# Properties of Glacial Ice and Glacier Classification

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## 1 Properties of glacial ice

Glaciers (and their various forms, including ice sheets and ice caps) are composed of ice that is moving under its own weight, which typically requires the ice to be ~30 m thick or more. Glacier ice is a crystalline solid and primarily originates from atmospherically-derived snow. At the highest elevations of a glacier, in the accumulation area, more snow falls annually than is lost by ablation, so over time the old snow gets buried and compacted by new snow above, causing it to increase in density. When snow has survived more than one melt season it is termed firn. Firn contains interconnected pores that allow for the movement of air and moisture, but as the firn continues to get buried and densifies, it finally reaches the pore close-off depth (Obbard et al., 2011). At this depth the pores become interconnected, which isolates air in individual bubbles, and defines the start of glacial ice. On glaciers with high annual snowfall and wet, dense snow packs (e.g., coastal Alaska), this metamorphism process occurs within just a few years, and the

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Glossary

- **Ablation** Any process which results in the loss of mass from a body of snow or ice. This typically occurs from surface melting, but can also occur due to processes such as wind redistribution, sublimation and iceberg calving.
- **Accumulation** Any process which results in the addition of mass to a body of ice or snow. 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 mass.
- **Cold ice (also referred to as polar ice)** Ice which is at a temperature below the pressure melting point, and which therefore contains little free liquid water.
- **Firn** Snow which has survived one summer melt season, but is not yet glacier ice. Typically found in the accumulation area of glaciers.
- **Ice age** Any period in Earth’s history when there is permanent snow or ice on the Earth’s surface; we are therefore currently in an ice age. Glacial/interglacial periods refer to relatively colder/warmer periods within an ice age, with the last glacial maximum occurring between approximately 26,500 and 19,000 years ago.
- **Temperate ice (also referred to as warm ice)** Ice which is at the temperature below the pressure melting point, and therefore contains liquid water.

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1 Properties of glacial ice

<p>| | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Properties of glacial ice</td>
</tr>
<tr>
<td>2</td>
<td>Glacier classification schemes</td>
</tr>
<tr>
<td>3</td>
<td>Morphological classification</td>
</tr>
<tr>
<td>3.1</td>
<td>Unconstrained by topography</td>
</tr>
<tr>
<td>3.1.1</td>
<td>Ice sheet</td>
</tr>
<tr>
<td>3.1.2</td>
<td>Ice cap</td>
</tr>
<tr>
<td>3.1.3</td>
<td>Ice stream</td>
</tr>
<tr>
<td>3.1.4</td>
<td>Outlet glacier</td>
</tr>
<tr>
<td>3.1.5</td>
<td>Ice shelf</td>
</tr>
<tr>
<td>3.2</td>
<td>Constrained by topography</td>
</tr>
<tr>
<td>3.2.1</td>
<td>Icefield</td>
</tr>
<tr>
<td>3.2.2</td>
<td>Valley glacier</td>
</tr>
<tr>
<td>3.2.3</td>
<td>Cirque glacier</td>
</tr>
<tr>
<td>3.2.4</td>
<td>Piedmont glacier</td>
</tr>
<tr>
<td>3.2.5</td>
<td>Hanging glacier</td>
</tr>
<tr>
<td>4</td>
<td>Thermal classification</td>
</tr>
<tr>
<td>4.1</td>
<td>Temperate (“warm”)</td>
</tr>
<tr>
<td>4.2</td>
<td>Cold (“polar”)</td>
</tr>
<tr>
<td>4.3</td>
<td>Polythermal</td>
</tr>
<tr>
<td>5</td>
<td>Conclusions</td>
</tr>
<tr>
<td>References</td>
<td>10</td>
</tr>
</tbody>
</table>

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transition to glacial ice can happen at depths of a few tens of meters or less. In contrast, on glaciers in polar environments (e.g., interior Antarctica), there is low annual snowfall and the dry, low density snowpack can take hundreds to thousands of years to turn to glacial ice, at depths of over a hundred meters (Cuffey and Paterson, 2010).

