

# Cryosphere

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First Edition, 2012

ISBN 978-81-323-1933-7

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*Published by:*

**Learning Press**

4735/22 Prakashdeep Bldg,

Ansari Road, Darya Ganj,

Delhi - 110002

Email: [info@wtbooks.com](mailto:info@wtbooks.com)

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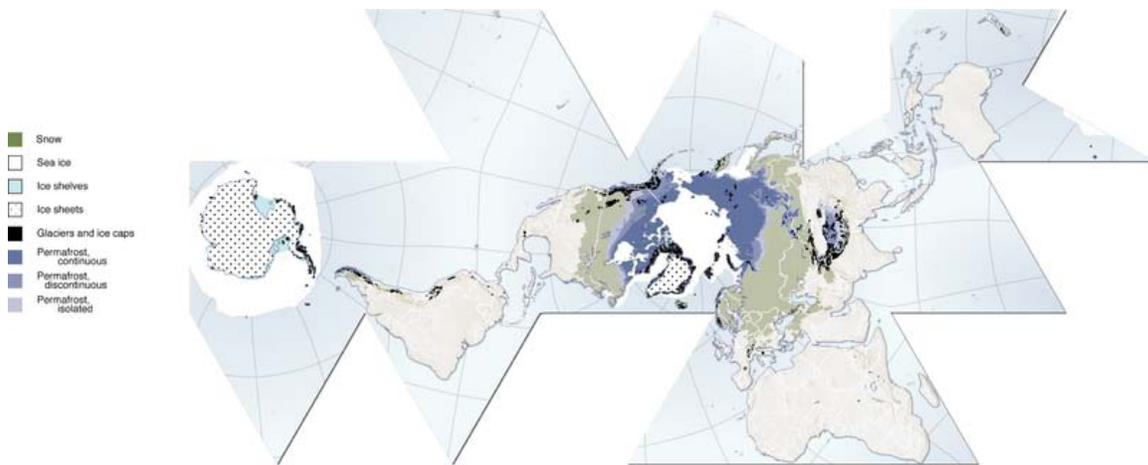
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# Introduction

The **cryosphere**, derived from the Ancient Greek word "κρύος" (*kyros* meaning "cold", "frost" or "ice"), is the term which collectively describes the portions of the Earth's surface where water is in solid form, including sea ice, lake ice, river ice, snow cover, glaciers, ice caps and ice sheets, and frozen ground (which includes permafrost). Thus there is a wide overlap with the hydrosphere. The cryosphere is an integral part of the global climate system with important linkages and feedbacks generated through its influence on surface energy and moisture fluxes, clouds, precipitation, hydrology, atmospheric and oceanic circulation. Through these feedback processes, the cryosphere plays a significant role in global climate and in climate model response to global change.



Overview of the Cryosphere and its larger components, from the UN Environment Programme Global Outlook for Ice and Snow.

## Structure

Frozen water is found on the Earth's surface primarily as snow cover, freshwater ice in lakes and rivers, sea ice, glaciers, ice sheets, and frozen ground and permafrost (permanently-frozen ground). The residence time of water in each of these cryospheric sub-systems varies widely. Snow cover and freshwater ice are essentially seasonal, and most sea ice, except for ice in the central Arctic, lasts only a few years if it is not seasonal. A given water particle in glaciers, ice sheets, or ground ice, however, may remain frozen for 10-100,000 years or longer, and deep ice in parts of East Antarctica may have an age approaching 1 million years.

Most of the world's ice volume is in Antarctica, principally in the East Antarctic Ice Sheet. In terms of areal extent, however, Northern Hemisphere winter snow and ice extent comprise the largest area, amounting to an average 23% of hemispheric surface area in January. The large areal extent and the important climatic roles of snow and ice, related to their unique physical properties, indicate that the ability to observe and model snow and ice-cover extent, thickness, and physical properties (radiative and thermal properties) is of particular significance for climate research.

There are several fundamental physical properties of snow and ice that modulate energy exchanges between the surface and the atmosphere. The most important properties are the surface reflectance (albedo), the ability to transfer heat (thermal diffusivity), and the ability to change state (latent heat). These physical properties, together with surface roughness, emissivity, and dielectric characteristics, have important implications for observing snow and ice from space. For example, surface roughness is often the dominant factor determining the strength of radar backscatter. Physical properties such as crystal structure, density, length, and liquid-water content are important factors affecting the transfers of heat and water and the scattering of microwave energy.

The surface reflectance of incoming solar radiation is important for the surface energy balance (SEB). It is the ratio of reflected to incident solar radiation, commonly referred to as albedo. Climatologists are primarily interested in albedo integrated over the shortwave portion of the electromagnetic spectrum (~0.3 to 3.5  $\mu\text{m}$ ), which coincides with the main solar energy input. Typically, albedo values for non-melting snow-covered surfaces are high (~80-90%) except in the case of forests. The higher albedos for snow and ice cause rapid shifts in surface reflectivity in autumn and spring in high latitudes, but the overall climatic significance of this increase is spatially and temporally modulated by cloud cover. (Planetary albedo is determined principally by cloud cover, and by the small amount of total solar radiation received in high latitudes during winter months.) Summer and autumn are times of high-average cloudiness over the Arctic Ocean so the albedo feedback associated with the large seasonal changes in sea-ice extent is greatly reduced. Groisman *et al.* (1994a) observed that snow cover exhibited the greatest influence on the Earth radiative balance in the spring (April to May) period when incoming solar radiation was greatest over snow-covered areas.

The thermal properties of cryospheric elements also have important climatic consequences. Snow and ice have much lower thermal diffusivities than air. Thermal diffusivity is a measure of the speed at which temperature waves can penetrate a substance. Snow and ice are many orders of magnitude less efficient at diffusing heat than air. Snow cover insulates the ground surface, and sea ice insulates the underlying ocean, decoupling the surface-atmosphere interface with respect to both heat and moisture fluxes. The flux of moisture from a water surface is eliminated by even a thin skin of ice, whereas the flux of heat through thin ice continues to be substantial until it attains a thickness in excess of 30 to 40 cm. However, even a small amount of snow on top of the ice will dramatically reduce the heat flux and slow down the rate of ice growth. The insulating effect of snow also has major implications for the hydrological cycle. In

non-permafrost regions, the insulating effect of snow is such that only near-surface ground freezes and deep-water drainage is uninterrupted.

While snow and ice act to insulate the surface from large energy losses in winter, they also act to retard warming in the spring and summer because of the large amount of energy required to melt ice (the latent heat of fusion,  $3.34 \times 10^5$  J/kg at  $0^\circ\text{C}$ ). However, the strong static stability of the atmosphere over areas of extensive snow or ice tends to confine the immediate cooling effect to a relatively shallow layer, so that associated atmospheric anomalies are usually short-lived and local to regional in scale. In some areas of the world such as Eurasia, however, the cooling associated with a heavy snowpack and moist spring soils is known to play a role in modulating the summer monsoon circulation. Gutzler and Preston (1997) recently presented evidence for a similar snow-summer circulation feedback over the southwestern United States.

The role of snow cover in modulating the monsoon is just one example of a short-term cryosphere-climate feedback involving the land surface and the atmosphere. From Figure 1 it can be seen that there are numerous cryosphere-climate feedbacks in the global climate system. These operate over a wide range of spatial and temporal scales from local seasonal cooling of air temperatures to hemispheric-scale variations in ice sheets over time-scales of thousands of years. The feedback mechanisms involved are often complex and incompletely understood. For example, Curry *et al.* (1995) showed that the so-called “simple” sea ice-albedo feedback involved complex interactions with lead fraction, melt ponds, ice thickness, snow cover, and sea-ice extent.

### **Lake ice and river ice**

Ice forms on rivers and lakes in response to seasonal cooling. The sizes of the ice bodies involved are too small to exert other than localized climatic effects. However, the freeze-up/break-up processes respond to large-scale and local weather factors, such that considerable interannual variability exists in the dates of appearance and disappearance of the ice. Long series of lake-ice observations can serve as a proxy climate record, and the monitoring of freeze-up and break-up trends may provide a convenient integrated and seasonally specific index of climatic perturbations. Information on river-ice conditions is less useful as a climatic proxy because ice formation is strongly dependent on river-flow regime, which is affected by precipitation, snow melt, and watershed runoff as well as being subject to human interference that directly modifies channel flow, or that indirectly affects the runoff via land-use practices.

Lake freeze-up depends on the heat storage in the lake and therefore on its depth, the rate and temperature of any inflow, and water-air energy fluxes. Information on lake depth is often unavailable, although some indication of the depth of shallow lakes in the Arctic can be obtained from airborne radar imagery during late winter (Sellman *et al.* 1975) and spaceborne optical imagery during summer (Duguay and Lafleur 1997). The timing of

breakup is modified by snow depth on the ice as well as by ice thickness and freshwater inflow.

## **Frozen ground and permafrost**

Frozen ground (permafrost and seasonally frozen ground) occupies approximately 54 million km<sup>2</sup> of the exposed land areas of the Northern Hemisphere (Zhang et al., 2003) and therefore has the largest areal extent of any component of the cryosphere. Permafrost (perennially frozen ground) may occur where mean annual air temperatures (MAAT) are less than -1 or -2°C and is generally continuous where MAAT are less than -7°C. In addition, its extent and thickness are affected by ground moisture content, vegetation cover, winter snow depth, and aspect. The global extent of permafrost is still not completely known, but it underlies approximately 20% of Northern Hemisphere land areas. Thicknesses exceed 600 m along the Arctic coast of northeastern Siberia and Alaska, but, toward the margins, permafrost becomes thinner and horizontally discontinuous. The marginal zones will be more immediately subject to any melting caused by a warming trend. Most of the presently existing permafrost formed during previous colder conditions and is therefore relic. However, permafrost may form under present-day polar climates where glaciers retreat or land emergence exposes unfrozen ground. Washburn (1973) concluded that most continuous permafrost is in balance with the present climate at its upper surface, but changes at the base depend on the present climate and geothermal heat flow; in contrast, most discontinuous permafrost is probably unstable or "in such delicate equilibrium that the slightest climatic or surface change will have drastic disequilibrium effects".

Under warming conditions, the increasing depth of the summer active layer has significant impacts on the hydrologic and geomorphic regimes. Thawing and retreat of permafrost have been reported in the upper Mackenzie Valley and along the southern margin of its occurrence in Manitoba, but such observations are not readily quantified and generalized. Based on average latitudinal gradients of air temperature, an average northward displacement of the southern permafrost boundary by 50-to-150 km could be expected, under equilibrium conditions, for a 1°C warming.

Only a fraction of the permafrost zone consists of actual ground ice. The remainder (dry permafrost) is simply soil or rock at subfreezing temperatures. The ice volume is generally greatest in the uppermost permafrost layers and mainly comprises pore and segregated ice in Earth material. Measurements of bore-hole temperatures in permafrost can be used as indicators of net changes in temperature regime. Gold and Lachenbruch (1973) infer a 2-4°C warming over 75 to 100 years at Cape Thompson, Alaska, where the upper 25% of the 400-m thick permafrost is unstable with respect to an equilibrium profile of temperature with depth (for the present mean annual surface temperature of -5°C). Maritime influences may have biased this estimate, however. At Prudhoe Bay similar data imply a 1.8°C warming over the last 100 years (Lachenbruch *et al.* 1982). Further complications may be introduced by changes in snow-cover depths and the natural or artificial disturbance of the surface vegetation.

The potential rates of permafrost thawing have been established by Osterkamp (1984) to be two centuries or less for 25-meter-thick permafrost in the discontinuous zone of interior Alaska, assuming warming from -0.4 to 0°C in 3–4 years, followed by a further 2.6°C rise. Although the response of permafrost (depth) to temperature change is typically a very slow process (Osterkamp 1984; Koster 1993), there is ample evidence for the fact that the active layer thickness quickly responds to a temperature change (Kane *et al.* 1991). Whether, under a warming or cooling scenario, global climate change will have a significant effect on the duration of frost-free periods in both regions with seasonally- and perennially-frozen ground.

# Chapter-1

## Snow

**Snow** is a type of precipitation within the Earth's atmosphere in the form of crystalline water ice, consisting of a multitude of snowflakes that fall from clouds. Since snow is composed of small ice particles, it is a granular material. It has an open and therefore soft structure, unless packed by external pressure. Snowflakes come in a variety of sizes and shapes. Types which fall in the form of a ball due to melting and refreezing, rather than a flake, are known as graupel, with ice pellets and snow grains as examples of graupel. Snowfall amount and its related liquid equivalent precipitation amount are determined using a variety of different rain gauges.

The process of precipitating snow is called **snowfall**. Snowfall tends to form within regions of upward motion of air around a type of low-pressure system known as an extratropical cyclone. Snow can fall poleward of their associated warm fronts and within their comma head precipitation patterns, which is called such due to its comma-like shape of the cloud and precipitation pattern around the poleward and west sides of extratropical cyclones. Where relatively warm water bodies are present, for example due to water evaporation from lakes, lake-effect snowfall becomes a concern downwind of the warm lakes within the cold cyclonic flow around the backside of extratropical cyclones. Lake-effect snowfall can be locally heavy. Thundersnow is possible within a cyclone's comma head and within lake effect precipitation bands. In mountainous areas, heavy snow is possible where upslope flow is maximized within windward sides of the terrain at elevation, if the atmosphere is cold enough.



Snow on trees, Germany

## Forms



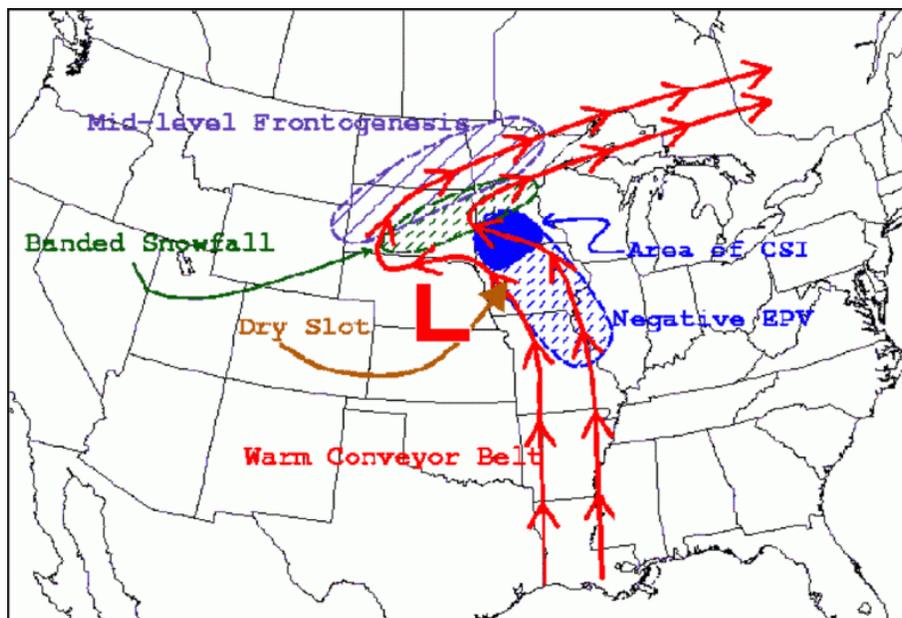
Photo of snow on the ground in the mountains of Southern California.

Once on the ground, snow can be categorized as powdery when fluffy, granular when it begins the cycle of melting and refreezing, and eventually ice once it packs down, after multiple melting and refreezing cycles, into a dense mass called snow pack. When powdery, snow moves with the wind from the location where it originally landed, forming deposits called snowdrifts which may have a depth of several meters. After attaching to hillsides, blown snow can evolve into a snow slab, which is an avalanche hazard on steep slopes. The existence of a snowpack keeps temperatures colder than they would be otherwise, as the whiteness of the snow reflects most sunlight, and the absorbed heat goes into melting the snow rather than increasing its temperature. The water equivalent of snowfall is measured to monitor how much liquid is available to flood rivers from meltwater which will occur during the upcoming spring. Snow cover can protect crops from extreme cold. If snowfall stays on the ground for a series of years uninterrupted, the snowpack develops into a mass of ice called glacier. Fresh snow absorbs sound, lowering ambient noise over a landscape because the trapped air between snowflakes attenuates vibration. These acoustic qualities quickly minimize, and reverse once a layer of freezing rain falls on top of snow cover. Walking across snowfall produces a squeaking sound at low temperatures.

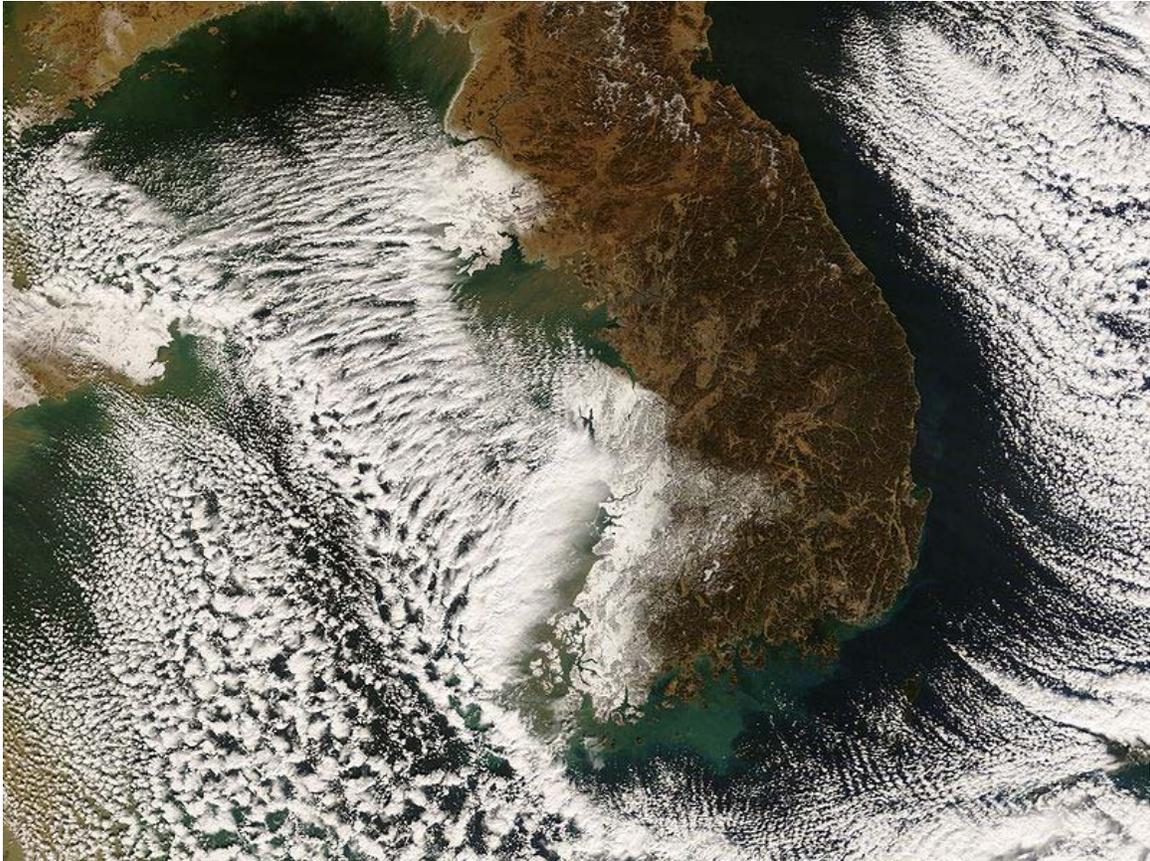
The energy balance of the snowpack itself is dictated by several heat exchange processes. The snowpack absorbs solar shortwave radiation that is partially blocked by cloud cover and reflected by snow surface. A long-wave heat exchange takes place between the snowpack and its surrounding environment that includes overlying air mass, tree cover and clouds. Heat exchange takes place by convection between the snowpack and the overlying air mass, and it is governed by the temperature gradient and wind speed. Moisture exchange between the snowpack and the overlying air mass is accompanied with latent heat transfer that is influenced by vapor pressure gradient and air wind. Rain on snow can add significant amounts of thermal energy to the snowpack. A generally insignificant heat exchange takes place by conduction between the snowpack and the ground. The small temperature change from before to after a snowfall is a result of the heat transfer between the snowpack and the air.

The term *snow storm* can describe a heavy snowfall while a *blizzard* involves snow and wind, obscuring visibility. *Snow shower* is a term for an intermittent snowfall, while *flurry* is used for very light, brief snowfalls. Snow can fall more than a meter at a time during a single storm in flat areas, and meters at a time in rugged terrain, such as mountains. When snow falls in significant quantities, travel by foot, car, airplane and other means becomes highly restricted, but other methods of mobility become possible: the use of snowmobiles, snowshoes and skis. When heavy snow occurs early in the fall, significant damage occurs to trees still in leaf. Areas with significant snow each year can store the winter snow within an ice house, which can be used to cool structures during the following summer. A variation on snow has been observed on Venus, though composed of metallic compounds and occurring at a substantially higher temperature.

## Cause



Preferred region of heavy snowfall ("Banded Snowfall") around the comma head of a wintertime low pressure area, shaded in green



Lake-effect snow bands near the Korean Peninsula

Extratropical cyclones can bring cold and dangerous conditions with heavy rain and snow with winds exceeding 119 km/h (74 mph), (sometimes referred to as windstorms in Europe). The band of precipitation that is associated with their warm front is often extensive, forced by weak upward vertical motion of air over the frontal boundary which condenses as it cools and produces precipitation within an elongated band, which is wide and stratiform, meaning falling out of nimbostratus clouds. When moist air tries to dislodge an arctic air mass, overrunning snow can result within the poleward side of the elongated precipitation band. In the Northern Hemisphere, poleward is towards the North Pole, or north. Within the Southern Hemisphere, poleward is towards the South Pole, or south.

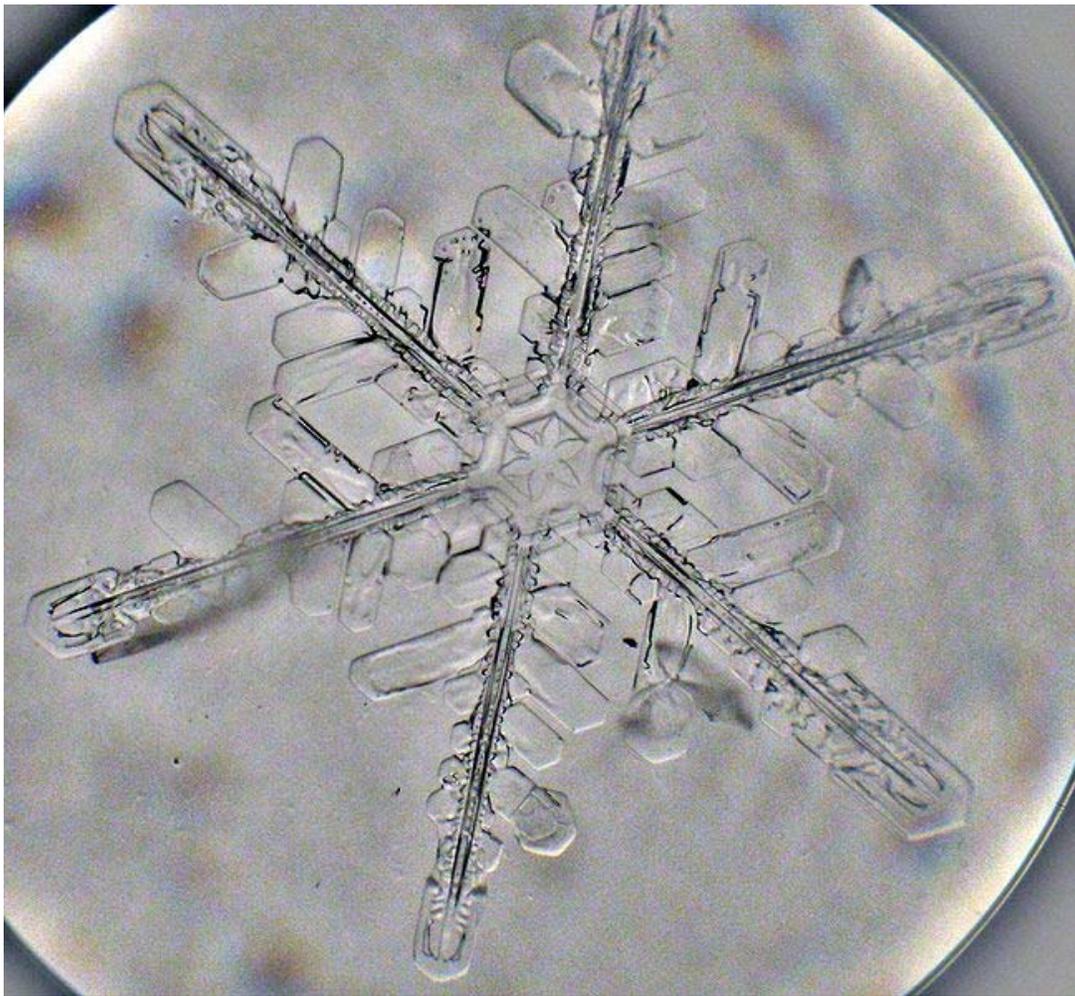
Within the *cold sector*, poleward and west of the cyclone center, small scale or mesoscale bands of heavy snow can occur within a cyclone's comma head pattern. The cyclone's comma head pattern is a comma-shaped area of clouds and precipitation found around mature extratropical cyclones. These snow bands typically have a width of 20 miles (32 km) to 50 miles (80 km). These bands in the comma head are associated with areas of frontogenesis, or zones of strengthening temperature contrast.

