UPPSALA
UNIVERSITET

2012

Report

Internal accumulation on glaciers: qualitative description and quantitative estimates

Sergey Marchenko
PhD Student Physical Geography

Contents

Abstract 3

Terminology and process domain identification 3

Processes involved in internal accumulation and links with other components of mass and energy balance 5

Review of published approaches to estimation of internal accumulation 10

Schemes not involving snow/firn properties 11

Schemes involving snow/firn properties: cold content and (or) pore space 12

Schemes designed for estimation of superimposed ice formation rate 16

Layered englacial models including description of internal accumulation 19

Main uncertainties in calculation of internal accumulation rates 23

Acknowledgements 24

Literature 25

Abstract

The aim of the report is to give a qualitative and quantitative overview of the problem of internal accumulation.

For that we will first present the problem of internal accumulation in a larger perspective and elaborate on the terminology involved defining the temporal and spatial domain of the process. Then we suggest a qualitative description of the internal accumulation distinguishing between different smaller scale processes process and analyze its relation with other components of the glacier mass balance. This is followed up by a review of the parameterizations offered by different authors up to date diving them in several groups according to their physical focus and concluding with identification of the main areas of uncertainty.

Terminology and process domain identification

The issue of melt and rain water refreezing in glaciers is a part of a bigger and more complicated glaciological problem which is usually known as the mass balance of glaciers. According to the Glossary of Glacier Mass Balance and Related Terms (Cogley et al, 2011, fig .1) ”refreezing of water within a glacier, between the summer surface and the bed, which goes undetected by measurements of surface mass balance” is called ”internal accumulation” whereas the more inclusive term “refreezing” is suggested to denote “internal accumulation plus refreezing within the snow” (*).

From positions of hydrology internal accumulation on a glacier defines the runoff coefficient of the glacier covered area – relation between amount of precipitation and runoff.

Having a substantial impact on the mass and energy balance of glaciers internal accumulation is important to be accounted for in other glaciological spheres. Through its control on englacial temperatures on which rheological properties of ice are dependent it is important for dynamics of glaciers. Internal accumulation might also have implications for remote sensing studies as it defines to a large extent stratigraphy of the upper part of a glacier and positioning of reflectors.

Following (Golubev, 1976) the annual cycle can be divided into 4 parts basing on the thermal regime of the upper part of the glacier that consists of snow and firn (everything that is above the impermeable ice layer): spring, summer, autumn and winter. Summer and winter are the seasons when the layer is entirely at the melting point or is entirely frozen respectively. During spring the layer defined above is being warmed from whatever temperature it acquired during winter up to the melting point. During the autumn season the layer on the top of the glacier is being cooled down to negative temperatures. Depending on local climatic conditions in particular year a glacier can experience or not experience all the above defined seasons. It may be convenient to divide the processes that are participating in refreezing of water into groups corresponding to the “thermal” season of the year when they occur.

*As a matter of fact the term “internal accumulation” refers to a mass that the glacier has “not lost” rather than a mass that the glacier has acquired. With respect to the glacier as a whole there is no additional mass involved in the process of internal accumulation, it is rather a relocation of mass. In that sense an alternative term might be offered – “decrement of ablation” which points to the fact the melt water that could have contributed to the runoff under different conditions was refrozen and in the particular case.

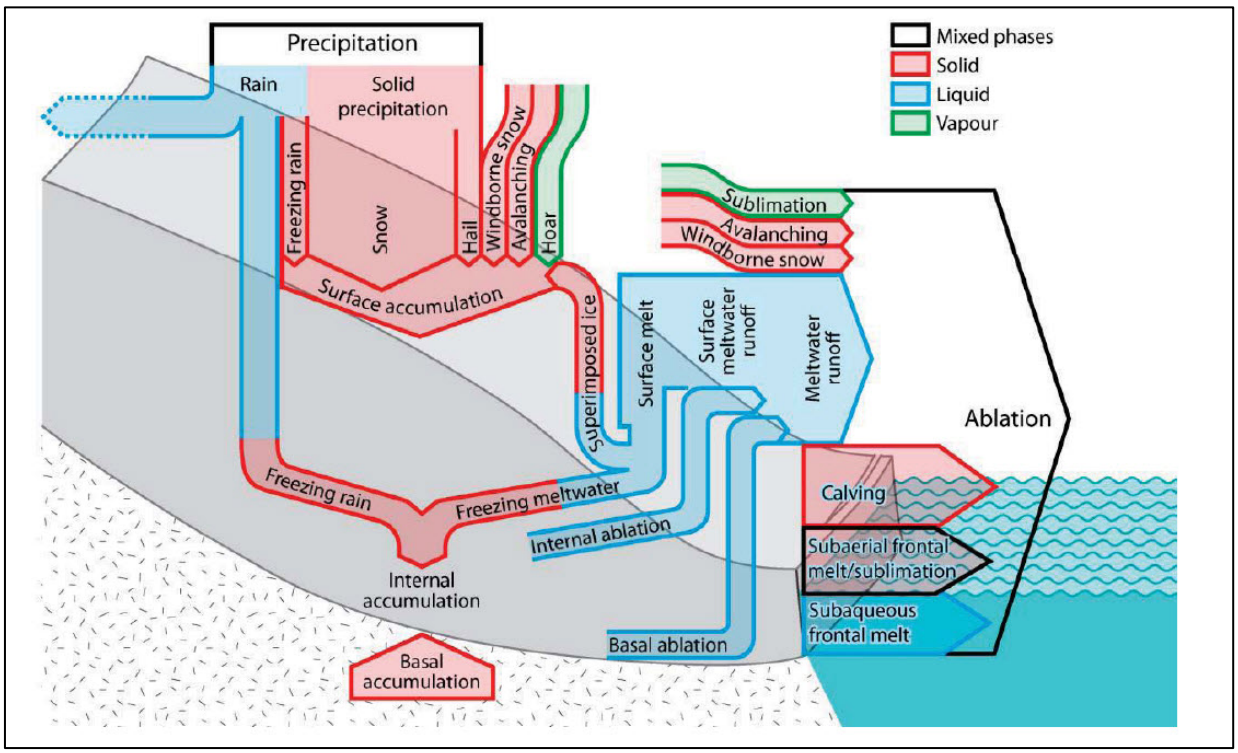

Figure 1. Components of the mass balance of a glacier. The arrows have arbitrary widths and do not indicate physical pathways of mass transfer. (Cogley et al, 2011)

For a layer of water to be frozen in or on a glacier specific conditions are needed. First of all presence of the very volume of water is required, it can appear as melt water of liquid precipitation. This puts the upper limit to the spatial domain where internal accumulation occurs. Guided by the fact that internal accumulation is a positive component of the glacier mass balance, the lower limit of the process distribution in terms of significance for mass balance is the equilibrium line of a glacier, though energy-wise it is also significant in the ablation area. Also cold conditions are required. Thus internal accumulation occurs under negative temperatures mainly in the porous upper part of a glacier and on its surface.

Processes involved in internal accumulation and links with other components of mass and energy balance

Components of mass balance of a glacier and internal accumulation in particular as well as components of the surface energy balance are controlled by the processes in the lower troposphere. This problem from the most general positions was thoroughly reviewed by Shumskii (1955, 1964), Benson (1960, 1961), and Muller (1962), as a result the study of ice formation zones or glacier facies was developed. An inter-comparison of the approaches is presented on fig. 2 below. Freezing of liquid water can occur in all ice formation zones, but the particular processes involved and their magnitude are controlled by the balance of mass and heat at the glacier surface. The relative importance of different processes and their connection with meteorological and glacial conditions is discussed below.

Under cold and relatively precipitation rich conditions of the upper elevation belts on glaciers and internal regions of the Greenland and Antarctic ice sheets very few melt water is available during the melt season and it entirely refreezes inside the upper porous layer of snow and firn. This includes the *recristallization, recristallization-infiltration and cold infiltration* according to the glacier zone classification from Shumskii (1955, 1964). Melt and rain water here affects the density and temperature of snow and firn increasing both and this is still important to be accounted for. Refreezing happens both during the spring season when water penetrates through the cold snow and firn and in autumn when cooling freezes suspended water. Process of water refreezing and release of the latent heat is especially important under conditions of warming climate when increasing melt rates may not result in increased run off because the energy will be spent on warming the glacier up. But once the cold content is used the part of the glacier will exit the cold infiltration zone domain and will start to contribute to the runoff.

Thus in the cold conditions of the upper three glacier zones internal accumulation takes place mainly during spring seasons. Additional volumes corresponding to the irreducible snow (firn) water content are frozen autumn. The conditions above are very important because they define the upper limit of the domain which is to be treated with great accuracy in terms of mass balance as not all mass that disappears from the surface refreezes, part of it runs of.

