Summer school in Glaciology
Fairbanks/McCarthy 7-17 June 2010

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1. GLACIER METEOROLOGY ENERGY BALANCE

Ice and snow melt at 0°C, but this does not necessarily mean that melting will occur with an air temperature ≥ 0°C. Glacier melting is determined by the energy balance at the glacier surface, where air temperature is only one factor among many. The energy balance is the balance between all positive and negative energy flows to the surface and is controlled by:

- meteorological conditions
- physical properties of the glacier surface.

The interactions between the atmosphere and glacier surface are complex due to complicated **feedback mechanisms**. The atmosphere affects the energy balance, while the atmosphere is affected in turn by the glacier due to the specific properties of snow and ice.

1.1 Properties of snow and ice

Snow and ice are distinguished from many other natural ground surfaces by the following properties:

- **the temperature at the surface cannot exceed 0°C**

  This affects the local climate in different ways:

  1. When the air temperature exceeds 0°C, the air temperature is colder close to the surface than it would otherwise be if the glacier were not present. The glacier cools the near-ground air.
  2. Large temperature gradients in the first few meters above the surface can occur during the summer, which leads to **stable atmospheric stratification**, i.e. strong temperature inversion close to the surface (Fig. 1). Thus, turbulence (and energy transfer) is reduced.
  3. Where stable conditions and an inclining glacier surface are present, **catabatic winds** (glacier winds) can occur. The main direction for catabatic winds is down the glacier. They are controlled by the extent of the near-surface temperature gradient and the steepness of the glacier.

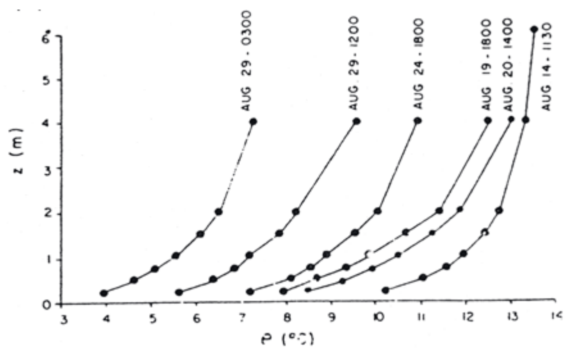

Fig. 1. Air temperature profile over a glacier surface.

- ➢ **low thermal conductivity of snow**
  Fresh snow at 0ºC = 0.08 W m^{-1} K^{-1}, old snow = 0.42 W m^{-1} K^{-1}, ice = 2.1 W m^{-1} K^{-1}
  - ➔ Snow is a good **insulator**; the thickness of the snow pack affects the cooling of ice beneath the snow.
- ➢ **transmission of short wave radiation**
  Approximately 1-2% of radiation passes through snow/ice (down to one meter in snow and a few meters in ice). The absorption decreases exponentially with depth (Fig. 2); the absorption can contribute to the heating of snow and can even cause sub-surface melting, which is the reason for the melting of a few centimetres beneath the surface of the blue ice zone in Antarctica.

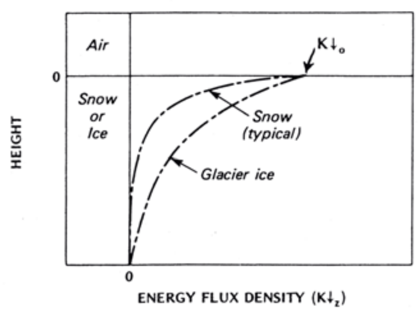

Fig. 2. Profile of solar radiation (K) in snow and ice (Oke 1987, p.85).

- ➢ **high and a strongly varying albedo**, i.e. a large fraction of incoming shortwave radiation is reflected.
- ➢ **high thermal emissivity**.

