Modeling supraglacial water routing and lake filling on the Greenland Ice Sheet

Alison F. Banwell, Neil S. Arnold, Ian C. Willis, Marco Tedesco, and Andreas P. Ahlstrøm (2012), Modeling supraglacial water routing area of the Paakitsoq region, west Greenland. The model rates forms a key development in successfully modeling the possible impacts of lake Marco Tedesco, Echelmeyer et al. calculates flow paths and water velocities over the snow-/ice-covered surface, routing the Seale et al. 2007 Greenland Ice Sheet (GrIS) is important as the average Neil S. Arnold, Ian C. Willis, measured over 117 changes in temperature, density and water content in the snow, firn and upper ice layers, and hence refreezing and therefore net runoff. The model is calibrated against field measurements of a filling lake in our study area during June 2011 and can be used to calculate the filling rate of the instrumented lake with a high degree of accuracy. The filling rate of the instrumented/modelled lake depends on melt and routing from the immediate lake catchment and from overflowing lakes in upstream catchments.


1. Introduction

[2] Assessing the impact of climate change on the Greenland Ice Sheet (GrIS) is important as the average temperature rise in the Arctic has increased at almost twice the global average rate in the past 100 years [Meehl et al., 2007]. Recent estimates from the Gravity Recovery and Climate Experiment (GRACE) indicate that the rate at which the GrIS is losing mass has accelerated from 137 Gt yr⁻¹ measured over 2002–2003, to 286 Gt yr⁻¹ measured over 2007–2009, and then to 430 Gt yr⁻¹ measured over 2010–2011 [Velicogna and Wahr, 2006; Velicogna, 2009; Box et al., 2011]. This rapid increase in mass loss is mainly a result of increased wastage at the margins [Alley et al., 2010; Chen et al., 2011; Seale et al., 2011; Zwally et al., 2011], due partly to increased iceberg calving and associated acceleration and thinning of marine-terminating outlet glaciers [Rignot and Kanagaratnam, 2006; Howat et al., 2011], but also to increased surface meltwater runoff affecting both marine- and land-terminating outlet glaciers [Mote, 2007; van den Broeke et al., 2009], which may be facilitated, at least in part, by increases in the melt area extent [Tedesco, 2007; Fettweis et al., 2010].

During the melt season, the ablation zone of the GrIS undergoes extensive melting of the upper layers of snow, firn and ice. Most of this melt becomes runoff [Box et al., 2006; Hanna et al., 2008] and may eventually drain into the ice sheet via fractures and moulins [Catania et al., 2008]. Some of the water may, temporarily at least, accumulate in topographic lows to form supraglacial lakes [McMillan et al., 2007; Selmes et al., 2011; Tedesco and Steiner, 2011]. These lakes form predominantly in compressive flow regimes [Das et al., 2008; Krawczynski et al., 2009], and are generally found in the same location from year to year [Thomsen et al., 1988; Echelmeyer et al., 1991] due to the strong influence of the basal topography on ice sheet surface topography [Box and Whitter, 2007; Lampkin and Vanderberg, 2011]. During the spring and early summer, lake areas and volumes typically increase. The size of a lake depends on (1) the size of the topographic depression; (2) lateral percolation of...
suggested that this mechanism may be responsible for seasonal variations in surface speeds of up to 200% on western margins of the GrIS [Joughin et al., 2008; Shepherd et al., 2009; Bartholomew et al., 2010; Palmer et al., 2011]. This surface melt-induced acceleration of flow [Zwally et al., 2002] is not explicitly accounted for in current predictive ice sheet models, leading to possible underestimation in Intergovernmental Panel on Climate Change (IPCC) projections of future sea level change [Rignot and Kanagaratnam, 2006; Meehl et al., 2007]. Additionally, if average air temperatures over Greenland continue to increase the intensity of surface melt, recent evidence suggests that lakes will drain by hydrofracture earlier in the melt season and that the lake population will extend to higher elevations, likely exposing an increased inland area of the ice sheet to sudden lake drainage events and subsequent hydraulic connections between surface and bed [Sundal et al., 2011; Liang et al., 2012].