Southwest of extratropical cyclones, curved cyclonic flow bringing cold air across the relatively warm water bodies can lead to narrow lake-effect snow bands. Those bands

bring strong localized snowfall which can be understood as follows: Large water bodies such as lakes efficiently store heat that results in significant temperature differences (larger than 13 °C or 23 °F) between the water surface and the air above. Because of this temperature difference, warmth and moisture are transported upward, condensing into vertically oriented clouds which produce snow showers. The temperature decrease with height and cloud depth are directly affected by both the water temperature and the large-scale environment. The stronger the temperature decrease with height, the deeper the clouds get, and the greater the precipitation rate becomes.

In mountainous areas, heavy snowfall accumulates when air is forced to ascend the mountains and squeeze out precipitation along their windward slopes, which in cold conditions, falls in the form of snow. Because of the ruggedness of terrain, forecasting the location of heavy snowfall remains a significant challenge.

## Snowflakes



Snowflake viewed in an optical microscope

Snow crystals form when tiny supercooled cloud droplets (about 10  $\mu\text{m}$  in diameter) freeze. These droplets are able to remain liquid at temperatures lower than  $-18\text{ }^{\circ}\text{C}$  ( $0\text{ }^{\circ}\text{F}$ ), because to freeze, a few molecules in the droplet need to get together by chance to form an arrangement similar to that in an ice lattice; then the droplet freezes around this "nucleus." Experiments show that this "homogeneous" nucleation of cloud droplets only occurs at temperatures lower than  $-35\text{ }^{\circ}\text{C}$  ( $-31\text{ }^{\circ}\text{F}$ ). In warmer clouds an aerosol particle or "ice nucleus" must be present in (or in contact with) the droplet to act as a nucleus. Ice nuclei are very rare compared to that cloud condensation nuclei on which liquid droplets form. Clays, desert dust and biological particles may be effective, although to what extent is unclear. Artificial nuclei include particles of silver iodide and dry ice, and these are used to stimulate precipitation in cloud seeding.

Once a droplet has frozen, it grows in the supersaturated environment, which is one where air is saturated with respect to ice when the temperature is below the freezing point. The droplet then grows by diffusion of water molecules in the air (vapor) onto the ice crystal surface where they are collected. Because water droplets are so much more numerous than the ice crystals due to their sheer abundance, the crystals are able to grow to hundreds of micrometers or millimeters in size at the expense of the water droplets by a process known as the Wegner-Bergeron-Findeison process. The corresponding depletion of water vapor causes the ice crystals grow at the droplets' expense. These large crystals are an efficient source of precipitation, since they fall through the atmosphere due to their mass, and may collide and stick together in clusters, or aggregates. These aggregates are snowflakes, and are usually the type of ice particle that falls to the ground. Guinness World Records list the world's largest snowflakes as those of January 1887 at Fort Keogh, Montana; allegedly one measured 38 cm (15 in) wide. Although the ice is clear, scattering of light by the crystal facets and hollows/imperfections mean that the crystals often appear white in color due to diffuse reflection of the whole spectrum of light by the small ice particles.

The shape of the snowflake is determined broadly by the temperature and humidity at which it is formed. The most common snow particles are visibly irregular. Planar crystals (thin and flat) grow in air between  $0\text{ }^{\circ}\text{C}$  ( $32\text{ }^{\circ}\text{F}$ ) and  $-3\text{ }^{\circ}\text{C}$  ( $27\text{ }^{\circ}\text{F}$ ). Between  $-3\text{ }^{\circ}\text{C}$  ( $27\text{ }^{\circ}\text{F}$ ) and  $-8\text{ }^{\circ}\text{C}$  ( $18\text{ }^{\circ}\text{F}$ ), the crystals will form needles or hollow columns or prisms (long thin pencil-like shapes). From  $-8\text{ }^{\circ}\text{C}$  ( $18\text{ }^{\circ}\text{F}$ ) to  $-22\text{ }^{\circ}\text{C}$  ( $-8\text{ }^{\circ}\text{F}$ ) the shape reverts back to plate-like, often with branched or dendritic features. At temperatures below  $-22\text{ }^{\circ}\text{C}$  ( $-8\text{ }^{\circ}\text{F}$ ), the crystal development becomes column-like, although many more complex growth patterns also form such as side-planes, bullet-rosettes and also planar types depending on the conditions and ice nuclei. If a crystal has started forming in a column growth regime, at around  $-5\text{ }^{\circ}\text{C}$  ( $23\text{ }^{\circ}\text{F}$ ), and then falls into the warmer plate-like regime, then plate or dendritic crystals sprout at the end of the column, producing so called "capped columns."

A snowflake consists of roughly  $10^{18}$  water molecules, which are added to its core at different rates and in different patterns, depending on the changing temperature and humidity within the atmosphere that the snowflake falls through on its way to the ground. As a result, it is extremely difficult to encounter two identical snowflakes. Initial attempts

to find identical snowflakes by photographing thousands their images under a microscope from 1885 onward by Wilson Alwyn Bentley found the wide variety of snowflakes we know about today. It is more likely that two snowflakes could become virtually identical if their environments were similar enough. Matching snow crystals were discovered in Wisconsin in 1988. The crystals were not flakes in the usual sense but rather hollow hexagonal prisms.

## Types



Hoar frost that grows on the snow surface due to water vapor moving up through the snow on cold, clear nights

**Types of snow** can be designated by the shape of its flakes, description of how it is falling, and by how it collects on the ground. A blizzard and snow storm indicate heavy snowfalls over a large area, snow squalls give heavy snowfalls over narrow bands, while flurries are used for the lightest snowfall. Types which fall in the form of a ball, rather than a flake, are known as graupel, with sleet and snow grains as types of graupel. Once on the ground, snow can be categorized as powdery when fluffy, granular when it begins the cycle of melting and refreezing, and crud or eventually ice once it packs down into a dense drift after multiple melting and refreezing cycles. When powdering, snow drifts

with the wind, sometimes to the depth of several meters. After attaching to hillsides, blown snow can evolve into a snow slab, which is an avalanche hazard on steep slopes.

### **While falling**



A snow flurry viewed from Buffalo City Hall

Snowfall's intensity is determined by visibility. When the visibility is over 1 kilometre (0.62 mi), snow is determined to be light. Moderate snow describes snowfall with visibility restrictions between .5 kilometres (0.31 mi) and 1 kilometre (0.62 mi). Heavy snowfall describes conditions when visibility is restricted below .5 kilometres (0.31 mi). The term thundersnow describes a thunderstorm which produces snow as the primary form of precipitation. A blizzard is a long-lasting snow storm with intense snowfall and

usually high winds. Particularly severe storms can create whiteout conditions where visibility is severely reduced. A snow storm is a long storm of relatively heavy snow, similar to a blizzard but without the wind requirement. A snowsquall is a brief, very intense snowstorm. A snow flurry describes a period of light snow with usually little accumulation with occasional moderate snowfall.

## Shapes

### Columns

A class of snowflakes that is shaped like a six sided column. One of the 4 classes of snowflakes.

### Dendrites

A class of snowflakes that has 6 points, making it somewhat star shaped. The classic snowflake shape. One of the 4 classes of snowflakes.

### Graupel

Precipitation formed when freezing fog condenses on a snowflake, forming a ball of rime ice. Also known as snow pellets.

### Ground blizzard

Occurs when a strong wind drives already fallen snow to create drifts and whiteouts.

### Lake-effect snow

Produced when cold winds move across long expanses of warmer lake water, picking up water vapor which freezes and is deposited on the lake's shores.

### Needles

A class of snowflakes that are acicular in shape (their length is much longer than their diameter, like a needle). One of the 4 classes of snowflakes.

### Rimed snow

Snowflakes that are partially or completely coated in tiny frozen water droplets called rime. Rime forms on a snowflake when it passes through a super-cooled cloud. One of the 4 classes of snowflakes.

### Sleet

In Canada and Britain, rain mixed with snow; Some Americans also refer to this as sleet, while others refer to sleet as ice pellets formed when snowflakes pass through a layer of warm air, partially or completely thaw, then refreeze upon passing through sufficiently cold air during further descent.



Snow on trees in DuBois, Pennsylvania.

## On the ground



Snow on the ground in Southern California.



Snow blowing from a roof in Ottawa, Ontario, Canada



### Snowfall in San Bernardino, California, United States

#### Artificial snow

Snow can be also manufactured using snow cannons, which actually create tiny granules more like soft hail (this is sometimes called "grits" by those in the southern U.S. for its likeness to the texture of the food). In recent years, snow cannons have been produced that create more natural-looking snow, but these machines are prohibitively expensive.

#### Blowing snow

Snow on ground that is being moved around by wind.

#### Chopped powder

Powder snow that has been cut up by previous skiers.

#### Corn

Coarse, granular wet snow. Most commonly used by skiers describing good spring snow. Corn is the result of diurnal cycle of melting and refreezing.

#### Cornice

An overhanging formation of windblown snow. Important in skiing and alpine climbing because the overhang can be unstable and hard to see from the leeward side.

#### Crud

This covers varieties of snow that all but advanced skiers find impassable. Subtypes are (a) windblown powder with irregularly shaped crust patches and ridges, (b) heavy tracked spring snow re-frozen to leave a deeply rutted surface strewn with loose blocks, (c) a deep layer of heavy snow saturated by rain (although this may go by another term). Crud is negotiated with a even weighting along the length of the skis, and smooth radius turns started, if necessary, with a pop or jump. When an advanced skier falls over on crud, it is probably because it is 'heavy crud', q.v.

#### Crust

A layer of snow on the surface of the snowpack that is stronger than the snow below, which may be powder snow. Depending on their thickness and resulting strength, crusts can be termed "supportable", meaning that they will support the weight of a human, "breakable", meaning that they will not, or "zipper", meaning that a skier can break and ski through the crust. Crusts often result from partial melting of the snow surface by direct sunlight or warm air followed by re-freezing.

#### Depth hoar

Faceted snow crystals, usually poorly or completely unbonded (unsintered) to adjacent crystals, creating a weak zone in the snowpack. Depth hoar forms from metamorphism of the snowpack in response to a large temperature gradient between the warmer ground beneath the snowpack and the surface. The relatively high porosity (percentage of air space), relatively warm temperature (usually near freezing point), and unbonded weak snow in this layer can allow various organisms to live in it.

#### Finger Drift

A narrow snow drift (30 cm to 1 metre in width) crossing a roadway. Several finger drifts in succession resemble the fingers of a hand.

#### Ice

Densely packed material formed from snow that doesn't contain air bubbles. Depending on the snow accumulation rate, the air temperature, and the weight of the snow in the upper layers, it can take snow a few hours or a few decades to form into ice.

#### Firn

Snow which has been lying for at least a year but which has not yet consolidated into glacier ice. It is granular.

#### Packed Powder

The most common snow cover on ski slopes, consisting of powder snow that has lain on the ground long enough to become compressed, but is still loose.

#### Packing snow

Snow that is at or near the melting point, so that it can easily be packed into snowballs and hurled at other people or objects. This is perfect for snow fights and other winter fun, such as making a snowman, or a snow fort.

#### Penitentes

Tall blades of snow found at high altitudes.

#### Pillow Drift

A snow drift crossing a roadway and usually 3 to 4.5 metres (10-15 feet) in width and 30 cm to 90 cm (1-3 feet) in depth.

#### Powder

Freshly fallen, uncompacted snow. The density and moisture content of powder snow can vary widely; snowfall in coastal regions and areas with higher humidity is usually heavier than a similar depth of snowfall in an arid or continental region. Light, dry (low moisture content, typically 4-7% water content) powder snow is prized by skiers and snowboarders. It is often found in the Rocky Mountains of North America and in most regions in Japan.



The textures of a snowdrift on the Long Mynd, Shropshire

**Slush**

Snow which partially melts upon reaching the ground, to the point that it accumulates in puddles of partially-frozen water.

**Snirt**

Snow covered with dirt, which occurs most often in spring, in Prairie States like North Dakota, where strong winds pick up black topsoil from uncovered farm fields and blow it into nearby towns where the melt rate is slower. The phenomenon is almost magical; one goes to sleep with white snow outside and awakens to black snow. Also, snow that is dirty, often seen by the side of roads and parking lots near areas that have been plowed.

**Snowdrift**

Large piles of snow which occur near walls and curbs, as the wind tends to push the snow up toward the vertical surfaces.

**Surface Hoar**

Faceted, corn-flake shaped snow crystals that are a type of frost that forms on the surface of the snow pack on cold, clear, calm nights. Subsequent snowfall can bury layers of surface hoar, incorporating them into the snowpack where they can form a weak layer. Sometimes referred to as hoar frost.

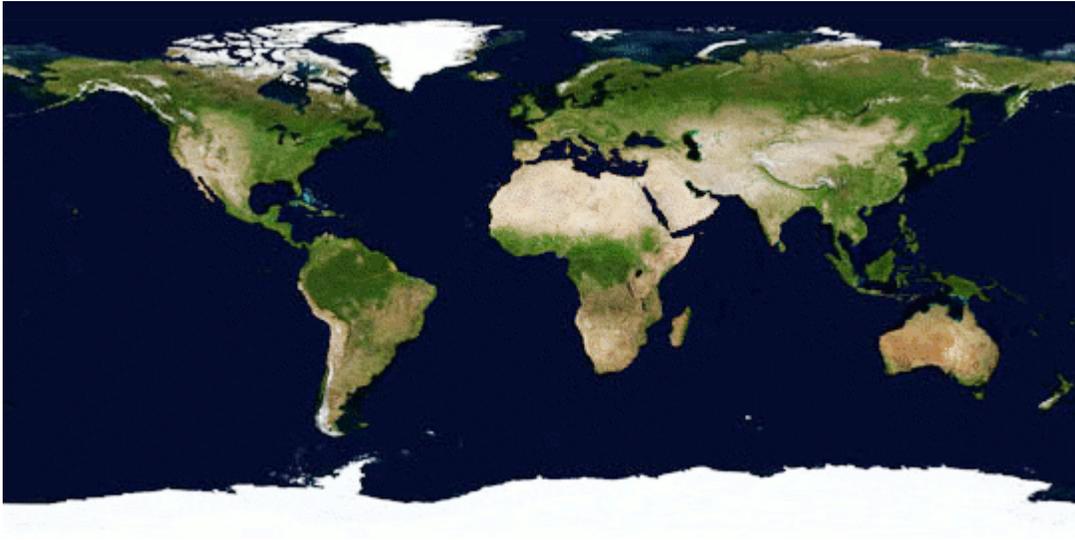
**Watermelon snow**

A reddish/pink-colored snow that smells like watermelons, and is caused by a red-colored green algae called *Chlamydomonas nivalis*.

## Wind slab

A layer of relatively stiff, hard snow formed by deposition of wind blown snow on the leeward side of a ridge or other sheltered area. Wind slabs can form over weaker, softer freshly fallen powder snow, creating an avalanche hazard on steep slopes.

## Density



An image(satellite images) showing seasonal snow changes



Firn

Snow remains on the ground until it melts or sublimates. Sublimation of snow directly into water vapor is most likely to occur on a dry and windy day such as when a strong downslope wind, such as a Chinook wind, exists. The *water equivalent* of a given amount of snow is the depth of a layer of water having the same mass and upper area. For example, if the snow covering a given area has a water equivalent of 50 centimeters (20 in), then it will melt into a pool of water 50 centimeters (20 in) deep covering the same area. This is a much more useful measurement to hydrologists than snow *depth*, as the density of cool freshly fallen snow widely varies. New snow commonly has a density of around 8% of water. This means that 33 centimeters (13 in) of snow melts down to 2.5 centimeters (1 in) of water. Cloud temperatures and physical processes in the cloud affect the shape of individual snow crystals. Highly branched or dendritic crystals tend to have more space between the arms of ice that form the snowflake and this snow will therefore have a lower density, often referred to as "dry" snow. Conditions that create columnar or plate-like crystals will have much less air space within the crystal and will therefore be denser and feel "wetter".

Once the snow is on the ground, it will settle under its own weight (largely due to differential evaporation) until its density is approximately 30% of water. Increases in density above this initial compression occur primarily by melting and refreezing, caused by temperatures above freezing or by direct solar radiation. In colder climates, snow lies on the ground all winter. By late spring, snow densities typically reach a maximum of 50% of water. When the snow does not all melt in the summer it evolves into firn, where individual granules become more spherical in nature, evolving into a glacier as the ice flows downhill.

## Acoustic properties



Snow falling in Scotland (21 seconds)

Under water, snowfall has a unique sound when compared to other forms of precipitation. Despite the different sizes and shapes on individual snowflakes, the sound made when individual flakes fall upon the surface of a freshwater lake are quite similar. On the ground, newly fallen snow acts as a sound-absorbing material, which minimizes sound over its surface. This is due to the trapped air between the individual crystalline flakes which act to trap sound waves and dampen vibrations. Once it is blown around by the wind and exposed to sunshine, snow hardens and its sound-softening quality diminishes. Snow cover as thin as 2 centimeters (0.79 in) thick changes the acoustic properties of a landscape. Studies concerning the acoustic properties of snow have revealed that loud sounds, such as from a pistol, can be used to measure snow cover permeability and depth. Within motion pictures, the sound of walking through cornstarch, salt, or cat litter has been found to be similar to that of walking through snow. When the temperature falls below  $-10\text{ }^{\circ}\text{C}$  ( $14.0\text{ }^{\circ}\text{F}$ ), snow will squeak when walked upon due to the crushing of the ice crystals within the snow. If covered by a layer of freezing rain, the hardened frozen surface acts to echo sounds, similar to concrete.

## Snowfall measurement



An ultrasonic snow depth sensor

The liquid equivalent of snowfall may be evaluated using a snow gauge or with a standard rain gauge having a diameter of 100 mm (4 in; plastic) or 200 mm (8 in; metal). Rain gauges are adjusted to winter by removing the funnel and inner cylinder and allowing the snow/freezing rain to collect inside the outer cylinder. Antifreeze liquid may be added to melt the snow or ice that falls into the gauge. In both types of gauges once the snowfall/ice is finished accumulating, or as its height in the gauge approaches 300 mm (12 in), the snow is melted and the water amount recorded.

Another type of gauge used to measure the liquid equivalent of snowfall is the weighing precipitation gauge. The wedge and tipping bucket gauges will have problems with snow measurement. Attempts to compensate for snow/ice by warming the tipping bucket meet with limited success, since snow may sublime if the gauge is kept much above the freezing temperature. Weighing gauges with antifreeze should do fine with snow, but again, the funnel needs to be removed before the event begins. At some automatic weather stations an ultrasonic snow depth sensor may be used to augment the precipitation gauge.

Spring snow melt is a major source of water supply to areas in temperate zones near mountains that catch and hold winter snow, especially those with a prolonged dry summer. In such places, water equivalent is of great interest to water managers wishing to predict spring runoff and the water supply of cities downstream. Measurements are made manually at marked locations known as *snow courses*, and remotely using special scales called *snow pillows*. Snow stakes and simple rulers can be used to determine the depth of the snow pack, though they will not evaluate either its density or liquid equivalent.

When a snow measurement is made, various networks exist across the United States and elsewhere where rainfall measurements can be submitted through the Internet, such as CoCoRAHS or GLOBE. If a network is not available in the area where one lives, the nearest local weather office will likely be interested in the measurement.

The world record for the highest seasonal total snowfall was measured in the United States at Mount Baker Ski Area, outside of the town Bellingham, Washington during the 1998–1999 season. Mount Baker received 2,896 cm (1,140 in) of snow, thus surpassing the previous record holder, Mount Rainier, Washington, which during the 1971–1972 season received 2,850 cm (1,120 in) of snow.

The Inuit carved snow goggles from caribou antlers to help prevent snow blindness. The goggles were curved to fit the user's face and had a large groove cut in the back to allow for the nose. A long thin slit was cut through the goggles to allow in a small amount of light, diminishing the amount of UV rays that get through. The goggles were held to the head by a cord made of caribou sinew.

## **Relation to river flow**

Many rivers originating in mountainous or high-latitude regions have a significant portion of their flow from snowmelt. This often makes the river's flow highly seasonal resulting in periodic flooding. In contrast, if much of the melt is from glaciated or nearly glaciated areas, the melt continues through the warm season, with peak flows occurring in mid to late summer.

## Effects on human society



A snow blockade in southern Minnesota, US in 1881

Substantial snowfall can disrupt public infrastructure and services, slowing human activity even in regions that are accustomed to such weather. Air and ground transport may be greatly inhibited or shut down entirely. Populations living in snow-prone areas have developed various ways to travel across the snow, such as skis, snowshoes, and sleds pulled by horses, dogs, or other animals and later, snowmobiles. Basic utilities such as electricity, telephone lines, and gas supply can also fail. In addition, snow can make roads much harder to travel and vehicles attempting to use them can easily become stuck.

The combined effects can lead to a "snow day" on which gatherings such as school, work, or church are officially canceled. In areas that normally have very little or no snow, a snow day may occur when there is only light accumulation or even the threat of snowfall, since those areas are unprepared to handle any amount of snow. In some areas, such as some states in the United States, schools are given a yearly quota of snow days (or "calamity days"). Once the quota is exceeded, the snow days must be made up. In other states, all snow days must be made up. For example, schools may extend the remaining school days later into the afternoon, shorten spring break, or delay the start of summer vacation.

Accumulated snow is removed to make travel easier and safer, and to decrease the long-term impact of a heavy snowfall. This process utilizes shovels, snowplows and is often assisted by sprinkling salt or other chloride-based chemicals, which reduce the melting temperature of snow. In areas with abundant snowfall, such as Northern Japan, people harvest snow and store it surrounded by insulation in ice houses. This allowed the ice to be used in summer for refrigeration or medical uses, which is one method of conserving electrical usage.



Snow in Old Fort, North Carolina caused by the 2009 Blizzard

## **Agriculture**

Snowfall can be beneficial to agriculture by serving as a thermal insulator, conserving the heat of the Earth and protecting crops from subfreezing weather. Some agricultural areas depend on an accumulation of snow during winter that will melt gradually in spring, providing water for crop growth. If it melts into water and refreezes upon sensitive crops, such as oranges, the resulting ice will protect the fruit from exposure to lower temperatures.

## Recreation



Making a giant snowball



Making a snow angel

Many winter sports, such as skiing, snowboarding, snowmobiling, and snowshoeing depend upon snow. Where snow is scarce but the temperature is low enough, snow cannons may be used to produce an adequate amount for such sports. Children and adults can play on a sled or ride in a sleigh. Although a person's footsteps remain a visible lifeline within a snow-covered landscape, snow cover is considered a general danger to hiking since the snow obscures landmarks and makes the landscape itself appear uniform.

One of the recognizable recreational uses of snow is in building snowmen. A snowman is created by making a man shaped figure out of snow - often using a large, shaped snowball for the body and a smaller snowball for the head which is often decorated with simple household items - traditionally including a carrot for a nose, and coal for eyes, nose and mouth; occasionally including old clothes such as a top hat or scarf.

Snow can be used to make snow cones, also known as snowballs, which are usually eaten in the summer months. Flat areas of snow can be used to make snow angels, a popular past-time for children.