1.2 Heating of snow

For snow and ice to melt, their temperature must first increase to the melting point, i.e. energy is required to warm snow from subfreezing temperatures to zero. Then snow/ice can begin to melt:

To warm 1 kg of snow 1 K requires 2009 J kg^{-1} K^{-1}; ice: 2097 J kg^{-1} K^{-1}
= **Specific heat capacity**
To melt 1 kg snow/ice requires 334 000 J kg^{-1}
= **Latent heat of fusion** (L_f)
the same amount of energy is releases by re-freezing
To sublimate 1 kg snow/ice requires 2 848 000 J kg^{-1}
= **Latent heat of sublimation** (L_f)

Release of the latent heat of fusion due to re-freezing causes heating of the snow at the beginning of the melt season to occur very quickly. Melt water produced at the surface, or rain, percolates down through the snow pack and re-freezes when it contacts snow that still has a negative temperature. Re-freezing of 1 g water releases so much energy that 160 g snow heats by 1 K. A snow pack of some meters can, by this way, warm from a number of degrees below freezing to zero within the period of a few days or hours depending upon the quantity of water and the cold content of the snowpack. Without the release of heat by re-freezing of water the warming of the snowpack takes much longer. (Fig. 3).

The energy required to heat a cold snow pack to 0°C is called **cold content**, *C*, and is a function of the temperature profile:

The **cold content** *C* of a column of snow/firn of depth *Z* below the surface is given by

$$C = -\int_0^Z \rho(z) c_p T(z) dz$$

ρ is the density at depth z (kg m^{-3}), c_p is the specific heat capacity (*energy required to warm up 1 kg by 1 K: snow: 2009 J kg^{-1} K^{-1}; ice: 2097 J kg^{-1} K^{-1}*), and T is the temperature at depth z (ºC). The cold content is equal to 0 if the snow is at 0°C at all depths. The snow or ice is then referred to as **temperate**.

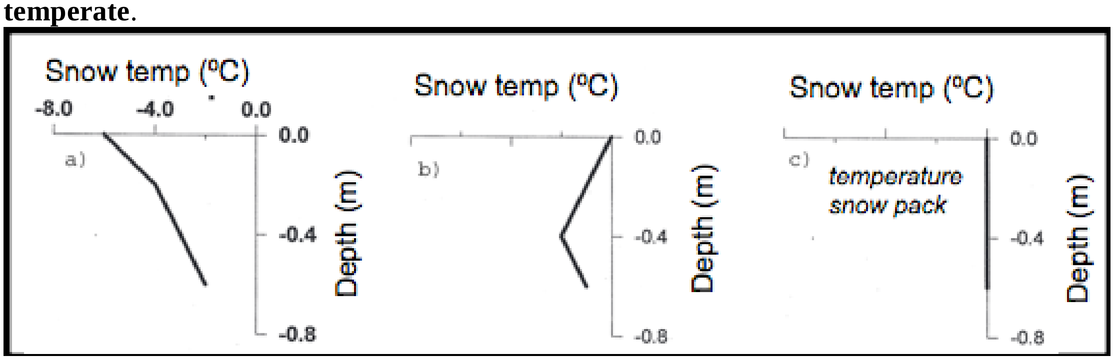

Fig. 3. Schematic illustration of the temperature distribution in snow: (a) The snow has cooled down from the surface, i.e. the snow temperature is coldest at the surface and the cold-wave is transported down by heat conduction. (b) A surplus of energy at the surface warms the surface snow to melting point. (c) Melt water percolates through snow and energy is released through re-freezing and the snow pack becomes temperate.

1.3 Energy balance

The energy balance is defined as the sum of all energy fluxes at the surface (Fig. 20):

$$Q_M = Q_N + Q_H + Q_L + Q_G + Q_R + \ldots$$

Q_M = energy available for melting
Q_N = net radiation
Q_H = sensible heat flux
Q_L = latent heat flux
Q_G = ground heat flux, change of the internal energy (temperature change)
Q_R = sensible heat flux of rain.

Melting, and all other fluxes that deliver energy to the surface, are defined as positive in glaciology (Note that the sign convention is different from meteorology). All fluxes are stated as Wm^{-2} (= $Js^{-1}m^{-2}$).
Sensible and latent heat fluxes are known as **turbulent heat fluxes**. They are determined by the temperature and humidity gradients between the air and surface, and by turbulence in the lower part of the atmosphere, i.e. by wind speed.