[5] Because of the important role of supraglacial lakes in the hydrology of the GrIS surface as well as the bed, the routing of meltwater to surface depressions, and thereby the filling, overflowing, and ultimately drainage of surface lakes by hydrofracture should be included in any future high-resolution model of ice dynamics applied to the ice sheet margin. To this end, we report here on the development and calibration of a water routing and lake filling model which uses the calculated surface runoff from an existing distributed, high spatial resolution model of surface mass balance (SMB) [Rye et al., 2010; Banwell et al., 2012] to calculate supraglacial lake filling rates over a 100 km² area of the Paakitsoq region of the west central GrIS. Although the dynamics of the ice sheet as a whole are likely to be most influenced by the cumulative effect of multiple lake drainage events across much greater spatial and temporal ranges than explored in this study, this work is an important intermediate step that is necessary before more simplified models of the same process can be developed over larger spatial scales and longer temporal scales. Further work will be undertaken in due course to validate the model against other independent data sets and to investigate the filling rates of lakes over a larger area and in future climate warming scenarios.

2. Study Site and Data

[6] Our study site, the “Ponting” area, is a 10 km × 10 km area located within the ~2300 km² Paakitsoq region in west central Greenland, northeast of Jakobshavn Isbrae (Figure 1). This area was chosen as a field site because satellite imagery showed the consistent filling and drainage of lakes in this region on an annual timescale at the time of our planned field season. The Ponting area is ~15 km inland from the ice sheet margin with ice elevations between ~750 m and ~980 m above sea level (masl). In this study we focus on “Lake Ponting,” located centrally within our model domain (69.589°N, -49.783°E, 962 masl) (Figure 2), ~10 km north of the JAR 1 Greenland Climate Network (GC-Net) automatic weather station [Steffen and Box, 2001]. We focus on the time period 9 to 30 June 2011, during which we have field measurements of snow depth and lake level, which we use for model calibration.

[7] The SMB model used to calculate melt/runoff (section 3.1) is driven with a full range of meteorological variables...
from the JAR 1 GC-Net station [Steffen and Box, 2001]. The
SMB model is calibrated using average daily snow surface
height measurements against four poles drilled into the ice
(located with 100 m of “Camp,” Figure 2) from 9 to 29 June
2011, during which time no precipitation fell. [10] The surface routing and lake filling model (SRLF model) (section 3.2) is calibrated against measured lake level
data. Two HOBO pressure sensors were installed in Lake
Ponting’s lake basin on 13 June 2011. The lake drained via
hydrofracture on 19 June 2011, giving 6 days of lake filling
data. One sensor was firmly secured to an aluminum pole
drilled into the ice at a height of ~0.5 m above the ice sur-
face. The second sensor was loosely attached to the pole so
that it rested on the ice but could slide down the pole as the
bottom of the lake melted, while remaining close to the pole.
This setup allowed for the ablation rate at the bottom of the
lake to be calculated from the difference between the time
series of lake depth recorded by the two sensors [Tedesco
et al., 2012]. Both sensors recorded pressure every five min-
utes on an internal data logger and were recovered following
lake drainage. The sensors have a water level accuracy of
0.5 cm and a resolution of 0.21 cm. Pressure data were
corrected for altitude and for barometric pressure changes
using data from a third sensor located ~1 km from the lake.
Measured lake depth data were converted to volume using a
depth-volume relationship derived from a digital elevation
model (DEM) of the area (see below). A handheld GPS
allowed us to measure the horizontal position of the sensors
in 5 m. Comparing this position with the
100 m resolution DEM suggested that the sensors were
installed in a DEM cell that was 0.7 m higher than the lowest
11 elevation DEM cell of the lake basin; we therefore add 0.7 m
to all the pressure sensor data so they are relative to the
deepest part of the lake, and we define the depth-volume
relationship for this lowest DEM cell.

In addition to converting lake depth to volume, a sur-
face DEM is also required by the SMB model to spatially
distribute meteorological data and compute topographic
shading, slope angles and aspects (section 3.1). The DEM is
subsequently used by the SRLF model to route the modeled
meltwater across the snow/ice surface to topographic lows to
form lakes (section 3.2). We use the ASTER Global Digital
Elevation Model (GDEM) which has a nominal grid size of
30 m (http://asterweb.jpl.nasa.gov/gdem.asp). We checked
the original GDEM for obvious artifacts in the area and none
were found. The GDEM quality files for the Paakitsoq
region show ASTER stacking numbers here lie between
8 and 12, yielding an accuracy of ±18.2 m (SD) <500 m
elevation, and ±13.8 m (SD) >500 m [MacFerrin, 2011].
The original data were smoothed with a 6 × 6 cell median
filter to remove small-scale noise then resampled to a 100 m
resolution using bilinear interpolation.

