Glacier Module 1 (August, 26-27, 2014)
Basics and processes at the surface

Learning Objectives

1. You know the basic glaciological terms and concepts.

2. You are able to explain and discuss the processes of transformation from snow to glacier ice.

3. You are able to recognize the most important phenomena at the glacier surface, to name them and to explain their formation/origination.

Terms and Concepts

Ice sheet, ice cap, glacier, accumulation area, ablation area, equilibrium line, glacier flow, (medial, lateral) moraine, glacier snout, glacier foreland, snow, snow crystal, hoar, firn, metamorphosis and its processes, snow profile, impurities in snow/firn/ice, density and density profile, sintering, transition from snow to ice, clathrate development, morphology of glacier ice (crystals, re-orientation and growth of crystals, air bubbles) supra glacial drainage, ogives, banding, moraines, debris cover, glacier table, cryoconite, sastrugi

References and Further Reading

Literature

Cogley et al. (2011): Glossary of glacier mass balance

Post & Chapelle (2000): phenomena on glacier surfaces, exciting pictures!

Hooke, R.L.B. (2005): transition from snow to glacier ice

Cuffey and Paterson (2010): energy balance, transition from snow to glacier ice

Fierz et al. (2009): snow profile

Weblinks

Glossary (Cogley et al., 2011): http://www.wgms.ch/downloads/Cogley_etal_2011.pdf

Snowcrystals: www.its.caltech.edu/~atomic/snowcrystals/

Additional Information

When considering and analysing processes at the glacier surface, glacier motion (cf. Module 2) can be neglected, i.e. a glacier is considered as a static entity. Basis for the spatial understanding of mass balances and temperatures of glaciers (cf. Module 3, 4) are the processes and phenomena at one single point of the glacier (without spatial context).

Transformation from snow to ice

The transformation of snow to ice occurs as the volume of air filled pores is reduced and the material increases in density. New snow has a very low density (100-200 kg m^{-3}) compared to glacier ice (830-917 km m^{-3}). Snow that has survived one or more melt seasons and has begun this transformation is known as firn and has a density of 300-700 kg m^{-3}. Firn becomes glacier ice when the interconnecting air passages between the grains are sealed off (so called “pore close-off”). Consequently the transition between firn and ice takes place at the so called “pore close-off depth”. In glacier ice, air is present only as bubbles and any increase in density results from compression of them.

The processes, by which snow is transformed into ice, and the time taken for the transformation to occur, depend on the climate. In cold polar regions and at high altitudes (no melting), the principle mechanisms are restructuring by wind, movement of crystals relative to one another, recrystallization by means of vapour transfer and shape and internal deformation of crystals.

The transformation of snow to ice is greatly accelerated when melting occurs at the surface or within the snowpack (low altitudes in low and mid-latitudes). Water percolates downwards through the snow, where it will refreeze if it comes into contact with cold snow or ice, forming superimposed ice.

Phenomena at the glacier surface and in the ice

Microscopic features: i.e. crystal characteristics, air bubbles, banding
Macroscopic features: i.e. moraines, crevasses, ogives, melt water phenomena

Ice crystals: Large crystal grow at small crystal costs. They recrystallize under high pressure and shear deformation (basal layer).

Air bubbles: The original spheric air bubbles can strongly deform. Thereby, the elongation can be a minimal for the total deformation.

Banding: in the ablation area, sedimentary layers and scars of crevasses can show a banding with a convex geometry in direction of the glacier flow. The differences in colour are based on different crystal sizes and air bubble content.

Moraines: debris, originating from weathering is falling from the surrounding rockwalls on the glacier surface and can regroup to: marginal moraines (lateral moraine → medial moraine, end moraine); ground moraine (material falling in the bergschrund + material from abrasion at the glacer bed); moraines on the surface (medial moraine → special phenomena originating from albedo and shading effects: glacier table, sand cones, cryoconite holes).

Crevasses: Cracks in the ice form perpendicular to the maximum tensile strain. Open crevasses are a clear sign of extending flow, zones of compressing flow do not have crevasses. Old scarred crevasses are outside of areas of extending flow. The most important crevasses: Randkluft (rock/ice) and Bergschrund (firn/ice); transverse and longitudinal crevasses, seracs, marginal and radial crevasses.