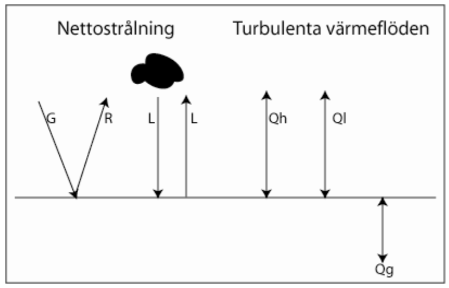

Fig. 4. Energy balance components

➢ Q_N, Net radiation

Net radiation is the balance between short wave and long wave radiation fluxes:

$$Q_N = G + R + L\downarrow + L\uparrow$$

G = **global radiation, shortwave incoming radiation**

- = incoming short wave radiation from the sun.
- wave length 0.15 - 4 μm.
- the component that reaches the earth's surface directly (**direct radiation**), is partly scattered in the atmosphere and reaches the surface as **diffuse radiation**; under cloud-free conditions the proportion of diffuse radiation is about 10-20%; under complete cloud cover, 100%.
- *G* is a function of the **solar constant** (average ~1368 Wm^{-2}, varies over the year as function of time), solar geometry (e.g. zenith angle), topographical effects (e.g. shadowing, inclination), cloud, and atmospheric composition.
- large spatial and time variations; is zero at night; high during the day.
- varies between 0 and >1000 Wm^{-2}.
- low global radiation in fog, high under clear-sky conditions.

R = **reflected global radiation**

- function of albedo (surface's reflectivity).
- cannot exceed global radiation (there cannot be more reflected radiation than is received).

L↓ = **incoming longwave radiation**

- long wave radiation is emitted by the atmosphere (primarily by water vapour, CO_2, and ozone).
- wave length 4 - 120 μm.
- function of air temperature and air humidity (cloudiness).
- higher air temperature, more cloud → higher L↓
- varies between about 250 and 350 Wm^{-2}

L↑ = **outgoing longwave radiation**

- long wave (heat-) radiation emitted by the earth's surface.

- function of the temperature of the ice/snow surface (Stefan-Boltzmans law: $L\uparrow = \varepsilon\sigma T^4 + (1-\varepsilon)L\downarrow$ ε = emissivity, σ = Stefan-Boltzmans constant).
- is constant 316 Wm^{-2} under melting conditions (surface temperature = 0ºC).
- cannot exceed 316 Wm^{-2} because ice/snow cannot be warmer than zero degrees.

$G + R$ = **shortwave radiation balance** = $G(1-\alpha)$ where α = albedo
- can not be negative, as $G \geq R$

$L\downarrow - L\uparrow$ = **longwave radiation balance**
- can be positive or negative

Net radiation can be positive or negative and varies typically between about -100 and 300 Wm^{-2}. Negative values during summer nights usually indicate clear skies and relatively low incoming long wave radiation (during nights, global radiation, and therefore also reflected global radiation, are either zero or negligible). During the night (when global radiation is zero) the net radiation is entirely determined by the longwave radiation balance.

Albedo

= the ratio of reflected and incoming short wave radiation: $\alpha = R/G$.

Snow, particularly fresh snow, has a very high albedo, i.e. a large part of the incoming global radiation is reflected, and the energy is not available for e.g. melting. Albedo is dependent upon the material properties of the surface (snow/ice, dirt content, water content (Table 1) and properties of incoming radiation, such as zenith angle and wave length, which are in turn affected by atmospheric conditions (e.g. cloud type and distribution).

Tab 1: Typical albedo values for different surfaces.

fresh snow	~ 0.7 - 0.9
firn	~ 0.5 - 0.6
ice	~ 0.3 - 0.4
grass	~ 0.1 - 0.3
forest	~ 0.1 - 0.2

- **Q_H, Sensible heat flux**

  = energy transfer of the air's heat.
  - Q_H is a function of:
    - **temperature gradient** (i.e. the temperature difference between the air temperature over the surface and the surface temperature).
    - **wind speed**
  - The higher the temperature gradient and the wind speed the larger the sensible heat flux.
  - Sensible heat flux is zero if the temperature gradient or wind speed is zero.

- **Q_L, Latent heat flux**
  - Energy released or consumed during a phase change, i.e. through:

    condensation (vapour to water, energy release)
    evaporation (water to vapour, energy consumption)
    sublimation (ice to vapour, energy consumption)
    resublimation (vapour to ice, energy release).