3. Methods

3.1. Surface Mass Balance Model

We model hourly melt and runoff using the high-res-
olution SMB model described by Rye et al. [2010], and
subsequently developed by Banwell et al. [2012]. The SMB
model consists of three coupled components: (1) an energy
balance component that calculates the energy exchange
between the glacier surface and the atmosphere; (2) an
accumulation routine; and (3) a subsurface component,
which simulates changes in temperature, density and water
content in the snow, firm and upper ice layers, and hence
refreezing and net runoff. Here, we describe the SMB model
only briefly, concentrating on the adaptations that have been
made to the model presented more fully by Rye et al. [2010]
and Banwell et al. [2012].

The SMB model uses a range of meteorological vari-
ables from the JAR1 GC-Net station notably: incoming
shortwave radiation (diffuse and direct), air temperature,
relative humidity, and wind speed readings at a nominal
height of 2 m above the ice surface. As incoming longwave
radiation data are not available for the Paakitsoq region, these
data were calculated using parameterizations based on the
work of Konzelmann et al. [1994] following Banwell et al.
[2012].

Following Banwell et al. [2012], the mass balance
year runs from 1 October to 30 September. We are particu-
larly interested in model output during the times of our
measurements, notably June 2011. The model also requires a
year of meteorological data used repetitively for 5 years for
spin-up purposes (see below). Due to instrument failure at
JAR 1 from 1 January 2011 to 17 May 2011, and the com-
mencement of our modeling work in late July 2011, we only
have meteorological data for 2011 from 18 May to 25 July
2011. We therefore synthesize a mass balance year from the

Figure 2. Landsat 7 ETM+ images of the Ponting area (green box in Figure 1) from (a) 17 June 2011 and
(b) 3 July 2011. The field camp is marked by the star.

...
The model is run for 5 years for spin-up purposes, then during the sixth year, we generate output for the third year with the SRLF model. Values for fresh snow albedo (0.80) and ice albedo (0.45) are set based on average measured albedo values at JAR 1 for the year of climate data, used by the SRLF model. These values are very similar to those set for fresh snow albedo (0.82) and ice albedo (0.48) in the study by Banwell et al. (2012) which calibrated the same SEB model for a larger 450 km$^2$ sub-section of the Paakitsoq region for two mass balance years (2000/2001 and 2004/2005).

The model requires a spin-up period of 5 years in order for the surface mass balance and the subsurface temperature and density profiles to attain equilibrium [Banwell et al., 2012]. Another key aim of the spin-up period is to produce a snowpack thickness on 9 June 2011 (of the main sixth-year model run) equal to that measured at “Camp” (Figure 2) on the same date. In order to achieve this, we prescribe the model with precipitation at a constant rate per hour, in meters water equivalent (mwe), on one random day per week from 1 October to 30 April inclusive (called “winter precipitation” hereafter). We prescribe winter precipitation in this way, rather than, for example, having it fall at lower rates continuously throughout the 7 months since it is more realistic and will allow more realistic subsurface temperature and density profiles to evolve over the winter.

During model calibration (see below) we establish the total amount of winter precipitation that is required for the model to produce a snowpack thickness on 9 June 2011. Following Banwell et al. (2012), the threshold temperature for precipitation falling as either snow or rain is set at 2°C.

The subsurface model is described fully by Rye et al. (2010), but briefly, it calculates temperature, density, and water content on a one-dimensional vertical grid extending at least 25 m from the surface into the ice sheet. Melterwater generated by the surface energy balance component percolates through the grid, with refreezing occurring where the temperature is below 0°C and the density is less than that of ice. The cell below receives any residual meltwater if either of these conditions is not met, or if there is excess meltwater after refreezing. Meltwater percolates until it reaches the impermeable snow/ice interface where superimposed ice may be formed. If the rate at which meltwater reaches this interface exceeds the rate of superimposed ice formation, then excess water will form runoff [Rye et al., 2010]. The hourly runoff in mmwe calculated by the SMB for each model grid cell is used as input to the SRLF model (section 3.2).