Newly fallen snow has a density of \(\sim 50–200 \text{ kg m}^{-3}\), with lower densities related to snow falling in colder and calmer conditions, and higher densities to snow falling in warmer and windier conditions. Density rapidly increases as new snow is buried, with firn typically having a density of \(400–830 \text{ kg m}^{-3}\) (Cuffey and Paterson, 2010). Pure glacial ice has a density of \(917 \text{ kg m}^{-3}\), although this varies slightly based on factors such as bubble content, debris content, depth and ice temperature. Ice crystals have a hexagonal structure caused by the tetrahedral arrangement of \(\text{H}_2\text{O}\) molecules that they are formed from. In glaciers, ice crystals typically grow in size with depth (i.e., pressure) as large grains consume small grains. The finest-grained ice layers contain ice crystals with a mean grain size of \(\sim 1 \text{ mm}\), although ranges of \(\sim 2 \text{ mm} \approx 2 \text{ cm}\) are more common as depth increases (Gow et al., 1997). Ice crystals grow preferentially along their long axis (c-axis), and become organized in the direction of ice motion due to the processes of recrystallization and ice deformation.

Geomorphologically, ice masses 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 of \(\sim 10\%\) of the Earth’s land surface (IPCC, 2019). This totals an area of \(\sim 16,000,000 \text{ km}^2\), with approximately \(97\%\) of this currently occupied by the Greenland and Antarctic ice sheets (Benn and Evans, 2010).

2 Glacier classification schemes

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 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 remainder of this chapter provides a review of these classification schemes, with a particular focus on providing examples of ice masses that fall into different categories. This review covers permanent terrestrial ice masses, including those attached to land. For a multilingual technical definition of all ice types that occur 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 (https://www.swisseduc.ch/glaciers, accessed June 16, 2020).

3 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., Benn and Evans, 2010; Singh et al., 2011). Terrestrial ice masses fall into two primary categories within the morphological classification scheme: those unconstrained by surrounding topography, and those constrained by it.

3.1 Unconstrained by topography

3.1.1 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 \(>3 \text{ km}\) thick ice (Bell et al., 2011). Ice masses need to cover an area of \(>50,000 \text{ km}^2\) to be classified as an ice sheet (Armstrong et al., 1973), and at present there are only two ice sheets on Earth: the Greenland and Antarctic ice sheets. Antarctica is the thickest of these, with a maximum ice depth of \(\sim 4.9 \text{ km}\) (Fretwell et al., 2013), compared to a maximum depth of \(\sim 3.3 \text{ km}\) for the Greenland Ice Sheet (Bamber et al., 2013). At the peak of the last glacial maximum, which lasted from approximately 26,500–19,000 years ago (Clark et al., 2009), 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.

3.1.2 Ice cap

Ice caps are somewhat similar to ice sheets in that they cover almost all underlying topography, but are smaller in size (\(<50,000 \text{ km}^2\)). 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; Fig. 1), Svalbard (e.g., Austfonna) and Iceland (e.g., Vatnajökull). These ice caps typically reach maximum ice thicknesses of \(\sim 500–1000 \text{ m}\); for example, Dowdeswell et al. (2004) measured a maximum ice thickness of \(880 \text{ m}\) on Devon Ice Cap, and Björnsson and Pálsson (2008) reported a maximum thickness of \(\sim 950 \text{ m}\) for Vatnajökull.
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 near the ice cap or ice sheet 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 all directions from the ice dome, and flow rates increasing toward the ice cap or ice sheet margins (Fig. 1; Van Wychen et al., 2014). The low-velocity areas that form the boundary between ice flowing into different drainage basins are termed ice divides (white lines in Fig. 1).

3.1.3 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 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 (Rignot et al., 2011). Velocities on ice streams reach up to ~1 km year\(^{-1}\), and they typically move at least an order of magnitude faster than the 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 suggests 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 bedrock (Blankenship et al., 2001).

There have been debates as to whether the location of ice streams is controlled by subglacial topography, but the evidence is ambiguous. For example, Joughin et al. (2001) discovered that some areas of the Northeast Greenland Ice Stream have a subglacial topographic expression, but in other locations under this ice stream there is little evidence of any subglacial channel. Bennett (2003) describes the Siple Coast ice streams as being the only contemporary examples of “pure” ice streams in Antarctica, as they lack any subglacial topographic control. However, subglacial troughs have a major influence on ice stream flow elsewhere in Antarctica, such as for the Ferrigno Ice Stream (Bingham et al., 2012). Adjacent ice streams have sped up and slowed down in the past in both the Siple Coast (Anandakrishnan and Alley, 1997; Bennett, 2003) 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 400 years of stagnation (Hulbe and Fahnestock, 2004).