- Q_L is a function of:
  - **vapour pressure gradient** (i.e. the difference between the vapour pressure in the air and at the surface); the maximum possible vapour pressure is a function of the temperature. The vapour pressure of a melting glacier surface is a constant 6.11 Pa, because the temperature is a constant 0ºC.
  - **wind speed**
- If the vapour pressure gradient is positive, i.e. the vapour pressure in the air is higher than the surface vapour pressure, the **latent heat flux** is **positive**; either **condensation** or **resublimation** occurs; thereby energy is released and available for example for melting.
- If the gradient is negative, i.e. the vapour pressure in the air is lower than the surface vapour pressure, the **latent heat flux** is **negative**; either **evaporation** or **sublimation** occurs; it requires energy that is then not available for e.g. melting. As about 8 times more energy is required for sublimation than for melting, the amount of energy that is required for melting is reduced significantly when sublimation occurs.
- The latent heat flux is zero if the vapour pressure gradient or wind speed is zero.

➢ **Q_L, Ground heat flux**

- is a function of change in cold content with time, *t*:

$$Q_G = \int_0^z \rho\left(z'\right) c_p \frac{\partial T}{\partial t} dz'$$

- Q_G is zero when the glacier and snow cover are temperate.
- A polythermal glacier with a cold surface layer has negative Q_G even during summer when the surface has attained melting point. On such glaciers, a certain percentage of the excess energy contributes to heating the cold surface layer while the rest goes to melting.

➢ **Q_R, Rain heat flux**

= transfer of the rain's heat energy

- function of rain intensity, R, and rain temperature, T_r.

$$Q_R = \rho_w c_w R(T_r - T_s)$$

c_p=specific heat of air, 1005 J kg^{-1} K^{-1} c_w specific heat of water, 4180 J kg^{-1} K^{-1}

- The significance of rain is often overestimated. In most cases rain is a very minor source of energy for melting, and contributes, on average, usually not more than a few percent of the energy for melting. A rainfall of 10 mm at 10ºC on a melting surface transfers, for example, energy equivalent to a melting of only 0.6 mm/day. Conversely, a number of centimetres or even 10s of centimetres can melt during a rain-free summer day. However, rain may have indirect effects that can accelerate melting through effects on other components of the energy balance. For example, the albedo of snow decreases when it becomes wet, resulting in increased absorption of short wave radiation and thus more melt.

Energy for melting

If the sum of net radiation and sensible and latent heat fluxes is positive on a temperate glacier, then the energy goes towards:

1. **heating** of snow/ice, when the snow/ice temperature at the surface is negative (the temperature must be 0^{o}C before snow melting can occur at the surface). The ground heat flux is negative.
2. **melting** when the snow/ice temperature at the surface is zero.

If the sum of net radiation and sensible and latent heat fluxes is negative, then the glacier cools. Melting is zero and the energy balance is closed by cooling of the glacier (positive ground heat flux).

The energy available for melt Q_M is converted to water equivalent melt, M, (in m w.e.) using the latent heat of fusion (L_f = 334 000 J kg^{-1}) and the density of water ρ_w:

$$M = \frac{Q_M}{\rho_w L_f}$$

⇨ Q_M = 100 Wm^{-2}, which is equivalent to M = 2.6 cm/day
⇨ Q_M = 384 Wm^{-2}, which is equivalent to M = 10 cm/day

Relative importance of energy balance components

Net radiation is usually dominant. For example, in the European Alps, net radiation usually dominates, contributing up to > 90% (Table 2). The turbulent heat fluxes are of increased importance in maritime regions (e.g. > 50%). On maritime glaciers more than half of the energy may be come from the turbulent fluxes.

Table 2. The relative importance of the energy balance components on different glaciers. Q_N is net radiation, $Q_H$sensible heat, Q_L latent heat, Q_G ground heat flux and Q_M is the energy available for melting given in % of total energy gain and energy loss. Values are average values of periods of some weeks or months over one or a number of melt seasons. The negative values for Q_L indicate sublimation, while positive values indicate condensation.

Glacier	Q_N	Q_H	Q_L	Q_G	Q_M
Aletschgletscher, Switzerland	92	8	-6	0	-94
Hintereisferner, Austria	90	10	-2	0	-98
Peytoglacier, Canada	44	48	8	0	-100
Storglaciären, Sweden	66	30	5	-3	-97

1.4 Interpretation of climate data

Climate data collected on a glacier allow some conclusions on conditions on the glacier, e.g. when it is melting, if the surface consists of snow or ice, etc. Some examples, referring to Fig. 5, are described below.

Wind direction:

The data show that westerly winds dominate. As Storglaciären is oriented in a west-east direction, this is downstream on the glacier, thus in the catabatic wind direction.