3.1. SMB Model Calibration Method

[15] The SMB model is calibrated using measured daily snow surface height data measured at Camp (Figure 2). We calibrate the SMB model based on (1) measured snowpack thickness on 9 June; (2) day on which superimposed ice becomes exposed; (3) total surface height decrease from 9 June until superimposed ice is exposed; and (4) average rate of snow surface height decrease over this time period. To best match these four criteria, we choose the most suitable values for two model parameters which we do not have suitable observations to constrain (1) total winter precipitation and (2) initial snow density (i.e., the density of snow which has just fallen onto the ice sheet surface). For total winter precipitation, we parameterize for a range of values from 300–400 kg m$^{-3}$ and for initial snow density we parameterize for a range of values between 300 kg m$^{-3}$ and 400 kg m$^{-3}$ (Table 1). Initial sensitivity tests involving a much wider range of values indicated that these ranges gave the best match between measured and modeled snow depth on 9 June and are also consistent with suggested ranges of values in the literature [Bassford, 2002; Bales et al., 2009; Burgess et al., 2010; Caffey and Paterson, 2010; Rye et al., 2010]. We appreciate that the range of values for initial snow density appears to be high, but as snow densification due to settling, compression and the action of the wind is not accounted for by the model, but is instead driven by melting and refreezing alone, a relatively high value for snow density is expected to be established during model calibration [Bassford, 2002; Wright, 2005; Rye et al., 2010]. For example, Banwell et al. (2012) established that the best value for initial snow density for both 2000/2001 and 2004/2005 was 400 kg m$^{-3}$. All combinations of parameter values at the given intervals in Table 1 are used for individual model runs. The measured and modeled ablation curves are plotted and compared qualitatively. The modeled curves that clearly show a bad match with measured data are immediately discarded. To determine the highest quantitative match, root mean square errors (RMSEs) between the measured and the modeled curves showing good visual matches with measurements are calculated.

3.2. Surface Routing and Lake Filling Model

[16] Lakes form in topographic hollows on the ice sheet surface. The rate at which they fill (and hence the water volume within any given lake at any given time) is controlled by the size of the supraglacial catchment which supplies the lake, the rate of water production within the catchment, and the rate of water flow within the catchment. We calculate the rate of water production using the SMB model (section 3.1), but the location and size of lakes, their catchment areas, and water routing within and between the catchments are controlled by the surface topography. The SRLF model we have developed consists of two main components. The first component takes a DEM of the surface and analyses the DEM to identify the topographic hollows which can contain lakes, the catchment areas which feed each lake, and the topological routing of water between catchments if the water level in any given lake reaches the overflow. The second component of the model calculates the time delay between melt production and that water entering...
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sinks. Full details of the algorithm are given by
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flood
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for water to cross the cell can be
hydrological sciences. However, most algorithms in com-
Modelled supraglacial catchments (colored
regions) for the Ponting area (green box in Figure 1), maxi-
mum possible lake extents (gray), catchment overflow points
(open circles), and topological links between catchments
(black lines).

the lake by calculating the route taken by water, and the
water flow velocity, within each catchment in order to cal-
culate input hydrographs for each lake.

2.2.1. Lake and Catchment Identification Algorithm
The identification of watersheds (and hence catch-
ment areas), flow accumulation (upstream areas) and flow
directions over a DEM are common operations within the
hydrological sciences. However, most algorithms in com-
mon use rely on the artificial filling of surface depressions
within the DEM (see Arnold [2010] for a review). As we are
specifically concerned with calculating the time-dependent
filling of these depressions with water to form lakes, we
need another approach. Thus, we use the algorithm developed
by Arnold [2010] for calculating lake and catchment
extent as this does not require the artificial filling of surface
sinks. Full details of the algorithm are given by Arnold
[2010], but briefly, the algorithm begins by calculating the
surface slope, and from this the direction of steepest descent,
for each cell within the DEM. Any cell with no lower
neighbor is defined as a sink, and becomes a potential
nucleus for a lake. The algorithm then searches the flow
direction matrix to find the set of DEM cells which ultima-
tely flow into each sink cell. This identifies a series of
separate catchment areas feeding each sink cell. The algo-
rum then searches each catchment boundary for the lowest
possible DEM cell over which water would pour as the sink
and any surrounding cells lower than their neighbors) flood
with water. The maximum areal extent of any lake within
each catchment can be calculated, as cells within the catch-
ment lower than the level of the outflow would flood with
water as the lake fills. The DEM also allows the lake hy-
sometry (depth/volume/area) to be calculated. This process
also allows the connectivity between catchments to be
identified, as the location of the outlet cell, and the DEM cell
into which water would pour, are known.