Snow can be used to alter the format of outdoor games such as Capture the flag, or for snowball fights. The world's biggest snowcastle, the SnowCastle of Kemi, is built in Kemi, Finland every winter. Since 1928 Michigan Technological University in

Houghton, Michigan has held an annual Winter Carnival in mid-February, during which a large Snow Sculpture Contest takes place between various clubs, fraternities, and organizations in the community and the university. Each year there is a central theme, and prizes are awarded based on creativity. Snowball softball tournaments are held in snowy areas, usually using a bright orange softball for visibility, and burlap sacks filled with snow for the bases.

## Damage



Damage caused by Lake Storm "Aphid" in October 2006

When heavy, wet snow with a snow-water equivalent (SWE) ratio of between 6:1 and 12:1 and a weight in excess of 10 pounds per square foot ( $\sim 50 \text{ kg/m}^2$ ) piles onto trees or electricity lines - particularly if the trees have full leaves or are not adapted to snow - significant damage may occur on a scale usually associated with hurricanes. An avalanche can occur upon a sudden thermal or mechanical impact upon snow that has accumulated on a mountain, which causes the snow to rush downhill en masse. Preceding an avalanche is a phenomenon known as an avalanche wind caused by the approaching avalanche itself, which adds to its destructive potential. Large amounts of snow which accumulate on top of man-made structures can lead to structural failure. During

snowmelt, acidic precipitation which previously fell into the snow pack is released, which harms marine life.

## **Design of structures considering snow load**

The design of all structures and buildings use the ground snow load to some extent by professional engineers and designers. In North America, the northern states will be designed to accommodate the live load design contributed by the ground snow load in a pounds per square foot (PSF) loading analysis. (*Snow loads are typically treated as 'dead loads' within the ASCE 7-latest edition.*) This load is typically the governing design factor on roofs and structural elements exposed to the effects of snow. Closer to the Equator, the snow load becomes less of a factor and snow may or may not be the governing factor. Ground snow in North America is provided by the American Society of Civil Engineers (ASCE7-latest edition) for most jurisdictions.

## **Extraterrestrial snow**

Very light snow is known to occur at high latitudes on Mars. A "snow" of hydrocarbons is also theorized to occur on Saturn's moon Titan.

While there is little or no water on Venus, there is a phenomenon which is quite similar to snow. The Magellan probe imaged a highly reflective substance at the tops of Venus's highest mountain peaks which bore a strong resemblance to terrestrial snow. This substance arguably formed from a similar process to snow, albeit at a far higher temperature. Too volatile to condense on the surface, it rose in gas form to cooler higher elevations, where it then fell as precipitation. The identity of this substance is not known with certainty, but speculation has ranged from elemental tellurium to lead sulfide (galena).

Snow cover has the second-largest areal extent of any component of the cryosphere, with a mean maximum areal extent of approximately 47 million km<sup>2</sup>. Most of the Earth's snow-covered area (SCA) is located in the Northern Hemisphere, and temporal variability is dominated by the seasonal cycle; Northern Hemisphere snow-cover extent ranges from 46.5 million km<sup>2</sup> in January to 3.8 million km<sup>2</sup> in August. North American winter SCA has exhibited an increasing trend over much of this century (Brown and Goodison 1996; Hughes *et al.* 1996) largely in response to an increase in precipitation. However, the available satellite data show that the hemispheric winter snow cover has exhibited little interannual variability over the 1972-1996 period, with a coefficient of variation (COV=s.d./mean) for January Northern Hemisphere snow cover of < 0.04. According to Groisman *et al.* (1994a) Northern Hemisphere spring snow cover should exhibit a decreasing trend to explain an observed increase in Northern Hemisphere spring air temperatures this century. Preliminary estimates of SCA from historical and reconstructed in situ snow-cover data suggest this is the case for Eurasia, but not for North America, where spring snow cover has remained close to current levels over most of this century. Because of the close relationship observed between hemispheric air

temperature and snow-cover extent over the period of satellite data (IPCC 1996), there is considerable interest in monitoring Northern Hemisphere snow-cover extent for detecting and monitoring climate change.

Snow cover is an extremely important storage component in the water balance, especially seasonal snowpacks in mountainous areas of the world. Though limited in extent, seasonal snowpacks in the Earth's mountain ranges account for the major source of the runoff for stream flow and groundwater recharge over wide areas of the midlatitudes. For example, over 85% of the annual runoff from the Colorado River basin originates as snowmelt. Snowmelt runoff from the Earth's mountains fills the rivers and recharges the aquifers that over a billion people depend on for their water resources. Further, over 40% of the world's protected areas are in mountains, attesting to their value both as unique ecosystems needing protection and as recreation areas for humans. Climate warming is expected to result in major changes to the partitioning of snow and rainfall, and to the timing of snowmelt, which will have important implications for water use and management. These changes also involve potentially important decadal and longer time-scale feedbacks to the climate system through temporal and spatial changes in soil moisture and runoff to the oceans.(Walsh 1995). Freshwater fluxes from the snow cover into the marine environment may be important, as the total flux is probably of the same magnitude as desalinated ridging and rubble areas of sea ice. In addition, there is an associated pulse of precipitated pollutants which accumulate over the Arctic winter in snowfall and are released into the ocean upon ablation of the sea-ice.

## Chapter-2

# Sea Ice



Change in extent of the Arctic Sea ice between March and September.

**Sea ice** is largely formed from seawater that freezes. Because the oceans consist of saltwater, this occurs below the freezing point of pure water, at about  $-1.8\text{ }^{\circ}\text{C}$  ( $28.8\text{ }^{\circ}\text{F}$ ).

Sea ice may be contrasted with icebergs, which are chunks of ice shelves or glaciers that calve into the ocean. Icebergs are compacted snow and hence are fresh water from the

beginning; sea ice loses its salt during its process of formation from sea water and thus eventually is fresh as well.

## **Types of sea ice**



Satellite image of Fennoscandia in winter. The Gulf of Bothnia and White Sea are covered with sea ice.

Land-fast ice, or simply fast ice, is sea ice that has frozen along coasts ("fastened" to them) or to the sea floor over shallow parts of the continental shelf, and extends out from land into sea. Unlike drift ice, it does not move with currents and wind.

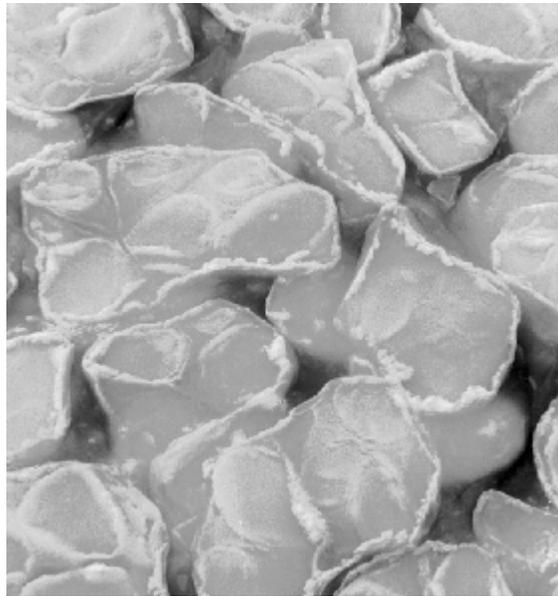
Drift ice consists of ice that floats on the surface of the water, as distinguished from the fast ice, attached to coasts. When packed together in large masses, drift ice is called *pack ice*. Pack ice may be either freely floating or blocked by fast ice while drifting past.

Pancake ice is sea ice broken into small round chunks looking like pancakes.

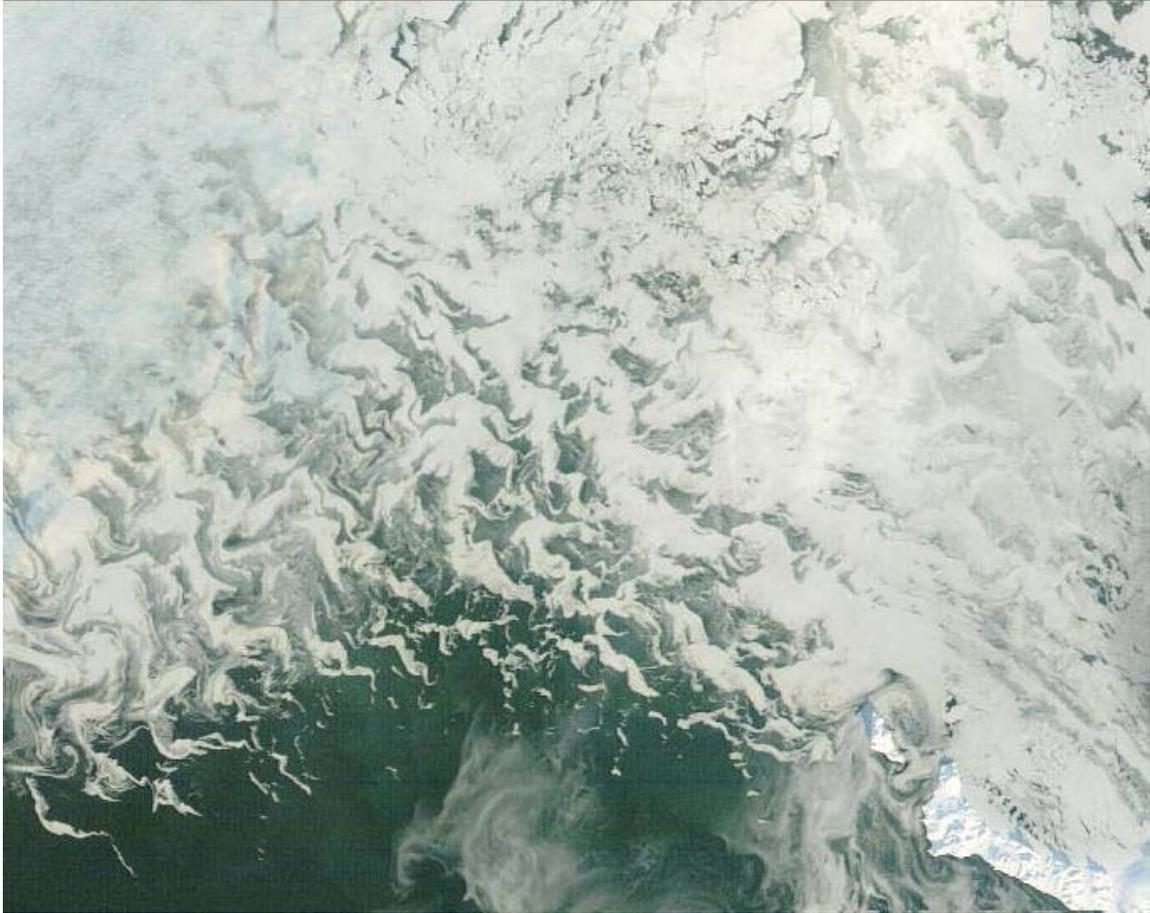
The most important areas of pack ice are the polar ice packs formed from seawater in the Earth's polar regions: the Arctic ice pack of the Arctic Ocean and the Antarctic ice pack of the Southern Ocean. Polar packs significantly change their size during seasonal changes of the year. Because of vast amounts of water added to or removed from the oceans and atmosphere, the behavior of polar ice packs have a significant impact of the global changes in climate.

An **ice floe** is a floating chunk of ice that is less than 10 kilometers (six miles) in its greatest dimension. Wider chunks of ice are called ice fields.

## **Formation of sea ice**



Pancake ice is sea ice that has been compressed by the action of waves on frazil ice. Plates are typically 1 - 3 meters across.



Satellite image of sea ice forming near St. Matthew Island in the Bering Sea.

Sea ice is a surprisingly complex composite composed primarily of pure ice in various states of crystallization along with air bubbles and included pockets of brine. Understanding its **growth processes** is important both for climate scientists for use in simulations as well remote sensing specialists since the composition and microstructural properties of the ice ultimately affect how it interacts with electromagnetic radiation.

Sea ice growth models for predicting the ice distribution and extent are also valuable for shipping concerns. An ice growth model can be combined with remote sensing measurements as an assimilation model as a means of generating more accurate ice charts.

## Overview

Several formation mechanisms of sea ice have been identified. At its earliest stages, sea ice consists of elongated, randomly oriented crystals. This is called frazil and mixed with water in the unconsolidated state is known as grease ice. If wave and wind conditions are calm, these crystals will consolidate at the surface and by selective pressure, begin to grow preferentially in the downward direction, forming nilas. In more turbulent

conditions, the frazil will consolidate by mechanical action to form pancake ice, which has a more random structure. Another common formation mechanism, especially in the Antarctic where precipitation over sea ice is high, is from snow deposition: on thin ice, the snow will weigh down the ice enough to cause flooding. Subsequent freezing will form ice with a much more granular structure.

One of the more interesting processes to occur within consolidated ice packs is changes in the saline content. As the ice freezes, most of the salt content gets rejected and forms highly saline brine inclusions between the crystals. With decreasing temperatures in the ice sheet, the size of the brine pockets decreases while the salt content goes up. Since ice is less dense than water, increasing pressure causes some of the brine to be ejected from both the top and bottom, producing the characteristic 'C'-shaped profile of first year ice. Brine will also drain through vertical channels, particularly in the melt season. Thus multi-year ice will tend to have both lower salinity and lower density than first-year ice.

## Vertical growth

The downward growth of consolidated ice in calm conditions is determined by the rate of heat transfer,  $Q^*$ , at the ice-water interface. If we assume that the ice is in thermal equilibrium both with itself and its surroundings and that the weather conditions are known, then we can determine  $Q^*$  by solving the following equation:

$$hQ^* = k(T_s - T_w)$$

for  $T_s$ , the surface temperature. The water temperature,  $T_w$ , is assumed to be at or near freezing, while the ice thickness,  $h$ , is assumed to be known and we can approximate the thermal conductivity,  $k$ , as an average over the layers (which have different salinities) or simply use the value for pure ice. The net heat flux is a total over four components:

$$Q^* = Q_E [e(T_s)] + Q_H(T_s) + Q_{LW}(T_s^4) + Q_{SW}$$

which are latent, sensible, longwave and shortwave fluxes, respectively. The equation can be solved using a numerical root-finding algorithm such as bisection: the functional dependencies on surface temperature are given, with  $e$  being the equilibrium vapor pressure.

While Cox and Weeks assume thermal equilibrium, Tanboe uses a more sophisticated thermodynamic model based on numerical solution of the heat equation. This would be appropriate when the ice is thick or the weather conditions are changing rapidly.

The rate of ice growth can be calculated from heat flux by the following equation:

$$g = \frac{Q^*}{L\rho}$$

where  $L$  is the latent heat of fusion for water and  $\rho$  is the density of ice. The growth rate in turn determines the saline content of the newly frozen ice. Empirical equations for determining the initial brine entrapment in sea ice have been derived by Cox and Weeks and Nakawo and Sinha and take the form:

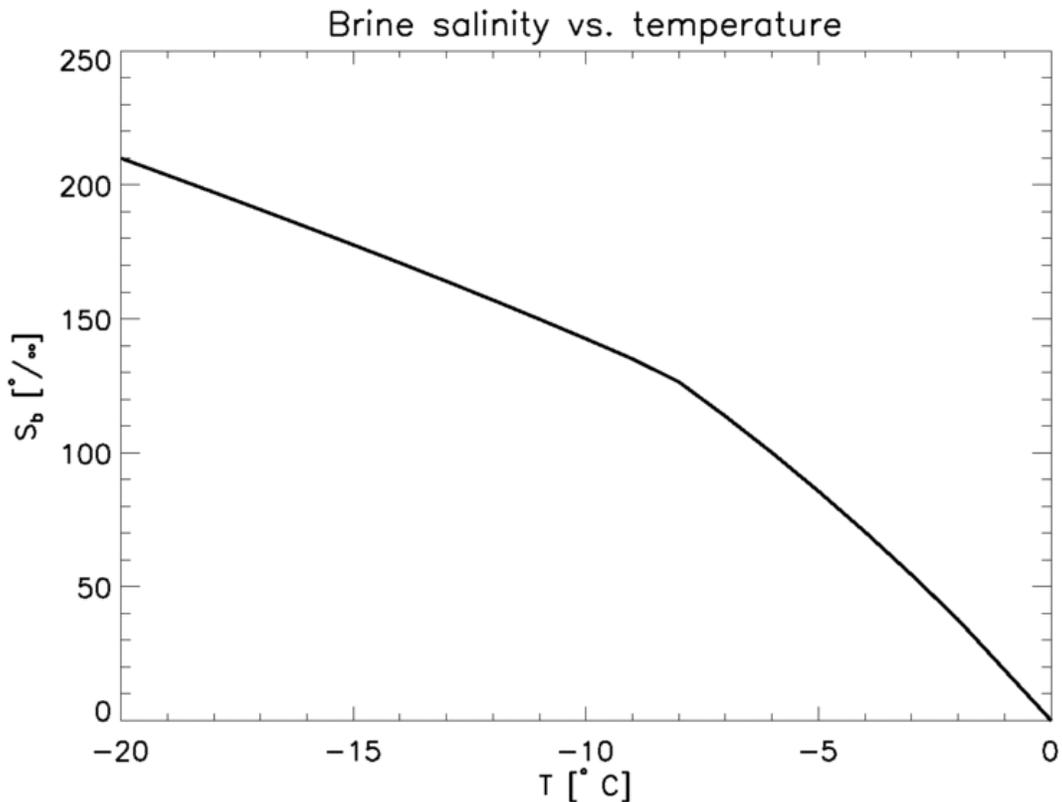
$$S = S_0 f(g)$$

where  $S$  is ice salinity,  $S_0$  is the salinity of the parent water and  $f$  is an empirical function of ice growth rate, e.g.:

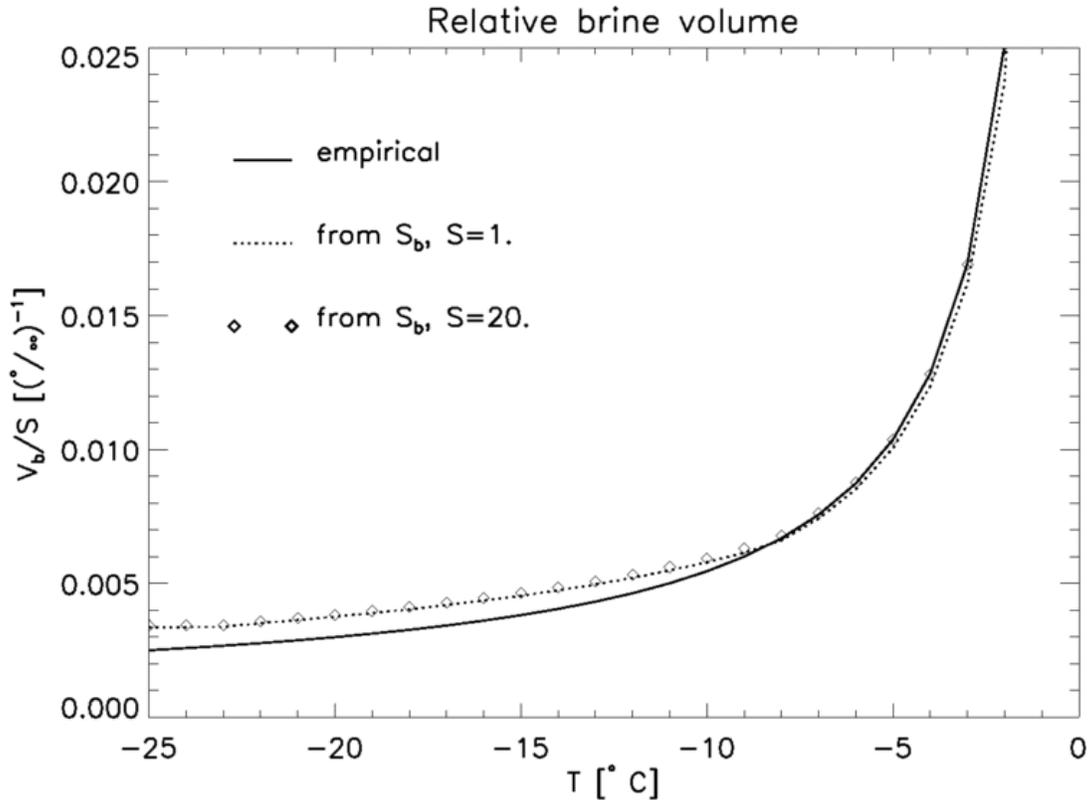
$$f(g) = \frac{0.12}{0.12 + 0.88 \exp(-4.2 \times 10^4 g)}$$

where  $g$  is in cm/s.

## Salt content



Brine salinity as a function of temperature



Ratio of brine volume to total salinity as a function of temperature

Brine entrapped in sea ice will always be at or near freezing since any departure will either cause some of the water in the brine to freeze, or melt some of the surrounding ice. Thus, brine salinity is variable and can be determined more or less exactly based strictly on temperature. References and contain empirical formulas relating sea ice temperature to brine salinity.

The relative brine volume,  $V_b$ , is defined as the fraction of brine relative to the total volume. It too is highly variable, however its value is more difficult to determine since changes in temperature may cause some of the brine to be ejected or move within the layers, particularly in new ice. Writing equations relating the salt content of the brine, the total salt content, the brine volume, the density of the brine and the density of the ice and solving for brine volume produces the following relation:

$$V_b = \frac{S \rho_i}{S_b \rho_b - S \rho_b + S \rho_i}$$

where  $S$  is sea ice salinity,  $S_b$  is brine salinity,  $\rho_i$  is the density of the ice and  $\rho_b$  is brine density. Compare with this empirical formula from Ulaby et al.:

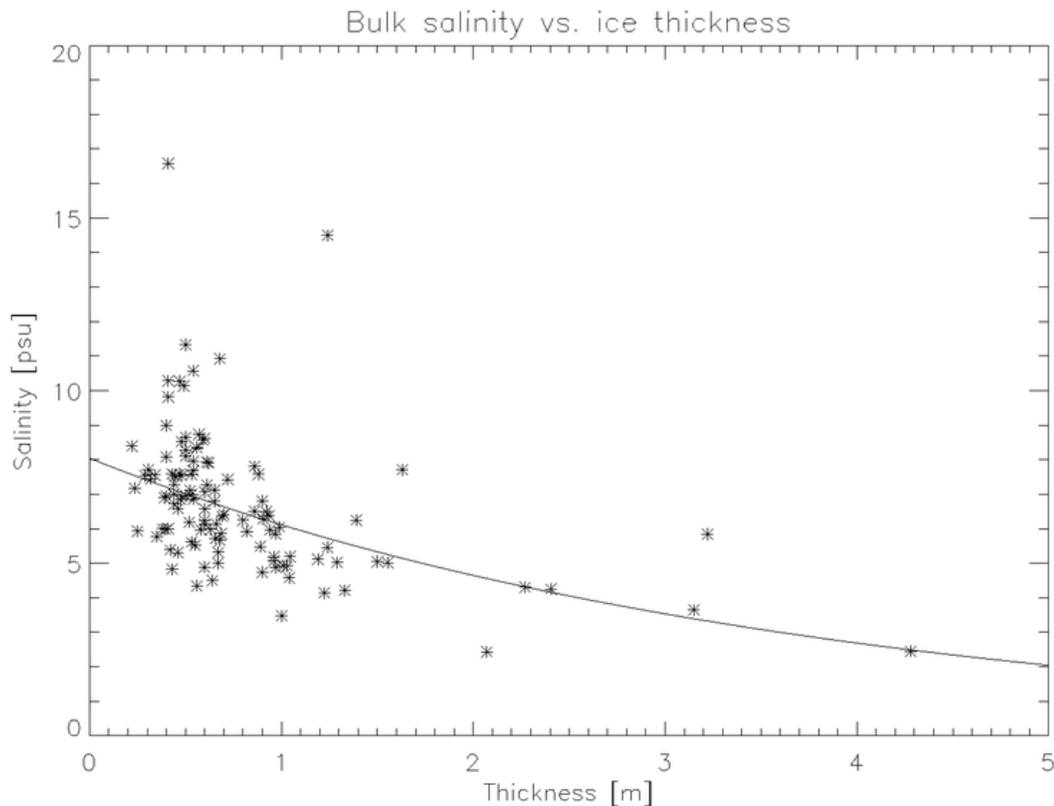
$$V_b = 10^{-3}S \left( -\frac{49.185}{T} + 0.532 \right)$$

where  $T$  is ice temperature in degrees Celsius and  $S$  is ice salinity in parts-per-thousand.