The effects of temperature rise and density increase are closely connected. Imagine that there is a volume V of snow with mass M, having temperature T_1 and a relatively small volume of water at 0°C having mass m is introduced.

The occurring process can be broken into 2 stages: release of the latent heat as a result of water freezing rises the temperature of snow from T_1 to T_2. After that M is at T_2 and m is still at 0°C (1) and heat transfer from m to M in the result of which they both have equal temperature T_3 (2).

To describe the first process we can write: $(T_2 - T_1) \cdot MC = mL$, where C - specific heat capacity of ice and L - latent heat of melt. And for the second: $T_3 = \left(T_{fp} - T_2\right)\frac{m}{2M} + T_2$, as there should be a linear dependence between m and T_3: if $m = 0$ then T_2 is not going to rise and $T_3 = T_2$, and if $m = M$ then $T_3 = \frac{T_2 + T_{fp}}{2}$, ($T_{fp}$ - temperature of water at freezing point).

Expressing T_2 from the eq. for the 1st process we have: $T_2 = T_1 + \frac{Lm}{CM}$ and putting it into the eq. for the second process: $T_3 = \left(T_{fp} - T_1 - \frac{Lm}{CM}\right)\frac{m}{2M} + \frac{Lm}{CM} + T_1$, which is the temperature of the snow after water refreezing.

The corresponding increase in density of snow or firn will be: $\Delta\rho = \rho_2 - \rho_1 = \left(M + m\right)/V - M/V = m/V$

As one goes towards warmer climatic conditions amount of melt water increases and it is also probable that precipitation will partly consist of liquid water. Runoff occurs when either the cold content or the pose space becomes a restriction for refreezing of the entire volume of the liquid water available.

In the first case the upper part of the snow-firn-ice column which is a subject for seasonal changes of temperature is not cooled enough during winter season to ensure refreezing of all the liquid water available. Then melt water will percolate deeper than the depth of zero year amplitudes warming the snow-firn column up to the melting point. After that the snow/firn pack can still hold some additional water and prevent it from running of by capillarity forces. But if melting of temperate snow/firn continues runoff is initiated and water will evacuated through englacial and subglacial conduits and channels. The glacier in this case is a warm glacier (or its part in question is warm).

In the second case the precipitation between sum of melt and rain water and amount of solid precipitation is such that all the pores are filled with refreezing water. It starts to accumulate on top of the impermeable horizon forming a slush layer. Assuming that the cold content of the active layer is large the slush layer will freeze from below while on top of that newly formed layer of superimposed ice runoff will occur.

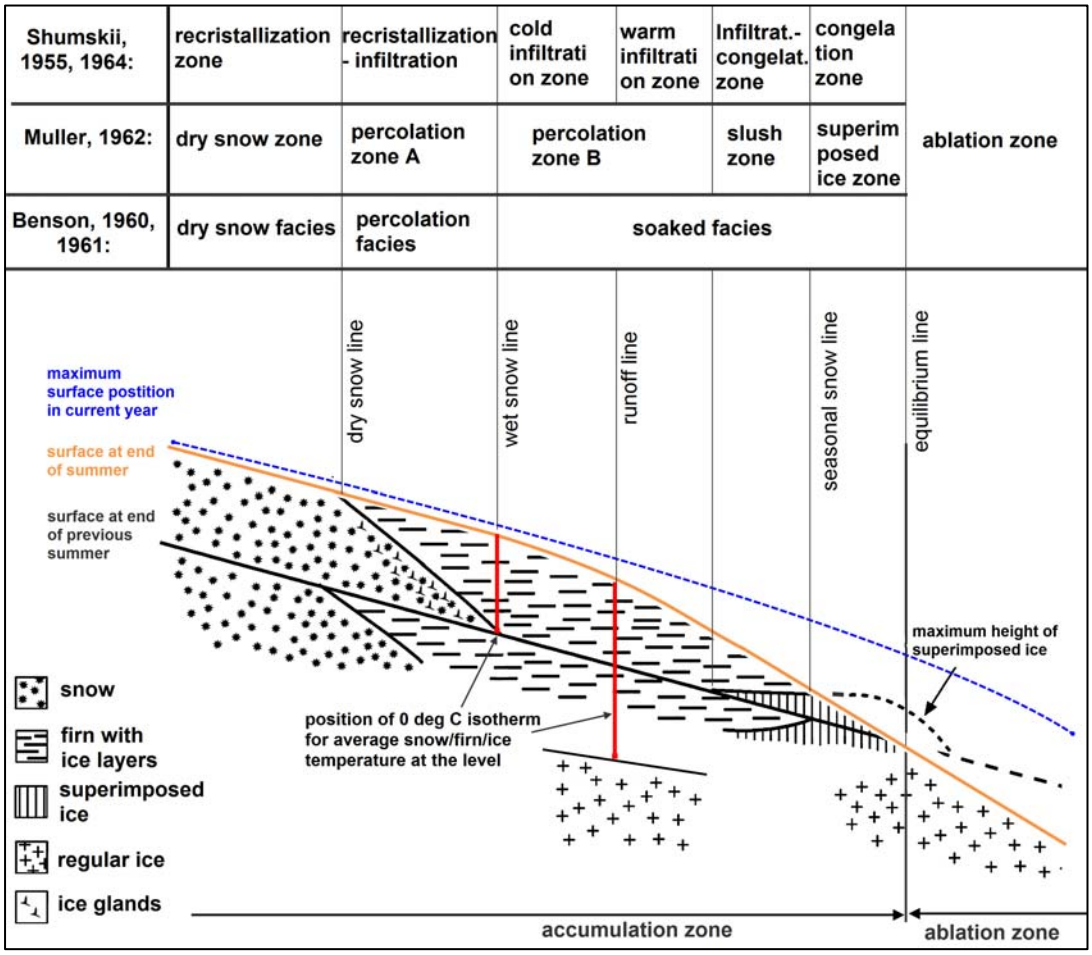

Figure 2. Zones of a glacier based on Shumskii (1955, 1964), Benson (1960, 1961) and Müller (1962).

Spatial variability of the meteorological parameters that define the accumulation of water in solid phase on a glacier and the cold content in the layer of seasonal temperature variability is such that in most cases at the lower limit of the cold infiltration zone the cold content is insufficient to freeze all the melt water while the pore space in firn layer underlying the seasonal snow is still not entirely filled. This can be explained by large increase in melt rates at lower elevations, milder winter seasons and still considerable precipitation.

Thus in the *warm infiltration zone* the cold content of the seasonally frozen layer is not sufficient to freeze all the water that infiltrates in the snow/firn column. In spring it infiltrates and warms up deeper horizons, which then sink down during next accumulation season to the level where they cannot be cooled in winter. So the glacier (or part of the glacier) below the lower surface of the active layer is at the melting point year round. This enables the melt water that percolates from the surface to contribute to the runoff as soon as it reaches the upper layer of the temperate firn or ice.

In spring liquid melt and rain water refreezes releasing the latent heat of fusion which will raise the temperature of the cold layer of snow and firn to the melting point and also increase its density. Water flow is usually non homogeneous. Observations show that instead of a uniform wetting front a complex system of flow fingers, ice glands and lenses is developed (Marsh and Woo, 1984; Boggild, 2000). As water flows through snow and firn part of the new “spring” ice (ice that accreted during “spring” on the grains of snow and firn) may melt away.

During autumn when the active layer in the warm infiltration zone is being cooled all the water that is present there refreezes, rising the density one more time. The amount of water freezing in winter should be equal to the maximum water holding capacity of snow and firn. The latter is dependent on the microstructure of snow and firn and their density. In case when runoff is restricted perhaps a volume of water exceeding the maximum WHC of snow or firn can be present in the snow firn column by the beginning of autumn cooling. There can be different causes for that. In case of plateau glaciers or flat parts of glaciers the hydraulic head can be not sufficient to overcome the obstacles that always exist in the snow-firn column (like buried wind crusts with uneven surface, icy layers).

Warm infiltration zone is usually the lowest in the accumulation zone in maritime conditions, where warm winters and big amounts of solid precipitation are observed. The thick layer of snow and firn resting upon ice contains a lot of pores so that the cold content is not sufficient to fill all pores with refrozen water at any level of the active layer. In the sense of geography that will be Iceland and maritime regions in Norway, partly glaciers on Spitsbergen.

In more continental environments which means that annual amplitudes of air temperatures are higher and precipitation is lower a different situation takes place. Instead of transition to ablation zone the *infiltration-congelation* is observed. In more continental climates relative decrease in amount of precipitation with altitude will be more sensitive than increase of temperatures. Lower rate of precipitation should result in increased densities and thermal conductivities of material in the active layer. This will enable the cold wave in winter to penetrate deeper into the glacier. At the same time pores in a thin layer of snow will be filled with refreezing snow much faster and an impermeable layer will be established.