3.2.2. Flow Delay Algorithm
The flow direction matrix calculated by the LCIA
allows the water flow path from any given DEM cell to the
sink cell (or lake) to be calculated. However, this informa-
tion by itself does not allow any flow delay to be calculated.
Thus, we link the LCIA with the flow delay algorithm ini-
tially developed by Arnold et al. [1998]. The FDA uses the
flow direction matrix, and the surface slope matrix, together
with assumptions about the physical processes controlling
water flow, to calculate a flow delay time between each
DEM cell and its sink cell. Over a glacier surface, water is
assumed to move across snow-covered cells by Darcian flow
in a saturated layer at the base of a seasonal snowpack
[Colbeck, 1978], or flow across “bare” ice cells in a supra-
glacial stream, governed by the Manning’s equation [Arnold
et al., 1998, equations 2 and 3, respectively]. Thus, for every
DEM cell, a “travel time” for water to cross the cell can be
calculated; this time depends on the slope of the cell, whether the cell is ice or snow (which governs the physical
processes assumed to control the flow), and the parameter
values which govern the water flow. For snow-covered cells,
the parameters are the snow porosity and permeability; for
ice cells the parameters are the assumed channel geometry
and roughness. By integrating the travel times downslope
along the calculated flow path, a total delay time from any
given cell to its sink can be calculated. This travel time will
vary as the snow cover across a catchment is lost over the
melt season, and as any lake within the catchment expands,
as this effectively shortens the path the water follows.

[20] Using the calculated delay times for each time step,
each hourly melt increment from the SMB is added to the
appropriate sink (or lake) cell(s) at the appropriate time step.
As the model run progresses, distinct input hydrographs for
each sink are produced. The total accumulated volume of
water within each lake at a given can then be calculated, and
from this (and the calculated lake hypsometry), the time step
at which the lake overflows its rim can be calculated; lakes
effectively fill with water from the original sink cell
“upward” by successive flooding of the next lowest DEM
cell(s) within the catchment until the water depth reaches the
level of the calculated outflow cell. Once a lake is full, any
further water inputs are passed into the downstream catch-
ment, as calculated by the LCIA. In this way, water can flow
in a series of “cascades” from its initial source cell, through a
series of full lakes, until it either reaches a lake which is yet
to overflow, or until it reaches the edge of the DEM domain.

[21] Preliminary model runs showed that water tended to
be delayed in the catchment for too long and did not fill the
instrumented lake fast enough when compared with mea-
urements. Field observations suggested that once the
snowpack had thinned to a threshold thickness, water starts
to flow quickly in the form of slush flows or in channels
incised into the saturated snowpack. Consequently, the
Table 1. Ranges From Which Parameter Values Were Chosen for Calibration of the SRLF Model

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Range Tested</th>
<th>Increment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow permeability (m²)</td>
<td>6 × 10⁻⁹ to 24 × 10⁻⁹</td>
<td>n/a</td>
</tr>
<tr>
<td>Effective snow porosity</td>
<td>0.50-0.75</td>
<td>n/a</td>
</tr>
<tr>
<td>k (snow permeability/effective snow porosity)</td>
<td>1.6 × 10⁻⁸ to 4.8 × 10⁻⁸</td>
<td>0.8 × 10⁻⁸</td>
</tr>
<tr>
<td>z (threshold snowpack thickness)</td>
<td>0.20-0.30</td>
<td>0.01</td>
</tr>
<tr>
<td>for Darcian to channelized flow switching (mwe)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*aHere n/a means not available.

algorithm was modified by introducing a parameter of threshold snowpack thickness at which flow switches from Darcian to channelized, rather than assuming that flow switching occurs once all the snow in a DEM cell has melted. This is a simple representation of the physical processes occurring on the surface of the GrIS that is closer to our observations, and allows more water to reach the lake basin more quickly. This threshold snow thickness parameter was tuned during model calibration (see below).

3.2.3. SRLF Model Calibration Method

The hourly runoff per model grid cell from the calibrated SMB model is used to drive the SRLF model. This model is calibrated through comparison of the modeled and measured lake filling data (i.e., the cumulative volume over time) for Lake Ponting. Through tuning model parameters, we aim to best match (1) the onset of meltwater arrival in the lake; (2) the initial lake filling rate before any overflow occurs; (3) the timing of the diurnal cycles in the initial filling rate; and (4) the lake filling rate once potential inflow from overflowing upstream lake(s) has occurred.