In new ice, the amount of brine ejected as the ice cools can be determined by assuming that the total volume stays constant and subtracting the volume increase from the brine volume. Note that this is only applicable to newly formed ice: any warming will tend to generate air pockets as the brine volume will increase more slowly than the ice volume decreases, again due to the density difference. Cox and Weeks provide the following formula determining the ratio of total ice salinity between temperatures,  $T_1$  and  $T_2$  where  $T_2 < T_1$ :

$$\frac{S(T_2)}{S(T_1)} = \frac{S_b(T_2) (1 - 1/\rho_i) \rho_b(T_2)}{S_b(T_1) \rho_b(T_1)} \exp \left\{ \frac{c}{\rho_i [S_b(T_1) - S_b(T_2)]} \right\}$$

where  $c=0.8 \text{ kg m}^{-3}$  is a constant. As the ice goes through constant warming and cooling cycles, it becomes progressively more porous through ejection of the brine and drainage through the resulting channels.



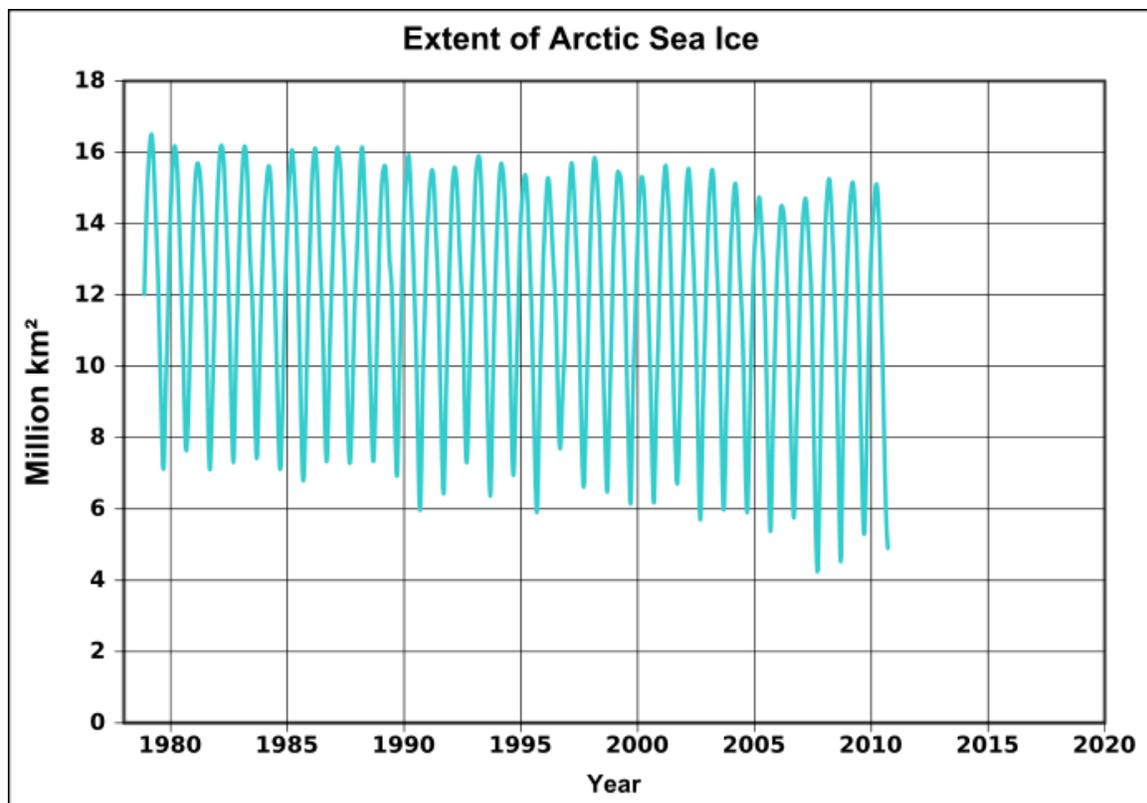
Plot of bulk salinity versus ice thickness for ice cores taken from the Weddell Sea.  
Courtesy Hajo Eicken

The figure above shows a scatter plot of salinity versus ice thickness for ice cores taken from the Weddell Sea, Antarctica, with an exponential fit of the form,  $S = \exp(ah + b)$ , overlaid, where  $h$  is ice thickness and  $a$  and  $b$  are constants.

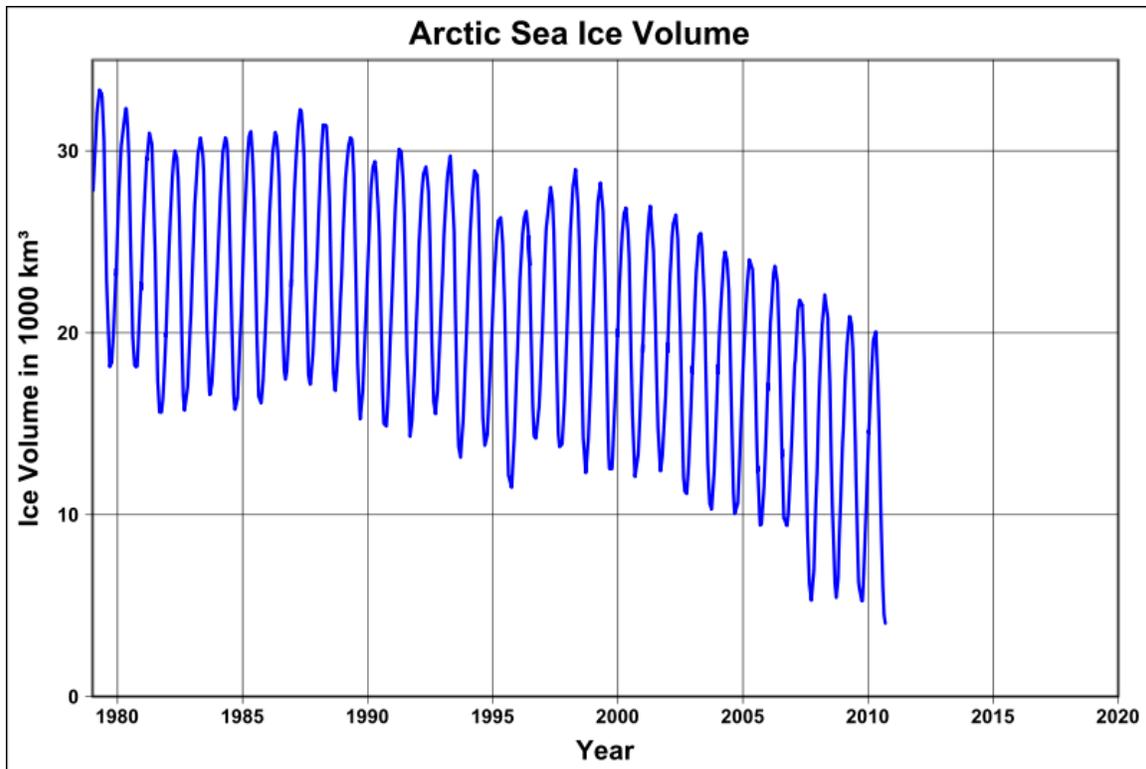
## Horizontal motion

The horizontal motion of sea ice is quite difficult to model because it is a non-Newtonian fluid. Sea ice will deform primarily at fracture points which in turn will form at the points of greatest stress and lowest strength, or where the ratio between the two is a maximum. Ice thickness, salinity and porosity will all affect the strength of the ice. The motion of the ice is driven primarily by ocean currents, though to a lesser extent by wind. Note that stresses will not be in the direction of the winds or currents, but rather will be shifted by coriolis effects—see, for instance, Ekman spiral.

## Yearly Freeze and Melt Cycle



Seasonal variation and long term decrease of arctic sea ice extent as determined by satellite measurements.



Seasonal variation and long term decrease of arctic sea ice volume as estimated by measurement backed numerical modelling.

Sea ice freezes and melts due to a combination of factors, including the age of the ice, air temperatures, and solar insolation. During the winter, the area of the Arctic ocean covered by sea ice increases, usually reaching a maximum extent during the month of March. As the seasons progress, the area covered in sea ice decreases, reaching its minimum extent in September most years. First-year ice melts more easily than older ice for two reasons: 1) First year ice is thinner than older ice, since the process of congelation growth has had less time to operate, and 2) first-year ice is less permeable than older ice, so summer meltwater tends to form deeper ponds on the first-year ice surface than on older ice, and deeper ponds mean lower albedo and thus greater solar energy capture.

### Monitoring and Observations

Satellite data has allowed the tracking of sea ice extents since 1978. During the warmest years, like the winter of 2005-2006, sea ice is observed to reach a winter maximum extent that is smaller than in the years before or after. The summer minimum Arctic ice extent for 2010 was the 3rd lowest over the period of satellite observations of the polar ice. The minimum record summer Arctic sea ice extent was in 2007.

## Ecology



ROV image of krill grazing under the ice.

Sea ice is part of the Earth's biosphere. Solid sea ice is permeated with channels filled with salty brine. These briny channels and the sea ice itself have its ecology, referred to as "sympagic ecology".

The decline of seasonal sea ice is putting the survival of Arctic species such as ringed seals and polar bears at risk.

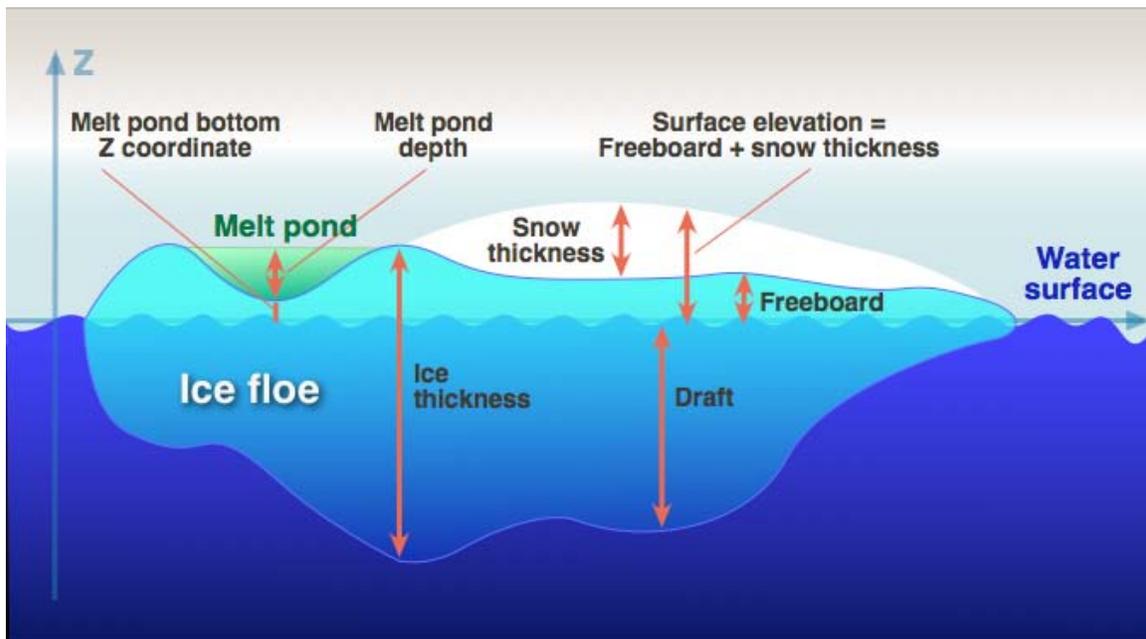
## Gallery



In the North Atlantic.



An icebreaker navigates through young (1 year old) sea ice.



Parameter to measure the size of a sea ice floe.

Sea ice covers much of the polar oceans and forms by freezing of sea water. Satellite data since the early 1970s reveal considerable seasonal, regional, and interannual variability in the sea-ice covers of both hemispheres. Seasonally, sea-ice extent in the Southern Hemisphere varies by a factor of 5, from a minimum of 3-4 million km<sup>2</sup> in February to a maximum of 17-20 million km<sup>2</sup> in September. The seasonal variation is much less in the Northern Hemisphere where the confined nature and high latitudes of the Arctic Ocean result in a much larger perennial ice cover, and the surrounding land limits the equatorward extent of wintertime ice. Thus, the seasonal variability in Northern Hemisphere ice extent varies by only a factor of 2, from a minimum of 7-9 million km<sup>2</sup> in September to a maximum of 14-16 million km<sup>2</sup> in March.

The ice cover exhibits much greater regional-scale interannual variability than it does hemispherical. For instance, in the region of the Seas of Okhotsk and Japan, maximum

ice extent decreased from 1.3 million km<sup>2</sup> in 1983 to 0.85 million km<sup>2</sup> in 1984, a decrease of 35%, before rebounding the following year to 1.2 million km<sup>2</sup>. The regional fluctuations in both hemispheres are such that for any several-year period of the satellite record some regions exhibit decreasing ice coverage while others exhibit increasing ice cover. The overall trend indicated in the passive microwave record from 1978 through mid-1995 shows that the extent of Arctic sea ice is decreasing 2.7% per decade. Subsequent work with the satellite passive-microwave data indicates that from late October 1978 through the end of 1996 the extent of Arctic sea ice decreased by 2.9% per decade while the extent of Antarctic sea ice increased by 1.3% per decade.

## Chapter-3

# Glacier



The Baltoro Glacier in the Karakoram, Kashmir. At 62 kilometres (39 mi) in length, it is one of the longest alpine glaciers on earth.



Ice calving from the terminus of the Perito Moreno Glacier, in western Patagonia, Argentina



The Aletsch Glacier, the largest glacier of the Alps, in Switzerland



Icebergs calved from outlet glaciers at Cape York, Greenland

A **glacier** is a large persistent body of ice. Originating on land, a glacier flows slowly due to stresses induced by its weight. The crevasses and other distinguishing features of a glacier are due to its flow. Another consequence of glacier flow is the transport of rock and debris abraded from its substrate and resultant landforms like cirques and moraines. A glacier forms in a location where the accumulation of snow and sleet exceeds its ablation (melting and sublimation) over many years, often decades or centuries. A glacier is distinct from sea ice and lake ice that form on the surface of bodies of water.

The word *glacier* comes from French. It is derived from the Vulgar Latin *\*glacia* and ultimately from Latin *glacies* meaning *ice*. The processes and features caused by glaciers and related to them are referred to as **glacial**. The process of glacier establishment, growth and flow is called **glaciation**. The corresponding area of study is called glaciology. Glaciers are important components of the global cryosphere.

On Earth, 99% glacial ice is contained within vast ice sheets in polar regions, but glaciers may be found in mountain ranges of every continent except Australia. In the tropics, glaciers occur only on high mountains.

Glacial ice is the largest reservoir of freshwater on Earth. Many glaciers store water during one season and release it later as meltwater, a water source that is especially important for plants, animals and human uses when other sources may be scant.

Because glacial mass is affected by long-term climate changes, e.g., precipitation, mean temperature, and cloud cover, glacial mass changes are considered among the most sensitive indicators of climate change and are a major source of variations in sea level.

## Types of glaciers



Mouth of the Schlatenkees Glacier near Innerschlöß, Austria

Glaciers are categorized in many ways including by their morphology, thermal characteristics or their behavior. **Alpine glaciers** form on the crests and slopes of mountains and are also known as "mountain glaciers", "niche glaciers", or "cirque glaciers". An alpine glacier that fills a valley is sometimes called a **valley glacier**. Larger glaciers that cover an entire mountain, mountain range, or volcano are known as an ice cap or ice field, such as the Juneau Icefield. Ice caps feed **outlet glaciers**, tongues of ice that extend into valleys below far from the margins of the larger ice masses.

The largest glacial bodies, ice sheets or continental glaciers, cover more than 50,000 km<sup>2</sup> (20,000 mile<sup>2</sup>). Several kilometers deep, they obscure the underlying topography. Only nunataks protrude from the surface. The only extant ice sheets are the two that cover most of Antarctica and Greenland. These regions contain vast quantities of fresh water. The volume of ice is so large that if the Greenland ice sheet melted, it would cause sea levels

to rise six meters (20 ft) all around the world. If the Antarctic ice sheet melted, sea levels would rise up to 65 meters (210 ft). **Ice shelves** are areas of floating ice, commonly located at the margin of an ice sheet. As a result they are thinner and have limited slopes and reduced velocities. **Ice streams** are fast-moving sections of an ice sheet. They can be several hundred kilometers long. Ice streams have narrow margins and on either side ice flow is usually an order of magnitude less. In Antarctica, many ice streams drain into large ice shelves. However, some drain directly into the sea, often with an ice tongue, like Mertz Glacier. In Greenland and Antarctica ice streams ending at the sea are often referred to as tidewater glaciers or outlet glaciers, such as Jakobshavn Isbræ (Kalaallisut: *Sermeq Kujalleq*).



Sightseeing boat in front of a tidewater glacier, Kenai Fjords National Park, Alaska

**Tidewater glaciers** are glaciers that terminate in the sea. As the ice reaches the sea pieces break off, or *calve*, forming icebergs. Most tidewater glaciers calve above sea level, which often results in a tremendous splash as the iceberg strikes the water. If the water is deep, glaciers can calve underwater, causing the iceberg to suddenly leap up out of the water. The Hubbard Glacier is the longest tidewater glacier in Alaska and has a calving face over 10 km (6 mi) long. Yakutat Bay and Glacier Bay are both popular with cruise ship passengers because of the huge glaciers descending hundreds of feet to the water. This glacier type undergoes centuries-long cycles of advance and retreat that are much less affected by the climate changes currently causing the retreat of most other glaciers. Most tidewater glaciers are outlet glaciers of ice caps and ice fields.

In terms of thermal characteristics, a *temperate* glacier is at melting point throughout the year, from its surface to its base. The ice of a *polar* glacier is always below freezing point from the surface to its base, although the surface snowpack may experience seasonal melting. A *sub-polar* glacier has both temperate and polar ice, depending on the depth beneath the surface and position along the length of the glacier.

## Formation

Glaciers form where the accumulation of snow and ice exceeds ablation. As the snow and ice thicken, they reach a point where they begin to move, due to a combination of the surface slope and the pressure of the overlying snow and ice. On steeper slopes this can occur with as little as 15 m (50 ft) of snow-ice. The snow which forms temperate glaciers is subject to repeated freezing and thawing, which changes it into a form of granular ice called firn. Under the pressure of the layers of ice and snow above it, this granular ice fuses into denser and denser firn. Over a period of years, layers of firn undergo further compaction and become glacial ice. Glacier ice has a slightly reduced density from ice formed from the direct freezing of water. The air between snowflakes becomes trapped and creates air bubbles between the ice crystals.

The distinctive blue tint of glacial ice is often wrongly attributed to Rayleigh scattering due to bubbles in the ice. The blue color is actually created for the same reason that water is blue, that is, its slight absorption of red light due to an overtone of the infrared OH stretching mode of the water molecule.

## Anatomy

The location where a glacier originates is referred to as the "glacier head". A glacier terminates at the "glacier foot", or terminus. Glaciers are broken into zones based on surface snowpack and melt conditions. The ablation zone is the region where there is a net loss in glacier mass. The equilibrium line separates the ablation zone and the accumulation zone. At this altitude, the amount of new snow gained by accumulation is equal to the amount of ice lost through ablation. The accumulation zone is the region where snowpack or superimposed ice accumulation persists.

A further zonation of the accumulation zone distinguishes the melt conditions that exist.

- The dry snow zone is a region where no melt occurs, even in the summer, and the snowpack remains dry.
- The percolation zone is an area with some surface melt, causing meltwater to percolate into the snowpack. This zone is often marked by refrozen ice lenses, glands, and layers. The snowpack also never reaches melting point.
- Near the equilibrium line on some glaciers, a superimposed ice zone develops. This zone is where meltwater refreezes as a cold layer in the glacier, forming a continuous mass of ice.

- The wet snow zone is the region where all of the snow deposited since the end of the previous summer has been raised to 0 °C.

The upper part of a glacier that receives most of the snowfall is called the *accumulation zone*. In general, the glacier accumulation zone accounts for 60-70% of the glacier's surface area, more if the glacier calves icebergs. The depth of ice in the accumulation zone exerts a downward force sufficient to cause deep erosion of the rock in this area. After the glacier is gone, its force often leaves a bowl or amphitheater-shaped isostatic depression ranging from large lake basins, such as the Great Lakes or Finger Lakes, to smaller mountain basins, known as *cirques*.

The "health" of a glacier is usually assessed by determining the glacier mass balance or observing terminus behavior. Healthy glaciers have large accumulation zones, more than 60% of their area snowcovered at the end of the melt season, and a terminus with vigorous flow.

Following the Little Ice Age, around 1850, the glaciers of the Earth have retreated substantially through the 1940s. A slight cooling led to the advance of many alpine glaciers from 1950-1985. However, since 1985 glacier retreat and mass balance loss has become increasingly ubiquitous and large.

## Motion



The Nadelhorn Glacier above Saas-Fee, Valais, Switzerland

Glaciers move, or flow, downhill due to the internal deformation of ice and gravity. Ice behaves like an easily breaking solid until its thickness exceeds about 50 meters (160 ft). The pressure on ice deeper than that depth causes plastic flow. At the molecular level, ice consists of stacked layers of molecules with relatively weak bonds between the layers. When the stress of the layer above exceeds the inter-layer binding strength, it moves faster than the layer below.

Another type of movement is through basal sliding. In this process, the glacier slides over the terrain on which it sits, lubricated by the presence of liquid water. As the pressure increases toward the base of the glacier, the melting point of water decreases, and the ice melts. Friction between ice and rock and geothermal heat from the Earth's interior also contribute to melting. This type of movement is dominant in temperate, or warm-based glaciers. The geothermal heat flux becomes more important the thicker a glacier becomes.

The rate of movement is dependent on the underlying slope, amongst many other factors.

### **Fracture zone and cracks**



Ice cracks in the Titlis Glacier



Signs warning of the hazards of a glacier in New Zealand

The top 50 meters of the glacier, being under less pressure, are more rigid; this section is known as the *fracture zone*, and mostly moves as a single unit, over the plastic-like flow of the lower section. When the glacier moves through irregular terrain, cracks up to 50 meters deep form in the fracture zone. The lower layers of glacial ice flow and deform plastically under the pressure, allowing the glacier as a whole to move slowly like a viscous fluid. Glaciers flow downslope, usually this reflects the slope of their base, but it may reflect the surface slope instead. Thus, a glacier can flow rises in terrain at their base. The upper layers of glaciers are more brittle, and often form deep cracks known as crevasses. The presence of crevasses is a sure sign of a glacier. Moving ice-snow of a glacier is often separated from a mountain side or snow-ice that is stationary and clinging to that mountain side by a *bergschrund*. This looks like a crevasse but is at the margin of the glacier and is a singular feature.

Crevasses form due to differences in glacier velocity. As the parts move at different speeds and directions, shear forces cause the two sections to break apart, opening the crack of a crevasse all along the disconnecting faces. Hence, the distance between the two separated parts, while touching and rubbing deep down, frequently widens significantly towards the surface layers, many times creating a wide chasm. Crevasses seldom are more than 150 feet (46 m) deep but in some cases can be 1,000 feet (300 m) or even deeper. Beneath this point, the plastic deformation of the ice under pressure is too great

for the differential motion to generate cracks. Transverse crevasses are transverse to flow, as a glacier accelerates where the slope steepens. Longitudinal crevasses form semi-parallel to flow where a glacier expands laterally. Marginal crevasses form from the edge of the glacier, due to the reduction in speed caused by friction of the valley walls. Marginal crevasses are usually largely transverse to flow.



Crossing a crevasse on the Easton Glacier, Mount Baker, in the North Cascades, USA

Crevasses make travel over glaciers hazardous. Subsequent heavy snow may form fragile snow bridges, increasing the danger by hiding the presence of crevasses at the surface. Below the equilibrium line, glacier meltwater is concentrated in stream channels. The meltwater can pool in a proglacial lake, a lake on top of the glacier, or can descend into the depths of the glacier via *moulins*. Within or beneath the glacier, the stream will flow in an englacial or sub-glacial tunnel. Sometimes these tunnels reemerge at the surface of the glacier.

