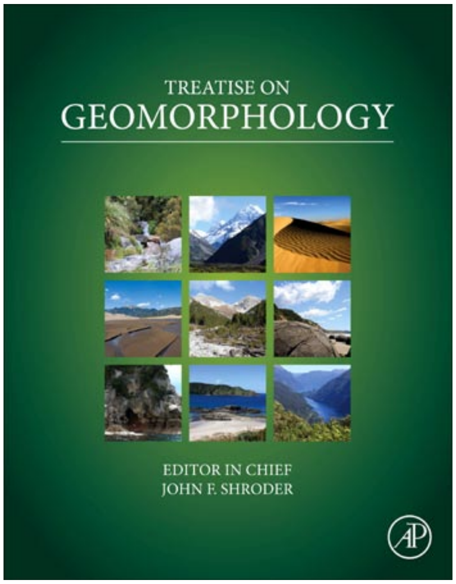

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8.4 Classification of Ice Masses

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Glossary

Ablation Any process that results in the loss of mass from a body of snow or ice. This typically occurs from surface melting in mid-latitude regions, but can also occur due to factors such as wind scouring, sublimation, and iceberg calving.

Accumulation Any process that results in the addition of mass to a body of ice of snow or ice. This most commonly occurs from snowfall, but may also occur due to other processes such as avalanching from surrounding slopes, wind drifting, or rainfall directly onto an ice or snow mass.

Cold ice Ice that is at a temperature below the pressure melting point, and which therefore contains little free liquid water.

Firn Any snow that has survived one summer melt season, but is not yet glacier ice.

Ice age A relatively cold period, when glaciers covered large portions of the Earth's surface. The most recent ice age is referred to as the Pleistocene and peaked approximately 20 000 years ago.

Temperate ice (also referred to as warm ice) Ice that is at the pressure melting point, and which can therefore contain substantial quantities of liquid water.

Abstract

Ice masses currently cover approximately 10% of the land surface of the Earth, and take on a wide variety of shapes, sizes, and internal characteristics. To aid in their description and in understanding the physical processes that control them, ice masses are typically classified using two different schemes: a morphological classification and a thermal classification. The morphological classification is primarily based on the relationship between ice masses and surrounding topography, as well as their size and form. The thermal classification is based on the temperature distribution within an ice mass, which can be either temperate (at the pressure melting point, ~ 0 °C), cold (below the pressure melting point), or polythermal (a mixture of both). In reality, many ice masses fit into different classifications along their length and exhibit a complex set of natural forms.

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8.4.1 Introduction

In the natural world, ice masses can take on a variety of forms that reflect their surrounding topography and climate, as well as their internal physical characteristics such as ice temperature, thickness, and subglacial water flow. Permanent ice masses currently exist in all of the major mountain ranges in the world, as well as over extensive lowland regions close to the poles, covering a total ~10% of the Earth's land surface (IPCC, 2007). This equates to a total of ~16 000 000 km^2, with approximately 97% of this area currently occupied by the Greenland and Antarctic Ice Sheets (Benn and Evans, 2010).

To aid in the description and understanding of permanent ice masses, two main classification schemes have been developed. The first of these is based on the surface morphology of an ice mass; in other words, the physical shape and geometry of an ice mass and how it relates to the surrounding topography. The second main classification scheme is based on the internal thermal characteristics of an ice mass, which is controlled by factors such as the surrounding climate, geothermal heat production, and insulation properties of the overlying ice cover. The purpose of this section is to review these classification schemes, with a particular focus on providing specific examples of ice masses that fall into different categories. The review here covers permanent terrestrial ice masses, including those attached to land. For a comprehensive multilingual definition of ice types occurring in the ocean, atmosphere, on land, and in relation to snow, the reader is referred to Armstrong et al. (1973). A good photo glossary of glacial features and landscapes is also available at the Swiss 'Glaciers Online' website.

8.4.2 Morphological Classification

The surface form of ice masses provides the classification scheme that has been traditionally used in glaciology, including by reference books ranging from the oldest (e.g., Agassiz, 1840) to more recent ones (e.g., Sugden and John, 1976; Sharp, 1988; Benn and Evans, 2010). Terrestrial ice masses fall into two primary categories within the morphological classification scheme: those unconstrained by surrounding topography and those constrained by it.

