggest that warmer (cooler) periods can cause short-term increases (decreases) in water pressure, and, by implication, sliding velocities. For example, short-term variations (over ~3–10 days) in subglacial water pressure occur in our modeled runoff series from early August onwards, even though the system has transitioned to conduits with a lower mean water pressure by then (Figure 6). This finding is consistent with previous modeling studies [e.g., *Schoof*, 2010; *Bartholomew et al*., 2012; *Banwell et al.,* 2013] that show how temporary imbalances between the rate of water delivery to the subglacial drainage system and its ability to evacuate the water are likely to result in short-term spikes in subglacial water pressure. We also find that these late melt season pressure variations are more pronounced in the moulins/junctions > 10 km from the margin (Figure 6b), than those closer to the margin (Figure 6a). This is because conduits under thicker ice rapidly close during times of low runoff inflow, lowering the capacity of the system, and thus enabling higher water pressures to be produced when inflow to the system increases. Finally, we find that the late melt season pressure variations are more pronounced for years later in the century and for more extreme RCP scenarios. For example, for 2050 and 2095, under RCP 8.5, the late melt season pressure fluctuations for moulins/junctions > 10 km of the margin are higher in amplitude than for other years and RCP scenarios (Figure 6b). This suggests that as runoff increases in future years, the higher late season pressure fluctuations may go some way to compensate for reduced ice velocities due to the earlier drop in the mean water pressure.

6. Conclusions

We have used a subglacial hydrology model, driven by output from a surface runoff and routing model, to simulate the likely responses of the subglacial drainage system at Paakitsoq (West Greenland) to climate warming during the 21st century. The surface runoff model calculates runoff using a PDD approach, and is driven by future climate scenarios for the 21st century based on the IPCC's RCPs 2.6, 4.5 and 8.5. Our main findings are:

- Under most future RCP scenarios, surface runoff increases throughout the 21st century. The exception to this is under RCP 2.6 where the modeled runoff decreases between 2050 and 2095, and the modeled runoff in 2025 is less than the modeled runoff for the baseline year (2000). The highest modeled runoff is for 2095 under RCP 8.5, when a ~7^{o}C warming results in a four-fold increase in runoff in the upper regions of the catchment.
- Although our model does not explicitly simulate the transition from a distributed to channelized subglacial drainage system, the season-long evolution of the conduits themselves increases the efficiency of the system from small, constricted conduits early in the melt seasons to larger, more efficient conduits later on. On a seasonal basis, we therefore capture the behavior of the drainage system inferred from previous observations [e.g. *Bartholomew et al.,* 2011a; *Hoffman et al*., 2011; *Moon et al*., 2014].
- The timing of the transition from a less efficient subglacial drainage system to a more efficient subglacial system for a marginal area of the GrIS (< 20 km of the ice margin) is dependent on the availability of surface runoff. As the century progresses, and/or as RCP scenarios become more extreme, runoff production generally increases, and the subglacial drainage system makes an earlier transition from a less efficient network operating at high water pressures to a more efficient network with lower water

pressures. An upglacier progression in the evolution of conduit efficiency is also observed throughout the summer.

- The earlier transition to an efficient subglacial drainage system throughout the 21st century (under most RCP scenarios) will likely cause an overall decrease in ice velocities for the marginal, Paakitsoq region of the GrIS. However, daily and weekly variations in surface runoff will cause short-term variations in subglacial water pressure, and by implication, ice velocities, even after the system has transitioned to the lower mid- to late-season pressure mean. These late season variations in subglacial water pressure are likely to become more pronounced as runoff increases during the 21st century, thus the associated velocity increases may go some way to compensate for the earlier increase in conduit efficiency.
- We suggest that for areas of the GrIS located further inland than our study region, where ice thicknesses are greater, an overall drop in average water pressure during the melt season may not occur. Instead, future increases in runoff availability may act to further increase subglacial water pressures, leading to increased basal sliding and ice velocities [e.g. *Doyle et al.,* 2014]. Future modeling work that facilitates the coupling of glacier hydrology and basal sliding and extends further into the ice sheet is required to test this hypothesis.

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Tables

RCP	**Year**	**% of time $P_w/P_i \geq 1$**
-	2000 (baseline)	11.6
2.6	2025	12.8
	2050	9.8
	2095	8.7
4.5	2025	11.1
	2050	9.3
	2095	8.7
8.5	2025	10.2
	2050	8.7
	2095	6.5

Table 1: The average percentage of time when $P_w/P_i \geq 1$ for the 10 selected moulins/junctions (see Section 4.2 and Figure 2) from 1 June to 31 August 2005, under RCP scenarios 2.6, 4.5 and 8.5, and for the years, 2025, 2050 and 2095. Also shown is the baseline year (2000).

Figures

Figure 1: Paakitsoq region (red box). Green outline shows the subglacial catchment feeding the Asiaq gauging station (green triangle for k = 0.95). Coordinates refer to UTM Zone 22°. The base Landsat 7 ETM+ image is dated 7 July 2001.

Figure 2: Conduit (black lines), moulin (black dots), junction (red dots), and outflow (green dots) locations overlaid onto the subglacial flow accumulation map for the subglacial catchment for k = 0.95 (see Figure 1 for location). Outflow locations not linked to upstream conduits indicate outflow from small marginal supraglacial catchments. The green triangle marks the Asiaq gauging station. Numbers in white indicate moulins/junctions that are referred to in the main text.

Figure 3: Total modeled runoff for the 3 RCPs (2.6 (blue), 4.5 (green), and 8.5 (red)) and 3 target years (2025, 2050, and 2095), for the period 1 June to 31 August 2005. Also shown is the total modeled runoff volume for the baseline year (2000, (gray)).

Figure 4: Daily modeled runoff volumes over the k = 0.95 catchment, for the baseline year (2000 (black)), and for the 3 RCPs (2.6 (blue), 4.5 (green), and 8.5 (red)) and 3 target years (2025, 2050, and 2095).

Figure 5: Total modeled surface runoff from 1 June to 31 August 2005 for the Paakitsoq region, shown for the baseline year (2000), and for 2095 in all three RCP scenarios (2.6, 4.5 and 8.5). The outline of the k = 0.95 surface catchment is shown.

Figure 6: Mean modeled P_w/P_i for selected moulins/junctions: a) < 10 km from the ice-sheet margin; and b) > 10 km from the margin, for the baseline year (2000 (black)), and for the 3 RCPs (2.6 (blue), 4.5 (green), and 8.5 (red)) and 3 target years (2025, 2050, and 2095).