## Speed

The speed of glacial displacement is partly determined by friction. Friction makes the ice at the bottom of the glacier move more slowly than the upper portion. In alpine glaciers, friction is also generated at the valley's side walls, which slows the edges relative to the center. This was confirmed by experiments in the 19th century, in which stakes were planted in a line across an alpine glacier, and as time passed, those in the center moved farther.

Mean speeds vary greatly. There may be no motion in stagnant areas, where trees can establish themselves on surface sediment deposits such as in Alaska. In other cases they can move as fast as 20–30 meters per day, as in the case of Greenland's Jakobshavn

Isbræ (Kalaallisut: *Sermeq Kujalleq*), or 2–3 m per day on Byrd Glacier, the largest glacier in the world in Antarctica. Velocity increases with increasing slope, increasing thickness, increasing snowfall, increasing longitudinal confinement, increasing basal temperature, increasing meltwater production and reduced bed hardness.

A few glaciers have periods of very rapid advancement called surges. These glaciers exhibit normal movement until suddenly they accelerate, then return to their previous state. During these surges, the glacier may reach velocities far greater than normal speed. These surges may be caused by failure of the underlying bedrock, the ponding of meltwater at the base of the glacier — perhaps delivered from a supraglacial lake — or the simple accumulation of mass beyond a critical "tipping point".

In glaciated areas where the glacier moves faster than one kilometer per year, glacial earthquakes occur. These are large scale tremblors that have seismic magnitudes as high as 6.1.

The number of glacial earthquakes in Greenland show a peak every year in July, August and September, and the number is increasing over time. In a study using data from January 1993 through October 2005, more events were detected every year since 2002, and twice as many events were recorded in 2005 as there were in any other year. This increase in the numbers of glacial earthquakes in Greenland may be a response to global warming.

Seismic waves are also generated by the Whillans Ice Stream, a large, fast-moving river of ice pouring from the West Antarctic Ice Sheet into the Ross Ice Shelf. Two bursts of seismic waves are released every day, each one equivalent to a magnitude 7 earthquake, and are seemingly related to the tidal action of the Ross Sea. During each event a 96 by 193 kilometer (60 by 120 mile) region of the glacier moves as much as .67 meters (2.2 ft) over about 25 minutes, remains still for 12 hours, then moves another half-meter. The seismic waves are recorded at seismographs around Antarctica, and even as far away as Australia, a distance of more than 6,400 kilometers. Because the motion takes place of such along period of time 10 to 25 minutes, it cannot be felt by scientists standing on the moving glacier. It is not known if these events are related to global warming

## **Ogives**

*Ogives* are alternating dark and light bands of ice occurring as narrow wave crests and wave valleys on glacier surfaces. They only occur below icefalls, but not all icefalls have ogives below them. Once formed, they bend progressively downglacier due to the increased velocity toward the glacier's centerline. Ogives are linked to seasonal motion of the glacier as the width of one dark and one light band generally equals the annual movement of the glacier. The ridges and valleys are formed because ice from an icefall is severely broken up, thereby increasing ablation surface area during the summertime. This creates a swale and space for snow accumulation in the winter, which in turn creates a ridge. Sometimes ogives are described as either wave ogives or band ogives, in which they are solely undulations or varying color bands, respectively.

## Geography



Black ice glacier near Aconcagua, Argentina

Glaciers occur on every continent and approximately 47 countries. Extensive glaciers are found in Antarctica, Chilean Patagonia, Canada, Alaska, Greenland and Iceland. Mountain glaciers are widespread, e.g., in the Andes, the Himalaya, the Rocky Mountains, the Caucasus, and the Alps. On mainland Australia no glaciers exist today, although a small glacier on Mount Kosciuszko was present in the last glacial period, and Tasmania was extensively glaciated. The South Island of New Zealand has many glaciers including Tasman, Fox and Franz Josef Glaciers. In New Guinea, small, rapidly diminishing, glaciers are located on its highest summit massif of Puncak Jaya. Africa has glaciers on Mount Kilimanjaro in Tanzania, on Mount Kenya and in the Ruwenzori Range.

Permanent snow cover is affected by factors such as the degree of slope on the land, amount of snowfall and the winds. As temperature decreases with altitude, high mountains — even those near the Equator — have permanent snow cover on their upper portions, above the snow line. Examples include Mount Kilimanjaro and the Tropical Andes in South America; however, the only snow to occur exactly on the Equator is at 4,690 m (15,387 ft) on the southern slope of Volcán Cayambe in Ecuador.

Conversely, areas of the Arctic, such as Banks Island, and the McMurdo Dry Valleys in Antarctica are considered polar deserts, as they receive little snowfall despite the bitter cold. Cold air, unlike warm air, is unable to transport much water vapor. Even during glacial periods of the Quaternary, Manchuria, lowland Siberia, and central and northern Alaska, though extraordinarily cold with winter temperatures believed to reach  $-100\text{ }^{\circ}\text{C}$  ( $-148\text{ }^{\circ}\text{F}$ ) in parts, had such light snowfall that glaciers could not form.

In addition to the dry, unglaciated polar regions, some mountains and volcanoes in Bolivia, Chile and Argentina are high (4,500 metres (14,800 ft) - 6,900 m (22,600 ft)) and cold, but the relative lack of precipitation prevents snow from accumulating into glaciers. This is because these peaks are located near or in the hyperarid Atacama desert.

## Glacial geology

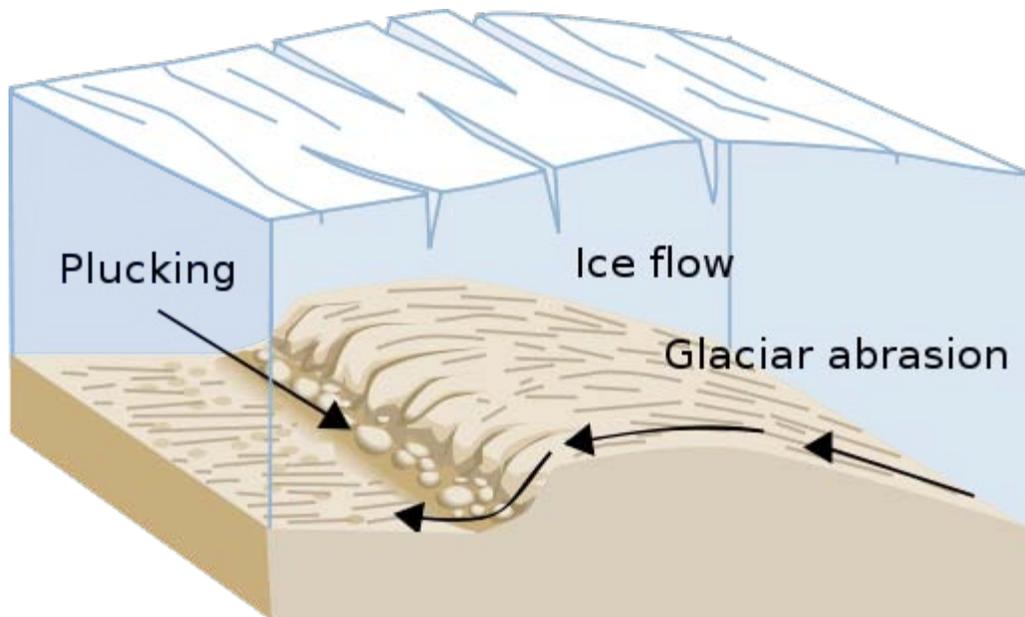


Diagram of glacial plucking and abrasion



Glacially plucked granitic bedrock near Mariehamn, Åland Islands

Rocks and sediments are added to glaciers through various processes. Glaciers erode the terrain principally through two methods: **abrasion** and **plucking**.

As the glacier flows over the bedrock's fractured surface, it softens and lifts blocks of rock that are brought into the ice. This process is known as plucking, and it is produced when subglacial water penetrates the fractures and the subsequent freezing expansion separates them from the bedrock. When the ice expands, it acts as a lever that loosens the rock by lifting it. This way, sediments of all sizes become part of the glacier's load. The rocks frozen into the bottom of the ice then act like grit in sandpaper.

Abrasion occurs when the ice and the load of rock fragments slide over the bedrock and function as sandpaper that smooths and polishes the surface situated below. This pulverized rock is called rock flour. The flour is formed by rock grains of a size between 0.002 and 0.00625 mm. Sometimes the amount of rock flour produced is so high that currents of meltwaters acquire a grayish color. These processes of erosion lead to steeper valley walls and mountain slopes in alpine settings, which can cause avalanches and rock slides. These further add material to the glacier.

Visible characteristics of glacial abrasion are glacial striations. These are produced when the bottom's ice contains large chunks of rock that mark scratches in the bedrock. By

mapping the direction of the flutes, researchers can determine the direction of the glacier's movement. Chatter marks are seen as lines of roughly crescent-shape depressions in the rock underlying a glacier, caused by the abrasion where a boulder in the ice catches and is then released repetitively as the glacier drags it over the underlying basal rock.

The rate of glacier erosion is variable. The differential erosion undertaken by the ice is controlled by six important factors:

- Velocity of glacial movement;
- Thickness of the ice;
- Shape, abundance and hardness of rock fragments contained in the ice at the bottom of the glacier;
- Relative ease of erosion of the surface under the glacier;
- Thermal conditions at the glacier base; and
- Permeability and water pressure at the glacier base.

Material that becomes incorporated in a glacier are typically carried as far as the zone of ablation before being deposited. Glacial deposits are of two distinct types:

- Glacial till: material directly deposited from glacial ice. Till includes a mixture of undifferentiated material ranging from clay size to boulders, the usual composition of a moraine.
- Fluvial and outwash: sediments deposited by water. These deposits are stratified through various processes, such as boulders' being separated from finer particles.

The larger pieces of rock which are encrusted in till or deposited on the surface are called "glacial erratics". They may range in size from pebbles to boulders, but as they may be moved great distances, they may be of drastically different type than the material upon which they are found. Patterns of glacial erratics provide clues of past glacial motions.

## **Moraines**

A **moraine** is any glacially formed accumulation of unconsolidated glacial debris (soil and rock) which can occur in currently glaciated and formerly glaciated regions, such as those areas acted upon by a past ice age. This debris may have been plucked off the valley floor as a glacier advanced or it may have fallen off the valley walls as a result of frost wedging or landslide. Moraines may be composed of debris ranging in size from silt-sized glacial flour to large boulders. The debris is typically sub-angular to rounded in shape. Moraines may be on the glacier's surface or deposited as piles or sheets of debris where the glacier has melted. Moraines may also occur when glacier- or iceberg-transported rocks fall into a body of water as the ice melts.



Lateral moraines of the reducing glacier in Engadin

# Types of moraines

## Lateral moraines



Lateral moraines above Lake Louise, Alberta, Canada.

Lateral moraines are parallel ridges of debris deposited along the sides of a glacier. The unconsolidated debris can be deposited on top of the glacier by frost shattering of the valley walls and from tributary streams flowing into the valley. The till is carried along the glacial margin until the glacier melts. Because lateral moraines are deposited on top of the glacier, they do not experience the postglacial erosion of the valley floor and therefore, as the glacier melts, lateral moraines are usually preserved as high ridges.



Moraines clearly seen on a side glacier of the Gorner Glacier, Zermatt, Switzerland. The **lateral moraine** is the high snow-free bank of debris in the top left hand quarter of the picture. The **medial moraine** is the double line of debris running down the centre-line of the glacier.

Lateral moraines stand high because they protect the ice under them from the elements, which causes it to melt or sublime less than the uncovered parts of the glacier. Multiple lateral moraines may develop as the glacier advances and retreats.

### **Ground moraines**



Ground moraine makes an irregular, rolling topography

Ground moraines are till covered areas with irregular topography and no ridges, often forming gently rolling hills or plains. It is accumulated at the base of the ice as lodgement till, but may also be deposited as the glacier retreats. In alpine glaciers, ground moraines are often located between the two lateral moraines. Ground moraine may be modified into drumlins by the overriding ice.

### **Rogen moraines**

Rogen moraines or Ribbed moraines are a type of basal moraines that forms a series of ribs perpendicular to the ice flow in an ice sheet. The depressions between the ribs are sometimes filled with water making the Rogen moraines look like tigerstripes on aerial photographs. Rogen moraines are named after Lake Rogen in Härjedalen, Sweden, the landform's type locality.



Medial moraines, Nuussuaq Peninsula, Greenland

### **End or terminal moraines**

End moraines, or terminal moraines, are ridges of unconsolidated debris deposited at the snout or end of the glacier. They usually reflect the shape of the glacier's terminus. Glaciers act much like a conveyor belt, carrying debris from the top of the glacier to the bottom where it deposits it in end moraines. End moraine size and shape is determined by whether the glacier is advancing, receding or at equilibrium. The longer the terminus of the glacier stays in one place the more debris will accumulate in the moraine. There are two types of end moraines; terminal and recessional. Terminal moraines mark the maximum advance of the glacier. Recessional moraines are small ridges left as a glacier pauses during its retreat. After a glacier retreats the end moraine may be destroyed by postglacial erosion.

### **Recessional moraine**

Recessional moraines are often observed as a series of transverse ridges running across a valley behind a terminal moraine. They form perpendicular to the lateral moraines that they reside between and are composed of unconsolidated debris deposited by the glacier. They are created during temporary halts in a glacier's retreat.

### **Medial moraine**

A medial moraine is a ridge of moraine that runs down the center of a valley floor. It is formed when two glaciers meet and the debris on the edges of the adjacent valley sides join and are carried on top of the enlarged glacier. As the glacier melts or retreats, the

debris is deposited and a ridge down the middle of the valley floor is created. The Kaskawulsh glacier in the Kluane National Park, Canada has a ridge of medial moraine 1 km wide.



The prominent dark streak at the left quarter is forming a medial moraine. This is seen as a mudflat at the water's surface. (Brüggen Glacier, Patagonia, Chile)

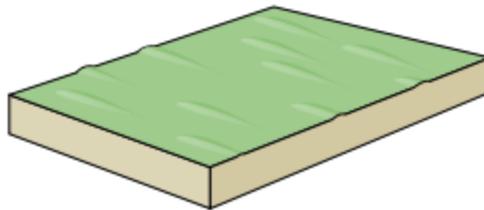
### **Supraglacial moraines**

Supraglacial moraines are created by debris accumulated on top of glacial ice. This debris can accumulate due to ice flow toward the surface in the ablation zone, melting of surface ice or from debris that falls onto the glacier from valley sidewalls.

### **Veiki moraine**

A Veiki moraine is a kind of hummocky moraine that forms irregular landscapes of ponds and plateaus surrounded by banks. It is formed by the irregular melting of an ice covered with a thick layer of debris. Veiki moraine is common in northern Sweden and parts of Canada.

### **Drumlins**



A drumlin field forms after a glacier has modified the landscape. The teardrop-shaped formations denote the direction of the ice flow.



Drowned drumlin in Clew Bay, County Mayo



Drumlin at Withrow Moraine and Jameson Lake Drumlin Field National Natural Landmark, Washington state



Drumlin in Cato, New York



*Bürglen* drumlin in Irgenhausen, Switzerland, where the Roman Irgenhausen Castrum is situated

A **drumlin** – from the Gaelic word *droimnín* ("little ridge"), first recorded in 1833 – is an elongated whale-shaped hill formed by glacial ice acting on underlying unconsolidated till or ground moraine. Its long axis is parallel with the movement of the ice, with the blunter end facing into the glacial movement. Drumlins are typically 1 to 2 km (0.6 to 1.2 mi) long, less than 50 m (165 ft) high and between 300 to 600 m (~0.25 mi) wide. Drumlins generally have a consistent ratio of 2:3.5 width to length dimensions. Drumlins are often in drumlin fields of similarly shaped, sized and oriented hills. Drumlins usually have layers indicating that the material was repeatedly added to a core, which may be of rock or glacial till. The composition of drumlins varies depending on the area in which they are found, and can consist of similar material to the till of the surrounding moraine or be comprised almost entirely of bedrock, sand and gravel or various mixtures thereof.

Drumlins are common in New York, the lower Connecticut River valley, eastern Massachusetts, the Monadnock Region of New Hampshire, Minnesota, Wisconsin, Alberta, Peterborough, Ontario, Southern Ontario, Nova Scotia, Poland, Estonia, around Lake Constance north of the Alps, County Monaghan, County Mayo, County Cavan and County Fermanagh in the northern provinces of Ireland, Greenland, Hindsholm in Denmark, Finland and Patagonia. Those in North America are regarded as a creation of the last Wisconsin ice age.

The islands of Boston Harbor Islands National Recreation Area are drumlins that became islands when sea levels rose as the glaciers melted. Clew Bay in Ireland is a good example of a 'drowned drumlin' landscape where the drumlins appear as islands in the sea, forming a 'basket of eggs' topography. Drumlins are typically aligned parallel to one another, usually clustered together in numbers reaching the hundreds or even thousands. These clusters can sometimes lead to the natural emergence and growth of complex water systems. In County Cavan, Ireland, there is a unique mesh of streams and rivers which feed into and out of 365 lakes in between drumlins; one lake for each day of the year.

Drumlin formation has recently been observed for the first time in Antarctica in the Rutford Ice Stream.

Drumlin soil classification is variable but often consists of a thin A soil horizon and a thin Bw horizon. The C horizon is close to the surface, and may be at the surface on an eroded drumlin.

## **Drumlin formation**

There are many theories as to the exact mode of origin and plenty of controversy among geologists interested in geomorphology. Some consider them a direct formation of the ice, while a theory proposed since the 1980s by John Shaw and others postulates creation by a catastrophic flooding release of highly pressurized water flowing underneath the glacial ice. Either way, they are thought to be a waveform (similar to ripples of sand at the bottom of a stream). It is not clearly understood whether drumlins are erosional or depositional features. It is also poorly understood why drumlins form in some glaciated areas and not in others. They are often associated with ribbed moraines.

## Glacial valleys



A glacial valley in the Mount Baker-Snoqualmie National Forest, showing the characteristic U-shape and flat bottom



Yosemite Valley from an airplane, showing the U-shape



This image shows the termini of the glaciers in the Bhutan Himalaya. Glacial lakes have been rapidly forming on the surface of the debris-covered glaciers in this region during the last few decades.

Before glaciation, mountain valleys have a characteristic "V" shape, produced by downward erosion by water. However, during glaciation, these valleys widen and deepen, forming a "U"-shaped glacial valley. Besides the deepening and widening of the valley, the glacier also smooths the valley due to erosion. In this way, it eliminates the spurs of earth that extend across the valley. Because of this interaction, triangular cliffs called truncated spurs are formed.

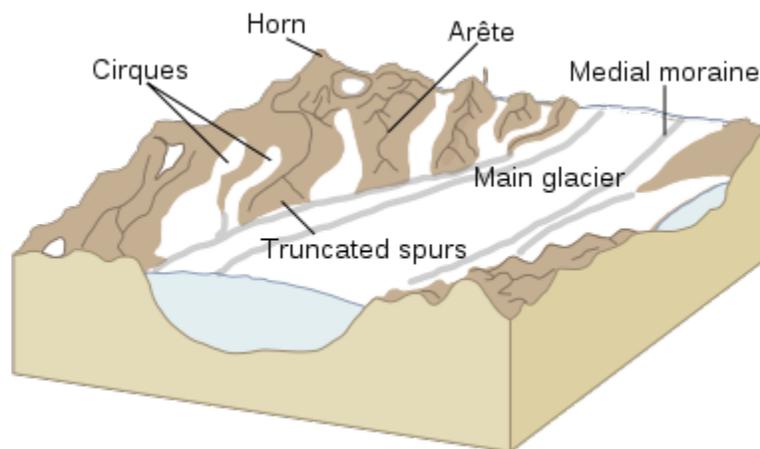
Many glaciers deepen their valleys more than their smaller tributaries. Therefore, when the glaciers recede from the region, the valleys of the tributary glaciers remain above the main glacier's depression, and these are called hanging valleys.

In parts of the soil that were affected by abrasion and plucking, the depressions left can be filled by lakes, called paternoster lakes.

At the 'start' of a classic valley glacier is the cirque, which has a bowl shape with escarped walls on three sides, but open on the side that descends into the valley. In the cirque, an accumulation of ice is formed. These begin as irregularities on the side of the mountain, which are later augmented in size by the coining of the ice. Once the glacier melts, these corries are usually occupied by small mountain lakes called tarns.

There may be two glacial cirques 'back to back' which erode deep into their backwalls until only a narrow ridge, called an arête is left. This structure may result in a mountain pass.

Glaciers are also responsible for the creation of fjords (deep coves or inlets) and escarpments that are found at high latitudes.



Features of a glacial landscape

### **Arêtes and horns (pyramid peak)**

An arête is a narrow crest with a sharp edge. The meeting of three or more arêtes creates pointed pyramidal peaks and in extremely steep-sided forms these are called horns.

Both features may have the same process behind their formation: the enlargement of cirques from glacial plucking and the action of the ice. Horns are formed by cirques that encircle a single mountain.

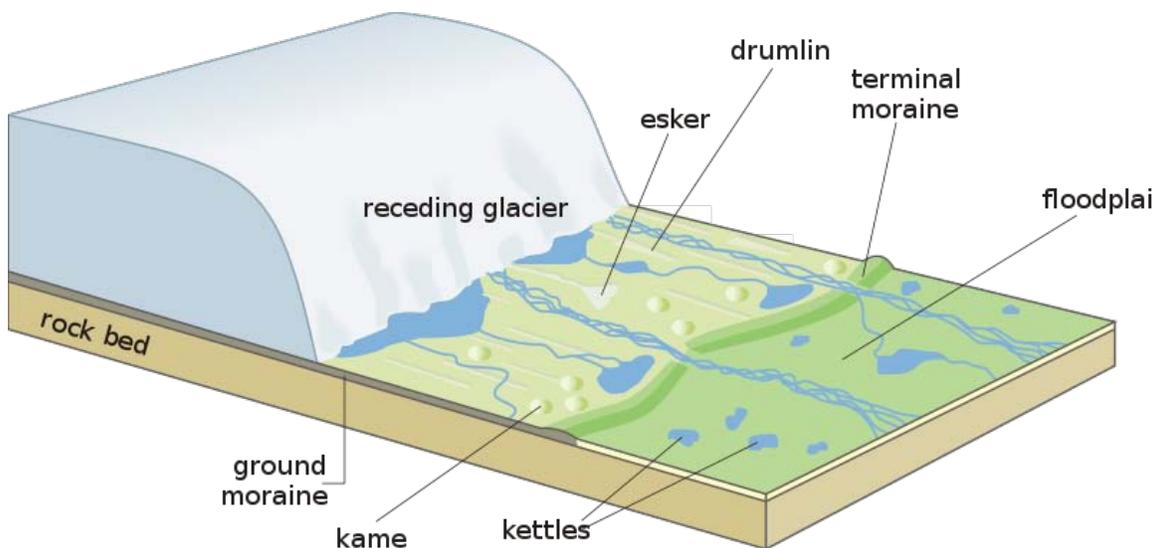
Arêtes emerge in a similar manner; the only difference is that the cirques are not located in a circle, but rather on opposite sides along a divide. Arêtes can also be produced by the collision of two parallel glaciers. In this case, the glacial tongues cut the divides down to size through erosion, and polish the adjacent valleys.

### **Roche moutonnée**

Some rock formations in the path of a glacier are sculpted into small hills with a shape known as *roche moutonnée* or "sheepback" rock. An elongated, rounded, asymmetrical, bedrock knob can be produced by glacier erosion. It has a gentle slope on its up-glacier side and a steep to vertical face on the down-glacier side. The glacier abrades the smooth slope that it flows along, while rock is torn loose from the downstream side and carried away in ice, a process known as 'plucking'. Rock on this side is fractured by a combination of various forces, such as water, ice in rock cracks, and structural stresses.

### **Alluvial stratification**

The water that rises from the ablation zone moves away from the glacier and carries with it fine eroded sediments. As the speed of the water decreases, so does its capacity to carry objects in suspension. The water then gradually deposits the sediment as it runs, creating an alluvial plain. When this phenomenon occurs in a valley, it is called a *valley train*. When the deposition is to an estuary, the sediments are known as "bay mud".