Fig. 1 Radial flow pattern displayed by Devon Ice Cap in the Canadian High Arctic, based on speckle tracking of Radarsat-2 image pairs collected in winter 2012. Surface elevations and ice thicknesses are greatest near the slow-moving central ice dome, with velocities highest along topographically constrained outlet glaciers. White lines indicate ice divides. Modified from Van Wychen W, Burgess DO, Gray L, Copland L, Sharp M, Dowdeswell J, and Benham T (2014) Glacier velocities and dynamic ice discharge from the Queen Elizabeth Islands, Nunavut, Canada. Geophysical Research Letters 41(2): 484–490. https://doi.org/10.1002/2013GL058558.
3.1.4 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 sheets and ice caps 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 recent research has focused on understanding the dynamics and variability of outlet glaciers and ice streams since they appear to provide a major control on broad-scale ice sheet/ice cap stability and mass balance (e.g., Pritchard et al., 2009; Rignot et al., 2011, 2019; Moon et al., 2014).

3.1.5 Ice shelf
Ice shelves comprise permanent floating ice masses that are attached to the coast (Fig. 2), and typically form the seaward extension of ice streams and outlet glaciers. This is particularly true in Antarctica where they comprise ~75% of the coastline (Shepherd et al., 2018), 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, 2500 m in depth, and covering a total area of >1,500,000 km², some 11% of the area of the Antarctic Ice Sheet (Fretwell et al., 2013; Schaffer et al., 2016). Ice shelves are also present along coastlines in the northernmost reaches of the Arctic (Copland and Mueller, 2017), 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, although many have recently broken up (Dowdeswell, 2017). The largest concentration of Arctic ice shelves occurs along northern Ellesmere Island, Canada, where 13 ice shelves had a total area of 535 km² in 2015 (Mueller et al., 2017). 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, with an example being the Ward Hunt Ice Rise on northern Ellesmere Island (Braun, 2017).

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 (Bernales et al., 2017). 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, 2017). The majority of mass

![Image](image_url)

**Fig. 2** Milne Shelf, Ellesmere Island, Canada, July 2014. Note the almost flat surface topography. Source: Luke Copland.
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; Canadian Ice Service, 2005). 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.

### 3.2 Constrained by topography

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

#### 3.2.1 Icefield

An icefield (also spelled ice field) forms the largest category of ice mass constrained by topography, and is somewhat similar to an ice cap in that they can reach substantial thicknesses (e.g., >1000 m for most of the central Northern Patagonia Icefield; Gourlet et al., 2016) 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; Figs. 3 and 4), southern South America (Northern and Southern Patagonia Icefield) and Ellesmere Island (Prince of Wales Icefield, Manson Icefield).

#### 3.2.2 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) (Figs. 4 and 5). 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 10–1000 m year\(^{-1}\) (although this can vary widely on both short and long timescales), with highest values where the ice is relatively thick and steep, and in maritime areas where there is highest snow accumulation (Waechter et al., 2015). 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 (Fig. 4), and are usually most clearly developed in the ablation area, where they can form substantial U-shaped valleys (Fig. 5). In places where

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**Fig. 3** St. Elias Icefield, Yukon, August 2016. Source: Luke Copland.
glaciers flow over bedrock obstacles they can become extremely steep and form icefalls, 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 can 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 which occupy depressions on shallower terrain and commonly receive much of their accumulation from snow avalanching and wind drifting (Benn and Evans, 2010).

### 3.2.3 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 (Figs. 4 and 6). They can either be entirely contained within the cirque depression or provide mass to downstream features such as icefalls and valley glaciers, and are very common in areas of glaciated mountainous terrain. They typically have a preferred north to northeasterly orientation in the northern hemisphere (and vice versa in the southern hemisphere) due to shading from the sun and atmospheric circulation patterns (Evans, 1977).

### 3.2.4 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 (Fig. 7). They typically occur at lower altitudes (i.e., in ablation areas), and usually 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 ~5000 km$^2$ (including tributaries; Sauber et al., 2005)

### 3.2.5 Hanging glacier

Hanging glaciers occur on steep terrain in mountainous areas, and frequently form from ice that spills out of cirques or clings to rock cliffs (Fig. 5). They are typically small in area, and provide a frequent source of avalanches to 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.

