**Glacier meteorology**

**Surface energy balance**

- *How does ice and snow melt?*
- *Where does the energy come from?*
- *How to model melt?*

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**Melting of snow and ice**

- *Ice and snow melt at 0°C (but not necessarily at air temperature \( \geq 0°C \))*
- *Depends on energy balance which is controlled by meteorological conditions → properties of the surface*

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**Properties of snow and ice**

- **Surface temperature can not exceed 0°C**
  - stable stratification in melt season
  - glacier wind (katabatic wind)
  - max surface vapor pressure 611 Pa
  - max longwave outgoing radiation = 316 Wm\(^2\)
- **Transmission of short wave radiation**
  - down to 10 m for ice and 1 m for snow
  - important for subsurface melting
- **High and variable albedo**

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**Warming of snow/ice**

- *First, snow/ice needs to be warmed to melting point*
- *Second, melt can occur*

1 g water refreezes \( \rightarrow \) 160 grams of snow will be warmed by 1 K

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**Cold content**

\[
C = -\int_0^z \rho(z)c_p T(z)dz
\]

**Ground heat flux**

\[
Q_G = \int_0^z \rho c_p \frac{\partial T}{\partial t}dz
\]
SURFACE ENERGY BALANCE

\[ Q_{atm} = Q_M + Q_G \]

Energy flux atmosphere to glacier surface

Energy available for melting

Change in internal energy (heating / cooling of the ice/snow)

Energy balance

\[ Q_M = Q_N + Q_H + Q_L + Q_R + Q_G + \ldots \]

- \( Q_N \): Net radiation
- \( Q_s \): Sensitive heat flux
- \( Q_L \): Latent heat flux
- \( Q_G \): Ground heat flux (heat flux in the ice/snow)
- \( Q_R \): Sensitive heat supplied by rain

ATMOSPHERE

- \( G \): Global radiation
- \( \alpha \): Albedo
- \( L \): Longwave incoming radiation
- \( L' \): Longwave outgoing radiation

GLACIER

- \( M \): Melt rate

\[ M = \frac{Q_M}{\rho L_f} \]

As temperature increases, the energy emitted increases, but the wavelength at which the peak radiations is emitted decreases.
GLOBAL RADIATION or shortwave incoming radiation

Top of atmosphere radiation

- **Solar constant** = amount of incoming solar electromagnetic radiation per unit area, measured on the outer surface of Earth's atmosphere, in a plane perpendicular to the rays.
- Roughly 1367 W/m², fluctuates by about 7% during a year (1412 W/m² to 1321 W/m²) → varying distance from the sun.

- Wavelength: 0.15-5 µm
- 2 components:
  - Direct / diffuse component:
  - Scattering by air molecules scattering and absorption by liquid and solid particles
  - Selective absorption by water vapor and ozone

**Direct solar radiation**

\[ I_c = I_0 \left( \frac{R_m}{R} \right)^2 \psi_a \frac{P}{P_0} \cos \theta \left[ \cos \beta \cos Z + \sin \beta \sin Z \cos(\varphi_{\text{sun}} - \varphi_{\text{slope}}) \right] \]

- \( I_c \): solar constant = 1368 W/m²
- \( R_m \): Earth-Sun radius (m = mean)
- \( \psi_a \): atmospheric transmissivity
- \( P \): atmospheric pressure (0 at sea level)
- \( \beta \): slope angle, \( \varphi_{\text{sun}} \), \( \varphi_{\text{slope}} \): solar and slope azimuth angle

Strong spatial variation

- **Controlling factors:**
  - site characteristics: slope, aspect, solar geometry
  - atmosphere: transmissivity (cloudiness, pollutants ...)
- \( I \) increases with:
  - decreasing angle of incidence
  - increasing transmissivity (affected by volcanoes, pollution, clouds)
  - increasing elevation (decreasing mass, less shading, multiple reflection)

Spatial variation of potential solar direct radiation

(clear-sky conditions)

\[ I_c = I_0 \left( \frac{R_m}{R} \right)^2 \psi_a \frac{P}{P_0} \cos \theta \left[ \cos \beta \cos Z + \sin \beta \sin Z \cos(\varphi_{\text{sun}} - \varphi_{\text{slope}}) \right] \]