Landscape produced by a receding glacier

Outwash plains and valley trains are usually accompanied by basins known as "kettles". These are glacial depressions produced when large ice blocks are stuck in the glacial alluvium. After they melt, the sediment is left with holes. The diameter of such depressions ranges from 5 m to 13 km, with depths of up to 45 meters. Most are circular in shape due to the melting blocks of ice becoming rounded. The lakes that often form in these depressions are known as "kettle lakes".

## **Deposits in contact with ice**

When a glacier reduces in size to a critical point, its flow stops, and the ice becomes stationary. Meanwhile, meltwater flows over, within, and beneath the ice leave stratified alluvial deposits. Because of this, as the ice melts, it leaves stratified deposits in the form of columns, terraces and clusters. These types of deposits are known as "deposits in contact with ice".

When those deposits take the form of columns of tipped sides or mounds, they are called *kames*. Some *kames* form when meltwater deposits sediments through openings in the interior of the ice. In other cases, they are just the result of fans or deltas towards the exterior of the ice produced by meltwater. When the glacial ice occupies a valley, it can form terraces or *kame* along the sides of the valley.

A third type of deposit formed in contact with the ice is characterized by long, narrow sinuous crests, composed fundamentally of sand and gravel deposited by streams of meltwater flowing within, or beneath the glacier. After the ice has melted, these linear ridges or eskers remain as landscape features. Some of these crests have heights exceeding 100 meters and their lengths surpass 100 km.

## **Loess deposits**

Very fine glacial sediments or rock flour is often picked up by wind blowing over the bare surface and may be deposited great distances from the original fluvial deposition site. These eolian loess deposits may be very deep, even hundreds of meters, as in areas of China and the Midwestern United States. Katabatic winds can be important in this process.

## **Transportation and erosion**

- **Entrainment** is the picking up of loose material by the glacier from along the bed and valley sides. Entrainment can happen by regelation or by the ice simply picking up the debris.
- **Basal Ice Freezing** is thought to be made by glaciohydraulic supercooling, though some studies show that even where physical conditions allow it to occur, the process may not be responsible for observed sequences of basal ice.
- **Plucking** is the process involves the glacier freezing onto the valley sides and subsequent ice movement pulling away masses of rock. As the bedrock is greater in strength than the glacier, only previously loosened material can be removed. It

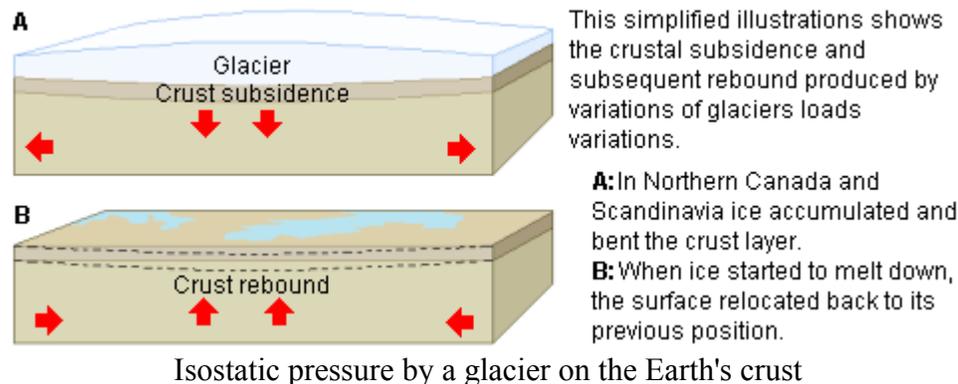
can be loosened by local pressure and temperature, water and pressure release of the rock itself.

- **Supraglacial debris** is carried on the surface of the glacier as lateral and medial moraines. In summer ablation, surface melt water carries a small load and this often disappears down crevasses.
- **Englacial debris** is moraine carried within the body of the glacier.
- **Subglacial debris** is moved along the floor of the valley either by the ice as ground moraine or by meltwater streams formed by pressure melting.

## Deposition

- **Lodgement till** is identical to ground moraine. It is material that is smeared on to the valley floor when its weight becomes too great to be moved by the glacier.
- **Ablation till** is a combination of englacial and supraglacial moraine. It is released as a stationary glacier begins to melt and material is dropped *in situ*.
- **Dumping** is when a glacier moves material to its outermost or lowermost end and dumps it.
- **Deformation flow** is the change of shape of the rock and land due to the glacier.

## Isostatic rebound



This rise of a part of the crust is due to an isostatic adjustment. A large mass, such as an ice sheet/glacier, depresses the crust of the Earth and displaces the mantle below. The depression is about a third the thickness of the ice sheet. After the glacier melts the mantle begins to flow back to its original position pushing the crust back to its original position. This post-glacial rebound, which lags melting of the ice sheet/glacier, is currently occurring in measurable amounts in Scandinavia and the Great Lakes region of North America.

An interesting geomorphological feature created by the same process, but on a smaller scale, is known as dilation-faulting. It occurs within rock where previously compressed rock is allowed to return to its original shape, but more rapidly than can be maintained without faulting, leading to an effect similar to that which would be seen if the rock were

hit by a large hammer. This can be observed in recently de-glaciated parts of Iceland and Cumbria.

## Chapter-4

# Ice Sheet

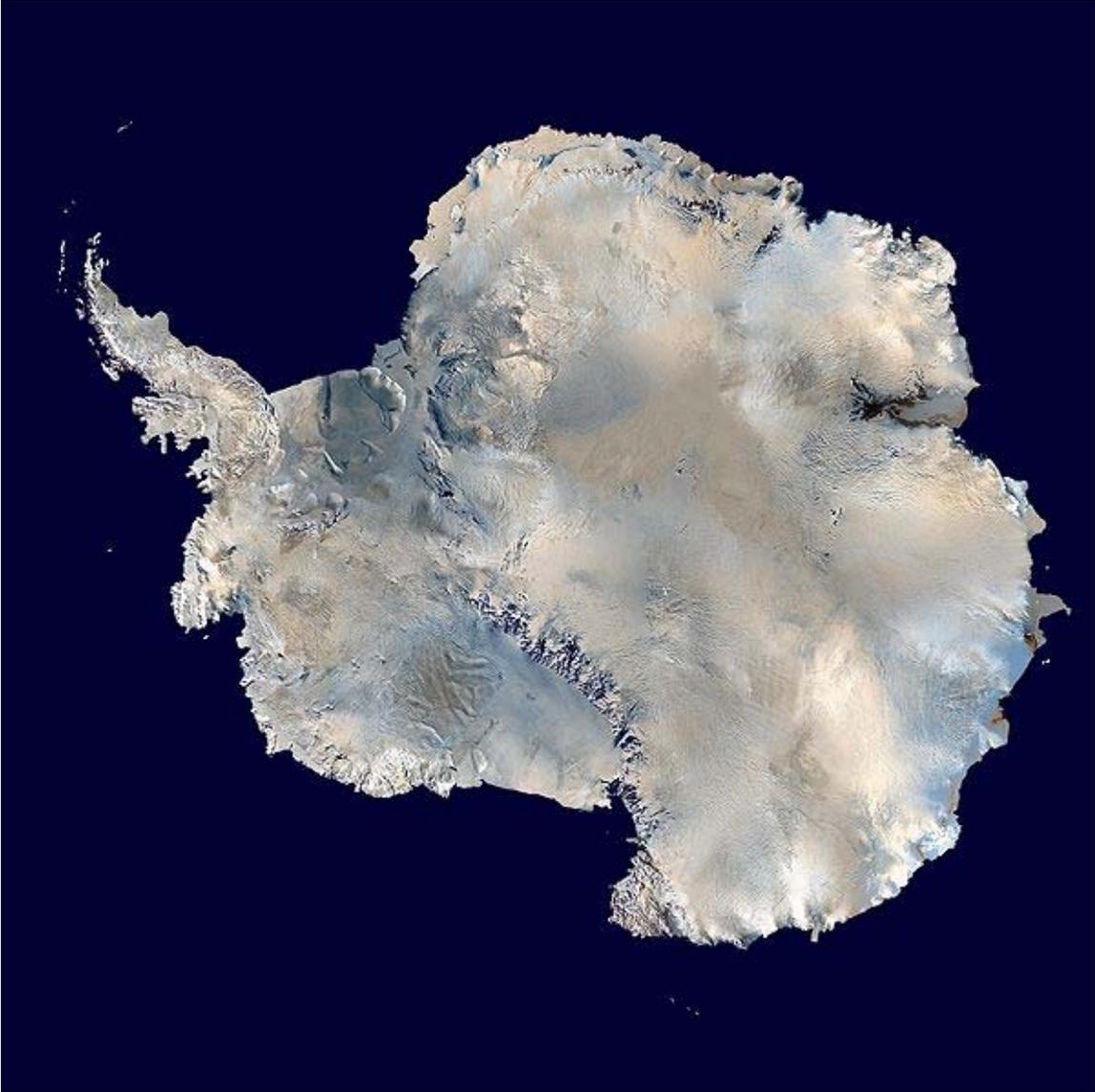
An **ice sheet** is a mass of glacier ice that covers surrounding terrain and is greater than 50,000 km<sup>2</sup> (20,000 mile<sup>2</sup>), thus also known as **continental glacier**. The only current ice sheets are in Antarctica and Greenland; during the last glacial period at Last Glacial Maximum (LGM) the Laurentide ice sheet covered much of Canada and North America, the Weichselian ice sheet covered northern Europe and the Patagonian Ice Sheet covered southern South America.

Ice sheets are bigger than ice shelves or alpine glaciers. Masses of ice covering less than 50,000 km<sup>2</sup> are termed an ice cap. An ice cap will typically feed a series of glaciers around its periphery.

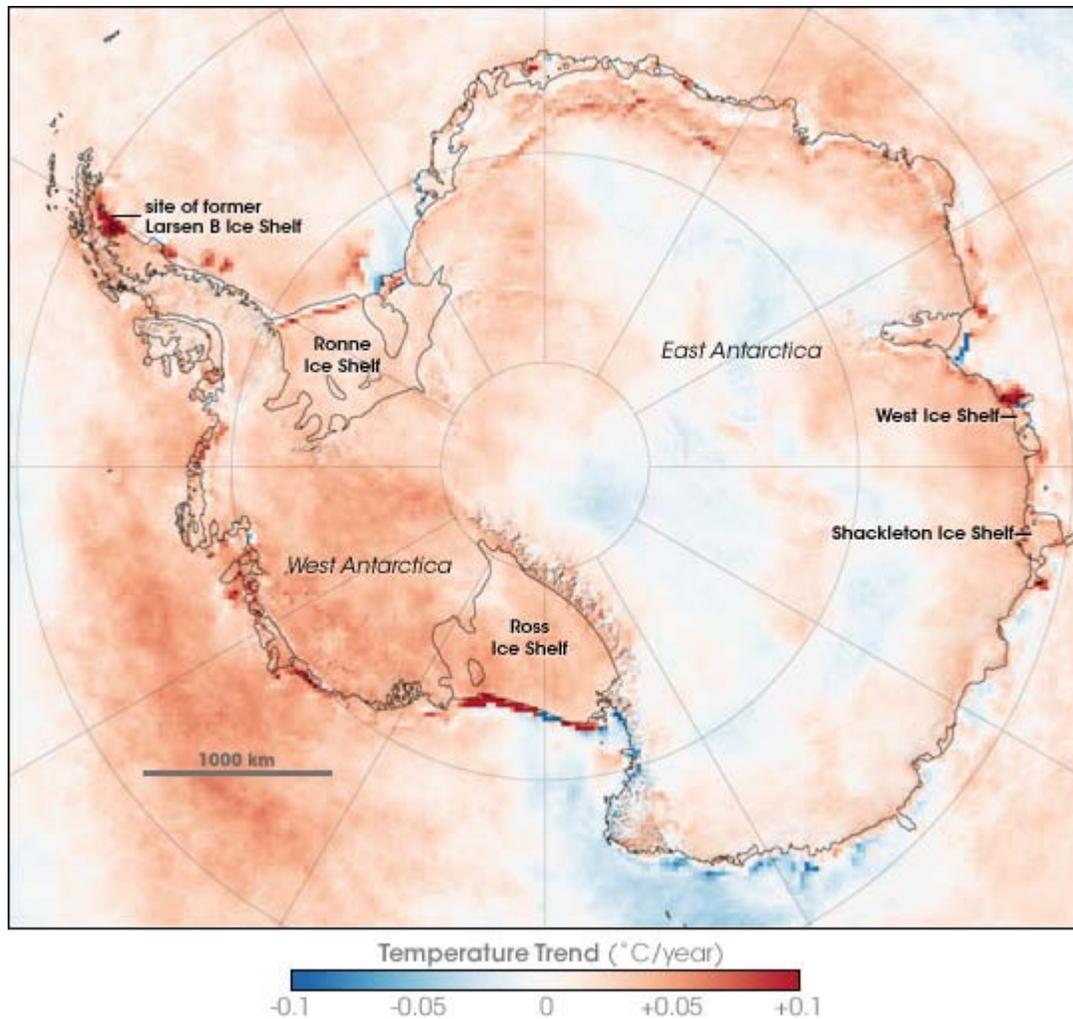
Although the surface is cold, the base of an ice sheet is generally warmer due to geothermal heat. In places, melting occurs and the melt-water lubricates the ice sheet so that it flows more rapidly. This process produces fast-flowing channels in the ice sheet — these are ice streams.

The present-day polar ice sheets are relatively young in geological terms. The Antarctic Ice Sheet first formed as a small ice cap (maybe several) in the early Oligocene, but retreating and advancing many times until the Pliocene, when it came to occupy almost all of Antarctica. The Greenland ice sheet did not develop at all until the late Pliocene, but apparently developed *very rapidly* with the first continental glaciation. This had the unusual effect of allowing fossils of plants that once grew on present-day Greenland to be much better preserved than with the slowly forming Antarctic ice sheet.

## Antarctic ice sheet



A satellite composite image of Antarctica



Antarctic Skin Temperature Trends between 1981 and 2007, based on thermal infrared observations made by a series of NOAA satellite sensors. Skin temperature trends do not necessarily reflect air temperature trends.

The **Antarctic ice sheet** is one of the two polar ice packs of the Earth. It covers about 98% of the Antarctic continent and is the largest single mass of ice on Earth. It covers an area of almost 14 million square km and contains 30 million cubic km of ice. That is, approximately 61 percent of all fresh water on the Earth is held in the Antarctic ice sheet, an amount equivalent to 70 m of water in the world's oceans. In East Antarctica, the ice sheet rests on a major land mass, but in West Antarctica the bed can extend to more than 2,500 m below sea level. The land in this area would be seabed if the ice sheet were not there.

The icing of Antarctica began with ice-rafting from middle Eocene times about 45.5 million years ago and escalated inland widely during the Eocene-Oligocene extinction event about 34 million years ago. CO<sub>2</sub> levels were then about 760 ppm and had been decreasing from earlier levels in the thousands of ppm. The glaciation was favored by an interval when the Earth's orbit favoured cool summers but Oxygen isotope ratio cycle

marker changes were too large to be explained by Antarctic ice-sheet growth alone indicating an ice age of some size. The opening of the Drake Passage may have played a role as well though models of the changes suggest declining CO<sub>2</sub> levels to have been more important.

Ice enters the sheet through precipitation as snow. This snow is then compacted to form glacier ice which moves under gravity towards the coast. Most of it is carried to the coast by fast moving ice streams. The ice then passes into the ocean, often forming vast floating ice shelves. These shelves then melt or calve off to give icebergs that eventually melt.

If the transfer of the ice from the land to the sea is balanced by snow falling back on the land then there will be no net contribution to global sea levels. A 2002 analysis of NASA satellite data from 1979-1999 showed that areas of Antarctica where ice was increasing outnumbered areas of decreasing ice roughly 2:1. The general trend shows that a warming climate in the southern hemisphere would transport more moisture to Antarctica, causing the interior ice sheets to grow, while calving events along the coast will increase, causing these areas to shrink. However more recent satellite data, which measures changes in the gravity of the ice mass, suggests that the total amount of ice in Antarctica has begun decreasing in the past few years. Another recent study compared the ice leaving the ice sheet, by measuring the ice velocity and thickness along the coast, to the amount of snow accumulation over the continent. This found that the East Antarctic Ice Sheet was in balance but the West Antarctic Ice Sheet was losing mass. This was largely due to acceleration of ice streams such as Pine Island Glacier. These results agree closely with the gravity changes.

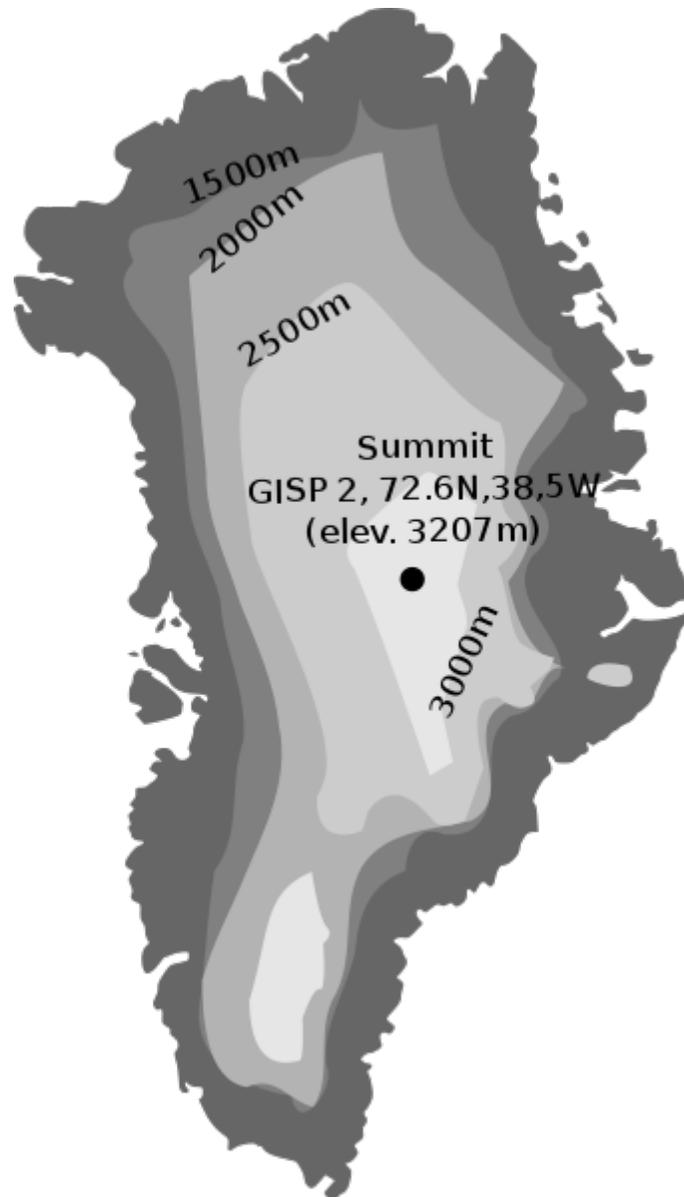
The continent-wide average surface temperature trend of Antarctica is positive and significant at  $>0.05^{\circ}\text{C}/\text{decade}$  since 1957. West Antarctica has warmed by more than  $0.1^{\circ}\text{C}/\text{decade}$  in the last 50 years, and this warming is strongest in winter and spring. Although this is partly offset by fall cooling in East Antarctica, this effect is restricted to the 1980s and 1990s.

Despite this warming total Antarctic sea ice anomalies have been steadily increasing since 1978 (NSIDC (2006)). 2007 showed the largest positive anomaly of sea ice in the southern hemisphere since records have been kept starting in 1979 and 2008 is currently on pace to surpass last years record. The atmospheric warming cannot be directly linked to the recent mass losses in West Antarctica. This mass loss is more likely to be due to increased melting of the ice shelves because of changes in ocean circulation patterns. This in turn causes the ice streams to speed up. The melting and disappearance of the floating ice shelves will only have a small effect on sea level, which is due to salinity differences. The most important consequence of their increased melting is the speed up of the ice streams on land which are buttressed by these ice shelves.

# Greenland ice sheet



Map of Greenland



Outline map of Greenland with ice sheet depths. GISP refers to a main site of the Greenland Ice Sheet Project, where a 3 km deep ice core was taken.

The **Greenland ice sheet** (Kalaallisut: *Sermersuaq*) is a vast body of ice covering 1,710,000 square kilometers (660,235 sq mi), roughly 80% of the surface of Greenland. It is the second largest ice body in the world, after the Antarctic Ice Sheet. The ice sheet is almost 2,400 kilometers (1,500 mi) long in a north-south direction, and its greatest width is 1,100 kilometers (680 mi) at a latitude of 77°N, near its northern margin. The mean altitude of the ice is 2135 meters. The thickness is generally more than 2 km (1.24 mi) and over 3 km (1.86 mi) at its thickest point. It is not the only ice mass of Greenland – isolated glaciers and small ice caps cover between 76,000 and 100,000 square kilometers (29,344 and 38,610 sq mi) around the periphery. Some scientists predict that global warming may be about to push the ice sheet over a threshold where the entire ice sheet

will melt in less than a few hundred years. If the entire 2,850,000 cubic kilometers (683,751 cu mi) of ice were to melt, it would lead to a global sea level rise of 7.2 m (23.6 ft). This would inundate most of the world's coastal cities and remove several small island countries from the face of the Earth, since island nations such as Tuvalu and Maldives have a maximum altitude below or just above this level.

The Greenland Ice Sheet is also sometimes referred to under the term *inland ice*, or its Danish equivalent, *indlandsis*. It is also sometimes referred to as an ice cap. "Ice sheet" is considered the more correct term, as "ice cap" generally refers to less extensive ice masses.

The ice in the current ice sheet is as old as 110,000 years. It is generally thought that the Greenland Ice Sheet formed in the late Pliocene or early Pleistocene by coalescence of ice caps and glaciers. It did not develop at all until the late Pliocene, but apparently developed very rapidly with the first continental glaciation.

The weight of the ice has depressed the central area of Greenland; the bedrock surface is near sea level over most of the interior of Greenland, but mountains occur around the periphery, confining the sheet along its margins. If the ice disappeared, Greenland would most probably appear as an archipelago, at least until isostasy lifted the land surface above sea level once again. The ice surface reaches its greatest altitude on two north-south elongated domes, or ridges. The southern dome reaches almost 3,000 meters (9,843 ft) at latitudes 63°–65°N; the northern dome reaches about 3,290 meters (10,794 ft) at about latitude 72°N. The crests of both domes are displaced east of the centre line of Greenland. The unconfined ice sheet does not reach the sea along a broad front anywhere in Greenland, so that no large ice shelves occur. The ice margin just reaches the sea, however, in a region of irregular topography in the area of Melville Bay southeast of Thule. Large outlet glaciers, which are restricted tongues of the ice sheet, move through bordering valleys around the periphery of Greenland to calve off into the ocean, producing the numerous icebergs that sometimes occur in North Atlantic shipping lanes. The best known of these outlet glaciers is Jakobshavn Isbræ (Kalaallit: *Sermeq Kujalleq*), which, at its terminus, flows at speeds of 20 to 22 metres or 65.6 to 72.2 feet per day.

On the ice sheet, temperatures are generally substantially lower than elsewhere in Greenland. The lowest mean annual temperatures, about  $-31\text{ }^{\circ}\text{C}$  ( $-23.8\text{ }^{\circ}\text{F}$ ), occur on the north-central part of the north dome, and temperatures at the crest of the south dome are about  $-20\text{ }^{\circ}\text{C}$  ( $-4\text{ }^{\circ}\text{F}$ ).

During winter, the ice sheet takes on a clear blue/green color. During summer, the top layer of ice melts leaving pockets of air in the ice that makes it look white.