We do not have suitable observations to constrain values for the three parameters which control supraglacial water flow velocity in the snowpack and so we perform multiple model runs and compare the results with measurements in order to identify suitable values. The three parameters are (1) snow permeability; (2) effective snow porosity; and (3) the threshold snowpack thickness (which we call z) for the Darcian to channelized flow switching. However, as snow permeability and effective snow porosity appear as a simple numerator and denominator, respectively, in Colbeck's [1978] equation [Arnold et al., 1998, equation 2], we combine them into one parameter, k, given by $k (\text{m}^2) = \text{snow permeability (m}^2/) \text{effective snow porosity.}$ Suitable ranges of values for snow permeability and effective snow porosity are given in Table 2. These ranges fall within those used by Marsh [1990], Arnold et al. [1998], and Willis et al. [2002] and give us a range of k values from 1.6 × 10⁻⁸ to 4.8 × 10⁻⁸ m². For z, initial sensitivity tests encompassing a range of values from 0.00 to 0.40 mwe indicated the most appropriate range of values to be between 0.20 and 0.30 mwe (Table 2). Initial sensitivity tests indicated that the model was quite insensitive to Manning’s roughness (n) and hydraulic radius (R). We therefore use constant values of $R = 0.035 \text{ m}$ and $n = 0.05 \text{ m}^{1/3} \text{s} [e.g., Arnold et al., 1998; Willis et al., 2002].

When the measured lake filling data for Lake Ponting are analyzed (Figure 5), we hypothesize that the dramatic increase in lake filling rate at ~12:30 LT on 18 June is due to the overflowing of one or both of the upstream lakes, Lakes X and Y (Figure 3). However, although this was the time at which the magnitude of meltwater discharge into Lake Ponting increased significantly, it is also possible that water started to overflow into Lake Ponting at a lower rate from the upstream lake(s) a few days earlier before a large, high capacity channel was able to form. We calibrate the SRLF model by performing model runs in order to quantitatively establish which combination of parameter values produces the best match between the early period of measured and modeled data. During these runs we focus on the 3.7 km² catchment area of Lake Ponting, without allowing for any inflow from the upstream Lakes X and Y. This enables us to best match the onset of meltwater arrival in the lake basin and the initial filling rate of Lake Ponting before the catchment area supplying Lake Ponting increases due to meltwater inflow from overflowing upstream lakes. Once the combination of parameter values for k and z which give the lowest RMSE between the early modeled and measured data has been determined, these values are used in subsequent model runs allowing for overflow from the upstream catchment(s), enabling us to observe the effect that inflow from the overflow of Lakes X and Y has on the filling rate of Lake Ponting.

4. Model Calibration

4.1. SMB Model Calibration

Measured snow depth was 0.99 m on 9 June. We measured a total snow height decrease of 0.64 m between 9 and 25 June when superimposed ice was exposed, suggesting that 0.35 m of superimposed ice formed at this site. Fifteen SMB runs covering the full range of parameter values in Table 1 were undertaken. A selection of graphs of modeled ablation using different parameter value combinations are plotted alongside the measured snow ablation graph in Figure 4. The model captures the magnitude of the total snow height decrease and the average ablation rate between 9 and 25 June. The model does not capture the minor variations in measured ablation rate during this time period likely due to local snow conditions at the measuring stakes. The model run which produces the best match with the measured data has the parameter values of initial snow density 350 kg m⁻³ and winter precipitation 0.44 m are used instead. Importantly, the slight lower density value of 350 kg m⁻³ (compared to 400 kg m⁻³) more accurately captures the slight increase in ablation rate on 13 June. These values are therefore set as parameter values in subsequent SMB model runs.

4.2. SRLF Model Calibration

Using hourly melt input per DEM cell from the calibrated SMB model, the SRLF Model is run 25 times in order to explore the full range of parameter values for z and k (Table 2). Graphs comparing the measured lake filling curve to various modeled lake filling curves are produced. By visually inspecting the graphs, it appears that the best fit between the modeled and measured data produced when...
parameter values of $z = 0.275$ mwe and $k = 4.0 \times 10^{-8}$ are used. As examples, Figure 5a shows the measured lake filling curve alongside modeled lake filling curves for various $k$ values given a $z$ value of 0.275 mwe, and Figure 5b shows modeled lake filling curves for various $z$ values given a $k$ value of $4.0 \times 10^{-8}$. However, this match is only good up until 16:00 LT on 16 June. Up until this point, (marked by a vertical dashed line in Figures 5a and 5b), the model run using these parameter values produces an RMSE between the measured and modeled data of $1.7 \times 10^4$ m$^3$ (i.e., 4% of the cumulative lake volume at 16:00 LT on 16 June).

