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The erosive effects of different temperature regimes on glaciers and ice sheets

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The temperature distribution within a glacier or an ice sheet governs its flow dynamics and its hydrology. The ice temperature also determines the shape and the extension of the glacier, to what extent it will affect its bed and the groundwater situation beneath the ice. In this paper I will summarize some common knowledge on the relationship between glaciers and temperature regime. And based on reasonable assumptions, I will argue for a realistic model for a past (and future) ice sheet covering Scandinavia.

In a maritime environment, such as the western part of Norway and the south coast of Iceland, the precipitation and ablation rates are high giving high rates of mass turnover in the glaciers. The mass surplus in the accumulation area is heated by refreezing meltwater every summer and thus temperate ice is formed. In the ablation area, the ice surface may be cooled by low winter temperatures but this frozen top layer melts off each year in the summer. This type of glacier thus remains *temperate.*

A temperate glacier is believed to be permeable to melt water. Water drains in a dendritic englacial drainage system, with small veins close to the ice surface and successively larger conduits with depth as a consequence of an increased waterflow and thus increased melt rate (Shreve 1972, Nye and Frank 1973). Estimations of the content of liquid water in a temperate glacier often ends at about 1%. The direction and inclination of the englacial water flow is governed by the ice surface slope. At the base of the glacier the water flow is directed by the force of gravity plus the weight of the overlying ice. This simple model for water flow fits well with observations but it implies a system in equilibrium, with a hydrostatic pressure at the bottom equal to the overburden pressure. In practice it means the influence of seasonal changes in melt water flow is negligible. Expressed differently we may say that the melt season with its high variability in meltwater flow is not of sufficient duration to influence the englacial drainage pattern significantly.

The temperate glacier thus slides over its substratum and erodes it. Diurnal and seasonal variations in subglacial water pressure influences the sliding speed significantly.

Glacial erosion is difficult to measure. The most common estimations are based on silt load measurements in proglacial streams. Assuming a uniform erosion rate beneath the glacier and a steady state, silt loads may be transformed into erosion rates but with large degrees of freedom. The erosion rate for Storglaciären in northern Sweden, is estimated to be 1 mm per year (Schneider and Bronge 1996, Holmlund *et al.* 1996). On softer bedrock, such as the volcanic rocks of Iceland, erosion rates of several millimetres per year have been calculated.

In a drier environment, precipitation and ablation rates are lower giving lower rates of mass turnover. It is common for the winter cooling depth to exceed the depth of summer melting causing a permanent sub frozen top layer in the ablation area of the glacier (Holmlund and Eriksson 1989, Björnsson *et al.* 1996). Depending on winter temperature and summer melt rates, this cold layer can be more or less thick on such *poly thermal glaciers.* In a dry polar climate the entire ice body may be at a temperature well below the freezing point, though melting occurs at the ice surface during the summer. These glaciers are referred to as *sub polar.* However, as melting occurs during the summer, it is common that temperate ice is produced at the higher reaches of the glacier while the entire ablation area is below the freezing point and frozen to its bed (Björnsson *et al.* 1996). A glacier which is entirely frozen does not slide over its substratum and there is no liquid water at the interface between bedrock and ice. As ice below the freezing point is impermeable there is no englacial drainage except for surface drainage channels which can, if they carry large quantities of water, melt down into the ice and form single conduits within the ice. Poly thermal glaciers are common in, for example, Northern Sweden and Spitsbergen where annual temperatures are a few degrees below freezing point. In colder environments the glaciers are sub polar, entirely frozen to their beds.

In the coldest environments, such as on top of the large ice sheets of Greenland and Antarctica, no melting occurs and thus the ice formed remains at a temperature equal to the annual mean temperature in the area. This is the *high polar* environment from which deep drilling projects have yielded comprehensive environmental and climatic. This is also the environment which best corresponds to the former ice sheet of Scandinavia, at least during its maximum phase.

The temperature regime within such an ice sheet is a function of the air temperature at the surface, the vertical ice movement (accumulation), advection of ice, heat conduction, internal friction and the geothermal gradient. If there were no ice movement we could simply extrapolate the geothermal gradient through the ice, but this situation is unrealistic for larger regions. It is unrealistic because the only place where there is no horizontal movement is at the ice divide, where the mass balance must be positive by necessity (to maintain the shape of the glacier), thus producing a vertical movement.

The temperature regime governs not only the erosion capacity of a glacier but also the viscosity of the ice. Ice at the pressure melting point deforms ten times more quickly than ice at -20°C. Physical factors which have a positive influence on basal temperatures are: low precipitation, thick ice, large geothermal gradient, fast ice movement, and, by the margins, high ablation rates. Frictional heating becomes increasingly important with depth as most of deformation occurs close to the bed.

The flow of an ice sheet is forced by gravity, and the potential energy released by the migration of ice masses from one altitude to a lower one is transformed to kinetic energy and heat by internal friction. This process is gradually more important with decreasing distance to the front. However, below the equilibrium line the emergence velocity (the upward movement which compensates for ablation at the surface) will favour geothermal heat conduction upwards in the ice, and thus warm the basal ice to the pressure melting point.

If we examine an ice sheet such as the Greenland ice sheet, ice is formed at the ice surface at temperatures 20-30 degrees below the freezing point. Where it is thickest, the ice may reach the pressure melting point at the bottom. Outside the deepest parts, the ice will freeze on to the bottom again because the ice thins and thus becomes cooler at its base. Along a flowline we may expect frozen conditions all the way to a marginal near position where internal heating may have raised the temperature to the pressure melting point. A special case is when ice drainage is concentrated into ice streams. The high rate of deformation in ice streams may warm the base far into the interior of the ice sheet, along the flow line of the ice stream. Finally at the margins there may exist a frozen rim, the depth of which is determined by the climate at the front.

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