Ogives: undulated structures with annual formation below ice falls. The massively stretched ice in the ice fall reveals a large surface and suffers in the summer months stronger mass loss compared to the winter months. In the compression zone below the ice fall the summer ice forms valleys.

Melt water phenomena: In zones of compressing flow the subsurface flow forms steep-sided channels. On cold (impermeable) ice ponds can form, which will deepen due to the lowered albedo and warm water temperature (entonnoirs). In crevassed zones, glacier moulins link the subsurface flow with the intra- and subglacial hydrological drainage system which ends at the glacier snout.

Glacier Module 2 (August, 26-27, 2014)
Glacier mass balance

Learning Objectives

1. You know the terms of mass balance.

2. You know the important methods to determine the mass balance of a glacier.

3. You know the most important sources of uncertainties.

Terms and Concepts

Mass balance (point, area, time), specific mass balance, methods to determine mass balance (and their sources of uncertainty), natural/stratigraphic budget year, hydrological year, mass balance gradient

References and Further Reading

Literature

Cogley et al. (2011): Glossary of glacier mass balance

Cuffey and Paterson (2010): glacier mass balance

Kaser at al. (2003): mass balance measurements with the glaciological method

Anonymous (1969): Mass-balance terms

Østrem & Brugman (1991): Glacier mass balance measurements, a manual.

Zemp et al. (2010): comparison of glaciological and geodetic method, discussion of uncertainties

Weblinks

World Glacier Monitoring Service (WGMS): www.wgms.ch background information’s for mass balance measurements, publications, mass balance data

Additional Information

Mass balance: definitions and processes

Mass balance is the change in the mass of a glacier, or part of the glacier, over a specific time period. The term mass budget is a synonym. The time period is often a year or a season. The definition of “year” depends on the method adopted for measurements of the balance.

Accumulation (c)

1. All processes that add the mass to the glacier
2. The mass gained by any of the processes (1.), expresses as a positive number:
    a. Snow fall
    b. Deposition of hoar, freezing rain, solid precipitation

c. Gain of windborne blowing snow and drifting snow
d. Avalanching
e. Basal freeze-on
f. Internal accumulation

Ablation (a)

1. All processes that reduce the mass of the glacier
2. The mass lost by means of any of the processes (1.), expressed as a negative number:
   a. Melting
   b. Calving
   c. Loss of windborne blowing and drifting snow
   d. Avalanching
   e. Sublimation

The mass balance is calculated by integrating accumulation (c) and ablation (a) over time (t) and the total glacier surface (S):

$$B = \int_S b dS$$

$$b = c + a = \int_{t_0}^{t_i} \dot{b}\, dt$$

$$c = \int_{t_o}^{t_i} \dot{c}\, dt$$

$$a = \int_{t_o}^{t_i} \dot{a}\, dt$$

B = mass balance, S = glacier surface, c = accumulation, a = ablation, t_0, t_i = start and end oft he considered time span and b = cumulative mass balance over the time period t_0 to t_1 at a specific point on the glacier surface.

In these equations, the minuscule letters refer to single points, the capital letters to the total glacier. The specific net-balance is defined to be the mass balance (B) divided by the glacier surface (S). The net balance of different glaciers can be compared and allows estimating the mass balance of unmeasured glaciers. To determine the mass balance of a glacier is difficult and contains many uncertainties. Therefore a combination of several different methods is often used.

Mass-balance profile is the variation b(z) of mass balance with altitude. The **mass balance gradient** is the of change of mass balance with altitude, that is, the derivative db/dz of the mass-balance profile b(z). If mass balance varies linearly with altitude, the mass-balance gradient will be constant with z; if not, the gradient will vary with z. The mass-balance gradient at the equilibrium-line altitude is called the activity index.

Mass-balance year

The time span, equal or approximately equal in duration to one calendar year, to which the annual mass balance in any time system refers.