8.4.2.1 Unconstrained by Topography

8.4.2.1.1 Ice sheet

Ice sheets are the largest class of ice mass, covering entire continents and subsuming almost all underlying topography (particularly in their central regions). This includes entire mountain ranges such as the Gamburtsev Mountains in East Antarctica, which are buried by >3-km-thick ice (Bell et al., 2011). Ice masses need to cover an area of >50 000 km^2 to be classified as an ice sheet (Armstrong et al., 1973), and at the present day there are only two ice sheets on the Earth: the Greenland and Antarctic Ice Sheets. Antarctica is the thickest of these, with a maximum ice depth of ~4.8 km (Lythe et al., 2001), compared to a maximum depth of ~3.3 km for the Greenland Ice Sheet (Bamber et al., 2001). At the peak of the last ice age, approximately 20 000 years ago, ice sheets covered almost 30% of the land surface of the planet. In the Northern Hemisphere, this included the Laurentide Ice Sheet over most of eastern North America, the Cordilleran Ice Sheet over most of western North America, and the Eurasian Ice Sheet over most of northern Europe.

8.4.2.1.2 Ice cap

Ice caps are somewhat similar to ice sheets in that they also cover almost all underlying topography, but are smaller in size (<50 000 km^2). As ice caps still require a substantial ice thickness to cover topography, they most commonly occur in high-latitude locations such as Ellesmere Island (e.g., Agassiz Ice Cap), Devon Island (e.g., Devon Ice Cap; **Figure 1**), Svalbard (e.g., Austfonna), and Iceland (e.g., Vatnajökull Ice Cap). These ice caps typically reach maximum ice thicknesses of ~500–1000 m; for example, Dowdeswell et al. (2004) found an ice thickness of 880 m on Devon Ice Cap, and Dowdeswell et al. (2008) found a maximum thickness of 565 m on Austfonna.

A particularly unique feature of both ice caps and ice sheets is that they typically display a convex surface form, with highest elevations and thickest ice commonly near the ice cap center, and progressive thinning toward the margins. These central high-elevation regions have been classified as 'ice domes' (Benn and Evans, 2010). They typically display a radial flow pattern, with ice movement emanating in almost directions from the ice dome, and flow rates increasing toward the ice cap or ice sheet margins (**Figure 1**). The low-velocity areas that form the boundary between ice flowing into different drainage basins are termed 'ice divides'.

8.4.2.1.3 Ice stream

Ice streams are defined as rapidly moving areas of an ice sheet that are bounded by slowly moving ice; they are therefore unconstrained by surrounding surface topography. They most commonly occur in areas away from ice domes due to their requirement for rapid ice movement, with examples including the ~700-km-long Northeast Greenland Ice Stream (Joughin et al., 2001), and the Siple Coast ice streams that flow into the Ross Ice Shelf, Antarctica (Bamber et al., 2000). Velocities on ice streams can reach 2 km yr^{-1} or more, which is typically at least an order of magnitude greater than the slow moving ice that surrounds them. This rapid change of motion typically creates a narrow shear margin, which is characterized by intense surface crevassing. Field, modeling, and remote-sensing evidence suggest that ice streams are typically warm bedded (i.e., contain liquid water) and underlain by deforming sediment, whereas the slowly moving ice around them is cold bedded and typically underlain by hard bedrock (Blankenship et al., 2001).

There have been a number of debates as to whether the location of ice streams is controlled by subglacial topography, but the evidence for this is ambiguous. For example, Joughin et al. (2001) found that some areas of the Northeast Greenland Ice Stream have a subglacial topographic expression, but in other locations there is little evidence of any kind of subglacial channel. Similarly, there appears to be little subglacial topographic control across most of the Siple Coast ice streams. There is also evidence that adjacent ice streams have sped up