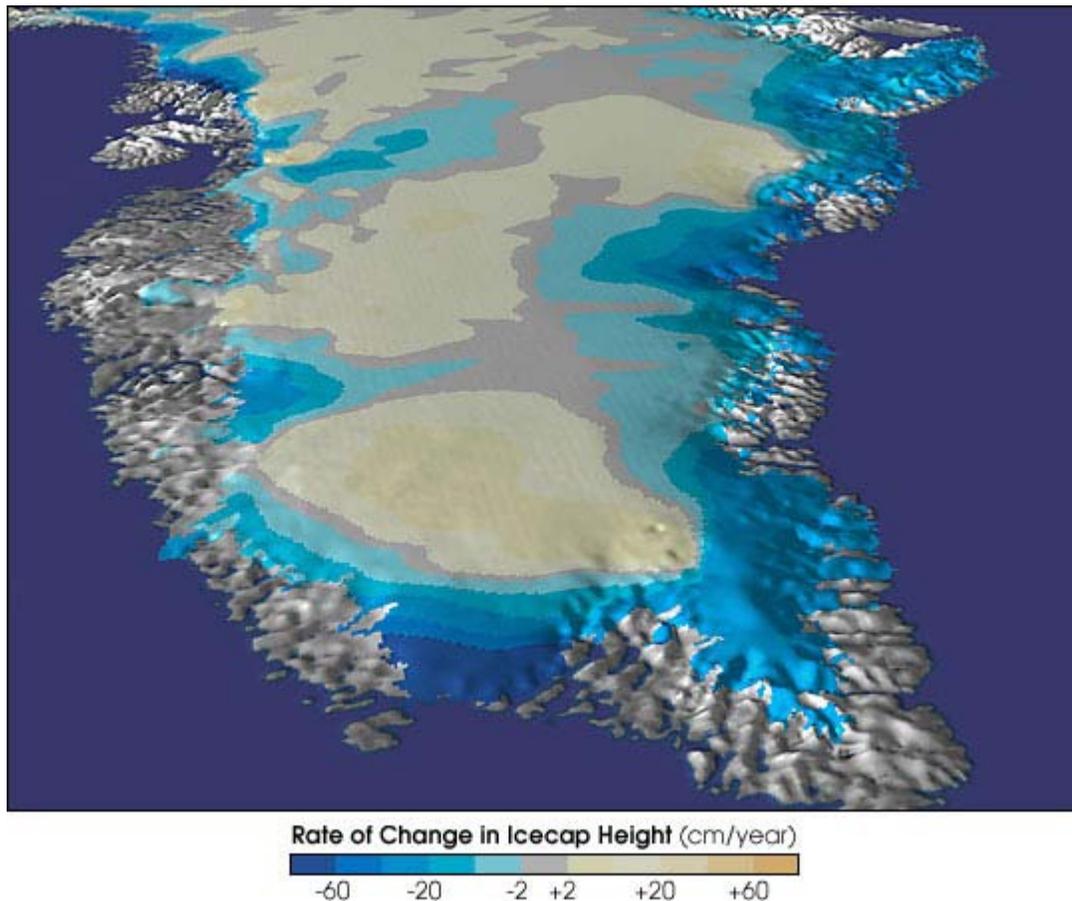


A helicopter taking off from the Greenland Ice Sheet

## **The ice sheet as a record of past climates**

The ice sheet, consisting of layers of compressed snow from more than a hundred thousand years, contains in its ice today's most valuable record of past climates. In the past decades, scientists have drilled ice cores up to 4 kilometers (2.5 mi) deep. Scientists have, using those ice cores, obtained information on (proxies for) temperature, ocean volume, precipitation, chemistry and gas composition of the lower atmosphere, volcanic eruptions, solar variability, sea-surface productivity, desert extent and forest fires. This variety of climatic proxies is greater than in any other natural recorder of climate, such as tree rings or sediment layers.

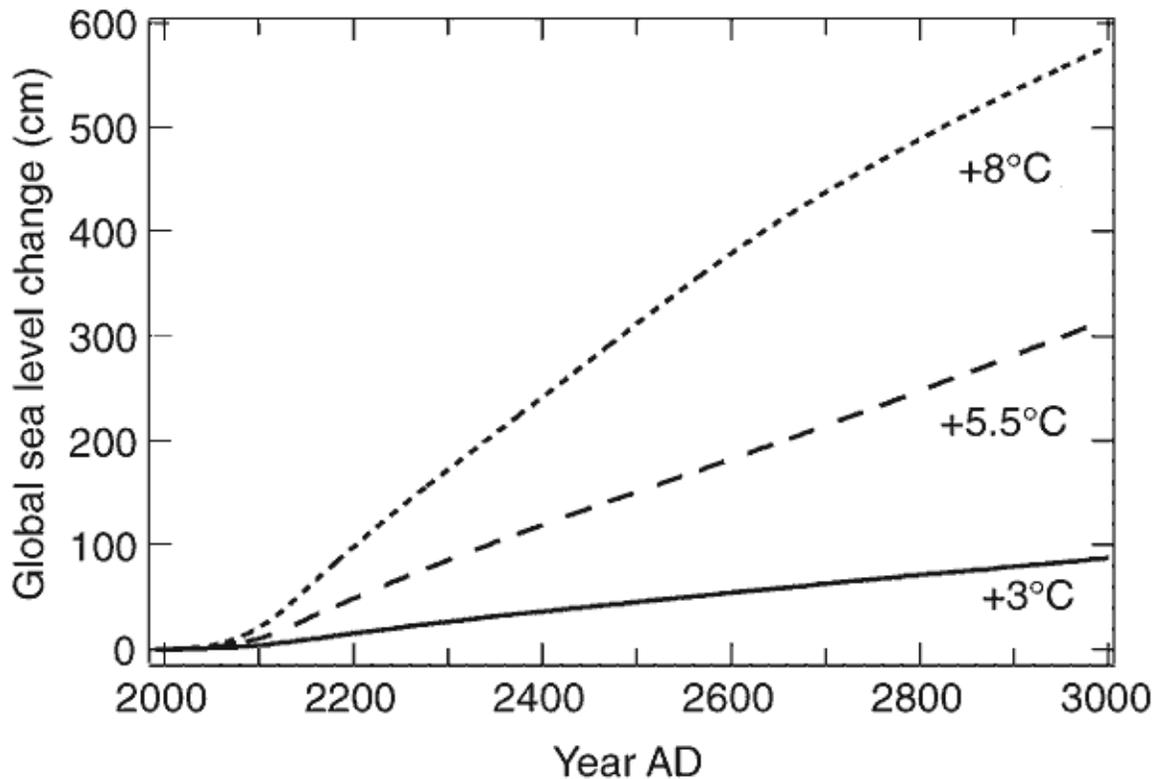
## The melting ice sheet



Rate of change in ice sheet height in cm per year

Positioned in the Arctic, the Greenland ice sheet is especially vulnerable to global warming. Arctic climate is now rapidly warming and much larger Arctic shrinkage changes are projected. The Greenland Ice Sheet has experienced record melting in recent years and is likely to contribute substantially to sea level rise as well as to possible changes in ocean circulation in the future. The area of the sheet that experiences melting has increased about 16% from 1979 (when measurements started) to 2002 (most recent data). The area of melting in 2002 broke all previous records. The number of glacial earthquakes at the Helheim Glacier and the northwest Greenland glaciers increased substantially between 1993 and 2005. In 2006, estimated monthly changes in the mass of Greenland's ice sheet suggest that it is melting at a rate of about 239 cubic kilometers (57 cu mi) per year. A more recent study, based on reprocessed and improved data between 2003 and 2008, reports an average trend of 195 cubic kilometers (47 cu mi) per year. These measurements came from the US space agency's GRACE (Gravity Recovery and Climate Experiment) satellite, launched in 2002, as reported by BBC. Using data from two ground-observing satellites, ICESAT and ASTER, a study published in *Geophysical*

Research Letters (September 2008) shows that nearly 75 percent of the loss of Greenland's ice can be traced back to small coastal glaciers.



Modelling results of the sea-level rise under different warming scenarios. The curve labels refer to the mean annual temperature rise over Greenland by 3000 AD. Note that the temperature projections shown are greater than globally averaged temperatures (by a factor of 1.2 to 3.1)

If the entire 2,850,000 km<sup>3</sup> (683,751 cu mi) of ice were to melt, global sea levels would rise 7.2 m (23.6 ft). Recently, fears have grown that continued global warming will make the Greenland Ice Sheet cross a threshold where long-term melting of the ice sheet is inevitable. Climate models project that local warming in Greenland will exceed 3 °C (5.4 °F) during this century. Ice sheet models project that such a warming would initiate the long-term melting of the ice sheet, leading to a complete melting of the ice sheet (over centuries), resulting in a global sea level rise of about 7 meters (23.0 ft). Such a rise would inundate almost every major coastal city in the World. How fast the melt would eventually occur is a matter of discussion. According to IPCC, the expected 3 degrees warming at the end of the century would, if kept from rising further, result in about 1 meter sea level rise over the next millennium (see image).

Some scientists have cautioned that these rates of melting are overly optimistic as they assume a linear, rather than erratic, progression. James E. Hansen has argued that multiple positive feedbacks could lead to nonlinear ice sheet disintegration much faster than claimed by the IPCC. According to a 2007 paper, "we find no evidence of millennial

lags between forcing and ice sheet response in paleoclimate data. An ice sheet response time of centuries seems probable, and we cannot rule out large changes on decadal time-scales once wide-scale surface melt is underway."



Satellite image of dark blue melt ponds

The melt zone, where summer warmth turns snow and ice into slush and melt ponds of meltwater, has been expanding at an accelerating rate in recent years. When the meltwater seeps down through cracks in the sheet, it accelerates the melting and, in some areas, allow the ice to slide more easily over the bedrock below, speeding its movement to the sea. Besides contributing to global sea level rise, the process adds freshwater to the ocean, which may disturb ocean circulation and thus regional climate.

## Recent ice loss events

- A major ice loss to northern Greenland's Petermann glacier occurred when the glacier lost 33 square miles (85 km<sup>2</sup>) of floating ice between 2000 and 2001.
- Between 2001 and 2005, a breakup of Sermeq Kujalleq erased 36 square miles (93 km<sup>2</sup>) from the ice field and raised awareness worldwide of glacial response to global climate change.
- In July 2008, researchers monitoring daily satellite images discovered that a 11-square-mile (28 km<sup>2</sup>) piece of Petermann broke away.
- Two years later, in August 2010, a sheet of ice measuring 260 square kilometres (100 sq mi) broke off from the Petermann Glacier. Researchers from the Canadian Ice Service located the calving from NASA satellite images taken on August 5. The images showed that Petermann lost about one-quarter of its 70 km-long (43 mile) floating ice shelf.

## Ice sheet acceleration

Two mechanisms have been utilized to explain the change in velocity of the Greenland Ice Sheets outlet glaciers. The first is the enhanced meltwater effect, which relies on additional surface melting, funneled through moulins reaching the glacier base and reducing the friction through a higher basal water pressure. (It should be noted that not all meltwater is retained in the ice sheet and some moulins drain into the ocean, with varying rapidity.) This idea, was observed to be the cause of a brief seasonal acceleration of up to 20 % on Sermeq Kujalleq in 1998 and 1999 at Swiss Camp. (The acceleration lasted two-three months and was less than 10% in 1996 and 1997 for example. They offered a conclusion that the “coupling between surface melting and ice-sheet flow provides a mechanism for rapid, large-scale, dynamic responses of ice sheets to climate warming”. Examination of recent rapid supra-glacial lake drainage documented short term velocity changes due to such events, but they had little significance to the annual flow of the large glaciers outlet glaciers. The second mechanism is a force imbalance at the calving front due to thinning causing a substantial non-linear response. In this case an imbalance of forces at the calving front propagates up-glacier. Thinning causes the glacier to be more buoyant, reducing frictional back forces, as the glacier becomes more afloat at the calving front. The reduced friction due to greater buoyancy allows for an increase in velocity. This is akin to letting off the emergency brake a bit. The reduced resistive force at the calving front is then propagated up glacier via longitudinal extension because of the backforce reduction. For ice streaming sections of large outlet glaciers (in Antarctica as well) there is always water at the base of the glacier that helps lubricate the flow. This water is, however, generally from basal processes, not surface melting.

If the enhanced meltwater effect is the key then since meltwater is a seasonal input, velocity would have a seasonal signal and all glaciers would experience this effect. If the force imbalance effect is the key the velocity will propagate up-glacier, there will be no seasonal cycle, and the acceleration will be focused on calving glaciers. Helheim Glacier, East Greenland had a stable terminus from the 1970s-2000. In 2001–2005 the glacier retreated 7 km (4.3 mi) and accelerated from 20 to 33 m or 65.6 to 108.3 ft/day, while

thinning up to 130 meters (430 ft) in the terminus region. Kangerdlugssuaq Glacier, East Greenland had a stable terminus history from 1960 to 2002. The glacier velocity was 13 m or 42.7 ft/day in the 1990s. In 2004–2005 it accelerated to 36 m or 118 ft/day and thinned by up to 100 m (328 ft) in the lower reach of the glacier. On Sermeq Kujalleq the acceleration began at the calving front and spread up-glacier 20 km (12 mi) in 1997 and up to 55 km (34 mi) inland by 2003. On Helheim the thinning and velocity propagated up-glacier from the calving front. In each case the major outlet glaciers accelerated by at least 50%, much larger than the impact noted due to summer meltwater increase. On each glacier the acceleration was not restricted to the summer, persisting through the winter when surface meltwater is absent.

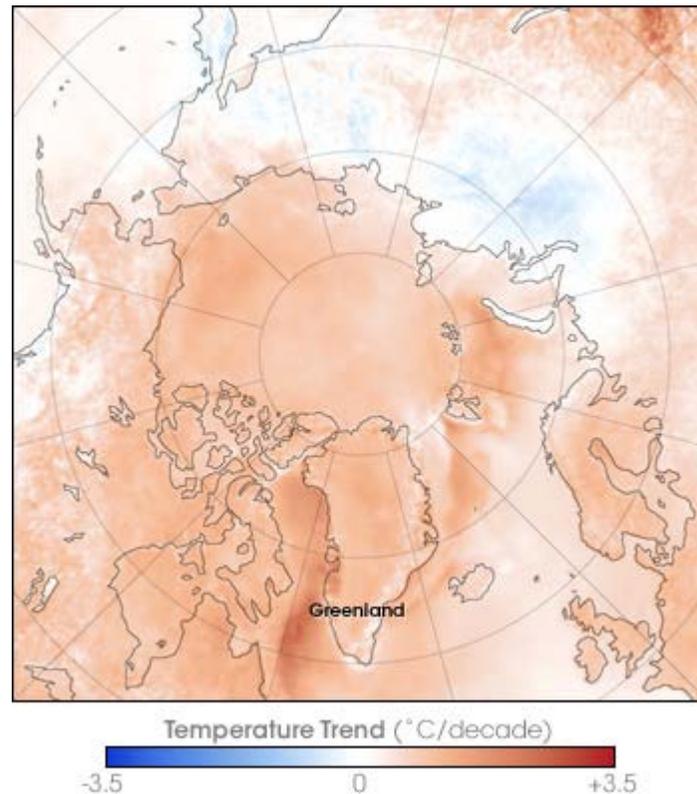
An examination of 32 outlet glaciers in southeast Greenland indicates that the acceleration is significant only for marine terminating outlet glaciers. That is glaciers that calve into the ocean. Further, noted that the thinning of the ice sheet is most pronounced for marine terminating outlet glaciers. As a result of the above, all concluded that the only plausible sequence of events is that increased thinning of the terminus regions, of marine terminating outlet glaciers, ungrounded the glacier tongues and subsequently allowed acceleration, retreat and further thinning. Enhanced meltwater induced acceleration does exist but is of a notably smaller magnitude and duration.

## **Increased precipitation**

Warmer temperatures in the region have brought increased precipitation to Greenland, and part of the lost mass has been offset by increased snowfall. However, there are only a small number of weather stations on the island, and though Satellite data can examine the entire island, it has only been available since the early 1990s, making trending difficult. It has been observed that there is more precipitation where it is warmer, up to 1.5 ma<sup>-1</sup> on the SE flank, and where cooler less or nil (25–80% of the island depending on the time of year). Actual figures for precipitation are available in "New precipitation and accumulation maps for Greenland", A. Ohmura and N. Reeh, Journal of Glaciology, 1991.

Data from NASA's Polar program confirms that the average elevation change above 2,000 m (6,562 ft) "was not significant".

## Rate of change



Arctic Temperature Trend 1981–2007

Several factors determine the net rate of growth or decline. These are

1. accumulation of snow in the central parts
2. melting of ice along the sheet's margins (runoff) and bottom,
3. iceberg calving into the sea from outlet glaciers also along the sheet's edges

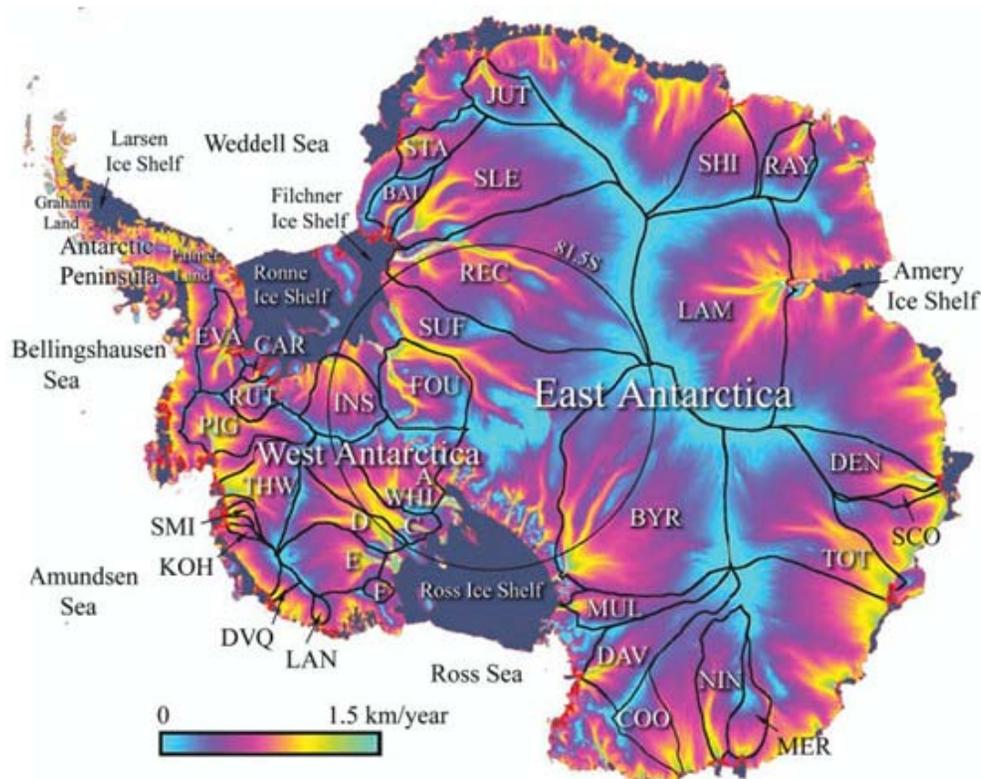
IPCC estimates in their third assessment report the accumulation to  $520 \pm 26$  Gigatonnes of ice per year, runoff and bottom melting to  $297 \pm 32$  Gt/yr and  $32 \pm 3$  Gt/yr, respectively, and iceberg production to  $235 \pm 33$  Gt/yr. On balance, the IPCC estimates  $-44 \pm 53$  Gt/yr, which means that the ice sheet may currently be melting. The most recent research using data from 1996 to 2005 shows that the ice sheet is thinning even faster than supposed by IPCC. According to the study, in 1996 Greenland was losing about  $96 \text{ km}^3$  or 23.0 cu mi per year in mass from its ice sheet. In 2005, this had increased to about  $220 \text{ km}^3$  or 52.8 cu mi a year due to rapid thinning near its coasts, while in 2006 it was estimated at  $239 \text{ km}^3$  (57.3 cu mi) per year. It was estimated that in the year 2007 Greenland ice sheet melting was higher than ever,  $592 \text{ km}^3$  (142.0 cu mi). Also snowfall was unusually low, which lead to unprecedented negative  $-65 \text{ km}^3$  ( $-15.6$  cu mi) Surface Mass Balance. If iceberg calving has happened as an average, Greenland lost 294 Gt of its mass during 2007 (one  $\text{km}^3$  of ice weights about 0.9 Gt).

According to the 2007 report from the IPCC, it is hard to measure the mass balance precisely, but most results indicate accelerating mass loss from Greenland during the 1990s up to 2005. Assessment of the data and techniques suggests a mass balance for the Greenland Ice Sheet ranging between growth of 25 Gt/yr and loss of 60 Gt/yr for 1961 to 2003, loss of 50 to 100 Gt/yr for 1993 to 2003 and loss at even higher rates between 2003 and 2005.

A paper on Greenland's temperature record shows that the warmest year on record was 1941 while the warmest decades were the 1930s and 1940s. The data used was from stations on the south and west coasts, most of which did not operate continuously the entire study period.

While Arctic temperatures have generally increased, there is some discussion over the temperatures over Greenland. First of all, Arctic temperatures are highly variable, making it difficult to discern clear trends at a local level. Also, until recently, an area in the North Atlantic including southern Greenland was one of the only areas in the World showing cooling rather than warming in recent decades, but this cooling has now been replaced by strong warming in the period 1979–2005.

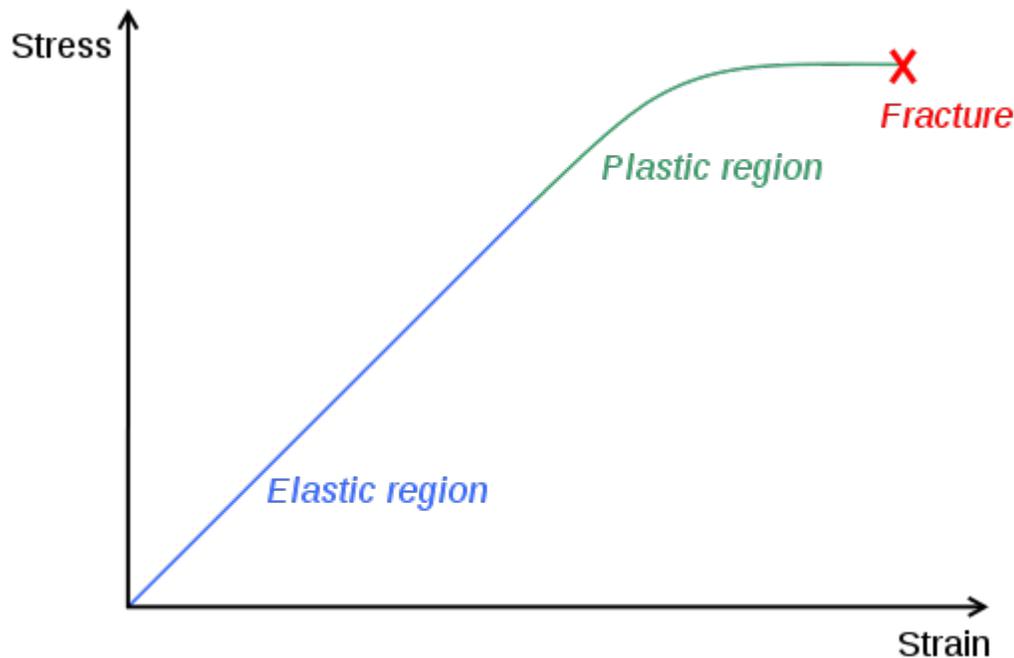
## Ice sheet dynamics



Glacial flow rate in the Antarctic ice sheet

**Ice sheet dynamics** describe the motion within large bodies of ice, such those currently on Greenland and Antarctica. Ice motion is dominated by the movement of glaciers, whose gravity-driven activity is controlled by two main variable factors: the temperature and strength of their bases. A number of processes alter these two factors, resulting in cyclic surges of activity interspersed with longer periods of inactivity, on both hourly and centennial time scales.

## Flow dynamics



The stress-strain relationship of plastic flow (teal section): a small increase in stress creates an exponentially greater increase in strain, which equates to deformation speed.

The main cause of flow within glaciers can be attributed to an increase in the surface slope, brought upon by an imbalance between the amounts of accumulation vs. ablation. This imbalance increases the shear stress on a glacier until it begins to flow. The flow velocity and deformation will increase as the equilibrium line between these two processes is approached, but are also affected by the slope of the ice, the ice thickness and temperature.