Water accumulating at the surface of impermeable ice will partly accrete to it building up a new layer of superimposed ice and partly will contribute to the runoff by this evacuating a substantial potion

of latent heat stored in the liquid water. So the cold content of the active layer can actually be smaller than the latent heat of fusion of melt and rain water available during the ablation period. Thus besides the refreezing processes characteristic for the warm infiltration zone which were mentioned above in the infiltration-congelation zone formation of superimposed ice during summer season occurs.

In case a glacier is situated in very continental conditions the transition from the cold infiltration zone to the infiltration-congelation zone can occur without the intermediate stage of the warm infiltration zone.

With transition to warmer conditions on lower elevation belts or more continental conditions the layer of firn left on the glacier by the end of the accumulation season decreases. So in the lowermost part of the accumulation zone ablation can affect not only the seasonal snow but also the newly formed layer of superimposed ice and in the following mass balance year snow is deposited right on the surface of superimposed ice. This happens in the *congelation zone* where predominant process involved in internal accumulation is formation of superimposed ice during summer. Other refreezing processes in spring and autumn are significant only for the energy budget but not for the mass balance.

It is to be noted that insignificance of internal accumulation for mass balance of a glacier or a part thereof one should understand only as an absence of direct affect of the process on relation between evident lowering of the glacier surface and the amount of runoff. It is obvious that if by the time melting starts the snow/firn/ice pack is temperate, then the entire heat flux from the atmosphere would be spent on lowering the glacier surface by melt. In most cases this is not the case and ablation is reduced by refreezing. Thus even in *ablation zone* it is important to account for refreezing of liquid water because without that the net effect of melt season may be overestimated.

*Table 1. Processes of internal accumulation in different glacier zones, environmental controls and importance for mass and energy budget of the glacier part in question. Seasons: spring/summer/autumn – temperature of the layer above impermeable ice is rising/0°C/falling. Zones: R - recristallization, RI - recristallization-infiltration, CI - cold infiltration, WI - warm infiltration, IC - infiltration-congelation, C – congelation. */* notation – "significance for mass balance"/"'significance for energy balance".*

Season		Spring	Summer	Autumn
Processes		freezing of water in pores of the snow/firn layer above the impermeable ice	formation of superimposed ice	freezing of capillary water in pores of the snow/firn layer above the impermeable ice
Primary controls		cold content of the porous layer, water mass	cold content of the active layer, water mass	amount of suspended water by the end of summer
Glacier zones Shumskii (1955, 1964)	R, R-I, CI	+/+	-/-	+/+
	WI	+/+	-/-	+/+
	IC	+/+	+/+	+/+
	C	-/+	+/+	-/+
	A	-/+	-/+	-/+

Depending in which of the zones described above a glacier or part thereof is, different processes will contribute to refreezing of water and their relevance for mass and energy balance of the glacier will also be different. This is summed up in Table 1. Changes in the boundaries between zones

may be registered using methods of remote sensing (Wolken et al, 2009). Under conditions of climate change boundaries among zones are likely to be shifted and it is important that models describing internal accumulation are able to reflect this adequately.

Review of published approaches to estimation of internal accumulation

Guided by the fact that refreezing of melt and rain water in glaciers is an important term in the mass balance equation and is expected to affect significantly the amount of water that will contribute to the runoff, a number of authors worked on finding a way to quantitatively describe the process. An extensive review and testing of different parameterizations of internal accumulation was carried out by Reijmer with co-authors (2011).

The complexity of the process along with the complexity of the porous snow/firn media where it occurs leaded to the fact that up to date there is no description of water refreezing in its full physical complexity. One important constraint is that internal accumulation can hardly be measured in the field and thus validation data for models is scarce. Moreover modeling of the process is expensive in terms of numerical calculations, which explains the emphasis on parameterizations rather than physical modeling of the process.

Further a review of eleven existing parameterization schemes is presented, it is not supposed to be complete and other approaches not mentioned below exist. Different schemes are grouped in four logical groups according to the physical basis behind the each approach:

1. Reeh (1991) suggested a simple parameterization for refreezing, while as Oerlemans (1991) proposes to calculate the parameter basing on climatic parameters. Wakahama et al. (1976), Bøggild (2007) and Wright et al. (2007) focused on formation of superimposed ice and presented models describing this process.
2. Driven by the fact that refreezing of the entire mass of melt and rain water available can be restricted either by the cold content of the active layer or by the pore space therein a number of authors chose one of the two (Braithwaite et al., 1994; Huybrechts and de Wolde, 1999) or both (Shumskii, 1955, 1964; Pfeffer et al., 1991; Janssens and Huybrechts, 2000) directions:
   - estimation of the pore space available for refreezing of water;
   - estimation of the cold content of the upper part of the glacier;

   These schemes are aimed at estimating the maximum amount of water that can be refrozen in a time period given certain conditions and all the excess water is allowed to run away. This point was very clearly expressed by Janssens and Huybrechts (2000) and then by Reijmer at al. (2011):

   $$E_r = \min[P_r, W_r],$$

   where E_r – effective retention mass or actual internal accumulation; P_r – potential refreezing mass which is the maximum amount of water that can be refrozen; W_r – available water mass, the latter is usually a function of the summer surface energy balance and can also include the rain water.
3. A more complicated layered englacial model describing evolution of temperature and density in the upper part of a glacier was presented by Greuell and Konzelmann (1994) and then used and developed in numerous studies.

Schemes not involving snow/firn properties

Oerlemans, 1991

A rather elegant approach is suggested to estimate the relation between runoff and refreezing of the melt water. Author avoids any physical properties of snow and the model does not distinguish between processes of water freezing in pores of snow/firn and superimposed ice formation.

The active layer is determined as the layer of snow, firn and ice in the upper part of the glacier which has the mass equal to the mass of a layer of pure ice of 2 m thickness. Its temperature by the end of winter season (Θ_{ice}) is equal to the mean annual air temperature. The surface energy balance is defined as $B = (1-\alpha)\cdot G + I_{in} + I_{out} + H_s + H_l$ ($[B] = \frac{W}{m^2}$). It is divided into two parts. One (H_{ice}) is spent on heating the active layer both by conductive heat transfer (less effective) and refreezing (more effective) and the other one (R) is spent on melting which in this case is equal to runoff. Thus: $B = H_{ice} + R$. The relation between R and H_{ice} is defined by the temperature of the active layer: $R = B \cdot \exp(\Theta_{ice})$, on the figure 3 below under the blue line is the fraction of B that will contribute to the runoff and above the line is the fraction of B that will be used to heat the active layer, both fractions are exponential function of temperature. So apparently melt and runoff seem to be possible even when the temperature of the active layer is negative, since Θ_{ice} is the bulk temperature of the whole active layer and in reality is expected to be significantly lower than the value in the lower part and higher in the upper part.

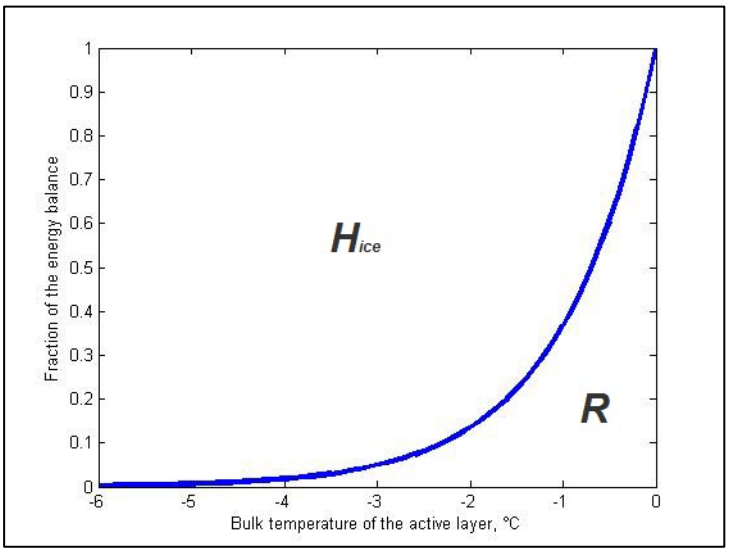

Figure 3. Temperature dependence of partitioning between heating (H_{ice}) and melting/runoff (R) parts of the radiation balance

In the article where this scheme is suggested there is no explicit description of the function that links H_{ice} and Θ_{ice} and would describe the heat transfer from the surface downwards. But it is said

that "*the empirical constants involved are chosen such that the result is in line with the findings of Ambach (1963)*".