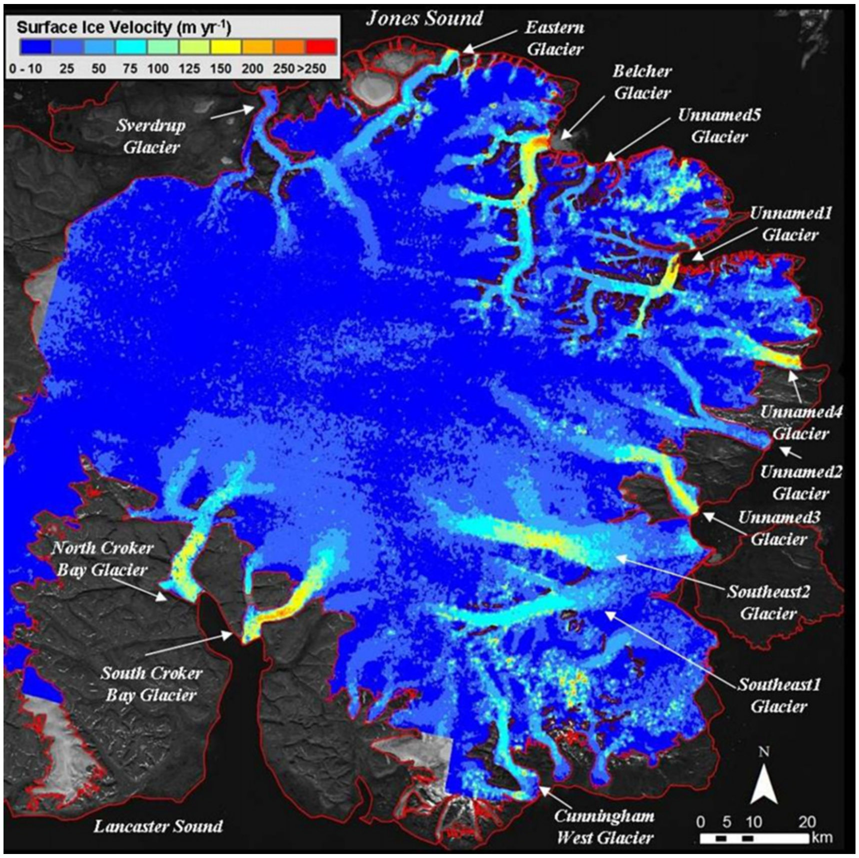

Figure 1 Radial flow pattern displayed by Devon Ice Cap in the Canadian High Arctic. Surface elevation and ice thicknesses are greatest near the slow-moving central ice dome, with velocities highest along topographically constrained outlet glaciers. Reproduced with permission from VanWychen, W., 2010. Spatial and temporal variations in Ice Motion, Belcher Glacier, Devon Island, Nunavut, Canada. MSc thesis, Department of Geography, University of Ottawa. http://www.ruor.uottawa.ca/en/

and slowed down in the past in both the Siple Coast (Anandakrishnan and Alley, 1997) and Weddell Sea (Vaughan et al., 2003) sectors of Antarctica, most likely driven by changes in subglacial water availability. For example, the Kamb Ice Stream (formerly Ice Stream 'C') appears to have stagnated in the mid-nineteenth century (Anandakrishnan et al., 2001), whereas the Whillans Ice Stream (formerly Ice Stream 'B') appears to have started rapid flow ~450 years ago after a 400-year period of stagnation (Hulbe and Fahnestock, 2004).

8.4.2.1.4 Outlet glacier

Outlet glaciers are defined by Armstrong et al. (1973) as "a valley glacier which drains an inland ice sheet or ice cap and flows through a gap in peripheral mountains," and are somewhat similar to ice streams in that they drain ice from a central ice dome. However, they are bounded by topographic valleys or troughs and therefore form a transitional ice mass between one that is unconstrained by topography versus one that is. On the large ice sheets, many outlet glaciers (e.g., Lambert Glacier, Antarctica; Kangerdlugssuaq Glacier, Greenland) originate as ice streams in their upper reaches. In combination, outlet glaciers and ice streams account for the bulk of the discharge from ice sheets (e.g., up to 90% for Antarctica; Bamber et al., 2000), with outlet glaciers providing the primary drainage for ice caps. As such, much modern research is focused on understanding the dynamics and variability of outlet glaciers and ice streams because they appear to provide a major control on broad-scale ice sheet/ice cap stability and mass balance (e.g., Rignot and Kanagaratnam, 2006; Pritchard et al., 2009).