Topographic shading Potential direct radiation

averaged over melt season

The large spatial variability of the direct component of global radiation in complex topography is responsible for much of the spatial variability in observed melt.
Modelled direct and diffuse radiation on Storglaciaren, Jun 7 – Sep 17, 1993

20-90 Wm²

- Clear-sky days = D approx. 15% of global radiation, Overcast days 100%
- D controlled by
  - atmospheric conditions (clouds)
  - spatially: albedo, skyview factor, V
- Less variable spatially than direct radiation

Extrapolation of global radiation

\[ G_{\text{gridcell}} = \frac{G_{\text{meas}}}{I_{\text{potential, measurementsite}}} \]

- Ratio is small under cloudy conditions
- Ratio is large under clear-sky conditions
- Ratio is proxy for cloudiness and assumed to be the same across glacier

ALBEDO

= average reflectivity over the spectrum 0.35-2.8 μm
= ratio between reflected and incoming solar radiation

Typical values:

- new snow 0.75 – 0.95
- old snow 0.4 – 0.7
- glacier ice 0.3 – 0.45
- soil, dark 0.1
- grass 0.2
- rain forest 0.15

= Large wavelength dependency
- Large variability in space and time
- Key variable in glacier melt modelling
- Ice albedo less variable than snow albedo, but often modified by sediment and debris cover
CONTROLS ON GLACIER ALBEDO

**Surface properties**
- **Grain size** (large grain size → decreases albedo)
- **Water content** (water increases grain size → decreases albedo)
- **Impurity content**
- **Surface roughness**
- **Crystal orientation/structure**

**Incident shortwave radiation**
- **Cloudiness** (increase since clouds preferentially absorb near-infrared radiation, higher fraction of visible light which has higher reflectivity); effect enhanced by multiple reflection, increase up to 15%
- **Zenith angle** (increase when sun is low, due to Mie scattering properties of the grains)

**How to model snow albedo?**
- **Radiative transfer models** including effects of grain size (most important control) and atmospheric controls
  - large data requirements, not practicable for operational purposes
- **Empirical relationships**
  - Aging curve approach: function of time after snowfall
    - \( \alpha = \alpha_s - \alpha_{\text{min}} + \alpha_{\text{add}} e^{-k n} \)
  - \( b, k = \) coefficients
  - \( \alpha_{\text{min}} = \) minimum snow albedo

**Snow albedo parameterisations**
1. U.S. Army Corps of Engineers (1956)
   \( \alpha_s = a_{\text{min}} + a_{\text{add}} e^{-k n} \)
   - function snow age \( n \) and temperature (through \( k \))
2. Brock et al. (2000)
   \( \alpha_s = a_{\text{1}} - a_{\text{2}} \log_{10} T_a \)
   - Snow albedo is computed as a function of accumulated maximum daily temperature since snowfall

*Figures and equations are placeholders for illustrative purposes.*

*Images and text excerpts are illustrative and do not reflect the actual content.*
Parallel-leveled instrument
Horizontally leveled

Longwave incoming radiation
- Wavelengths 4-120 µm
- Emitted by atmosphere (water vapor, CO₂, ozone)
- Function of air temperature and humidity (cloudiness)
- High values compared to shortwave radiation
- Longwave rad. balance < 0, when fog → ca 0 W/m²
- Spatial variation: Topographic effects:
  - reduced by obscured sky
  - enhanced by radiation from slopes and air inbetween
- Climate change: Temp increase or more cloudiness → more L↓

\[ L↓ = ε_{\text{eff}} \sigma T^4 \]
The longwave incoming radiation is the largest contribution to melt (~ 70%) \[ L_{\downarrow} = \varepsilon_{\text{eff}} \sigma T^4 \]

About 70% of the longwave incoming radiation originates from within the first 100 m of the atmosphere.

Variations of screen-level temperatures can be regarded as representative of this boundary layer.