Reeh, 1991

An approach is usually recognized as P_{max} is widely used for estimating the amount of water that does not contribute to the runoff, mainly because of its evident simplicity.

The fraction of the meltwater that refreezes inside a glacier is prescribed in the beginning of the melt season to be equal to 0.6 of the amount of the precipitation, everything else is allowed to run away from the glacier. The value for P_{max} was chosen arbitrarily, the requirement was to match the modeling results with other similar estimates.

Schemes involving snow/firn properties: cold content and (or) pore space

Shumskii, 1955

In his book on principles of structural glaciology P. A. Shumskii offered a definition of the conditions under which runoff can occur from a glacier or its part. Consequently inverting the question one can define the conditions under which the maximum amount of water can be refrozen inside a snow-firn column. Following the qualitative description of these conditions given in the previous section one can conclude that for a portion of melt or rain water not to be refrozen inside the firn it is needed that the preceding portions of water have not eliminated neither the cold content of the firn by releasing the latent heat nor the pore space by filling it. The author also offers a more complete and accurate description of the conditions.

The cold content condition can be written as:

$$\int_{z_0}^{z_s} CT\rho_z dz > (M + P_l)\cdot L ,$$

where z - depth, z_0 – depth of zero annual temperature amplitudes, $z_s = 0$ – surface, T - temperature, C – heat capacity of ice, ρ_z – snow or firn density at depth z, M – mass of melt water, P_l – mass of liquid precipitation, L – latent heat of fusion of water.

A strict formulation for the pore space condition is also derived. If the density of snow at the surface is ρ_0, then a mass P_s kg of that snow will have the volume of $V_0 = P_s/\rho_0$ Let us assume that P_s is the mass of annual solid precipitation on a unit area. The corresponding volume of pores will be:

$$V_p = \frac{P_s}{\rho_0} - \frac{P}{\rho_i} = \frac{P_s\left(\rho_i - \rho_0\right)}{\rho_0 \cdot \rho_i}$$

where ρ_i is the density of ice.

Then taking into consideration that the mass of water (from melt M and rain P_l) should be less than the mass of water that can be frozen in the available pore space the following can be written:

$$M + P_l < P_s \cdot \rho_l \cdot \frac{\left(\rho_i - \rho_0\right)}{\rho_0 \cdot \rho_i} ,$$

where ρ_l is the density of water, which is a known constant ($\rho_l = 1000\,kg/m^3$), density of ice here can be substituted by the density at which pores in firn close, which is close to $830\,kg/m^3$. Assuming that the average density at the surface is $\rho_0 = 300\,kg/m^3$ $\rho_0 = 300$ kg/m^3 the last equation is:

$$M + P_l < 2.12 \cdot P_s$$

Water appearing in a cold pack of snow or firn increases its density. This is done by two different processes. One is refreezing of water (increasing the mass) and the other is settling of snow and firn (decreasing the volume). The latter process occurs without presence of melt water too but at a much slower rate. Temperature change inside snow accounts only refreezing, so the value for ρ_0 should be the density up to which the surface density of a snow will increase solely due to settling when it submerges to the depth of water infiltration while being buried by layers of subsequent accumulation. Assuming that $\rho_0 = 400\,kg/m^3$: $M + P_l < 1.3 \cdot P_s$; if $\rho_0 = 500\,kg/m^3$: $M + P_l < 0.8 \cdot P_s$.

Pfeffer et al, 1991

For quantitative estimation of the contribution of the Greenland ice sheet to the sea level rise authors introduce the notion of the runoff limit, which refers to the altitudinal level above which entire amount of annual melt refreezes in the glacier and below which additional water is allowed to run away. The limit is defined in terms of spatially and temporally varying annual melt M and annual accumulation C by the following condition:

$$M \geq \frac{c}{L} C \left| T_f \right| + \left(C - M \right) \frac{\rho_{pc} - \rho_c}{\rho_c},$$

Where c – heat capacity of ice, L – latent heat of fusion of ice, T_f – initial firn temperature, ρ_{pc} – density of pore closure in firn, ρ_c – initial firn density.

Thus the first term on the right-hand-side of the equation is the cold content of the current year's accumulation, and the second term describes the mass of melt water required to fill the pore space of the accumulation layer reduced by the ablation. This suggests a combined approach designed to account in a way for both cold-content and pore space restrictions to runoff. All additional melt water that is generated after the condition suggested above is satisfied contributes to the runoff. It is to be noted the approach as it is presented is numerically rather light. Moreover authors calculate the value for $\frac{M}{C}$ from the equation with the following values for the parameters (uniform over the ice sheet): $\rho_{pc} = 830\,kg/m^3$, $\rho_c = 300\,kg/m^3$, $T_f = -15°C$ and suggest that the runoff limit can be defined by an even simpler expression: $\frac{M}{C} = 0.7$. Yet this simplicity is reached at the expense of method performance as it does not account for the possibility of the melt water to penetrate deeper than the depth the current season's accumulation, nor it allows for superimposed ice formation.

Janssens and Huybrechts, 2000

Authors test different model but only one of them (*iv*) was not completely taken from a different study, though essentially the same approach as in (Pfeffer et al, 1991) is used with the only exception that the temperature of the active layer T_f was equaled to the mean annual temperature which was not taken uniform over the ice sheet but could be changed as a function of latitude, altitude and time.

Braithwaite et al, 1994

The concept for the scheme is the same as the second run off condition by Shumskii: assuming that the cold content of a snow-firn-ice column is significant there should be a certain limit to the amount of water that can be refrozen in the available pore space in the time period of one year. So the maximum amount of water that can be refrozen before melt water will start to form runoff is defined before the actual start of melting. The limiting factor here is the volume of pores.

The approach allows contains two tuning parameters: density at the surface and the density of firn when it becomes impermeable to water. But the nature of the tuning parameters is such that they demand a field validation rather than guessing and are unique for each specific site and moment in time. Considering that the authors used the refreezing model to study variations of near surface firn density in the lower accumulation area of west Greenland (Pakitsoq at 70°N) the assumption of a high cold content of the active layer by the end of the winter is most likely to be valid. But this is not a general rule and in other regions the scheme will not be acceptable. Even though the scheme for estimation of conditions for initiation of runoff repeats the approach by Shumskii (1955, 1964) below we present the logical line of the authors as it is elegant and is in a way different from that by Shumskii.

Assume (fig. 4) a surface layer of firn including several annual layers is wetted during the summer (year 1) by melt water which reaches to a maximum depth of h in the autumn (year 1). The density at the bottom of the wetted layer is ρ_2. The wet firn freezes in the winter (between year 1 and year 2) and an accumulation C of new snow, of thickness h_0 and density ρ_1, is added to the firn by spring (year 2) . Melting M at the surface of the new snow in summer (year 2) and refreezing of the melt water results in a new surface layer of wet firn in the autumn (year 2) . For a steady state, this new firn layer in the autumn of year 2 has the same thickness h and same density gradient $\rho_1 - \rho_2$ as the surface firn layer in the autumn of year 1. The glacier surface rises by $h_0 = C/\rho_1$ in local coordinates (i.e. as measured against an accumulation stake) due to snow accumulation C, and then falls again by M/ρ_1 due to melting M at the surface. The net change of surface elevation (in local coordinates) from autumn to autumn is then $\frac{C-M}{\rho_1}$ and the bottom of the wetted layer, i.e. where the density becomes ρ_2, rises by the same amount that represents the thickness of an annual layer. As there is no run-off, and neglecting density variations below the wetted layer, the net result of an annual cycle of accumulation, melting and refreezing is to add a new firn layer, of density ρ_2 and thickness

$\frac{C-M}{\rho_0}$, to the ice-sheet surface. From conservation of mass, the material in this layer must be equal to the snow accumulation:

$$C = \frac{\rho_2 (C-M)}{\rho_0}.$$

With some re-arrangement, the density of firn just below the wetted layer is:

$$\rho_2 = \frac{\rho_0}{1-\frac{M}{C}}.$$

Alternatively, the melting $\frac{M}{C}$ required to achieve the density ρ_2 is:

$$\frac{M}{C} = \frac{\rho_2 - \rho_0}{\rho_2},$$

which is given by Shumskii (1964, p. 416). The density ρ_2 is said to be increasing with melt M.