8.4.2.1.5 Ice shelf

Ice shelves comprise permanent floating ice masses that are attached to the coast (**Figure 2**), and typically form the seaward extension of ice streams and outlet glaciers. This is particularly true in Antarctica where they comprise ~30% of the coastline (Sugden and John, 1976), with the majority of the discharge from the ice sheet passing through them. Antarctic ice shelves are some of the largest glaciological features on Earth, reaching >1000 km in length and >450 000 km^2 in area (e.g., Ross Ice Shelf). Ice shelves are also present along coastlines in the northernmost reaches of the Arctic, but are typically orders of magnitude smaller than those in the Antarctic. For example, some small ice shelves appear to exist around the margins of Severnaya Zemlya and Franz Josef Land in the Russian High Arctic (Dowdeswell, in press), whereas the largest concentration of Arctic ice shelves occurs along

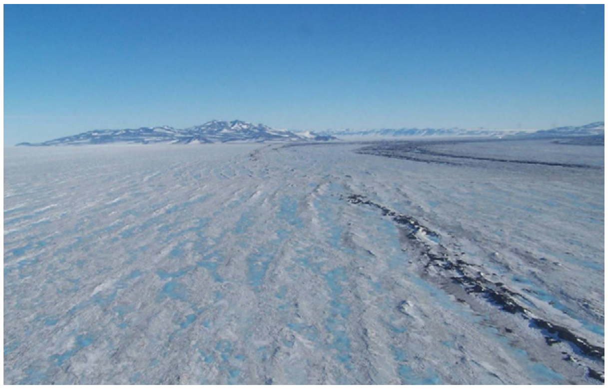

Figure 2 View across the McMurdo Ice Shelf, Antarctica, November 2004. Note the almost flat surface topography; dark material comprises surface moraines that indicate ice motion toward the camera.

Figure 3 View across the St. Elias Icefield, Yukon, July 2010.

Figure 4 View up the temperate Grosser Aletschgletscher, Switzerland, August 2010. This ice mass forms the longest valley glacier in the European Alps at a total length of ~23 km. Note the hanging glaciers in the upper right corner of the image.

northern Ellesmere Island, with a total area of 720 km^2 at the end of summer 2008 (Mueller et al., 2008). In some locations, ice shelves can build to a great enough thickness that they ground on underlying topography and create a dome-like structure that rises above the surrounding ice shelf; this feature is termed an 'ice rise'.

In Antarctica, ice shelves are typically fed from upstream outlet glaciers and ice streams, although significant sources of mass input may also come from snow accumulation on their surface (particularly on very large ice shelves such as the Ross; Crary et al., 1962), as well as the basal freeze-on of ocean water (Fricker et al., 2001). In the Arctic, some ice shelves on northern Ellesmere Island receive input from surrounding glaciers (e.g., Milne Ice Shelf), whereas others have built up from the *in situ* thickening of sea ice and surface snow accumulation with no known glacial input (e.g., Ward Hunt Ice Shelf) (Jeffries, 1992). The majority of mass loss from ice shelves occurs from the calving of ice at their termini; these large, freely floating ice masses are termed 'icebergs' in the Antarctic and 'ice islands' in the Arctic, and are typically defined as having a height of >5 m above sea level (Armstrong et al., 1973). Icebergs also refer to the production of any floating ice from glaciers, whether in freshwater or the ocean, and whether in the Arctic, Antarctic, or elsewhere.

8.4.2.2 Constrained by Topography

This category forms the second main group of ice masses classified by morphology and comprises the bulk of ice masses found in mountainous regions around the world.

8.4.2.2.1 Icefield

An icefield (also spelled ice field) forms the largest category of ice mass constrained by topography, and icefields are somewhat similar to ice caps in that they can reach substantial thicknesses (up to ~1000 m) and typically drain in a radial pattern from a high-elevation region. They do not share the same dome-like surface as ice caps, however, as they have generally low relief surfaces dissected by areas of bedrock that reach above their surface as 'nunataks'. Icefields typically form across broad high-elevation regions in areas such as Yukon/Alaska (e.g., St. Elias Icefield; **Figure 3**), southern South America (e.g., Northern and Southern Patagonia Icefield), and Ellesmere Island (e.g., Prince of Wales Icefield).

8.4.2.2.2 Valley glacier

A valley glacier consists of a moving ice mass which is constrained by surrounding topography and overlooked by rock cliffs (Sugden and John, 1976; **Figure 4**). Note that motion is required for an ice mass to be defined as a glacier, which typically means that glaciers have a minimum thickness of ~30 m. Thinner stationary ice will simply form a snow patch or ice patch. Typical motion on valley glaciers varies from approximately 10 to 1000 m yr^{-1} (although this can vary widely on both short and long timescales), with highest values where the ice is relatively thick and steep. Surface slope provides one of the main drivers for ice motion, so to maintain movement valley glaciers typically decrease in surface elevation along their length. They commonly receive their source from icefields or cirques, and are usually most clearly developed in the ablation area of ice masses, where they can form substantial U-, or parabolic-shaped valleys (**Figure 4**). In places where glaciers flow over bedrock obstacles, they can become extremely steep and form ice falls, which are typically heavily crevassed on their surface.