In the **stratigraphic system** the annual mass balance is the change of mass during the period between formation of two successive minima in the sequence of annual cycles of mass growth and decline. These minima are usually reached at different times in successive years, and the duration of the mass-balance year may therefore vary irregularly and substantially in duration from year to year. Point mass balances can be determined unambiguously in the stratigraphic system, but glacier-wide determinations require the assumption that the diachronous character of the summer surface can be neglected.

In the **fixed-date system** the first day of the mass-balance year is always on the same calendar date, which is typically chosen to coincide with the start of the local hydrological year. Due to logistical constraints it is often impossible to conduct field surveys on these exact dates. Therefore the data need to be corrected, which is often done by estimating ablation and accumulation between the survey date and the fixed date using meteorological data from a nearby weather station or a database of upper-air measurements.

In the **floating-data system** the mass-balance year is defined by the calendar dates of the two successive surveys, which may vary from year to year.

In the **combined system** two time systems of mass-balance measurement are combined, usually the stratigraphic system with either the fixed-date system or the floating-date system.

Different methods to determine the mass balance

Direct glaciological method

A method of determining mass balance in-situ on the glacier surface by measurements of accumulation and ablation, generally including measurements at stakes (ablation), in snow pits (accumulation) and snow probing in the accumulation area. The measurements may also rely on depth probing and density sampling of the snow and firn, and coring. They are made at single points, the results from a number of points being extrapolated and integrated (by using contours of equal balances) to yield the surface mass balance over a larger area such as an elevation band or the entire glacier. The most important value is the net balance which can be compared with other glaciers. The source of uncertainties mainly originates from the set up of the network of stakes and pits, the density assumptions, the inter-/extrapolation techniques, as well as the survey dates and related time systems which can strongly differ between measurement programs.

Geodetic method

Geodetic method is an umbrella-term for any method for determining mass balance by repeated mapping of glacier surface elevations to estimate the volumetric balance; cartographic method and topographic method are synonyms. The conversion of volume change to mass change (i.e. the mass balance) requires information on the density of the mass lost or gained, or an assumption about the time variations in density. Elevation changes are commonly measured using repeated altimetry, photogrammetry or ground surveys. In the past, glacier mapping relied on ground surveying with theodolites and similar instruments, but global navigation satellite system receivers are now common, offering more rapid and more accurate coverage. The entire glacier surface may be mapped, but discrete elevation measurements, for example along a central flowline, are often extrapolated to the full glacier surface. Major source of uncertainties are the assumptions about density of firn and ice required to convert volume change to mass change.

Hydrological method

A method of determining the mass balance indirectly by solving the water balance for the change in storage ΔW in a drainage basin:

$\Delta W = P - E - Q$,

with P the precipitation, E the evapotranspiration and Q the discharge, each of these quantities being a total over a specific time period. In practical work the hydrological method can be applied only to an entire drainage basin. It does not provide any information on the spatial distribution or gradients. The quantity ΔW will include changes in storage in lakes, seasonal snowpatches, soil and aquifers as well as in the glacier. Each of these changes must be accounted for to isolate the mass balance of the glacierized part of the catchment, but the changes in storage other than in the glacier and the snow cover are often assumed to be negligible over annual periods.

Index method

There is an assumed function between the equilibrium line altitude at the end of the budget year and the glacier mass balance. With such empirical relations the mass balance can be estimated based on mapped equilibrium lines (from aerial photo) with small effort. As a rule of thumb an accumulation area ratio (AAR) of 0.6 has been approved. The values for AAR_0 scatter considerable (0.5-0.75) and without calibration the method is only useful for trend estimations.

Linear mass balance model

This method is based on the assumption that the mass balance at one point and time can be described by a variable depending on the location, a time-dependent variable and a random error. It is furthermore assumed that the mass balance at one single point (sometimes also a combination of points) is linearly linked to the total mass balance of a glacier. When over a longer time period of at least a few years the total mass balance of a glacier is known from other sources (usually from geodetic observations) then the annual glacier wide mass balance can be estimated using the longer term mean value of the local measurement and the deviations from this longer term mean in the individual years. The accuracy of the method strongly depends on the representativity of the chosen local measurements and the validity of the assumption of a linear dependence.