### 4 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 from boreholes (e.g., Blatter, 1987), but borehole drilling is typically a time-consuming and expensive process. Many indirect measures are therefore used to determine internal and basal ice temperatures, including modeling (e.g., Gilbert et al., 2014), ground-penetrating radar (e.g., Björnsson et al., 1996), glacier

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**Fig. 6** Cirque glaciers and icefalls at an altitude of ~3000 m in the St. Elias Mountains, Yukon. Source: Luke Copland.
hydrology and dynamics (e.g., Irvine-Fynn et al., 2011), and analysis of the glacial sedimentary record (e.g., Hambrey and Glasser, 2012).

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 viscosity of ice increases by a factor of five between ~−10 °C and ~−25 °C (Cuffey and Paterson, 2010). Ice masses which have a frozen bed typically undergo slow motion due to internal ice deformation alone. In 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 (Phillips et al., 2010). It is clear that ice temperature is also important in controlling other glaciological processes such as glacier surging (Fowler et al., 2001), subglacial sediment production (Hodson et al., 1997), and landscape development (Dyke, 1993).

In general, ice masses are grouped into one of three different thermal classifications: 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.

### 4.1 Temperate (“warm”)

In this category, ice is at the pressure melting point throughout, except for a near-surface layer up to ~20 m thick that is subject to seasonal temperature variations (Cuffey and Paterson, 2010). 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 (Fig. 5). Beyond the obvious requirement for a warm climate with an extended period of positive 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 build-up of cold ice at low elevations. In the same climate regime, thicker glaciers tend to be warmer than thinner glaciers as ice acts as an insulator that prevents cold surface temperatures from reaching the glacier bed.

![Fig. 7](image_url) An example of polythermal piedmont glaciers issuing from small ice caps and outlet glaciers near the east coast of Ellesmere Island, Canada. Source: air photo A16616-059 from the National Air Photo Library, Ottawa. © Her Majesty the Queen in Right of Canada.
4.2 Cold ("polar")

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, with little to no basal motion. Cold glaciers are typically thin and occur 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 \(-17\,\text{°C}\), the glaciers are frozen to their beds, and sublimation accounts for up to 80\% of their total annual ablation (Lewis et al., 1998).

4.3 Polythermal

Polythermal glaciers contain a mixture of both temperate and cold ice, and occur 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). Fig. 8 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 firm turns to ice and is carried to lower elevations by ice motion. It has a net vertical downward component in the accumulation area and an upward component in the ablation area. 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 a and b are generally referred to as "predominantly 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 Fig. 8 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,

![Fig. 8 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 and Hutter K (1991) Polythermal conditions in Arctic glaciers. Journal of Glaciology 37: 261–269, with permission of the International Glaciological Society.](image-url)
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. For example, modeling by Pattyn (2010) suggests that 55% of the grounded part of the Antarctic Ice Sheet is at the pressure melting point. They are thus best classified as being polythermal. Besides modeling, evidence for basal melting is provided by features such as the widespread occurrence of subglacial drainage (Wright et al., 2012; Moon et al., 2014), Antarctic subglacial lakes (Wright and Siegert, 2012), extensive basal freeze-on in East Antarctica (Bell et al., 2011), and borehole drilling into temperate basal ice (Engelhardt et al., 1990; Bentley and Koci, 2007).

A few studies have demonstrated the existence of polythermal glaciers at high altitude in the Himalaya-Karakoram-Tibet region. For example, Miles et al. (2018) used borehole temperatures to show the presence of polythermal ice in the lower ablation area of Khumbu Glacier, Nepal, with this part of the glacier consisting of ~56% temperate ice and much of the remainder within 0.8 °C of the melting-point. Temperature measurements at depths of up to 92 m, on glaciers ranging in altitude from 3815 to 6303 m in Tibet and areas to the north, recorded mean temperatures of ~6.6 to ~11.3 °C (Liu et al., 2009). However, more measurements are needed to understand the distribution of internal and basal ice temperatures across this region, particularly because changes in thermal regime have been linked to changes in the occurrence of glacier surges in locations such as the Karakoram (Hewitt, 2005).

5 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.

References