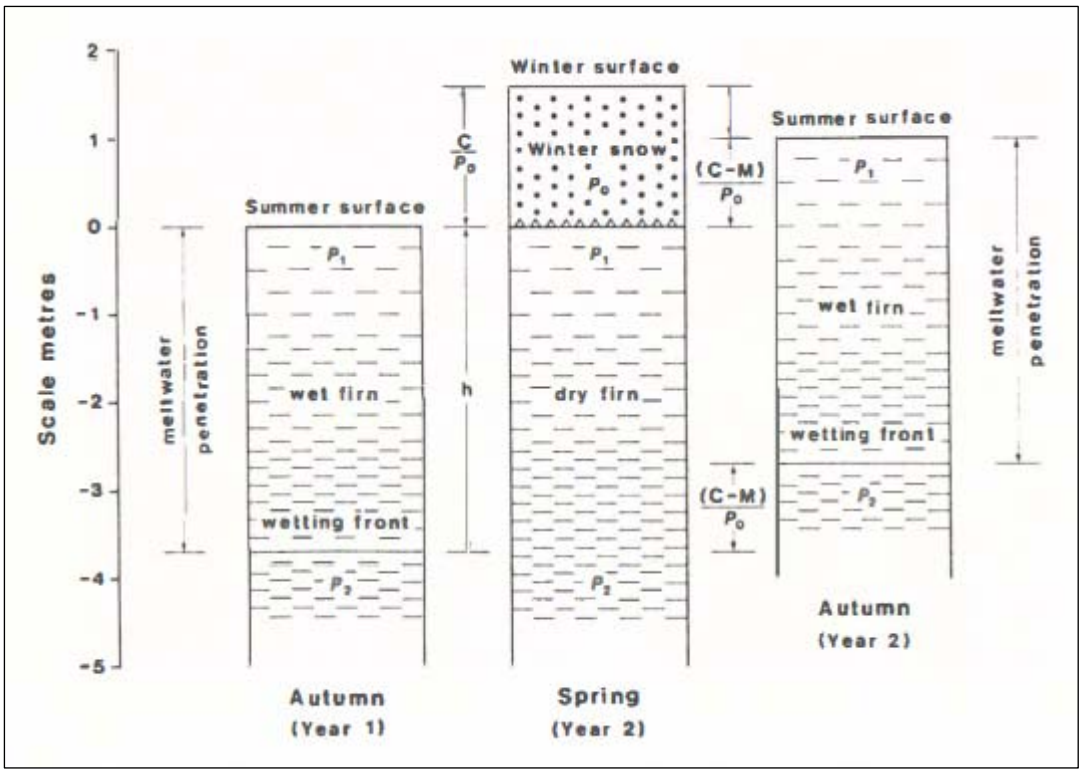

Figure 4. Principles of the simple refreezing model Elevations are measured in local coordinates. (Braithwaite et al, 1994)

By using the following values: $\rho_1 = 375 kg/m^3$ and $\rho_2 = 890 kg/m^3$ in the last formula the authors calculated that if the amount of melt water exceeds 58% of the annual precipitation then any additional water will contribute to the runoff. If not then an annual layer of accumulation looses mass in its first year due to melting at the surface, and then gains the same mass in subsequent years due to refreezing of melt water. Until the depth where firn is transformed to ice is greater than the depth of water penetration there is no runoff.

Huybrechts and de Wolde, 1999

The presented parameterization is based on the same principles as the one by Pfeffer with co-authors (1991) but neglected with the second term on the right hand side of the equation which describes the pore space of the seasonal snow layer. In the model, the amount of refreezing depends on the cold content of the upper ice sheet layers, which puts an upper limit on the production of superimposed ice ($[I] = m/year$).

$$I = \frac{T_{air} \cdot dC}{L}$$

The maximum amount of superimposed ice is equivalent to the latent heat released to raise the temperature of the uppermost 2 m of the ice sheet surface (d) from the mean annual temperature (T_{air}) to the melting point.

Yet it is not quite obvious why the term superimposed ice is used to refer to the mass of melt and rain water that is refrozen on the Greenland Ice Sheet. It is most probably that the term was used erroneously as the physical meaning of the equation used us closer to estimation of refreezing of melt and rain water in the pores of snow and firn in spring when the pack is being warmed up.

Schemes designed for estimation of superimposed ice formation rate

Wakahama et al, 1976

A physically-oriented approach is suggested for describing the process of formation of superimposed ice. It is argued that the major control on the growth of superimposed ice is the rate at which heat energy can be conducted down from the firn – ice interface. This can be obtained by solving the equation of one dimensional heat conduction: $\rho_i C_i \left(\frac{\partial T}{\partial t} \right) = \frac{\partial}{\partial t} \left(k_i \frac{\partial T}{\partial z} \right)$, under special boundary conditions (for notation here and after see the list below).

1) Since in every time step a new layer will be accreted on the existing layer of superimposed ice its thickness will increase. The boundary between two medium (ice and snow) will be moving upwards. The mass of water that has to be frozen in a unit volume of snow for it to become superimposed ice is: $\rho_{si} - \rho_{ws}(1-w)$. It is not quite obvious why authors include the term $(1-w)$ which decreases the mass of water to be refrozen by the amount of free water, this mass also has to be refrozen. It might be that the suggestion is that the structure of superimposed ice includes free water in liquid phase and the quantity of the water is equal to w. And $L \cdot \left(\rho_{si} - \rho_{ws}(1-w) \right)$ is the resulting amount of heat that has to be conducted away.

Thus $\frac{\partial H}{\partial t} = \left(k_i \frac{\partial T}{\partial z} \right) \cdot \frac{1}{L} \cdot \frac{1}{L \cdot \left(\rho_{si} - \rho_{ws}(1-w) \right)}$ – is the change in thickness of the layer of superimposed ice in time; which is also the velocity of movement of the boundary between two medias – superimposed ice and snow.

2) $T = T_0 = 0\ ^\circ C$ at the upper limit of the domain – surface of the superimposed ice;

3) $T = T_b = const$ at the lower limit of the domain – depth where temperature does not change in a seasonal cycle.

Initial temperature distribution in the ice is prescribed to be linear distribution between T_0 and T_b.

The symbols above are defined as follows:

$\rho_i = 917\, kg/m^3$ – density of ice,

C_i – heat capacity of ice,

T – temperature,

t – time,

k_i – thermal conductivity of ice,

z – vertical coordinate (positive upwards),

$\rho_{si} = 890\, kg/m^3$ – density of the superimposed ice,

$\rho_{ws} = 470\, kg/m^3$ – density of wet snow above the superimposed,

$w = 0.15$ – fraction of the liquid water content of snow right above the superimposed ice,

$L = 333.6\, J/kg$ – latent heat of melt for ice (the value used is 333.6 J/kg),

H – thickness of the layer of superimposed ice.

It is not evident in what way the model accounts for different rate of melt (and/or rain) water supply Q, though later in the text of the article results for different values of Q are shown. It is likely that the model implies that the entire mass of water once produced on top of a temperate snowpack is immediately infiltrated down to the superimposed ice-snow interface. Then it is compared with the maximum mass of water that can be frozen in the time step: it can not be bigger. If the condition is met entire melt water volume is refrozen, if not then excess water drains away. In that form the scheme is close to the P-max schemes presented above, though the potential refreezing is calculated not for the entire season but for each of the small time steps and it is dependent (though not directly) on the actual changing meteorological conditions.

The model appeared to produce very good results and correlates well with both the field data of the authors and their laboratory experiments. It was used in several later works for calculating the loss of melt water as superimposed ice. An improvement of the calculation scheme was offered by Boggild (2007).

Wright, 2007

The suggested scheme enables one to calculate the maximum amount of water that can be refrozen at glacier during summer season (i. e. superimposed ice formation) basing on the knowledge of annual and winter air temperatures. So this is a variation on the pattern of P-max approach. The scheme is based on the field measurements at the glacier Midre Lovenbreen in Spitsbergen, where

during the 2002 melt season a series of temperature measurements in the active layer was performed along with meteorological observations.

The distribution of temperature with depth in the end of the spring is parameterized as:

$$T(z) = \left(\frac{z}{d_{ice}} - 1\right) \cdot \left(\overline{T}_a - \overline{T}_w\right) + \overline{T}_a$$

And in the end of summer as:

$$T(z) = -\sqrt{\overline{T_a^2} \cdot \left(1 - \frac{(z - d_{ice})^2}{d_{ice}^2}\right)},$$

where z is the vertical coordinate, d_{ice} is the thickness of thermally active layer on the glacier, $\overline{T_a}$ is the mean annual air temperature and $\overline{T_w}$ – mean annual winter temperature.

On the figure 5 the two temperature distributions are shown along with temperature measurements on the glacier. It is also to be noted that the first temperature profile was measured at the moment when the whole snow pack was warmed up to the melting point but before the start of superimposed ice formation. The second profile was measured after the end of superimposed ice formation.