Beyond this point, the gradient of the measured lake filling data starts to show a slight increase and deviates from the almost linear gradient of the modeled lake filling graph. These two parameter values are therefore used in subsequent model runs.

### 5. Results and Discussion

As previously mentioned, our measured volume data for Lake Ponting indicates a sudden rise in filling rate at 12:30 LT on 18 June (Figure 5). Furthermore, field observations on 19 June (a few hours after Lake Ponting had drained by hydrofracture), showed a large incised channel, containing a river routing water into the Lake Ponting basin coming from the direction of Lakes X and Y (Figures 2 and 3).

This channel had not been visible during our first visit to the basin on 13 June. We hypothesize, therefore, that the sudden increase in lake filling rate at 12:30 LT on 18 June was due to the initiation of, or sudden rise in, meltwater inflow from one or more upstream lakes. Our measurements and field observations do not allow us to determine whether the river was routing water just from Lake X, or whether Lake Y was also overflowing in to Lake X so that Lake Ponting was receiving water from both upstream catchments as well as its own after 18 June 2011.

Having calibrated the SRLF model for the Lake Ponting basin only, without allowing for potential inflow from overflowing upstream lakes, we now rerun the model using the optimal parameter set for the whole 10 km Lake Ponting area, allowing for lake overflow from basin to basin. The output from this model run is shown together with that from the original run without overflow and the measured lake volume data in Figure 6. We successfully model (1) the onset of meltwater arrival in the lake; (2) the initial lake filling rate before any overflow occurs; (3) the timing of the diurnal cycles in the initial filling rate; and (4) to an extent, the lake filling rate once overflow from upstream Lakes X and Y has occurred. The main discrepancies between the modeled and measured lake volume data are related to the timings in the overflow of the two upstream lakes: X and Y.

The graph of modeled Lake Ponting volume allowing for Lakes X and Y overflow starts to deviate from the graph of modeled lake volume without upstream lake overflow at 16:00 LT on 16 June (Figure 6). This is when the modeled Lake X volume reached a maximum depth of $\sim 1.5$ m and started to overflow into Lake Ponting. As the graph of measured lake volume also deviates from the graph of modeled lake volume without overflow at this time, is it likely that Lake X also started to overflow at this time in reality? The modeled Lake Y volume reached a maximum depth of $3.8$ m and started to overflow into Lake X (and on to Lake Ponting) at 16:00 LT on 17 June. This is just over 24 h before the time (12:30 LT on 18 June) when we infer Lake Y to have overflowed into Lake X (and on to Lake Ponting) from the measured data. After this time, the modeled rate of filling of Lake Ponting is slightly less than the measured rate. This supports our hypothesis that water from both Lakes X and Y was necessary in order to produce the measured increase in the filling rate of Lake Ponting after 12:30 LT on 18 June. However, the measured rate may be slightly higher than the modeled filling rate due to channel
incision into ice/snow, thereby allowing a higher discharge of water to flow from Lake Y into Lake X, and on into Lake Ponting. This process is not accounted for by the model. Thus, it is likely that Lake X overflowed into Lake Ponting at 14:00 LT on 16 June (creating a slight rise in Lake Ponting’s filling rate), before Lake Y then overflowed into Lake X, and on into Lake Ponting, at 12:30 LT on 18 June (creating a much more significant rise in Lake Ponting’s filling rate). In our model, Lake X first overflows into Lake Ponting in agreement with the measured data, but then Lake Y overflows into Lake X on 17 June. Thus, there are discrepancies between our model and measurements, not in the magnitude of Lake Ponting volume increase after the overflow of Lake X and Y, but in the timings of the overflow of Lake Y.

In order to provide additional evidence to help constrain the timings of lake overflows, we consult the first available Landsat image after the 18 June 2011, from 3 July 2011 (Figure 2b). When this image is compared to the 17 June Landsat image (Figure 2a), Lake X appears to have completely drained by 3 July, whereas Lake Y is now significantly smaller (estimated to be 0.08 km$^2$ on 3 July compared to ~0.39 km$^2$ by 17 June). We can therefore infer that in reality, it was likely that Lake Y did not drain via hydrofracture, but instead probably overflowed into Lake X until the lake level reached the height of the catchment overflow point. Consistent with both measured and modeled lake filling data, we therefore infer that the rapid increase in Lake Ponting’s filling rate on 18 June was due to both the overflow of Lake X and Lake Y (with the overflow of Lake Y likely occurring after the overflow of Lake X). Other sources of time-coincident satellite imagery at appropriate spatial resolutions needed to examine these lake drainage dynamics were not available.