The smallest glaciers are termed 'ice aprons' and consist of thin snow and ice accumulations that adhere to the side of a mountain and commonly form the source of avalanches (Armstrong et al., 1973). Ice aprons typically occur at high altitudes near the tops of mountains, and contrast with glacierets that occupy depressions on shallower terrain and commonly receive much of their accumulation from snow avalanching and wind drifting (Benn and Evans, 2010).

8.4.2.2.3 Cirque glacier

Cirque glaciers form in bowl-like depressions near the tops of mountains called cirques, which are typically characterized by a flat floor and steep sides (**Figure 5**). They can either be entirely contained within the cirque depression or provide mass to downstream features such as ice falls and valley glaciers, and are very common in areas of glaciated mountainous terrain. They have a preferred northerly and northeasterly orientation in the Northern Hemisphere (and vice versa in the Southern Hemisphere) due to shading from the sun.

8.4.2.2.4 Piedmont glacier

Piedmont glaciers, also called 'piedmont lobes', form where constrained valley glaciers and outlet glaciers reach open lowland areas and are able to spread out laterally, whereby they take on a characteristic tear-drop or lobate shape due to the flow of the ice (**Figure 6**). They typically occur at lower altitudes (i.e., in ablation areas) and generally form glacier termini. They are common in high-latitude regions such as Ellesmere Island, but are relatively uncommon in steep mountainous areas such as the Andes, Alps, and Rockies. The Malaspina Glacier in SE Alaska forms the largest piedmont glacier in the world with a width of ~65 km and area of ~3900 km^2.

8.4.2.2.5 Hanging glacier

Hanging glaciers occur on steep terrain in mountainous areas, and generally form from ice that spills out of cirques or clings to rock cliffs (**Figure 4**). They are typically small in area and provide a frequent source of avalanches to lower altitude regions below. In some cases, hanging glaciers can physically separate a glacier into an upper accumulation area and a lower ablation area fed almost entirely by ice avalanching from above.

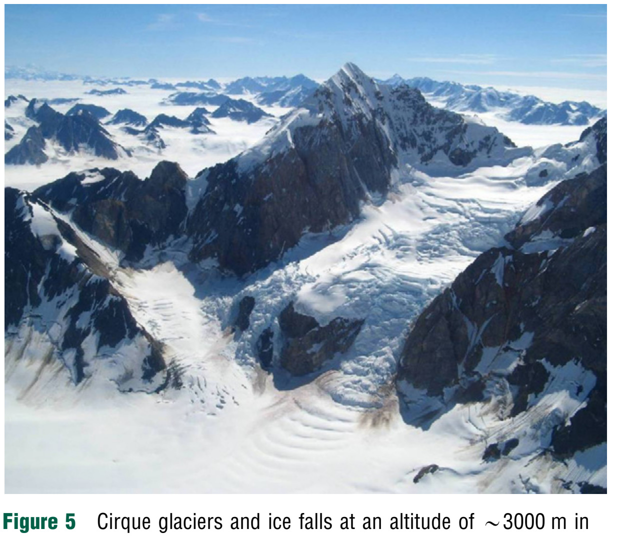

Figure 5 Cirque glaciers and ice falls at an altitude of ~3000 m in the St. Elias Mountains, Yukon.

8.4.3 Thermal Classification

In addition to classifying ice masses based on their morphology, they can also be classified based on their thermal regime. The internal and basal temperature of ice masses can be determined directly with the use of boreholes (e.g., Blatter, 1987), but borehole drilling is generally a time-consuming and expensive process. Many indirect measures are therefore used to determine internal ice temperatures, including modeling (e.g., Blatter and Haeberli, 1984), ground-penetrating radar (e.g., Björnsson et al., 1996), and spatial and temporal variability in motion at the glacier surface (e.g., Copland et al., 2003).

The concept of pressure melting point is important when considering the thermal regime of glaciers, as the melting point of water decreases with increasing pressure. The depression in melting point occurs at a rate of 0.072 °C per million pascals (Benn and Evans, 2010), which is equivalent to ~110 m of ice. This means that the pressure melting point is ~ −2.5 °C at depths of ~4000 m in the deepest parts of the Antarctic Ice Sheet.