Glacier Module 3 (August, 26-27, 2014)
Glacier motion: deformation and sliding

Learning Objectives

1. You know about the principle of mass conservation (steady state geometry), which explains the general flow field of an idealised glacier.

2. You are able to explain the physical processes of glacier flow and its two components: deformation and sliding.

Terms and Concepts

Shape of a glacier in a steady state (profiles of surface and bed), dependences between climate / mass balance / geometry, flow lines, extending and compressing flow, submerging and emerging flow, crevasses, bergschrund, deformation of ice (Glen’s flow law, basal shear stress, shape factor), water pressure, effective pressure and basal sliding.

References and Further Reading

Literature

Hooke (2005): Glacier flow, deformation of ice, sliding (chapters 4, 5, 9, 10)

Cuffey and Paterson (2010): Glacier flow, deformation of ice, sliding (chapters 3, 7, 8, 10)

Benn and Evans (2010): Glacier flow, deformation of ice, sliding (chapters 4, 5)

Van der Veen (1999): Glacier dynamics

Willis (1995): Review on glacier motion

Additional Information

Consideration of continuity

Assuming a steady state of a glacier (mass balance = 0, geometry and density = constant) for every single glacier cross section the following equation is valid:

$$Q = \bar{u}F = \int_{H_0} bs\,dH = -\int_{H_u} bs\,dH$$

Q = flow rate and $\bar{u}$ = mean flow velocity above the cross section area F at the cross profile, H_0 and H_u = elevation ranges above and below the cross section, b = budget, s = area ratio of the glacier.

At the equilibrium line (Index *e*) the flow rate is

$$Q_e = \overline{u}_e F_e = |B_c| = |B_a|$$

and related to the mass turnover of an entire glacier. For a given glacier surface, maritime glaciers with high mass balance gradients flow fast (high mass turnover), glaciers in a dry continental climate, flow slowly (low mass turnover with small mass balance gradients). For a given climate and mass balance gradient large glaciers flow faster than small ones.

In the accumulation area Q and therefore $\overline{u}$ are increasing towards the equilibrium line, reach their maximum at the equilibrium line and decrease in the ablation area towards the glacier front. In flow direction mean flow velocity normally increases towards the equilibrium line (extending flow) and decreases in the ablation area (compressing flow). Corresponding to these two longitudinal deformations there is lateral contraction in the accumulation area (converging flow) and a lateral stretch in the ablation area (diverging flow). A cross section in the accumulation area shows a concave but in the ablation area a convex surface, as the ice moves in direction of the steepest slope. Therefore and because of the systematic varying ice thickness, the displacement vectors in the ice is more strongly inclined than at the surface (exception: equilibrium line!).The vertical component of the glacier motion compensates the accumulation by submerging flow and the ablation by emerging flow. Things deposited at the highest altitude of a glacier are translated in the deepest layers to the glacier snout, where the ice is oldest. The ice at the equilibrium line is the youngest. Ice in mountain glaciers is about 0–10^3 years old, in large ice sheets 0-10^6 years.

Deformation und sliding

The flow velocity at the glacier surface (u_s) includes a component of sliding at the glacier bed (u_b) and a component of ice deformation (u_d):

$$u_s = u_b + u_d$$

u = (s = surface-, b = basal-, d = deformation-) velocity

Basal sliding u_b is negligible for cold glaciers. Within temperate glaciers the velocity ratio (u_b/u_s) varies, but can be set to about 0.5 in a large-scale / long-term mean. If the proportion of sliding is extremely high the glacier is unstable.

Deformation of ice

In the laboratory creep experiments with ice probes can be performed with a constant stress σ or a constant strain rate $\dot{\varepsilon}$. The experiment with constant stress reveals that the strain rate depending on the applied stress increases, after an elastic reaction, in the first phase (primary creep) with decreasing velocity, in the second phase (secondary creep or steady state creep) at constant velocity and at the end (tertiary creep) with progressive growing velocity until the probe fails. Glacier flow in nature corresponds to the secondary creep (constant strain rate $\dot{\varepsilon}$ at constant stress). Plotting strain rate $\dot{\varepsilon}$ from the secondary creep as a function of stress σ gives a power law; this is the Glen’s flow law:

$$\dot{\varepsilon} = A\sigma^n$$

$\dot{\varepsilon}$ = strain rate, σ = stress, A and n = flow law parameter

The flow law parameter A and n (≈ 3) are determined from drill hole and tunnel experiments, where A is depending on:

- temperature (warm = "soft", cold = "rigid")
- the unfrozen water content
- the size of ice crystals (small crystals = "soft")
- the concentration of impurities (influences water content, internal friction and crystal size).