There are two possible explanations for the source of the errors between the timings of the measured and modeled lake overflows. First, as already mentioned, these errors may be due to the model’s inability to simulate the process of opening and growth of overflow channels from lakes. Second, the discrepancies between the timings of the measured...
671 and modeled lake overflows may be due to inaccuracies in
672 the ASTER GDEM. As a test of this latter hypothesis, we
673 carry out additional model runs with altered DEM topogra-
674 phy and find that if the average depth of Lake Y was 0.87 m
675 deeper (giving Lake Y an extra 270,000 m$^3$ of volume in
676 addition to its current maximum volume of 820,000 m$^3$),
677 Lake Y would overflow at 12:30 LT on 18 June, in agree-
678 ment with the measured data. This modeled curve is plotted
679 in Figure 6 (purple line).

680 6. Summary and Conclusions

681 [33] Using measured snow surface height data, we suc-
682 cessfully calibrate a distributed, high-resolution surface
683 mass balance (SMB) model for a small (100 km$^2$) subset of
684 the larger (2300 km$^2$) Paakitsoq region of west central
685 Greenland for June 2011. Key SMB model parameter values
686 are (1) total winter precipitation and (2) initial snow density.
687 Values for 2011 are within the ranges found when the model
688 was calibrated against longer time series of measured surface
689 height and albedo over the mass balance years 2000/2001
690 and 2004/2005 [Banwell et al., 2012].

691 [34] We model the routing of this runoff across snow-/ice-
692 covered cells to topographic depressions which fill to form
693 supraglacial lakes, which can overflow into their down-
694 stream catchment(s) once full. This surface routing/lake
695 filling (SRLF) model is calibrated using measured volume
696 data from supraglacial Lake Ponting. The key SRLF model
697 parameters are (1) snow permeability, (2) effective snow
698 porosity, and (3) threshold snowpack thickness (z) for the
699 Darcian/channelized flow switching.

700 [35] We successfully model (1) the onset of meltwater
701 arrival in the lake, (2) the initial lake filling rate before any
702 overflow occurs, and (3) the timing of the diurnal cycles in
703 the initial filling rate; and iv) to an extent, the lake filling rate
704 once overflow from upstream Lakes X and Y has occurred.
705 Our modeled data also confirm that the rapid rise in the
706 measured filling rate of Lake Ponting was due to the over-
707 flow of upstream Lake Y, which flowed into Lake X (which
708 had already overflowed), then on into Lake Ponting, so that
709 Lake Ponting suddenly received water from both upstream
710 catchments once Lake Y overflowed its basin. There are
711 discrepancies of around a day or so between the timings of
712 the modeled and measured lake overflows. These discrep-
713 ancies could be explained by the model’s inability to
714 simulate the process of opening and growth of overflow
715 channels and consequent changes in water velocity, and/or
716 inaccuracies in the DEM which could alter calculated lake
717 volumes and hence the timing of overflow events.

718 [36] By linking a surface hydrology model with melt input
719 from a calibrated mass balance model we show that we are
720 able to model the filling rate of a supraglacial lake in a subset
721 of the Paakitsoq region in west Greenland with high accu-
722 racy. We also demonstrate that water inflow from overflow-
723 ing lakes in surrounding catchments can play a key role
724 in increasing the filling rate of a lake, and we are able to
725 model the timings of these overflow events with relatively
726 high accuracy given the quality of the available surface
727 topographic data sets. As the rapid drainage of some of these
728 supraglacial lakes by hydrofracture is thought to play a key
729 role in the establishment of a hydraulic connection between
730 the surface and subglacial drainage systems, advancing our
731 ability to accurately model the temporal and spatial vari-
732 ability of lake volumes will ultimately improve our ability to
733 predict changes in ice sheet dynamics, mass balance and sea
734 level contributions.

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Figure 6. Measured volume over time for Lake Ponting (red) alongside graphs of modeled volume over time for runs with (blue), and without (green), lake overflow. The purple line represents the modeled volume over time for Lake Ponting for the case when the basin of Lake Y is adjusted as described in the text.
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