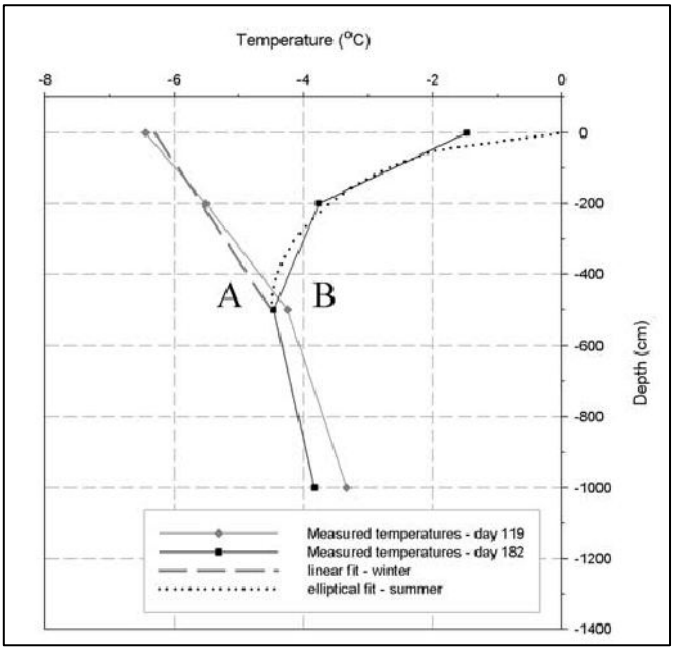

Figure 5. Temperature profiles measured within the glacier ice of Midre Lovenbreen at stake NP7 during the 2002 melt season. Temperature profiles are shown for the period before the start of superimposed ice formation (profile A, day 119) and for after superimposed ice formation has ceased (profile B, day 182). Also shown are the model fits to each measured profile on which the Wright Pmax model is based.

The primary source of heat that forces the temperature profile to evolve during summer is the release of latent heat of fusion as the meltwater accretes and forms a layer of superimposed ice. So the amount of energy needed to heat up the active layer to the temperature conditions measured in the end of summer should correspond to the amount of superimposed ice formed.

Integration of the area between the two lines can be expressed in terms explained above as:

$$E_{ice} = \frac{c_i d_{ice}}{2 \cdot \left(\left(1 - \frac{\pi}{2}\right) \overline{T_a} - \overline{T_w} \right)}.$$

P-max for every year can be calculated on the basis of the following equation:

$$P_{\max B} = \frac{E_{ice}}{L_f} = \frac{c_i d_{ice}}{2 \cdot L_f \cdot \left(\left(1 - \frac{\pi}{2}\right) \overline{T_a} - \overline{T_w} \right)}$$

From the measurements on the glacier the model is best fitted to the data using the following values $\overline{T_a} = -4.5\ °C$, $\overline{T_a} = -6.3\ °C$ and $d_{ice} = 500\,cm$.

It is to be noted that this model was developed solely for using in the mass balance model of a glacier primarily nourished by superimposed ice. Validation was done on only one glacier.

The model does not include any parameterization of water refreezing in the pores of snow and firn because for that study it was not very relevant due to local conditions of physical geography. Yet the logic offered can be used to describe both processes of refreezing of water – in pores of snow and firn (in spring) and as superimposed ice (in summer). For that the first of the functions of temperature distribution with depth should describe the situation for the end of the winter season.

Layered englacial models including description of internal accumulation

A number of layered models describing evolution of snow and firn pack was offered. These schemes being more demanding for extensive field data and computational recourses allow for much complete and precise treatment of physical properties and processes occurring in snow and firn pack. These models all use similar parameterizations of the subsurface processes and simulate snow properties such as density, temperature, liquid water content of snow, some also account for microstructure of snow. They have proved to reproduce results of observations reasonable well provided that sufficient calibration was applied and an extensive and reliable dataset is used for input.

One example is a physical model SNOWPACK developed in SLF for the Swiss avalanche warning system (Bartelt, Lehning, 2002). Though it was not created for the purpose and wasn't properly validated for conditions of high latitudes the model was applied for simulation of superimposed ice formation on Kongsvegen glacier in Spitsbergen (Obleitner and Lehning, 2004). After modification of the water transport routine of the model to also treat effects of water ponding the overall model was producing remarkably good results.

A less detailed and physically extensive but more general and flexible approach for estimation of the rates of internal accumulation was offered by Greuell and Konzelmann (1994). The model calculates temperature (T), water content (w) and density (ρ) in a snow/firn/ice pack on a vertical irregular 1D grid extending from the surface to depths of first tens of meters. It is driven by and coupled to a separate surface energy balance model with other three parameters: surface temperature (T_0), subsurface energy flux (Q_G) and energy available for melt (Q_m).

Originally the model was applied to study the energy and mass balance at the Swiss ETH (Federal Institute of Technology) camp on Greenland (Greuell and Konzelmann, 1994). Later it was applied and

further developed by a number of authors including Bassford (2002), Bougamont et al. (2005), Bassford et al (2006), Reijmer and Hock (2008), Ettema et al. (2010), and recently by Wheler and Flowers (2011), Reijmer et al. (2012) and van Pelt et al. (2012). The model is usually referred to as SOMARS (Simulation Of glacier surface Mass balance And Related Sub-surface processes). The overall logic remained the same for each implementations of the model though some authors used additional modules and different parameterizations for several processes. Below we would like to present a general scheme of the model as presented in Bassford (2002) and also Bassford et al. (2006) as it includes also description of superimposed ice formation.

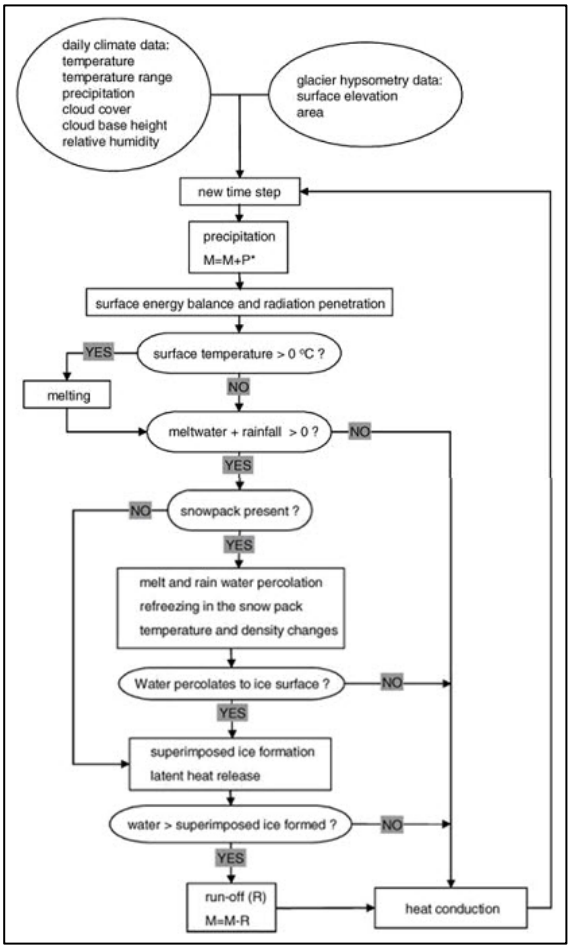

Figure 6. Schematic structure of the surface mass balance model. M – transient mass balance, P - precipitation rate, R – runoff.*

Changes in the temperature of cells are calculated using the core thermodynamic equation of the model (for notation see table 2 below):

$$\rho \cdot c_i \frac{dT_i}{dt} = \frac{\partial}{\partial z}\left(K \frac{\partial T_i}{\partial z} \right) + \frac{\partial Q_t}{\partial z} - \frac{\partial}{\partial z}\left(M \cdot L_i \right) + \frac{\partial}{\partial z}\left(F \cdot L_i \right),$$

Table 2. Notation for the principle equation of the englacial model

sign	units	description
ρ	kg/m^3	density
$c_i = 2012$	$J/(K \cdot kg)$	specific heat capacity of snow/firn/ice
T_i	K	snow/firn/ice temperature
t	s	time
z	m	vertical coordinate
K	$J/(m \cdot K \cdot s)$	effective heat conductivity of snow/firn/ice
Q_t	$J/(m^2 \cdot s)$	energy flux between the surface and the atmosphere,
M	$kg/(m^2 \cdot s)$	melt rate
$L_i = 333.5 \cdot 10^{-3}$	J/kg	latent heat of fusion of snow/ice
F	$kg/(m^2 \cdot s)$	freezing rate

The thermodynamic equation is claimed to be valid for the upper 20 m of a glacier because the influence of advection and conduction parallel to the surface as well as energy produced by ice deformation are negligible. Effect of the vertical advections is accounted for by the model intrinsically because Lagrangian coordinated are used.