When the amount of strain (deformation) is proportional to the stress being applied, ice will act as an elastic solid. Ice will not flow until it has reached a thickness of 30 meters (98 ft), but after 50 meters (164 ft), small amounts of stress can result in a large amount of strain, causing the deformation to become a plastic flow rather than elastic. At this point the glacier will begin to deform under its own weight and flow across the landscape. According to the Glen-Nye Flow law, the relationship between stress and strain, and thus the rate of internal flow, can be modeled as follows:

$$\Sigma = k\tau^n$$

*Where*

$\Sigma$  = shear strain (flow) rate

$\tau$  = stress

$n$  = a constant between 2-4 (typically 3 for most glaciers) that increases with lower temperature

$k$  = a temperature-dependent constant

The lowest velocities are near the base of the glacier and along valley sides where friction acts against flow, causing the most deformation. Velocity increases inward toward the center line and upward, as the amount of deformation decreases. The highest flow velocities are found at the surface, representing the sum of the velocities of all the layers below.

Glaciers may also move by basal sliding, where the base of the glacier is lubricated by meltwater, allowing the glacier to slide over the terrain on which it sits. Meltwater may be produced by pressure-induced melting, friction or geothermal heat.

The top 50 meters of the glacier form the fracture zone, where ice moves as a single unit. Cracks form as the glacier moves over irregular terrain, which may penetrate the full depth of the fracture zone.

## Glacial bottom processes



A cross-section through a glacier. The base of the glacier is more transparent as a result of melting.

Most of the important processes controlling glacial motion occur in the ice-bed contact—even though it is only a few meters thick. Glaciers will move by sliding when the basal shear stress drops below the shear resulting from the glacier's weight.

$$\tau_D = \rho g h \sin \alpha$$

where  $\tau_D$  is the driving stress, and  $\alpha$  the angle of repose.

$\tau_B$  is the basal shear stress, a function of bed temperature and softness.

$\tau_F$ , the shear stress, is the lower of  $\tau_B$  and  $\tau_D$ . It controls the rate of plastic flow, as per the figure (inset, right).

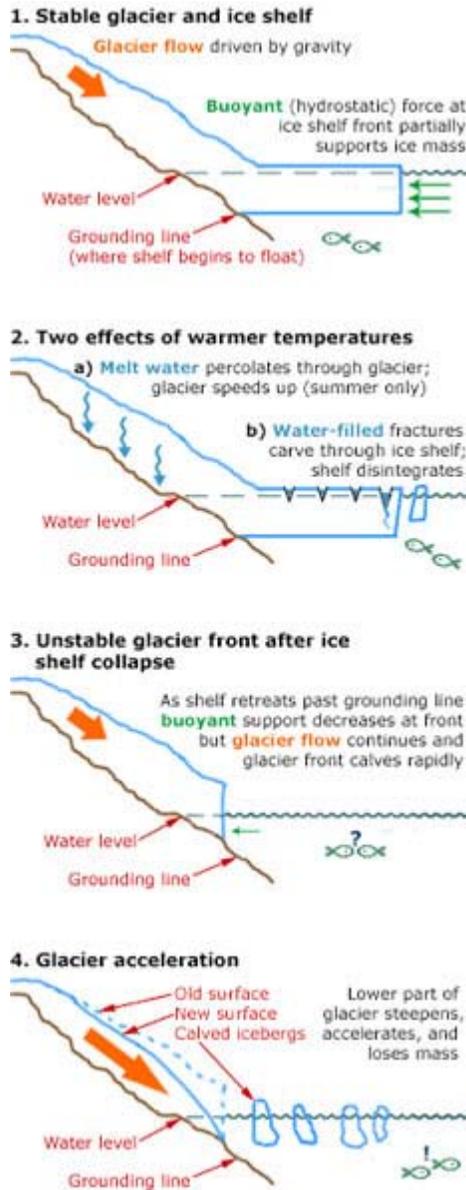
For a given glacier, the two variables are  $\tau_D$ , which varies with  $h$ , the depth of the glacier, and  $\tau_B$ , the basal shear stress.

### **Basal shear stress**

The basal shear stress is a function of three factors: the bed's temperature, roughness and softness.

Whether a bed is hard or soft depends on the porosity and pore pressure; higher porosity decreases the sediment strength (thus increases the shear stress  $\tau_B$ ). If the sediment strength falls far below  $\tau_D$ , movement of the glacier will be accommodated by motion in the sediments, as opposed to sliding. **Porosity** may vary through a range of methods.

- Movement of the overlying glacier may cause the bed to undergo dilatancy; the resulting shape change reorganises blocks. This reorganises closely packed blocks (a little like neatly folded, tightly packed clothes in a suitcase) into a messy jumble (just as clothes never fit back in when thrown in in a disordered fashion). This increases the porosity. Unless water is added, this will necessarily reduce the pore pressure (as the pore fluids have more space to occupy).
- Pressure may cause compaction and consolidation of underlying sediments. Since water is relatively incompressible, this is easier when the pore space is filled with vapour; any water must be removed to permit compression. In soils, this is an irreversible process.
- Sediment degradation by abrasion and fracture decreases the size of particles, which tends to decrease pore space, although the motion of the particles may disorder the sediment, with the opposite effect. These processes also generate heat, whose importance will be discussed later.



### Factors controlling the flow of ice

A soft bed, with high porosity and low pore fluid pressure, allows the glacier to move by sediment sliding: the base of the glacier may even remain frozen to the bed, where the underlying sediment slips underneath it like a tube of toothpaste. A hard bed cannot deform in this way; therefore the only way for hard-based glaciers to move is by basal sliding, where meltwater forms between the ice and the bed itself.

Bed softness may vary in space or time, and changes dramatically from glacier to glacier. An important factor is the underlying geology; glacial speeds tend to differ more when they change bedrock than when the gradient changes.

As well as affecting the sediment stress, **fluid pressure** ( $p_w$ ) can affect the friction between the glacier and the bed. High fluid pressure provides a buoyancy force upwards on the glacier, reducing the friction at its base. The fluid pressure is compared to the ice overburden pressure,  $p_i$ , given by  $\rho gh$ . Under fast-flowing ice streams, these two pressures will be approximately equal, with an effective pressure ( $p_i - p_w$ ) of 30 kPa; i.e. all of the weight of the ice is supported by the underlying water, and the glacier is afloat.

A number of factors can affect **bed temperature**, which is intimately associated with **basal meltwater**. The melting point of water decreases under pressure, meaning that water melts at a lower temperature under thicker glaciers. This acts as a "double whammy", because thicker glaciers have a lower heat conductance, meaning that the basal temperature is also likely to be higher.

Bed temperature tends to vary in a cyclic fashion. A cool bed has a high strength, reducing the speed of the glacier. This increases the rate of accumulation, since newly fallen snow is not transported away. Consequently, the glacier thickens, with three consequences: firstly, the bed is better insulated, allowing greater retention of geothermal heat. Secondly, the increased pressure can facilitate melting. Most importantly,  $\tau_D$  is increased. These factors will combine to accelerate the glacier. As friction increases with the square of velocity, faster motion will greatly increase frictional heating, with ensuing melting - which causes a positive feedback, increasing ice speed to a faster flow rate still: west Antarctic glaciers are known to reach velocities of up to a kilometre per year. Eventually, the ice will be surging fast enough that it begins to thin, as accumulation cannot keep up with the transport. This thinning will increase the conductive heat loss, slowing the glacier and causing freezing. This freezing will slow the glacier further, often until it is stationary, whence the cycle can begin again.

Supraglacial lakes represent another possible supply of liquid water to the base of glaciers, so they can play an important role in accelerating glacial motion. Lakes of a diameter greater than  $\sim 300$  m are capable of creating a fluid-filled crevasse to the glacier/bed interface. When these crevasses form, the entirety of the lake's (relatively warm) contents can reach the base of the glacier in as little as 2–18 hours - lubricating the bed and causing the glacier to surge.

Finally, **bed roughness** can act to slow glacial motion. The roughness of the bed is a measure of how many boulders and obstacles protrude into the overlying ice. Ice flows around these obstacles by melting under the high pressure on their lee sides; the resultant meltwater is then forced down a steep pressure gradient into the cavity arising in their stoss, where it re-freezes. Cavitation on the stoss side increases this pressure gradient, which assists flow.

## Erosional effects



Differential erosion enhances relief, as clear in this incredibly steep-sided Norwegian fjord.

Because ice can flow faster where it is thicker, the rate of glacier-induced erosion is directly proportional to the thickness of overlying ice. Consequently pre-glacial low hollows will be deepened and pre-existing topography will be amplified by glacial action, while nunataks, which protrude above ice sheets, barely erode at all - erosion has been estimated as 5 m per 1.2 million years! This explains, for example, the deep profile of fjords, which can reach a kilometer in depth as ice is topographically steered into them. Being the principal conduits for draining ice sheets, fjords' extension inland increases the

rate of ice sheet thinning. It also makes the ice sheets more sensitive to changes in climate and the ocean.

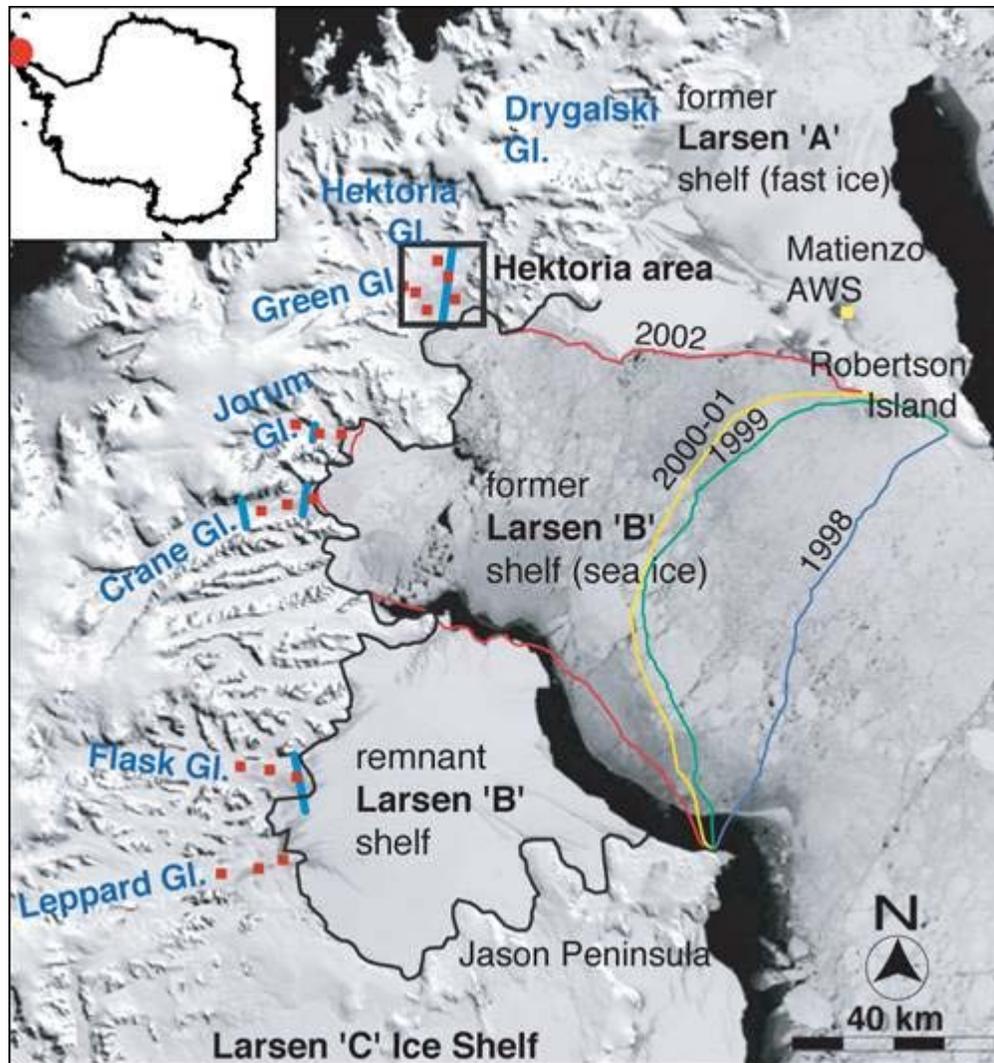
## **Pipe and sheet flow**

The flow of water under the glacial surface can have a large effect on the motion of the glacier itself. Subglacial lakes contain significant amounts of water, which can move fast: cubic kilometres can be transported between lakes over the course of a couple of years.

This motion is thought to occur in two main modes: **pipe flow** involves liquid water moving through pipe-like conduits, like a sub-glacial river; **sheet flow** involves motion of water in a thin layer. A switch between the two flow conditions may be associated with surging behaviour. Indeed, the loss of sub-glacial water supply has been linked with the shut-down of ice movement in the Kamb ice stream. The subglacial motion of water is expressed in the surface topography of ice sheets, which slump down into vacated subglacial lakes.

## **Boundary conditions**

The interface between an ice stream and the ocean is a significant control of the rate of flow.



The collapse of the Larsen B ice shelf had profound effects on the velocities of its feeder glaciers.

**Ice shelves** - thick layers of ice floating on the sea - can stabilise the glaciers that feed them. These tend to have accumulation on their tops, may experience melting on their bases, and calve icebergs at their periphery. The catastrophic collapse of the Larsen B ice shelf in the space of three weeks during February 2002 yielded some unexpected observations. The glaciers that had fed the ice sheet increased substantially in velocity. This cannot have been due to seasonal variability, as glaciers flowing into the remnants of the ice shelf (Flask, Leppard) did not accelerate.

Ice shelves exert a dominant control in Antarctica, but are less important in Greenland, where the ice sheet meets the sea in fjords. Here, melting is the dominant ice removal process, resulting in predominant mass loss occurring towards the edges of the ice sheet, where icebergs are calved in the fjords and surface meltwater runs into the ocean.

**Tidal effects** are also important; the influence of a 1 m tidal oscillation can be felt as much as 100 km from the sea. On an hour-to-hour basis, surges of ice motion can be modulated by tidal activity. During larger spring tides, an ice stream will remain almost stationary for hours at a time, before a surge of around a foot in under an hour, just after the peak high tide; a stationary period then takes hold until another surge towards the middle or end of the falling tide. At neap tides, this interaction is less pronounced, without tides surges would occur more randomly, approximately every 12 hours.

Ice shelves are also sensitive to basal melting. In Antarctica, this is driven by heat fed to the shelf by the circumpolar deep water current, which is 3 °C above the ice's melting point.

As well as heat, the sea can also exchange salt with the oceans. The effect of latent heat, resulting from melting of ice or freezing of sea water, also has a role to play. The effects of these, and variability in snowfall and base sea level combined, account for around 80 mm a<sup>-1</sup> variability in ice shelf thickness.

## Long term changes

Over long time scales, ice sheet mass balance is governed by the amount of sunlight reaching the earth. This variation in sunlight reaching the earth, or insolation, over geologic time is in turn determined by the angle of the earth to the sun and shape of the Earth's orbit, as it is pulled upon by neighboring planets; these variations occur in predictable patterns called Milankovitch cycles. Milankovitch Cycles dominate climate on the Glacial/Interglacial timescale, but there exist variations in ice sheet extent that are not linked directly with insolation.

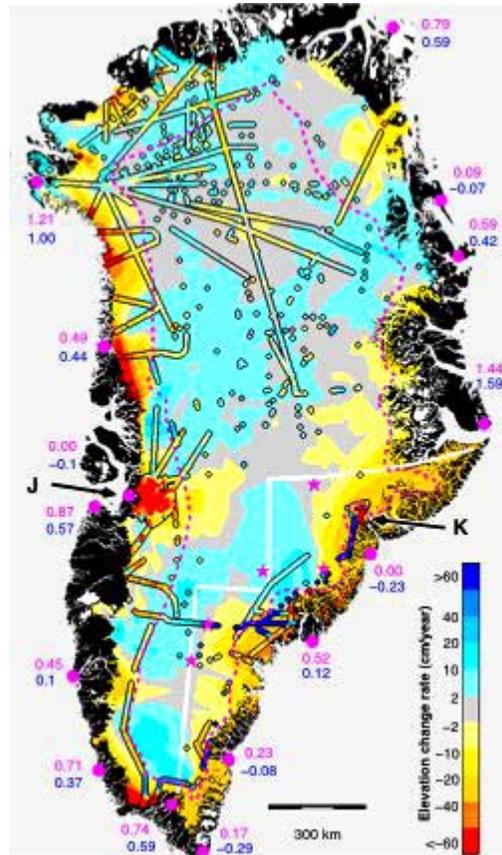
For instance, during at least the last 100,000 years, portions of the ice sheet covering much of North America, the Laurentide ice sheet broke apart sending large flotillas of icebergs into the North Atlantic. When these icebergs melted they dropped the boulders and other continental rocks they carried, leaving layers known as ice rafted debris. These so-called Heinrich events, named after their discoverer Hartmut Heinrich, appear to have a 7,000-10,000 year periodicity, and occur during cold periods within the last interglacial.

Internal ice sheet "binge-purge" cycles may be responsible for the observed effects, where the ice builds to unstable levels, then a portion of the ice sheet collapses. External affects might also play a role in forcing ice sheets. Dansgaard-Oeschger events are abrupt warmings of the northern hemisphere occurring over the space of perhaps 40 years. While these D-O events occur directly after each Heinrich event, they also occur more frequently - around every 1500 years; from this evidence, paleoclimatologists surmise that the same forcings may drive both Heinrich and D-O events.

**Hemispheric Asynchrony in Ice Sheet Behavior** has been observed by linking short term spikes of methane in Greenland ice cores and Antarctic ice cores. During Dansgaard-Oeschger events, the northern hemisphere warmed considerably, dramatically increasing the release of methane from wetlands, that were otherwise tundra during

glacial times. This methane quickly distributes evenly across the globe, becoming incorporated in Antarctic and Greenland ice. With this tie, paleoclimatologists have been able to say that the ice sheets on Greenland only began to warm after the Antarctic ice sheet had been warming for several thousand years. Why this pattern occurs is still open for debate.

## Effects of climate change on ice sheet dynamics



Rates of ice sheet thinning in Greenland

The implications of the current climate change on ice sheets are difficult to constrain. It is clear that increasing temperatures are resulting in reduced ice volumes globally. (Due to increased precipitation, the mass of parts of the Antarctic ice sheet may currently be increasing, but the total mass balance is unclear.)

Since the surging nature of ice sheet motion is a relatively recent discovery, and is still a long way from being entirely understood, no models have yet made a comprehensive evaluation of the effects of climate change. However, it is clear that climate change will act to destabilise ice sheets by a number of mechanisms.

Rising sea levels will reduce the stability of ice shelves, which have a key role in reducing glacial motion. Some Antarctic ice shelves are currently thinning by tens of

metres per year, and the collapse of the Larsen B shelf was preceded by thinning of just 1 metre per year. Further, increased ocean temperatures of 1 °C may lead to up to 10 metres per year of basal melting. Ice shelves are always stable under mean annual temperatures of −9 °C, but never stable above −5 °C; this places regional warming of 1.5 °C, as preceded the collapse of Larsen B, in context.

Increasing global temperatures take around 10,000 years to directly propagate through the ice before they influence bed temperatures, but may have an effect through increased surfacial melting, producing more supraglacial lakes, which may feed warm water to glacial bases and facilitate glacial motion.

Also, in areas of increased precipitation, such as Antarctica, the addition of mass will increase rate of glacial motion, hence the turnover in the ice sheet.

Observations, while currently limited in scope, do agree with these predictions of an increasing rate of ice loss from both Greenland and Antarctica.

A possible positive feedback may result from shrinking ice caps, in volcanically active Iceland at least. Isostatic rebound may lead to increased volcanic activity, causing basal warming - and, through CO<sub>2</sub> release, further climate change.

## **Predicted effects of global warming**

The Greenland, and probably the Antarctic, ice sheets have been losing mass recently, because losses by melting and outlet glaciers exceed accumulation of snowfall.

According to the Intergovernmental Panel on Climate Change (IPCC), loss of Antarctic and Greenland ice sheet mass contributed, respectively, about  $0.21 \pm 0.35$  and  $0.21 \pm 0.07$  mm/year to sea level rise between 1993 and 2003.

The IPCC projects that ice mass loss from melting of the Greenland ice sheet will continue to outpace accumulation of snowfall. Accumulation of snowfall on the Antarctic ice sheet is projected to outpace losses from melting. However, loss of mass on the Antarctic sheet may continue, if there is sufficient loss to outlet glaciers. In the words of the IPCC, *"Dynamical processes related to ice flow not included in current models but suggested by recent observations could increase the vulnerability of the ice sheets to warming, increasing future sea level rise. Understanding of these processes is limited and there is no consensus on their magnitude."* More research work is therefore required in order to improve the reliability of predictions of ice-sheet response on global warming.

Ice sheets and glaciers are flowing ice masses that rest on solid land. They are controlled by snow accumulation, surface and basal melt, calving into surrounding oceans or lakes and internal dynamics. The latter results from gravity-driven creep flow ("glacial flow") within the ice body and sliding on the underlying land, which leads to thinning and

horizontal spreading. Any imbalance of this dynamic equilibrium between mass gain, loss and transport due to flow results in either growing or shrinking ice bodies.

Ice sheets are the greatest potential source of global freshwater, holding approximately 77% of the global total. This corresponds to 80 m of world sea-level equivalent, with Antarctica accounting for 90% of this. Greenland accounts for most of the remaining 10%, with other ice bodies and glaciers accounting for less than 0.5%. Because of their size in relation to annual rates of snow accumulation and melt, the residence time of water in ice sheets can extend to 100,000 or 1 million years. Consequently, any climatic perturbations produce slow responses, occurring over glacial and interglacial periods. Valley glaciers respond rapidly to climatic fluctuations with typical response times of 10–50 years. However, the response of individual glaciers may be asynchronous to the same climatic forcing because of differences in glacier length, elevation, slope, and speed of motion. Oerlemans (1994) provided evidence of coherent global glacier retreat which could be explained by a linear warming trend of 0.66°C per 100 years.

While glacier variations are likely to have minimal effects upon global climate, their recession may have contributed one third to one half of the observed 20th Century rise in sea level (Meier 1984; IPCC 1996). Furthermore, it is extremely likely that such extensive glacier recession as is currently observed in the Western Cordillera of North America, where runoff from glacierized basins is used for irrigation and hydropower, involves significant hydrological and ecosystem impacts. Effective water-resource planning and impact mitigation in such areas depends upon developing a sophisticated knowledge of the status of glacier ice and the mechanisms that cause it to change. Furthermore, a clear understanding of the mechanisms at work is crucial to interpreting the global-change signals that are contained in the time series of glacier mass balance records.

Combined glacier mass balance estimates of the large ice sheets carry an uncertainty of about 20%. Studies based on estimated snowfall and mass output tend to indicate that the ice sheets are near balance or taking some water out of the oceans. Marinebased studies suggest sea-level rise from the Antarctic or rapid ice-shelf basal melting. Some authors (Paterson 1993; Alley 1997) have suggested that the difference between the observed rate of sea-level rise (roughly 2 mm/y) and the explained rate of sea-level rise from melting of mountain glaciers, thermal expansion of the ocean, etc. (roughly 1 mm/y or less) is similar to the modeled imbalance in the Antarctic (roughly 1 mm/y of sea-level rise; Huybrechts 1990), suggesting a contribution of sea-level rise from the Antarctic.

Relationships between global climate and changes in ice extent are complex. The mass balance of land-based glaciers and ice sheets is determined by the accumulation of snow, mostly in winter, and warm-season ablation due primarily to net radiation and turbulent heat fluxes to melting ice and snow from warm-air advection, (Munro 1990). However, most of Antarctica never experiences surface melting. Where ice masses terminate in the ocean, iceberg calving is the major contributor to mass loss. In this situation, the ice margin may extend out into deep water as a floating ice shelf, such as that in the Ross Sea. Despite the possibility that global warming could result in losses to the Greenland

ice sheet being offset by gains to the Antarctic ice sheet, there is major concern about the possibility of a West Antarctic Ice Sheet collapse. The West Antarctic Ice Sheet is grounded on bedrock below sea level, and its collapse has the potential of raising the world sea level 6–7 m over a few hundred years.

Most of the discharge of the West Antarctic Ice Sheet is via the five major ice streams (faster flowing ice) entering the Ross Ice Shelf, the Rutford Ice Stream entering Ronne-Filchner shelf of the Weddell Sea, and the Thwaites Glacier and Pine Island Glacier entering the Amundsen Ice Shelf. Opinions differ as to the present mass balance of these systems (Bentley 1983, 1985), principally because of the limited data. The West Antarctic Ice Sheet is stable so long as the Ross Ice Shelf is constrained by drag along its lateral boundaries and pinned by local grounding.