Knowledge of the thermal regime of ice masses is important as temperature strongly determines their dynamic behavior. For example, the deformation rate of ice reduces by a

Figure 6 An example of polythermal piedmont glaciers issuing from small ice caps and outlet glaciers near the east coast of Ellesmere Island, Canada. http://airphotos.nrcan.gc.ca/contact_e.php. Copyright Her Majesty the Queen in Right of Canada.

factor of 5 between −10 and −25 °C (Paterson, 1994). Ice masses that have a frozen bed undergo slow motion due to internal ice deformation alone. By contrast, ice masses at the pressure melting point can have free water at their beds and typically experience higher surface velocities due to the addition of basal sediment deformation (where the ice lies on a soft bed) and basal sliding. Ice masses with beds at the pressure melting point typically exhibit high spatial and temporal variability in their surface motion in relation to variability in water inputs and associated subglacial water pressure (Willis, 1995). This means that variability in meltwater production may lead to changes in glacier thermal regime and associated ice motion (Zwally et al., 2002). It is clear that ice temperature is also important in controlling other glaciological processes such as glacier surging (Clarke, 1976), outbreak floods (Skidmore and Sharp, 1999), and glacial erosion (Dyke, 1993).

In general, ice masses are grouped into one of three different thermal classes: temperate, cold, or polythermal. Ice masses from the smallest glacierets to the largest ice sheets can exhibit any one (or more) of these temperature regimes.

8.4.3.1 Temperate ('Warm')

In this category, ice is at the pressure melting point throughout, except for a surface layer up to ~15 m thick that is subject to seasonal temperature variations. Water is able to freely exist within and at the bed of these ice masses, which means that they can undergo basal motion and move relatively rapidly. Temperate glaciers most commonly occur in relatively low-latitude and low-altitude mountainous regions such as the Rockies, New Zealand, western Norway, and the European Alps (Figure 4). Beyond the obvious requirement for a warm climate with an extended period of summer temperatures to remove any winter cold penetration, temperate glaciers also occur in regions with high mass turnover. This is because high snowfall rates tend to insulate underlying ice from cold winter temperatures, and high-surface melt rates lead to latent heat release as meltwater percolates into snow and refreezes. In addition, high ablation rates lead to the rapid removal of any winter buildup of cold ice at low elevations. In the same climate regime, thicker glaciers also tend to be warmer at their base than thin glaciers as ice acts as an insulator that prevents cold surface temperatures from reaching the glacier bed.

8.4.3.2 Cold

In cold ice masses, the ice is below the pressure melting point throughout, meaning that water is not able to freely exist within them and at their base. This results in slow surface motion rates which are dominated by internal deformation. Cold glaciers are typically thin and exist in cold climates where there are low rates of mass turnover, such as the Dry Valleys of Antarctica. In the Dry Valleys, mean annual temperatures are around −17 °C, the glaciers are frozen to their beds, and sublimation accounts for up to 80% of their total annual ablation (Lewis et al., 1998).

8.4.3.3 Polythermal

Polythermal glaciers contain a mixture of both temperate and cold ice, and are found in high-latitude and high-altitude regions throughout the world. They can take on a variety of thermal structures, with free water only able to exist in their temperate parts (Blatter and Hutter, 1991). **Figure 7** sketches some of these thermal patterns, with types b and c occurring in relatively warmer climates characterized by sufficient melting at high altitudes to raise their accumulation areas to the pressure melting point due to latent heat release from the refreezing of percolating meltwater. Once warmed, this firn turns to ice and is then carried to lower elevations by ice motion. The near surface of the ablation zone in these glaciers tends to be cold due to accumulation rates that are insufficient to insulate the ice from winter cooling, and because water is not able to percolate the impermeable ice in these regions. Glacier types b and c are often referred to as 'predominantly