Longwave outgoing radiation
\[ L_{\uparrow} = \varepsilon \sigma T^4 + (1 - \varepsilon) L_{\downarrow} \]

- Longwave reflectance of snow: < 0.05
- Emissivity \( \varepsilon > 0.95 \)
- Often assumed \( \varepsilon = 1 \)

\( L_{\uparrow} = 315.6 \text{ W m}^{-2} \)
- \( \sigma = 5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4} \)
- Assume black body radiation \( \Rightarrow \varepsilon = 1 \)
Radiation messages
Net radiation is negative during clear-sky nights:
low atmospheric emissivity → low incoming longwave radiation → cooling of the air
Net radiation is higher during cloudy nights (high atmospheric emissivity → high incoming longwave radiation → warming of the air
Longwave incoming radiation (ca. 300 W/m²) much larger than shortwave incoming radiation (150-250 W/m²), but less variation

Turbulent heat fluxes
Driven by temperature and moisture gradients between air and surface and by turbulence as mechanism of vertical air exchange

Turbulent heat fluxes =
Ability to transfer x gradient of relevant property

Eddy diffusivity

Turbulent heat fluxes
Driven by temperature and moisture gradients between air and surface and by turbulence as mechanism of vertical air exchange

Sensible heat flux
- Function of temperature gradient
- Function of wind speed
Latent heat flux (the energy released or absorbed during a change of state)
- Function of vapour pressure gradient
- Function of wind speed
Fluxes also affected by
- Surface roughness
- Atmospheric stability
Turbulent heat fluxes

\[ Q_H = \rho c_p C_H u (\theta_z - \theta_s) \]

\[ Q_E = \rho L_v C_E u (q_z - q_s) \]

- Bulk aerodynamic method

Exchange coefficient is function of surface roughness and stability function (empirical expressions to define stability functions)

Fluxes affected by
- Surface roughness
- Atmospheric stability

Sublimation

\[ Q_s = L_s \frac{\partial P}{\partial z} \left[ \ln \frac{ z }{ z_0 } - \psi_M(z/L) \right] \left[ \ln \frac{ z }{ z_{z0} } - \psi_E(z/L) \right] \]

- To melt 1 kg snow/ice requires 334 000 J kg\(^{-1}\)
- Latent heat of fusion
- To sublimate 1 kg of snow requires 2 848 000 J kg\(^{-1}\)
- Latent heat of sublimation (8x \(L_f\) !!!)

Latent heat flux

\[ Q_M = Q_N + Q_H + Q_L + Q_C + Q_R + ... \]

Positive vapor gradient

\[ \rightarrow \text{Condensation} \]

To melt 1 kg snow/ice requires 334 000 J kg\(^{-1}\) = Latent heat of fusion

Negative vapor gradient

\[ \rightarrow \text{Sublimation} \]

To sublimate 1 kg snow/ice requires 2 848 000 J kg\(^{-1}\) = Latent heat of sublimation
Typical for during dry periods in the outer tropics
Sublimation occurs
\[ L_s = 8^\circ L_a \] 8x less ablation than under wet conditions for same energy input

**Sensible heat flux by rain**

\[ Q_R = \rho c_p R (T_r - T_s) \]

- Negligible compared to 30-180 W/m² net radiation averaged over longer periods
- But rainfall has indirect effects: changes albedo, mechanical removal of snow...

**What data needed for computing an energy balance?**

1) **Temp**
   - sensible heat flux
   - latent heat flux
   - rain heat flux
   - (longwave incoming radiation)
2) **Humidity**
   - latent heat flux
   - (longwave incoming radiation)
3) **Wind speed**
   - sensible heat flux
   - latent heat flux
4) **Global radiation**

**Temporal variation of energy balance components**

Westfork Glacier, Alaska Range
Energy partitioning (%)

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Net rad</th>
<th>Sensible</th>
<th>Latent</th>
<th>Ground heat</th>
<th>Melt</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aletschgletscher, Switzerland</td>
<td>92</td>
<td>8</td>
<td>-6</td>
<td>0</td>
<td>-94</td>
</tr>
<tr>
<td>Hintereisferner, Austria</td>
<td>90</td>
<td>10</td>
<td>-2</td>
<td>0</td>
<td>-98</td>
</tr>
<tr>
<td>Peytoglacier, Canada</td>
<td>44</td>
<td>48</td>
<td>8</td>
<td>0</td>
<td>-100</td>
</tr>
<tr>
<td>Storglaciären, Sweden</td>
<td>66</td>
<td>30</td>
<td>5</td>
<td>-3</td>
<td>-97</td>
</tr>
</tbody>
</table>