Albedo:

Between May and the beginning of July, values vary between 0.7 and 0.8, i.e. 70-80% of the incoming short wave radiation is reflected. These high values indicate that the surface around the climate station is snow covered. In mid July, the albedo decreases considerably over several days and reaches a value around 0.3, which indicates that the snow is melting and the ice surface is being exposed. The increased albedo at the end of July indicates the occurrence of a summer snowfall. Low air temperatures around 0ºC and reduced global radiation values that indicate cloudiness support this conclusion.

Global radiation:

The global radiation typically displays cyclical diurnal variations with high values in the middle of the day and values close to zero at night. Also observe the large variations that occur from day to day. High peak values and regular diurnal cycles indicate cloud-free conditions (e.g. 5 July). Much lower values, such as on the 16th and 17th of July, indicate foggy days.

Incoming longwave radiation:

The radiation varies less than global radiation and is also active at night. Low values, e.g. in mid May, are caused by low air temperatures and little cloud cover. The latter is also indicated by high global radiation. The highest longwave radiation is attained in mid July in conjunction with high air temperatures, air humidity, and precipitation. Low global radiation values also indicate cloudy and foggy conditions. Under cloudy/foggy conditions, part of the reduction of energy from short wave solar radiation is compensated by an increase in long wave radiation.

Outgoing longwave radiation:

The radiation attains a maximum value of 316 Wm^{-2}, at which the surface is melting, i.e. the surface temperature has reached melting point. As 0ºC is the maximum temperature that snow and ice can attain, no further radiation can be emitted. At the beginning of the measurement series, the radiation is below the maximum value, which means that surface melting is not occurring. However, diurnal variations, with lower values at night and higher values during day time, show that the surface temperature varied, but stayed under 0ºC. During August, mostly at night, outgoing longwave radiation sporadically drops below the maximum value, indicating cooling of the glacier surface, and in this case, no surface melting occurred.

Net radiation:

Net radiation exhibits large diurnal variations, which are controlled mostly by the cyclical variations in global radiation. Net radiation is most positive during the summer but at the beginning and end of the melting season negative values occur, as the short wave radiation balance is lower due to high albedo and incoming long wave radiation is lower due to colder air temperatures.