Ice is almost incompressible, therefore the terms “extending flow” and “compressing flow” are only valid for the flow direction and have to be compensated by lateral compression “converging flow” or lateral extension “diverging flow”, respectively. If reaching a critical stress (ca. 0.5 bar) and strain rates (ca. 10^{-2} to 10^{-3} per year) crevasses can form.

The decisive stress of glacier flow is the shear stress at the glacier bed (τ_b):

$$\tau_b = f\rho g h \sin\alpha$$

with f = shape factor for the friction at the valley side, ρ = density of ice, g = gravitational acceleration, α = surface slope of the glacier und h = ice thickness

This results, integrated over the total glacier thickness in:

$$u_d = u_s - u_b = \frac{2A}{n+1}(\rho g \sin\alpha)^n h^{n+1}$$

u = (s = surface-, b = basal-, d = deformation-) velocity, A and n = flow law parameter, h = ice thickness

If the ice thickness can be measured (radar, seismic soundings) the shear stress τ_b can be determined (α has to be averaged over a reference distance of 5-10 times the ice thickness). For larger glaciers these values are around 1 to 1.5 bar (100 - 150 kPa). If perfect plasticity is assumed and τ_b is constant, the product $h\sin\alpha$ is also constant: a glacier is thin where the surface is steep and thick where the surface slope is small. Small and continental glaciers with a low mass turnover flow with small shear stress; whereas maritime and large glaciers with a high mass turnover have large values of basal shear stress. In steep parts (crevasses and seracs) the flow velocity and therefore also the shear stress is higher than in flat areas.

Basal sliding

Sliding subsummates all movement processes taking place at the glacier bed, i.e. at the interface ice/bedrock. The complexity of these activities and the limited access to the glacier bed make it difficult to understand all details. The knowledge about the basal boundary conditions for glacier flow is still sparse and waits to be solved. The sliding component of the glacier flow can be measured or estimated by:

- borehole inclometry (problem: deformation of the borehole)
- borehole camera (problem: turbid water, deformation of the borehole)
- sliding at the glacier margins (problem: representativeness)
- short-term velocity movement variations

Essential for friction and sliding at the glacier bed is the difference between ice and water pressure (P_i and P_w = measured with a piezometer in boreholes and moulins).

For numeric calculations a generic sliding law is used:

$$u_b = C\frac{\tau_b^m}{(P_i - P_w)}$$

P_i = ice pressure, P_w = ware pressure, C = parameter of roughness, m and p = constant

Instead of studying glacier lying directly on bed rock, studies were set up to determine the hydraulic-mechanical characteristics of glaciers settled on soft deformable moraines and to consider the influence of friction from debris at the glacier bed.

Glacier Module 4 (August, 26-27, 2014)
Glacier temperatures and thermal regimes

Learning Objectives

1. You can explain the thermal regime of a glacier and their characteristics.

2. You know about the different zones of firn facies.

3. You are able to explain the characteristics, relevance and distribution of temperatures in glacier ice and firn and you can discuss the relevant processes and possible impacts of an increase in temperature.