A zero heat flux is assumed at the lower boundary of the grid. Part of the short-wave radiation penetrates the surface into deeper layers of snow, firn and ice, influencing on the englacial temperature and the surface energy balance - surface receives less energy for melting while the underlying layers are heated.

Melting is considered after calculating of the energy exchange between the glacier surface and the atmosphere. If the temperature of the uppermost grid cell is raised above $0°C$, then it is set to $0°C$ and the excess energy is used for surface melting.

Subsurface melting is not calculated explicitly in the model but may occur as a result of radiation penetration. For conservation of energy, subsurface cells with a temperature $> 0°C$ are set to $0°C$ and the extra energy is used to heat the uppermost grid cell. If melt water enters a cell with a temperature $< 0°C$ then refreezing takes place. The maximum refreezing rate that can occur in a cell at a point in time is limited by the amount of latent energy required to raise the cell temperature to $0°C$. If the amount of melt water percolating into a cell is less that its capacity, then all of the melt water refreezes. If the quantity of melt water exceeds the capacity then excess melt water will be issued to the cell below. Refreezing results in an increase in cell density and temperature. If sufficient refreezing occurs in a cell to increase its density to that of ice ($910\,kg/m^3$) then an impermeable ice layer forms in the snowpack.

Melt water continues to move down through the grid until either all of the melt water refreezes or an impermeable ice cell is encountered, such as an ice layer within the snowpack or the underlying continuous glacier ice, on which superimposed ice forms. This process is described using the logic from (Wakahama et al., 1976).

There are three additional modules of the model that are worth mentioning and were presented in the original study (Greuell and Konzelmann, 1994) and also in (Reijmer and Hock, 2008).

- One of them is the dry snow densification which occurs independently of densification due to refreezing (which is usually much faster) and dependent on the initial density of the grid cell, time, temperature accumulation.
- The other module accounts for the liquid water held in porous snow and firn by capillary forces. This allows the model the account for internal accumulation during autumn and also slightly reduces the speed of water penetration in spring.
- And the third one regulates slush and runoff formation from the water that percolated through the successive vertical layers and ahs reached impermeable ice. On top of the ice layer, water can accumulate and form a slush layer where all pore spaces are occupied by water. Water is removed from the slush layer as a function of the surface slope β (Reijmer and Hock, 2008). The amount of water leaving the slush layer corresponds to the model ablation. The fraction of water in the slush layer that is lost per time-step increases with increasing β. The speed at which runoff occurs is assumed to be greater at the surface than within the snowpack. But authors note that up to date no measurements are available to verify parameters used in the scheme.

Grid design and time stepping logic significantly differed between different realizations of the original model as it was applied for a wide range of glacial conditions and purposes. Design of the grid system for each situation is non-trivial because of the moving upper boundary caused by snow accumulation and surface melting, densification. But in all cases the following patterns were preserved: Each grid element has a size (dz_i), a density (ρ_i) and water content (w_i). The temperature (T_i) is attributed to the grid points, which are situated at the centre of the grid elements. The size of the uppermost grid element varies between 3 and 12 cm. The element size increases linearly (or exponentially) with depth and reaches 3 m at a depth of 25 m. Initially, numerous annual layers each containing one or more grid elements may be specified, the requirement is set that interannual surfaces always coincide with the boundary between two grid elements. Essentially, grid elements do not move relative to material points. This has the advantage that mass cannot artificially diffuse through the grid enabling the simulation of the formation of superimposed ice, for instance. The sizes of the grid elements are continually changing owing to densification while accumulation, condensation and evaporation increase or decrease the size of the uppermost grid element. Grid elements that become too small during a simulation are fused with neighbouring ones and elements that become too large are split up into two new ones.

The time step is variable. It is automatically computed depending on the actual energy flux coming in from the atmosphere, on the size of the grid elements, on density and on conductivity. Possible time steps range between 1 and 30 min. After 60 min of simulation the length of the time step is recomputed.

This model was successfully applied for mass balance studies on Greenland ice sheet: (Greuell and Konzelmann, 1994), (Bougamont et al., 2005), (Ettema et al., 2010), (Reijmer et al., 2012); ice caps of Severnaya Zemlya (Bassford, 2002), (Bassford et al., 2006); Spitsbergen glaciers Lomonosovfonna, Kongsvegen: (Erath, 2005), (Schrott, 2006), (Krismer, 2009), (van Pelt et al., 2012); Storglaciaren in northern Sweden (Reijmer and Hock, 2008) and Gonjek Range, Southwest Yukon, Canada (Wheler and Flowers, 2011).

An inter-comparison study of different models was done by Reijmer with co-authors (2012). I comprised all of the above mentioned models. The parameterizations were forced and coupled to a regional atmospheric climate model (RACMO2) and applied to the Greenland ice sheet. Also another RCM with an incorporated layered snow model (RCM – MAR, snow model - CROCUS) was included in the comparison.

The annual, period average (1958–2008) and ice sheet averaged amount of refreezing calculated with the different parameterizations differed up to a factor of two with the layered models. The spatial fields showed large differences as well, especially in the lower areas of the ice sheet (up to a factor 5), which was also noted in an earlier comparison study by Janssens and Huybrechts (2000). Results suggested large sensitivity of internal accumulation rates to the value of the active layer depth that is included in most models and determines the cold content that can be spent for refreezing of the melt and rain water. It was also noted that all parameterizations could be tuned within realistic limits, to produce ice sheet and annual average amount of refreezing similar to that generated by layered models (RACMO2+SOMARS, MAR+CROCUS), but this did not necessarily result in better spatial correspondence.

Main uncertainties in calculation of internal accumulation rates

As a conclusion we would like to revise the main difficulties that were met by authors studying spatial and temporal variability of internal accumulation on glaciers and that need to be addressed in further work. The review is not supposed to complete and exhaustive only major knowledge gaps are listed.

Firstly it is to be noted that there are very dew direct estimations of internal accumulation rates to which modeling results can be compared to. This is understandable if one considers the spatial and temporal domain where the process occurs. In terms of mass balance it is significant in a 10 m thick layer in the whole accumulation area and in terms of energy balance it is to be accounted for in ablation area as well. Refreezing occurs in spring, summer and autumn seasons. The direct way to measure internal accumulation rates would be to perform a long and extensive repeated measurements series of snow/firn/ice density in the thermally active layer of glacier (around 10 m) at several locations on a glacier (at least in every glacial zone), which is close to impossible. Another constraint is that the measurements should introduce as less as possible disturbance in the snow/firn/ice pack as otherwise it would affect water flow, refreezing and compaction of the media. One possible solution would be to use thermistor multiple strings or remote sensing techniques such as GPR studies.

Another important issue connected with precise description temperature evolution inside a snow/firn/ice pack is the uncertainties in heat conductivity of that porous media (Sturm et al., 1997).

Usually it is substituted with effective heat conductivity which is described as a function of density and (or) temperature, microstructure. The uncertainties are the largest (30%) for seasonal snow and are less considerable for pure ice (Greuell, Konzelmann, 1994).

Densification of snow and firn is a very important but still poorly understood process. It's relative importance is much less in the ablation zone and congelation zone of glaciers where seasonal snow s deposited directly on an impermeable layer of superimposed ice. But a dominant part of the internal accumulation domain lays outsides these zones and densification processes not directly connected with refreezing of melt and rain water.

Water flow in cold snow and firn masses often occurs not homogenously but rather in pattern that is denoted in literature as preferential flow, water piping, flow finger formation, etc (Marsh P., and Woo M-K., 1984). This effect is important from hydrological and thermodynamical points of view because by means of this process additional masses of water and thus of latent heat of fusion are delivered at greater depths, where penetration of wetting front is not feasible because usually not enough water is available and its penetration rate is rather slow, by this preferential flow decreases the depth of penetration of the water front. This will result in smoothing of vertical temperature gradient and thus less heat will be lost during cooling of the surface layer of a glacier in winter. In case preferential flow forms channels with increased water permeability of snow and firn that extend to depths of more then approximately 10 m, then it is very likely that the water not be frozen during the next winter acting as a stable heat source (Humphrey et al., 2012). One of the most efficient ways to study the characteristics of preferential flow in glaciers is to measure the temperature evolution in the surface layer of a glacier. Then sudden spikes at depths lower then corresponding position of the wetting front would be an indication of flow finger formation close to the sensor that has registered the temperature instability. To increase probability to "catch" a flow pipe formation several strings with sensors measuring temperature can be installed (Conway H., Benedict R., 1994).