**Summary Energy balance**

- **Radiation** can be measured directly; longwave often not measured, needs parameterization f(T, e, n); extrapolation of global and longwave radiation relatively straightforward, but modeling albedo is major uncertainty;
- **Turbulent fluxes** generally need parameterization f(T, e, u) at both point scale and larger scale; problems determining the exchange coefficients and stability functions, issues with validity of underlying theory; large uncertainty (but often only smaller contributor to total melt energy);
- Often **net radiation** dominant source of energy, longwave radiation largest source (often the double the amount of shortwave incoming);
- **Variations from day to day** often determined by variations in turbulent heat fluxes
- Spatial variations in energy balance determined by shortwave incoming radiation
- **Energy by rain** generally very small
- Direction of vapor pressure gradient important → sublimation reduces energy available for melt significantly
- **Ice heat flux**: cooling by conduction, but warming of snow mainly by refreezing of melt/rain water; flux important for cold/polythermal glaciers

**Melt energy – weather patterns**

\[ Q_M = Q_N + Q_H + Q_L + \ldots \]

day-to-day variability in melt is often determined by the turbulent fluxes

**How to model melt?**

1. **Physically based energy-balance models**: each of the relevant energy fluxes at the glacier surface is computed from physically based calculations using direct measurements of the necessary meteorological variables
2. **Temperature-index or degree-day models**: melt is calculated from an empirical formula as a function of air temperature alone
Melt modelling

Data requirements
Model sophistication

Temperature index
Energy balance

Input data:
Air temp
Air temp humidity
wind speed
radiation

Mass balance models

Model type
Spatial discretization

0-Dim
Fully distributed
Elevation bands

Temp-index regression
Temp-index or simplified energy balance
Energy balance

Increasing model sophistication

Temperature-index melt models

• Assume a relationship between air temperature and melt: \( M = f(T) \), \( M = f(T^+) \)

Relationship melt - air temperature

Positive degree-day sum
\( PDD = \sum T^+ \)

Relationship melt – degree-day sum

Data by R. Braithwaite
Physical basis of temp-index models

*Air temperature directly affects several components of the surface energy balance*

\[ L \downarrow = \varepsilon \sigma T^4 \]

\[ f(T) \]

\[ f(e(T)) \]

\[ Q_M = G(1 - \alpha) + L \downarrow - L \uparrow + Q_H + Q_L \]

- \( L \downarrow \) is the largest contribution to melt (~ 70%) 
  (Ohmura, 2001, Physical basis of temperature-index models)

Case studies (Sicart et al., 2008, JGR):
Correlation between energy fluxes and air tempe
- \( L \) has low variability compared to other fluxes
- \( L \) is poorly correlated to air temperature when cloud variations dominate its variability (usual in mountains)

Degree-day model

\[ M = f_{\text{ice/snow}} \times T^+ \]

Typical values

<table>
<thead>
<tr>
<th>Snow</th>
<th>Ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>~3-5 mm/d/K</td>
<td>~5-10 mm/d/K</td>
</tr>
</tbody>
</table>

Degree-day

The name of a derived unit, the K d, equal in magnitude to a 1 K departure of temperature, above or below a reference temperature, sustained for a period of 1 day.

Different choices of the reference temperature -> e.g. heating degree-day, freezing degree-day).

In glaciology: positive degree-day (relative to the reference temperature 0 °C).

Degree-day factor

In a positive degree-day model, the coefficient of proportionality between ablation \( a \) and the positive degree-day sum.

Degree-day factors

<table>
<thead>
<tr>
<th>Glacier</th>
<th>[mm/day/K]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aletsch Glacier, Switzerland</td>
<td>5.3</td>
</tr>
<tr>
<td>John Evans Glacier, Canada</td>
<td>5.5</td>
</tr>
<tr>
<td></td>
<td>4.1</td>
</tr>
<tr>
<td></td>
<td>2.7</td>
</tr>
<tr>
<td>Alfotbreen, Norway</td>
<td>4.5</td>
</tr>
<tr>
<td>Storglaciaren, Sweden</td>
<td>3.2</td>
</tr>
<tr>
<td>Dokriani Glacier, Himalaya</td>
<td>5.7</td>
</tr>
<tr>
<td>Yala Glacier, Himalaya</td>
<td>11.6</td>
</tr>
<tr>
<td>Glacier AX010</td>
<td>11.6</td>
</tr>
<tr>
<td>Thule Ramp, Greenland</td>
<td>12</td>
</tr>
<tr>
<td>Camp IV-EGIG, Greenland</td>
<td>18.6</td>
</tr>
<tr>
<td>GIMES profile, Greenland</td>
<td>8.7</td>
</tr>
<tr>
<td>Qamanarsup sermia</td>
<td>2.8</td>
</tr>
</tbody>
</table>