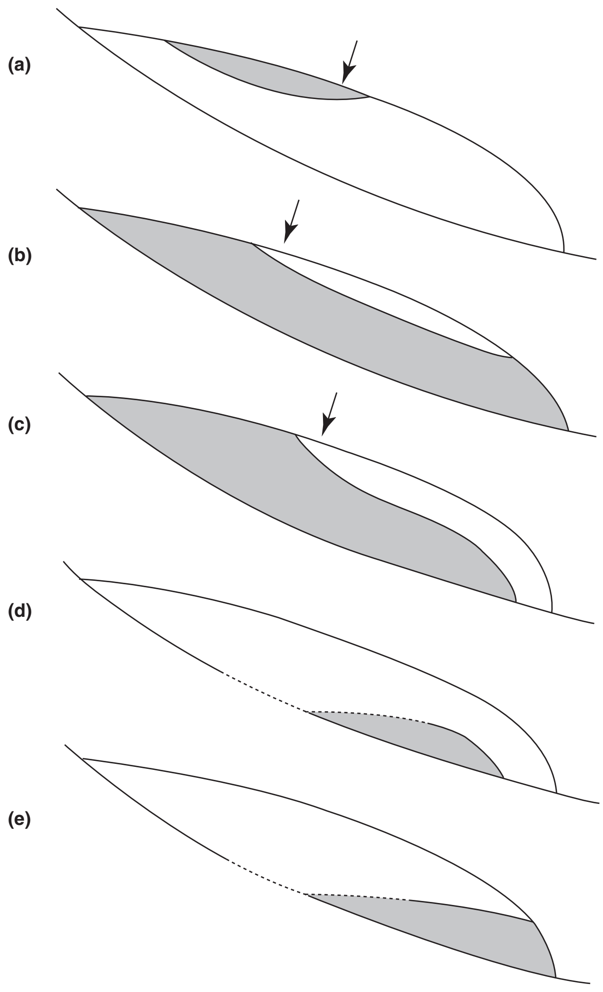

Figure 7 Various types of thermal structures within polythermal glaciers. Shading indicates temperate areas, white indicates cold areas, and dashed lines indicate melting conditions along cold-temperate transitions. Arrows indicate approximate position of the equilibrium line altitude. Adapted from Blatter, H., Hutter, K., 1991. Polythermal conditions in Arctic glaciers. Journal of Glaciology 37, 261–269, with permission from Journal of Glaciology.

warm polythermal', and occur most commonly in polar maritime climates such as Svalbard (Björnsson et al., 1996) and inland areas of Scandinavia (e.g., Storglaciären, Sweden; Holmlund and Eriksson, 1989).

Glacier types d and e in **Figure 7** typically occur in colder, drier climates than types b and c. In these situations, there is insufficient percolation and refreezing in the accumulation area to raise it to the melting point, so the cold ice from this region is carried to lower elevations. In the ablation area the surface remains cold, but the bed is able to warm from geothermal heating, strain heating, and the input of surface meltwater via features such as crevasses and moulins. These types of glaciers are typically referred to as 'predominantly cold polythermal', and are common throughout regions such as the Canadian High Arctic (Blatter and Hutter, 1991). For example, John Evans Glacier on Ellesmere Island alternates between type d in the winter and type e in the summer, and contains large volumes of subglacial meltwater across its terminus that can lead to rapid variations in horizontal and vertical motion (Copland et al., 2003). Glacier type a also falls into the category of predominantly cold polythermal, but would occur in warmer climates than types d and e, where there is sufficient melting to raise the snow to the pressure melting point in the lower parts of the accumulation area.

Recent studies of the Antarctic and Greenland ice sheets reveal that they have a more complex thermal structure than previously believed, with cold surface temperatures but large areas of the bed at the pressure melting point. They are thus best classified as being polythermal. Evidence for basal melting is provided by features such as the widespread occurrence of Antarctic subglacial lakes (Siegert et al., 2005), extensive basal freeze-on in East Antarctica (Bell et al., 2011), borehole drilling into wet sediments on Whillans Ice Stream (Engelhardt et al., 1990), and modeling in Greenland (Funk et al., 1994).

8.4.4 Conclusions

In the natural world, it should be remembered that ice masses exist in a wide variety of forms that do not always fall easily into the categories described above. This broad continuum means that transitions occur between all of the categories listed here, and that a single ice mass may fall into several different categories (particularly at different points along its length). For example, an ice particle traced from the center of Antarctica might pass through an ice sheet, ice stream, outlet glacier, and ice shelf before it reaches the ocean. Even within a relatively short mountain glacier, an ice particle might pass from a cirque to an icefall to a valley glacier and then to a piedmont glacier. Ice masses classified into one category by the morphological classification scheme can also be classified into different categories based on the thermal scheme (and vice versa). For example, valley glaciers may be classified as temperate, cold, or polythermal based on the climate regime that they exist in. Despite these complexities, the classification of ice masses is important for aiding in their descriptions, understanding how they form, and providing knowledge about the controls on their dynamics and changes.

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