It is also very likely that the snow and firn water holding capacities are very important parameters for estimating the magnitude of refreezing. It is obvious that it almost totally controls ice formation of infiltration ice during "autumn" (when the active layer cools down). In several studies of water flow in snow (Golubev, 1976; Marsh and Woo, 1984) and refreezing schemes (for example Reijmer and Hock, 2008) it is suggested that for a volume of water (be it a wetting front or a flow finger) to move from one layer to another it is necessary that the layer is not only warmed up to the melting point but also its water content reaches the maximum water holding capacity.

Acknowledgments

This report is a contribution of the Nordic Centre of Excellence SVALI, Stability and Variations of Arctic Land Ice, funded by the Nordic Top-level Research Initiative (TRI). The author is also grateful to his supervisors Veijo Pohjola and Carl Boggild for helpful revision.

Literature

Bartelt P. and M Lehning. 2002 A physical SNOWPACK model for the Swiss avalanche warning. Part I. Numerical model. *Cold Reg. Sci. Technol.*, 35(3), 123–145;

Bassford, R.P., 2002. Geophysical and numerical modelling investigations of the ice caps in Severnaya Zemlya. (PhD thesis, University of Bristol, U.K.);

Bassford, R. P., M. J. Siegert, J. A. Dowdeswell, J. Oerlemans, A. F. Glazovsky, and Y. Y. Macheret, 2006 Quantifying the mass balance of ice caps on Severnaya Zemlya, Russian High Arctic. I: Climate and mass balance of the Vavilov Ice Cap, Arct. Antarct. Alp. Res., 38(1), 1– 12;

Benson C. S., 1960 Stratigraphic studies in the snow and firn of the Greenland ice sheet, Thesis in partial fulfillment of the requirements for the degree of doctor of philosophy, California Institute of Technolohy, Pasadena, California;

Benson, C.S., 1961 Stratigraphic studies in the snow and firn of the Greenland Ice Sheet. Folia Geogr. Danica 9, 13–37;

Boggild, 2000 Preferential flow and meltwater retention in cold snow packs in West Greenland, Nordic Hydrology 31(45);

Bøggild C., 2007 Simulation and parameterization of superimposed ice formation, Hydrol. Process.21, 1561–1566;

Bougamont, M., Bamber, J. L., and Greuell, 2005 W.: A surface mass balance model for the Greenland ice sheet, J. Geophys. Res., 110, doi:10.1029/2005JF000348;

Braithwaite R. J., Laternser M., Pfeffer W. T., 1994 Variations of near surface firn density in the lower accumulation area of the Greenland ice sheet, Pâkitsoq, West Greenland, J. of Glaciology V. 40, No 136;

Cogley, J.G., R. Hock, L.A. Rasmussen, A.A. Arendt, A. Bauder, R.J. Braithwaite, P. Jansson, G. Kaser, M. Moller, L. Nicholson and M. Zemp, 2011 Glossary of Glacier Mass Balance and Related Terms, IHP-VII Technical Documents in Hydrology No. 86, IACS Contribution No. 2, UNESCO-IHP, Paris;

Conway H. and Benedict R., 1994 Infiltration of water into snow, Water Resources Research 30-3, p. 641-649;

Erath S, 2005 MS Simulation of the mass and energy balance at Kongsvegen 2001-2003 (Ms. Thesis, Dep. of meteorology and geophysics, University of Innsbruck, Austria);

Ettema, J., van den Broeke, M. R., van Meijgaard, E., van de Berg, W. J., Box, J. E., and Steffen, K., 2010 Climate of the Greenland ice sheet using a high-resolution climate model – Part 1: Evaluation, The Cryosphere, 4, 511–527, doi:10.5194/tc-4-511-2010;

Golubev G. N., 1976 Gidrologiya lednikov (Hydrology of glaciers), Gidrometeoizdat, Leningrad (in Russian);

Greuell W. and Konzelman T., 1994 Numerical modelling of the energy balance and the englacial temperature of the Greenland ice sheet. Calculations for the ETH-Camp location (West Greenland, 1155 m a. s. l.), Global Planet. Change, 9, 91–114, doi:10.1016/0921- 8181(94)90010-8;

Humphrey, N. F., J. T. Harper, and W. T. Pfeffer, 2012 Thermal tracking of meltwater retention in Greenland's accumulation area, J. Geophys. Res., 117, F01010, doi:10.1029/2011JF002083;

Huybrechts P. and de Wolde J., 1999 The Dynamic Response of the Greenland and Antarctic ice sheets to Multiple-century climate warming, J. Climate, 12, 2169–2188;

Janssens I. and Huybrechts P., 2000 The treatment of meltwater retention in mass-balance parameterizations of the Greenland ice sheet, Ann. Glaciol., 31, 133–140;

Krismer T., 2009 Local and spatial mass balance modelling on an Arctic glacier, Kongsvegen, Spitsbergen (Ms. Thesis, Dep. of meteorology and geophysics, University of Innsbruck, Austria);

Marsh P., and Woo M-K., 1984 Wetting front advance and freezing of meltwater within a snow cover, 1. Observations in the Canadian Arctic, *Water Res. Res., Vol. 20,(12),* pp. 1853-1864;

Müller, F., 1962 Zonation in the accumulation area of the glaciers of Axel Heiberg Island, N.W.T., Canada. J. Glaciol. 4, 302–313;

Obleitner F. and M Lehning, 2004 Measurement and simulation of snow and superimposed ice at the Kongsvegen glacier, Svalbard (Spitzbergen). *J. Geophys. Res.*, 109(D4), D04106.(10.1029/2003JD003945.);

Oerlemans J., 1991 The mass balance of the Greenland ice sheet: sensitivity to climate change as revealed by energy balance modelling, The Holocene, 1, 40–49;

van Pelt W. J. J., Oerlemans J., Reijmer C. H., Pohjola V. A., Pettersson R., and van Angelen J. H., 2012 Simulating melt, runoff and refreezing on Nordenskioldbreen, Svalbard, using a coupled snow and energy balance model, The Cryosphere, 6, 641–659, doi:10.5194/tc-6-641-2012;

Pfeffer W., Meier M., and Illangasekare T., 1991 Retention of Greenland runoff by refreezing: implications for projected future sea level change, J. Geophys. Res., 96;

Reeh N., 1991 Parameterization of melt rate and surface temperature on the Greenland ice sheet, Polarforschung, 59, 113–128;

Reijmer C. and Hock R., 2008 Internal accumulation on Storglaciaren, Sweden, in a multi-layer snow model coupled to a distributed energy- and mass-balance model, J. Glaciol., 54, 61–72;

Reijmer C. H., van den Broeke M. R., Ettema J. and Stap L. B., 2012 Refreezing on the Greenland ice sheet - a comparison of parameterizations, The Cryosphere, 6, 743–762, doi:10.5194/tc-6-743-2012;

Schneider T. and Jansson P., 2004 Internal accumulation in firn and its significance for the mass balance of Storglaciaren, Sweden, J. Glaciol., 50, 25–34;

Schrott D., 2006 Flachenhafte Modellierung der Energie- und Massenbilanz am Hintereisferner - das Haushaltsjahr 2003/04 (Ms. Thesis, Dep. of meteorology and geophysics, University of Innsbruck, Austria) (In German);

Shumskii P. A. 1955 Osnovi strukturnogo ledovedeniya (Principles of structural glaciology), Akademia Nauk, Moscow (in Russian);

Shumskii P. A. 1964 Principles of structural glaciology, translation by David Kraus, Dover Publications, Inc., New York;

Sturm M., Holmgren J., Konig M., Morris K., 1997 The thermal conductivity of seasonal snow. J. Glaciol.-43-143, 26–41;

Trabant D. and Mayo L., 1985 Estimation and effects of internal accumulation on five glaciers in Alaska, Ann. Glaciol., 6, 113–117;

Wakahama G., Kuroiwa D., Hasemi T. and Benson C.S., 1976 Field observations and experimental and theoretical studies on the superimposed ice of McCall Glacier, Alaska. J. Glaciol., 16(74), 135–149;

Wheeler B. A. and Flowers G. E., 2011 Glacier subsurface heat-flux characterizations for energy-balance modelling in the Donjek Range, southwest Yukon, Canada, J. Glaciol.-57-201, 121–133;

Wolken, G. J., M. Sharp, and L. Wang (2009), Snow and ice facies variability and ice layer formation on Canadian Arctic ice caps, 1999–2005, J. Geophys. Res., 114, F03011, doi:10.1029/2008JF001173;

Wright A., Wadham J., Siegert M., Luckman A., Kohler J., and Nuttall A., 2007 Modeling the refreezing of meltwater as superimposed ice on a high Arctic glacier: A comparison of approaches, J. Geophys. Res., 112, F04016, doi:10.1029/2007JF000818;