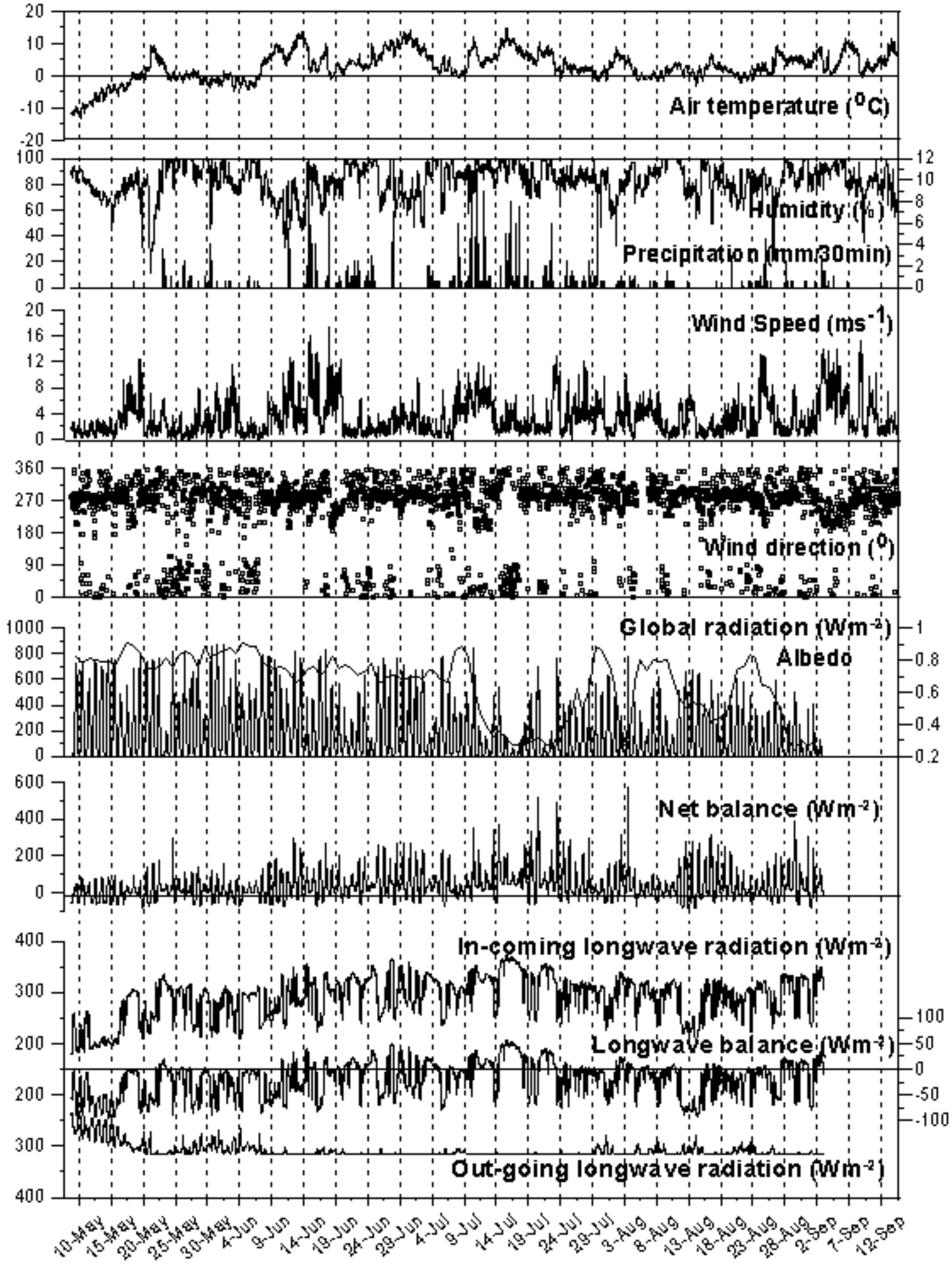

Fig. 5. Hourly values for meteorological variables and energy balance components on Storglaciären.

1.5 Modeling melt

Energy balance methods

Melting can be modelled through calculating the energy balance, i.e. the energy that is available for melting. The energy balance of a temperate glacier can be calculated with the help of the following meteorological data at one level (e.g. 2 m) on the glacier:

- net radiation
- air temperature (-> used to compute the sensible heat flux)
- air humidity (-> used to compute the latent heat flux)
- wind speed (-> used to compute the sensible and latent heat fluxes).

The energy balance components (if not measured directly) can be approximated with the help of appropriate parameterisations, i.e. as functions of variables that control the component and that can be measured simply. This usually includes a simplification of the physical processes.

Instead of net radiation, the different radiation components can be measured separately. In practise, parameterisations are most often used because radiation measurements are expensive and difficult to obtain under the difficult conditions that typically prevail on a glacier. So incoming longwave radiation can, for example, be parameterised with the help of air temperature, humidity, and cloud data. A large number of the parameterisations for the different components of the energy balance have been developed.

Temperature index methods

Energy balance methods are physically based but require input data that are often not available. Therefore, in practise, the temperature index method is usually employed. The method is based upon an empirical relation between melting and air temperature. Despite the fact that it is the energy balance that determines how much can melt, it has been shown that over longer periods (>days/weeks) melting correlates well with air temperature. This is due to the fact that most energy balance components directly or indirectly depend upon air temperature. For example, the turbulent flux, incoming longwave radiation, and rain heat flux are directly a function of air temperature, while the air temperature is in turn indirectly affected by solar radiation. The simplest method is what is known as the **degree-day method**, where melt, M, is proportional to the positive temperature. Daily melt is calculated multiplying a degree-day factor by the positive temperatures. The degree-day factor is the coefficient of proportionality $f = -a/\varphi$ between *surface ablation a* (which is negative) and the positive degree-day sum φ over any period.

The degree-day factor parameterizes all of the details of the energy balance that results in ablation by melting and possibly sublimation, and is therefore a simplification. It is usually treated as one or more constants; in particular, it is different for snow and for glacier ice, because the ice is generally less reflective than snow. It is usually expressed in mm w.e. K^{-1} d^{-1} or kg m^{-2} K^{-1} d^{-1}. Melting is assumed to be zero when the air temperature is ≥ 0ºC. Different variations of the traditional degree-day method have been developed. Many hydrological models (e.g. the HBV-model) employ this method.

1.6 References

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