Terms and Concepts

Zones of firn facies, MAAT, superimposed ice, thermal regime of glaciers (cold, polythermal, temperate), seasonal temperature fluctuations (effects of snow and melting, percolation and re-freezing of melt water), temperature profile at the surface in cold firn, palaeo temperatures in deep temperature profiles, temperature profiles in polythermal glaciers, morphology of ice (air bubbles, crystals), phenomena of melt water

References and Further Reading

Literature

Haeberli, W. (1975): Temperatures of firn and ice in the Alps

Suter, S. (2001): cold firn, measuremants and modelling

Shumskii, P.A. (1964): zones of firn facies

Müller, F. (1962): zones of firn facies

Cuffey and Paterson (2010): temperatures of ice, zones of firn facies

Hooke (2005): temperatures of ice

Additional Information

Ice temperatures and zones of firn facies

Temperate ice consists of ice, air and water in equilibrium of phases at the pressure melting point. A plus or minus flux of energy leads to a change of the relative rate between water and ice, but does not change the temperature. In cold ice almost no fluid water can be found and an energy flux causes a significant change of temperature. Concerning the thermal regime, glaciers can be divided into the following classes: cold glaciers (cold ice throughout), polythermal glaciers (containing cold and temperate ice) and temperate glaciers (temperate ice throughout). The pressure melting point in a glacier is lowered about -0.7°C per 1000 m ice overburden.

With increasing mean annual air temperature (MAAT) the characteristics of firn and ice changes, leading to zones of typical firn facies:

In the **dry-snow zone / recrystallization zone** no melting occurs, even in summer. The mean firn temperature corresponds to the MAAT and is below -15 to -20°C. The accumulation is marginal (cm to dm water equivalent per year) and the densification without the influence of water is very slow. The transition from firn to ice is located at a depth of 50-100 m. Such areas can be found in the Antarctica, in the centre of Greenland and at very cold and high mountain summits (e.g. Mt. Mc. Kinley).

In the **infiltration-recrystallization zone** melting occurs, but the melt water refreezes within the uppermost annual layer. Due to the transport of latent heat the mean firn temperature is higher than the MAAT and lies between -10 and -20°C. The accumulation and densification rates are higher than in the recrystallization zone. The transition from firn to ice is located at a depth of 30-50 m. This type of firn facies is common on glaciers close to the poles and on higher mountain summits (e.g. Mont Blanc, Monte Rosa). The melt water has the potential to mix up the signature of a layer (chemical structures, particles). But as the melt water barely percolates more than one annual layer, firn and ice from this facies zone is still suitable to serve as a climate archive.

In the **cold infiltration zone (percolation zone A)** the melt water percolates through several annual layers, but the firn temperature does not reach 0°C. The mean firn temperature is below 0°C and above about -10°C. This zone can be found on sub-arctic accumulation areas and higher mountain summits (e.g. summits above 4000 m in the Alps).

In the **warm infiltration zone (percolation zone B)** the melt water can leave the temperate firn pack (mass loss). Accumulation and densification rate reach the highest values. The transition from firn to ice is located in a depth of 20-30 m. Temperate firn is typical for low-altitude maritime mountain ranges with a MAAT > -10°C, e.g. glacier areas at 3000-4000 m altitude in the Alps.

Because the freezing of one gram of water releases the same amount of heat that is required to raise the temperature of 160 grams of snow by one degree, refreezing of melt water is the most important factor in warming the snow. As summer advances, successively deeper layers of snow are raised to the melting point. The amount of melt water produced during a summer normally increases with decrease of elevation.

The lower boundary of the firn region is the snow line. In dry, cold regions between snow line and equilibrium line (where accumulation and ablation are equal, balance b = 0) there is often an **infiltration-congelation zone** located. In this zone superimposed ice is formed by freezing of large amounts of melt water in the firn to continuous mass. Below the equilibrium line the ablation area of a glacier is located.

These considerations pertain close to the glacier surface and are based on Shumskii (1964) and Müller (1962). Cuffey and Paterson (2010) use the term wet snow zone for the cold/warm percolation zone.

The temperatures at greater depths of glaciers are influenced by the glacier flow. The vertical flow component in the accumulation area transports firn and ice from the surface to depths and in the ablation region material from the depth to the surface. Hence in cold and polythermal glaciers complex temperature distributions can result due to the transport of cold ice from the accumulation area to the lower elevations. Thermal palaeo-effects (i.e. the temperature increase within the 20th century) are conserved in the accumulation areas (submerging flow, accumulation) but eliminated in the ablation areas (emerging flow, ablation). The temperature of glacier ice is easy to measure and a parameter which integrates all related processes. Temperatures of ice are measured with thermistors in boreholes and temperatures of firn in pits.

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