Spatial and diurnal variation
Derived from energy balance modeling

Hock, 1999, J.Glacial
SPATIAL VARIABILITY OF DEGREE-DAY FACTORS

Calculation of degree-day factors for various points on the Greenland ice sheet with an atmospheric and snow model (thesis Filip Lefebvre)

Performance of degree-day model

Melt = DDF_{ice/snow} * T^*
Modified temperature-index model

Including potential direct solar radiation

- Classical degree-day factor
  \[ M = DDF_{\text{ice/snow}} \times T^+ \]

- Including pot. direct radiation
  \[ M = (M_f + a_{\text{ice/snow}} \times I) \times T^+ \]

Model introduces
- a spatial variation in melt factors
- a diurnal variation in melt factors

Simulated cumulative melt
Summer 1994

Model comparison

Gornergletscher outburst floods

Huss et al., 2007, J. Glaciol.

Photo: Shin Sugiyama
Extended temperature-index models including other data than temperature

Degree-day model

\[ M = DDF_{\text{ice/snow}} \times T^+ \]

Extended temperature-index model including radiation

\[ M = a \times R + \text{melt factor} \times T \]

Energy balance model

\[ Q_M = G(1-\alpha) + L \downarrow - L \uparrow + Q_H + Q_L \]

→ Gradual transition from degree-day models to energy balance-type expressions by increasing the number of climate input variables

Simplified energy balance model

\[ M = a \times R + \text{melt factor} \times T \]

→ Gradual transition from degree-day models to energy balance-type expressions by increasing the number of climate input variables

Distributed temp-index model by Pellicciotti et al, 2005, J. Glaciol.

\[ M = TF \cdot T + AF \cdot (1-\alpha) \cdot I \]

Temperature

Albedo

Incoming shortwave radiation

Melt is function of
- Global radiation, and thus topography (in particular shading)
- Surface properties and particularly albedo
- Temperature extrapolation

Model only requires air temperature

Global radiation and albedo parameterized

Pellicciotti et al, 2005, J. Glaciology
**Temperature-index versus energy balance**

<table>
<thead>
<tr>
<th>Temperature index</th>
<th>Energy balance</th>
</tr>
</thead>
<tbody>
<tr>
<td>• Wide availability of Temp-data</td>
<td>• Physical based – describe physical processes more adequately</td>
</tr>
<tr>
<td>• Easy interpolation and forecasting</td>
<td>• Projections more reliable</td>
</tr>
<tr>
<td>• Good model performance</td>
<td>• Easy interpolation and forecasting</td>
</tr>
<tr>
<td>• Computational simplicity</td>
<td>• Wide availability of Temp-data</td>
</tr>
<tr>
<td>• Empirical, not physically based</td>
<td>• Physical based – describe physical processes more adequately</td>
</tr>
<tr>
<td>• DDF vary, works on ‘average conditions’</td>
<td>• Projections more reliable</td>
</tr>
<tr>
<td>• Does not work in tropics</td>
<td>• Easy interpolation and forecasting</td>
</tr>
<tr>
<td>• Model parameter stability under different climate conditions?</td>
<td>• Wide availability of Temp-data</td>
</tr>
<tr>
<td></td>
<td>• Computational simplicity</td>
</tr>
</tbody>
</table>

*Both approaches are needed, use depends on application and data availability*

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**SUMMARY**

- Both temp-index and energy balance models are useful tools, choice depends on data availability
- Awareness of limitations
- Need for more approaches of intermediate complexity and moderate data input
- Both temp-index and energy balance models need calibration (parameter tuning)

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**Energy balance on Storglaciären**

**Energy balance:**


**Temperature-index methods:**


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**Literature**

- Energy balance:
  
- Temperature-index methods: