Glaciers
(Types, Motion, Morphology & Geology)

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Introduction

The Baltoro Glacier in the Karakoram, Kashmir, Northern Pakistan. 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.

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.

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 °C (−148 °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.

Transportation and erosion

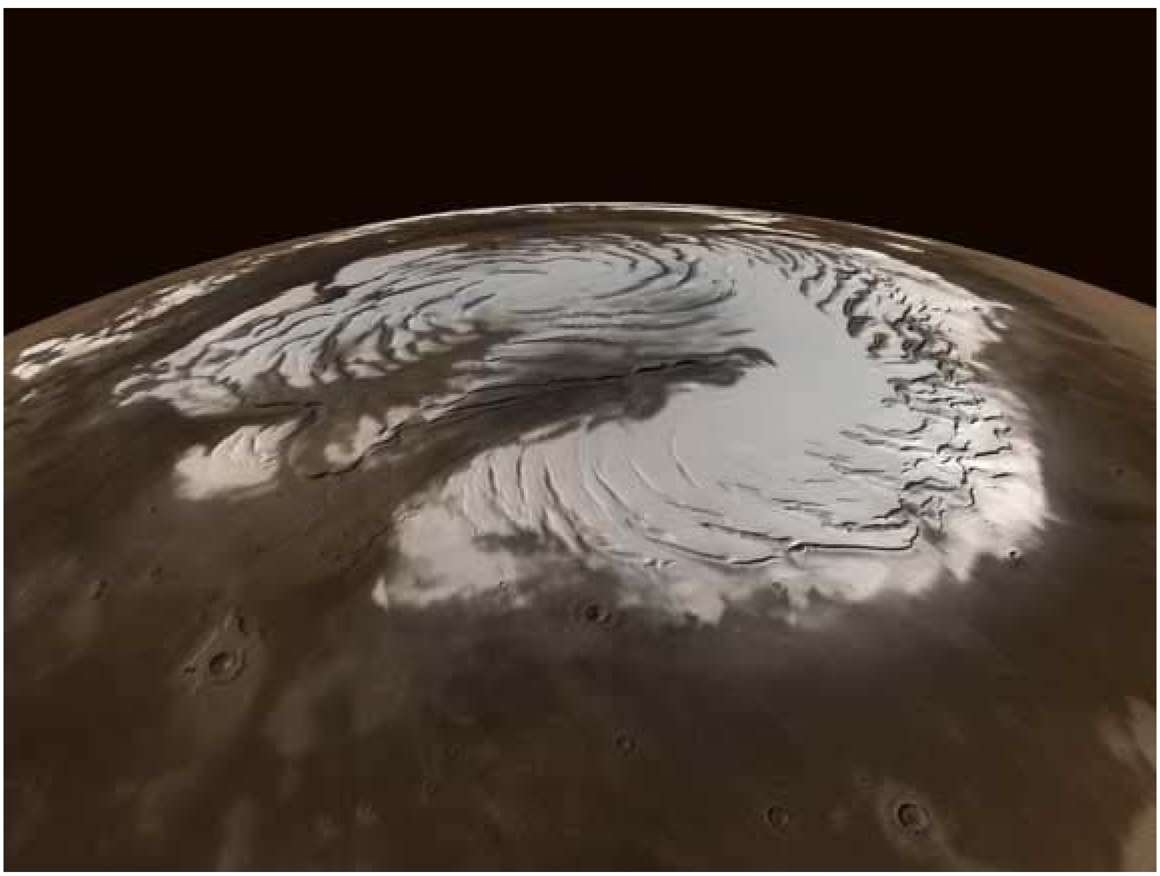
- **Entrainment** is the picking up of loose material by the glacier from along the bed and valley sides. Entrainment can happen by regelation or by the ice simply picking up the debris.
- **Basal Ice Freezing** is thought to be to be made by glaciohydraulic supercooling, though some studies show that 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.

Glaciers on Mars

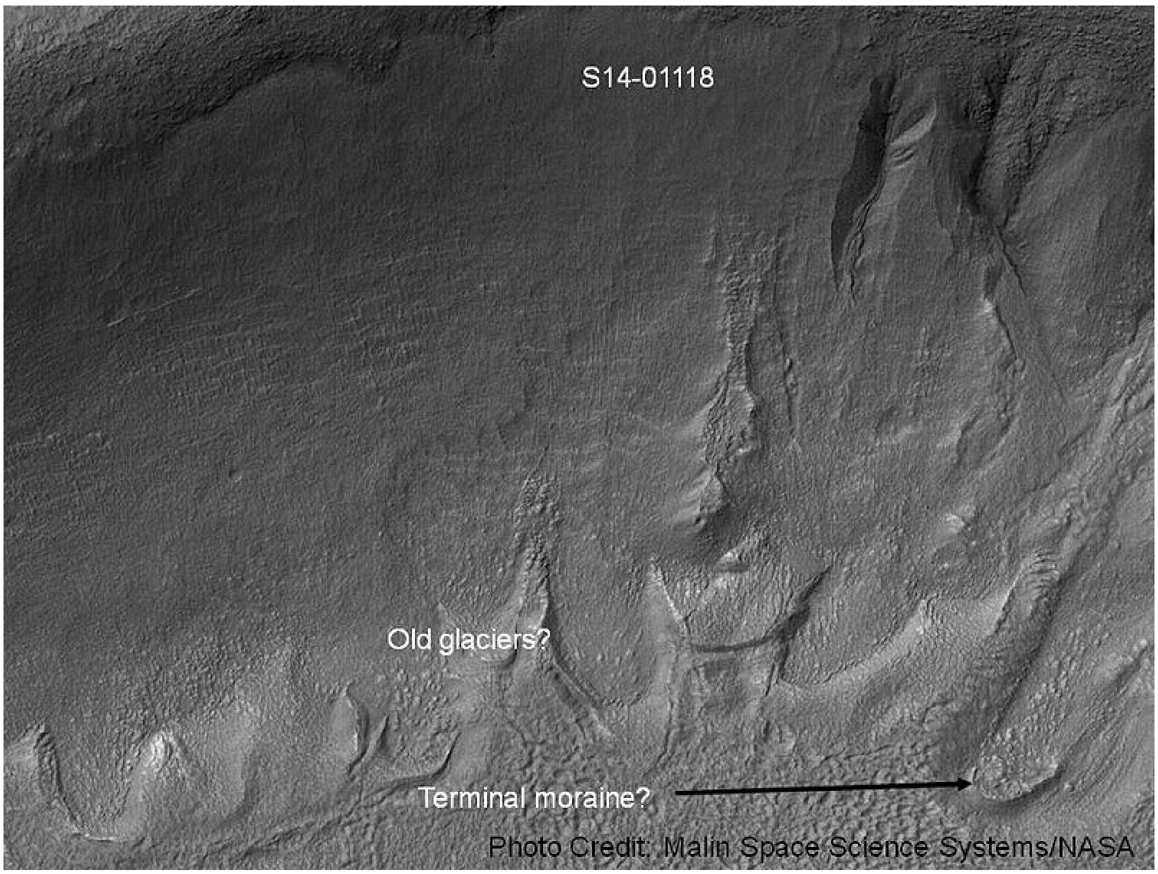


Northern polar icecap on Mars

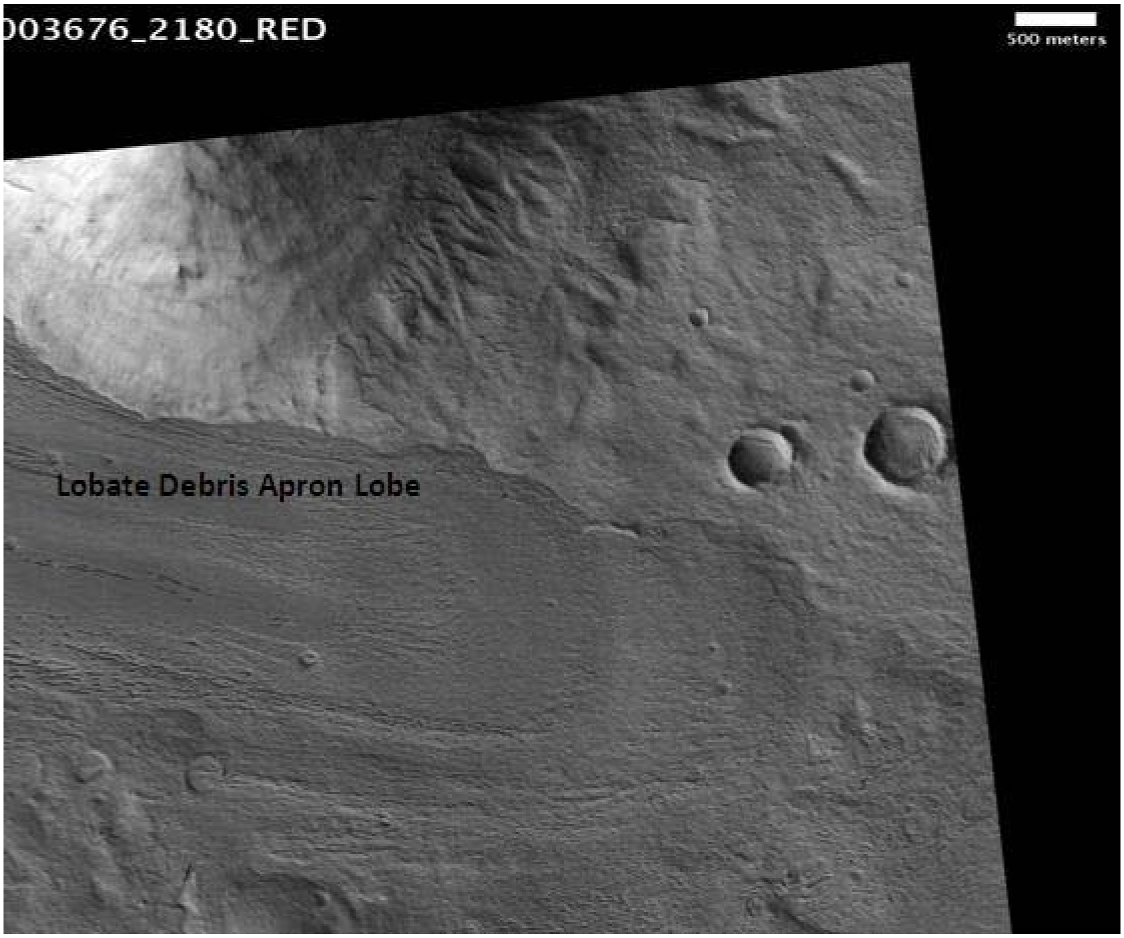
Elsewhere in the solar system, the vast polar ice caps of Mars rival those of the Earth and show glacial features. Especially the south polar cap is compared to glaciers on Earth.

Other glacial features on Mars are glacial debris aprons and the lineated valley fills of the *fretted terrain* in northern Arabia Terra. Topographical features and computer models indicate the existence of more glaciers in Mars' past.

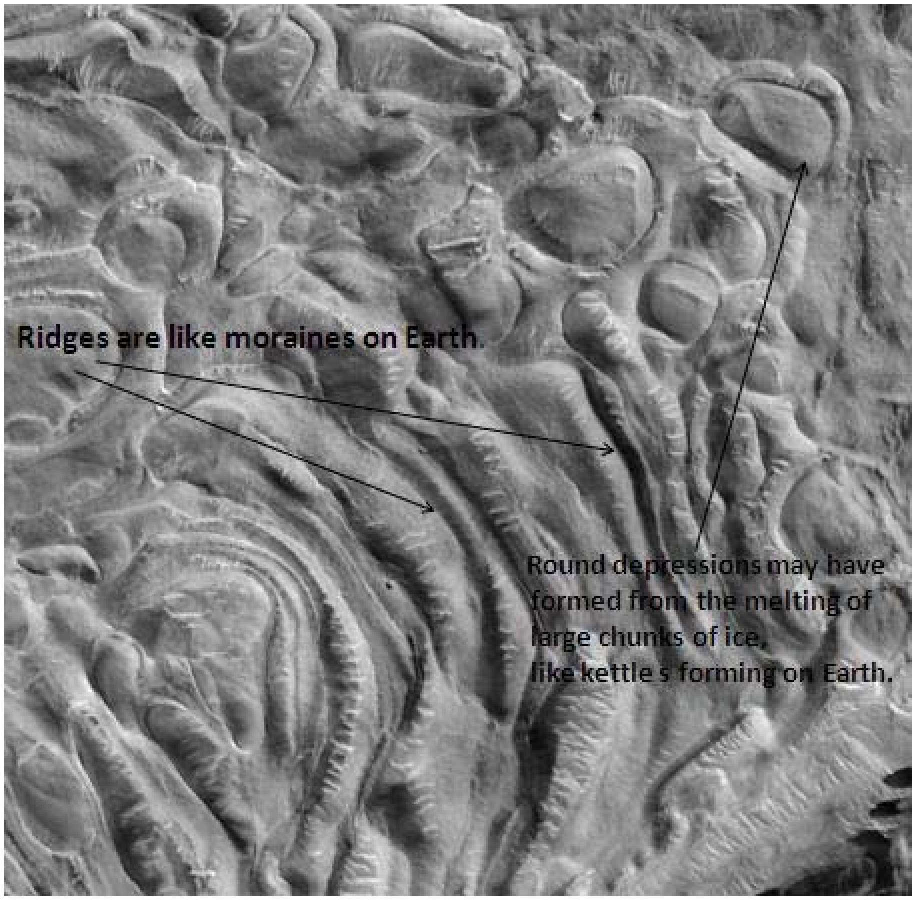
Martian glaciers are affected by the thin atmosphere of Mars. Because of the low atmospheric pressure, ablation near the surface is solely due to sublimation, not melting. As on Earth, many glaciers are covered with a layer of rocks which insulates the ice. A radar instrument onboard the Mars Reconnaisance Orbiter found ice under a thin layer of rocks in formations called Lobate Debris Aprons (LDA's).



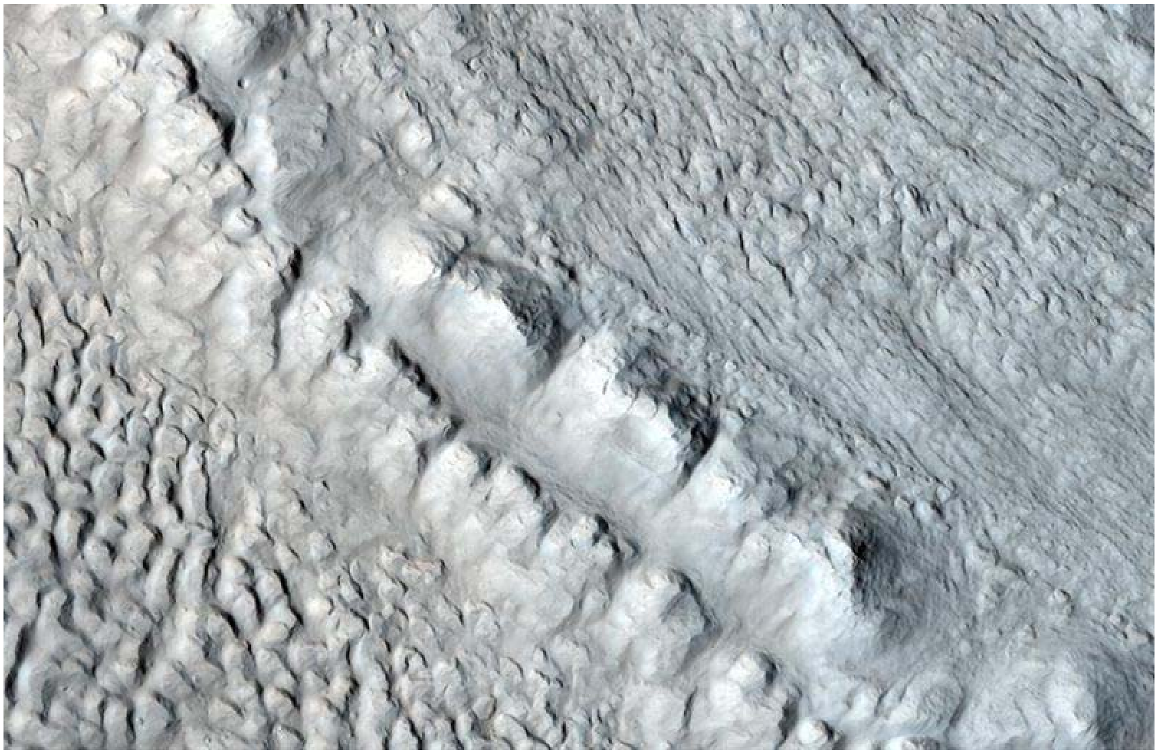
Gullies in a crater in Eridania, north of the large crater Kepler. Also, features that may be remains of old glaciers are present. One, to the right, has the shape of a tongue.



Lobate Debris Apron in Phlegra Montes, Cebrenia quadrangle. The debris apron is probably mostly ice with a thin covering of rock debris, so it could be a source of water for future Martian colonists. Scale bar is 500 meters long.



Moreux Crater moraines and kettle holes, as seen by HiRISE



Enlargement of area in rectangle of the previous image. On Earth the ridge would be called the terminal moraine of an alpine glacier. Picture taken with HiRISE under the HiWish program.

Chapter-1

Types of Glaciers and Morphology

Mouth of the Schlatenkees Glacier near Innergschlöß, 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^2 (20,000 $mile^2$). 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.

Glacier Morphology

Franz Josef Glacier in New Zealand

Glacier morphology, or the form a glacier takes, is influenced by temperature, precipitation, topography, and other factors. Types of glaciers range from massive ice sheets, such as the Greenland ice sheet or those in Antarctica, to small cirque glaciers perched on a mountain. Glaciers types can be grouped into two main categories, based on whether or not ice flow is constrained by the underlying bedrock topography.

Unconstrained

Ice sheets and ice caps

Ice sheet

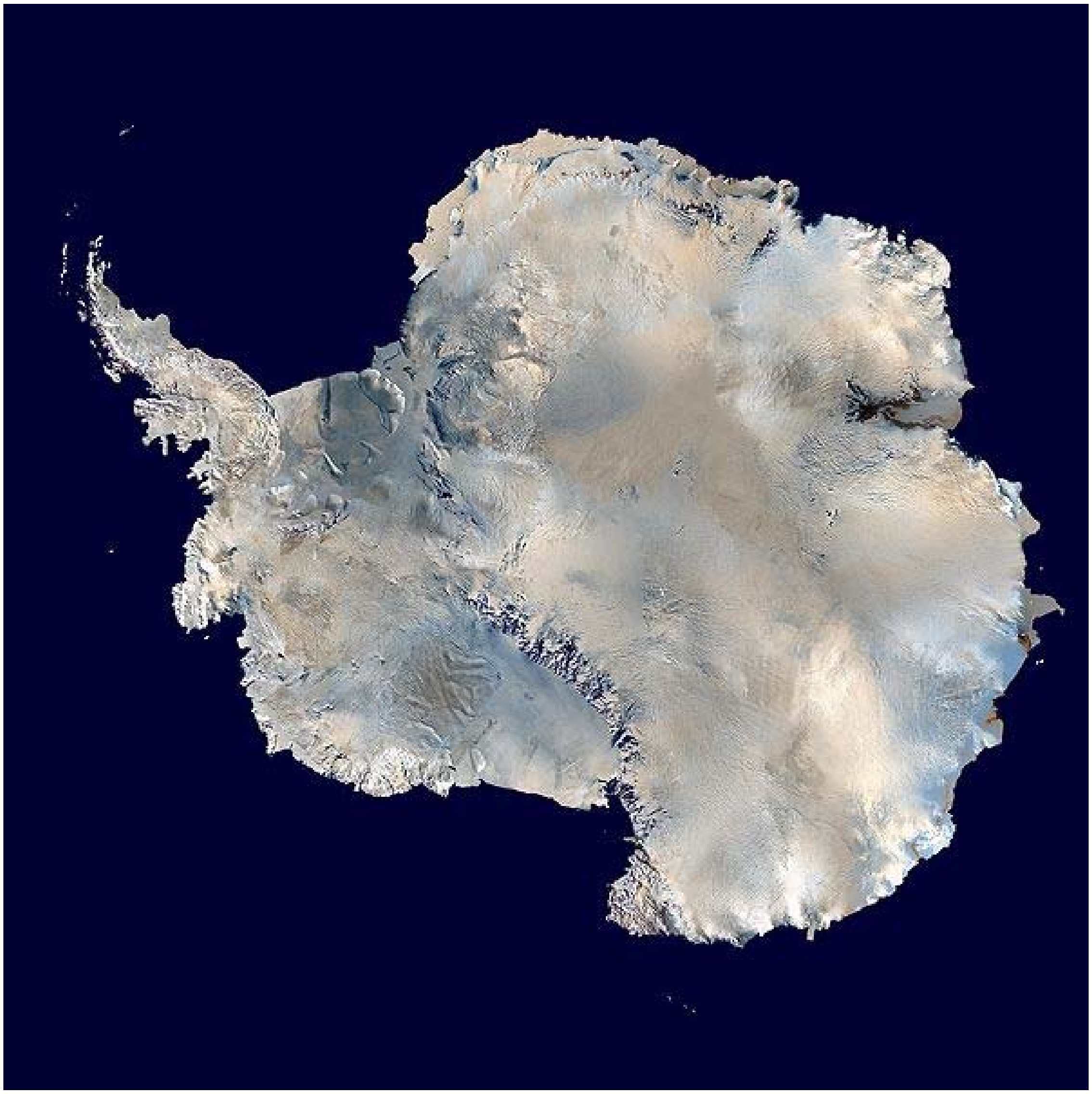
An **ice sheet** is a mass of glacier ice that covers surrounding terrain and is greater than 50,000 km^2 (20,000 $mile^2$), thus also known as **continental glacier**. The only current ice sheets are in Antarctica and Greenland; during the last glacial period at Last Glacial Maximum (LGM) the Laurentide 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^2 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

The Antarctic ice sheet is the largest single mass of ice on Earth. It covers an area of almost 14 million km^2 and contains 30 million km^3 of ice. Around 90% of the fresh water on the Earth's surface is held in the ice sheet, and, if melted, would cause sea levels to rise by 61.1 meters. The continent-wide average surface temperature trend of Antarctica is positive and significant at >0.05°C/decade since 1957.

The Antarctic ice sheet is divided by the Transantarctic Mountains into two unequal sections called the East Antarctic ice sheet (EAIS) and the smaller West Antarctic Ice Sheet (WAIS). The EAIS rests on a major land mass but the bed of the WAIS is, in places, more than 2,500 meters below sea level. It would be seabed if the ice sheet were not there. The WAIS is classified as a marine-based ice sheet, meaning that its bed lies below sea level and its edges flow into floating ice shelves. The WAIS is bounded by the Ross Ice Shelf, the Ronne Ice Shelf, and outlet glaciers that drain into the Amundsen Sea.

Greenland ice sheet

Map of Greenland

The Greenland ice sheet occupies about 82% of the surface of Greenland, and if melted would cause sea levels to rise by 7.2 metres. Estimated changes in the mass of Greenland's ice sheet suggest it is melting at a rate of about 239 cubic kilometres (57.3

cubic miles) per year. These measurements came from NASA's Gravity Recovery and Climate Experiment (GRACE) satellite, launched in 2002, as reported by BBC News in August 2006.

Ice sheet dynamics

Ice movement is dominated by the motion of glaciers, whose activity is determined by a number of processes. Their motion is the result of cyclic surges interspersed with longer periods of inactivity, on both hourly and centennial time scales.

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 ± 0.35 and 0.21 ± 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 cap

Vatnajökull, Iceland

An **ice cap** is an ice mass that covers less than 50 000 km² of land area (usually covering a highland area). Masses of ice covering more than 50 000 km² are termed an ice sheet.

Ice caps are not constrained by topographical features (i.e., they will lie over the top of mountains) but their *dome* is usually centred on the highest point of a massif. Ice flows away from this high point (the ice divide) towards the ice cap's periphery.

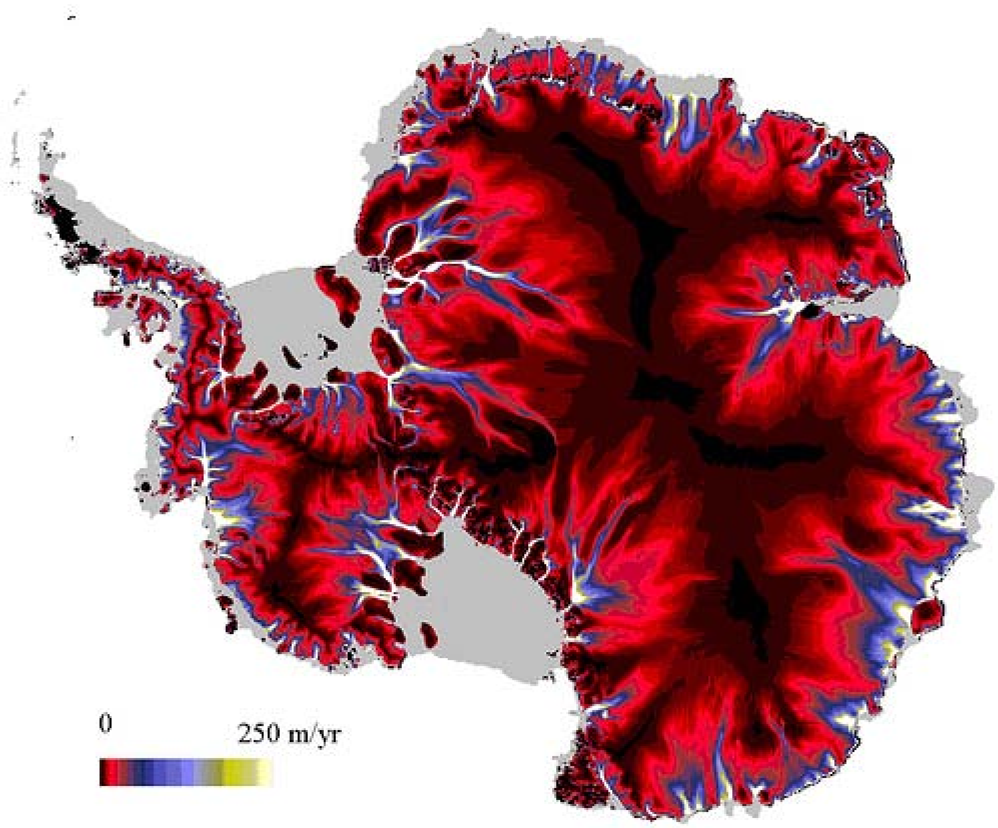
Ice caps have significant effects on the geomorphology of the area they occupy. Plastic moulding, gouging and other glacial erosional features become present upon the glacier's retreat. Many lakes, such as the Great Lakes in North America, as well as numerous valleys have been formed by glacial action over hundreds of thousands of years.

On Earth, there are about 30 million km³ of total ice mass. The average temperature of an ice mass ranges between -20°C and -30°C. The core of an ice cap exhibits a constant temperature that ranges between -15°C and -20°C.

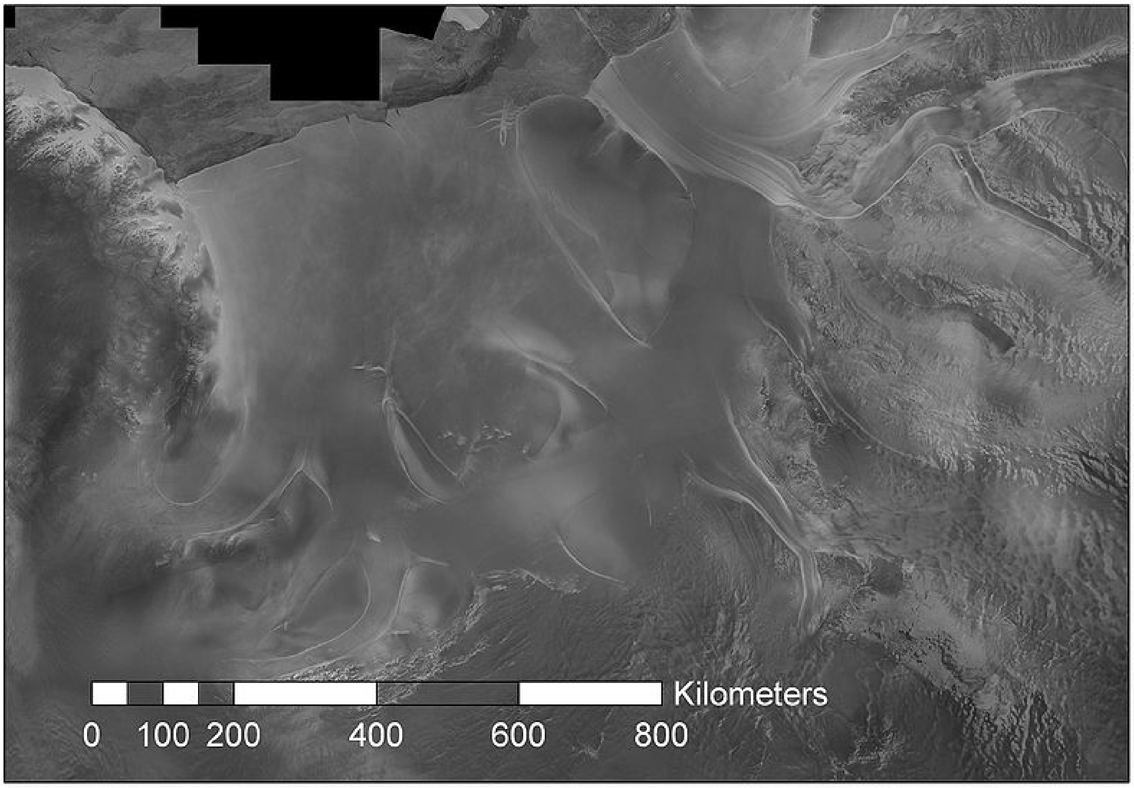
A high-latitude region covered in ice, though strictly not an ice cap (since they exceed the maximum area specified in the definition above), are called polar ice caps; the usage of this designation is widespread in the mass media and arguably recognized by experts.

Vatnajökull is an example of an ice cap in Iceland.

Ice streams



Velocity map of Antarctica. Ice streams can be seen with increasing speeds (blue-yellow-white) flowing toward the coast.



Radarsat image of ice streams flowing into the Filchner-Ronne Ice Shelf

An **ice stream** is a region of an ice sheet that moves significantly faster than the surrounding ice. Ice streams are a type of glacier. They are significant features of the Antarctic where they account for 10% of the volume of the ice. They are up to 50 km wide, 2 km thick, can stretch for hundreds of kilometres, and account for most of the ice leaving the ice sheet.

The speed of an ice stream can be over 1,000 meters per year, an order of magnitude faster than the surrounding ice. The shear forces at the edge of the ice stream cause deformation and recrystallization of the ice, making it softer, and concentrating the deformation in narrow bands or shear margins. Crevasses form, particularly around the shear margins.

Most ice streams have some water at their bases, which lubricates the flow. The type of bedrock also is significant. Soft, deformable sediments result in faster flow than hard rock.

Antarctica

The Antarctic Ice Sheet is drained to the sea by several ice streams. The largest in East Antarctica is Lambert Glacier. In West Antarctica the large Pine Island and Thwaites

Glaciers are currently the most out of balance, with a total net mass loss between them of 85 gigatonnes per year measured in 2006.

It has been suggested that the Antarctic ice sheet is losing mass. The past and ongoing acceleration of ice streams and outlet glaciers is considered to be a significant, if not the dominant cause of this recent imbalance.

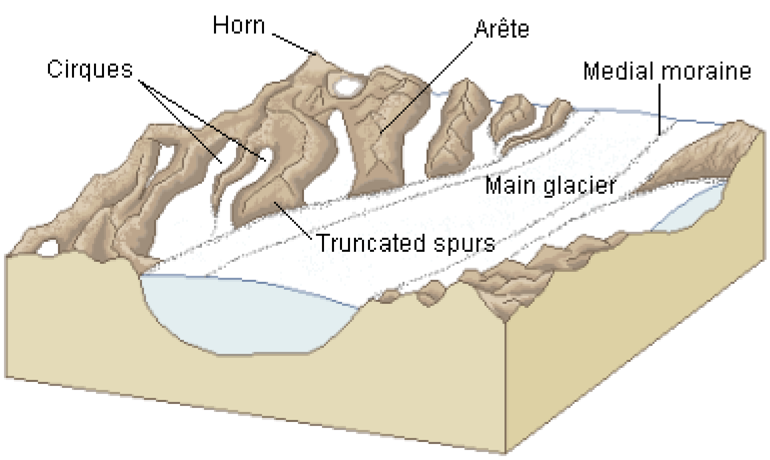
Greenland

Ice streams that drain the Greenland ice sheet into the sea include Helheim Glacier, Jakobshavn Isbræ and Kangerdlugssuaq Glacier.

Ice domes

Ice domes located in the accumulation zone in the higher altitude portions.

Constrained



Features of a glacial landscape

Icefield

Harding Icefield, Kenai National Wildlife Refuge

An **ice field** (also spelled **icefield**) is an area less than 50,000 km² (19,305 mile²) of ice often found in the colder climates and higher altitudes of the world where there is sufficient precipitation. It is an extensive area of interconnected valley glaciers from which the higher peaks rise as nunataks. Ice fields are larger than alpine glaciers, smaller than ice sheets and similar in area to ice caps.

Formation

Ice fields are formed by a large accumulation of snow which, through years of compression and freezing, turns into ice. Due to ice's susceptibility to gravity, ice fields usually form over large areas that are basins or atop plateaus thus allowing a continuum of ice to form over the landscape and not be interrupted by glacial channels. Glaciers often form on the edges of ice fields serving as gravity-propelled drains on the ice field which is in turn replenished by the ice field's snowfall.

While an ice cap is unconstrained by topography, an icefield is. An ice field is also distinguishable from an ice cap because it does not have a dome-like form (Summerfield 1991).

Ice fields of the world

Antarctica

While not technically an ice field (a continental-sized ice sheet rather), Antarctica is extensively covered by ice.

Asia

There are a handful of ice fields in the Himalayas and Altay Mountains (the border range between the Central Asian Republics and China). One unexpected ice field is located in Yolyn Am, a mountain valley located in the northern end of the Gobi Desert.

Oceania

There are no ice fields in Australia.

New Zealand has

- Garden of Eden ice field
- Garden of Allah ice field Map of the gardens
- Olivine Ice PlateauGoogle map reference

Reference: USGS SATELLITE IMAGE ATLAS OF GLACIERS OF THE WORLD, GLACIERS OF IRIAN JAYA, INDONESIA, AND NEW ZEALAND,GLACIERS OF NEW ZEALAND By TREVOR J.H. CHINN

Europe

The only large ice fields in continental Europe are in Norway (e.g., Dovre and Jotunheimen), but these are much smaller than their Canadian or Alaskan counterparts. There are a handful of small ice fields, also, in the southern Alps. Iceland also features a large ice field that covers a high percentage of the island.

North America

One of the more famous North American ice fields is the Columbia Icefield located in the Rocky Mountains between Jasper and Banff, Alberta. Despite its fame, it is actually a comparatively small ice field relative to the America cordillera.

A large number of particularly expansive ice fields lie in the Coast Mountains, Alaska Range, and Chugach Mountains of Alaska, British Columbia, and the Yukon Territory. The 6,500 km² Stikine Icecap (located between the Stikine and Taku Rivers) and the 2,500 km² Juneau Icefield (located between Lynn Canal and the Taku River) both straddle the British Columbian-Alaskan border. Farther north, the Kluane Icecap — which feeds the immense Malaspina and Hubbard Glaciers as well as the Bagley Icefield — sits upon the British Columbia-Yukon Territory-Alaska border and surrounds most of the Saint Elias Mountains as well as both Mount Saint Elias and Mount Logan; it extends as far west as the Copper River.

There are also large ice fields located in the Kenai Peninsula-Chugach Mountains area, such as the Sargent Icefield and the Harding Icefield. Throughout the Alaska Range there also large icefields (including one surrounding Denali), which are mostly unnamed.

South America

In South America, there are two main ice fields, *Campo de Hielo Norte* (translates to Northern Ice Field or Northern Patagonian Ice Field), in Chile, and *Campo de Hielo Sur* (translates to Southern Ice Field or Southern Patagonian Ice Field), shared by Chile and Argentina. There is also a small ice field on the western (Chilean) portion of Tierra del Fuego proper.

An icefield covers a relatively large area, usually located in mountainous terrain. The underlying topography controls or influences the form that an icefield takes. Often, nunataks poke through the surface of icefields. Examples of icefields include the Columbia Icefield in the Canadian Rockies and the Northern and Southern Patagonian Ice Field in Chile.

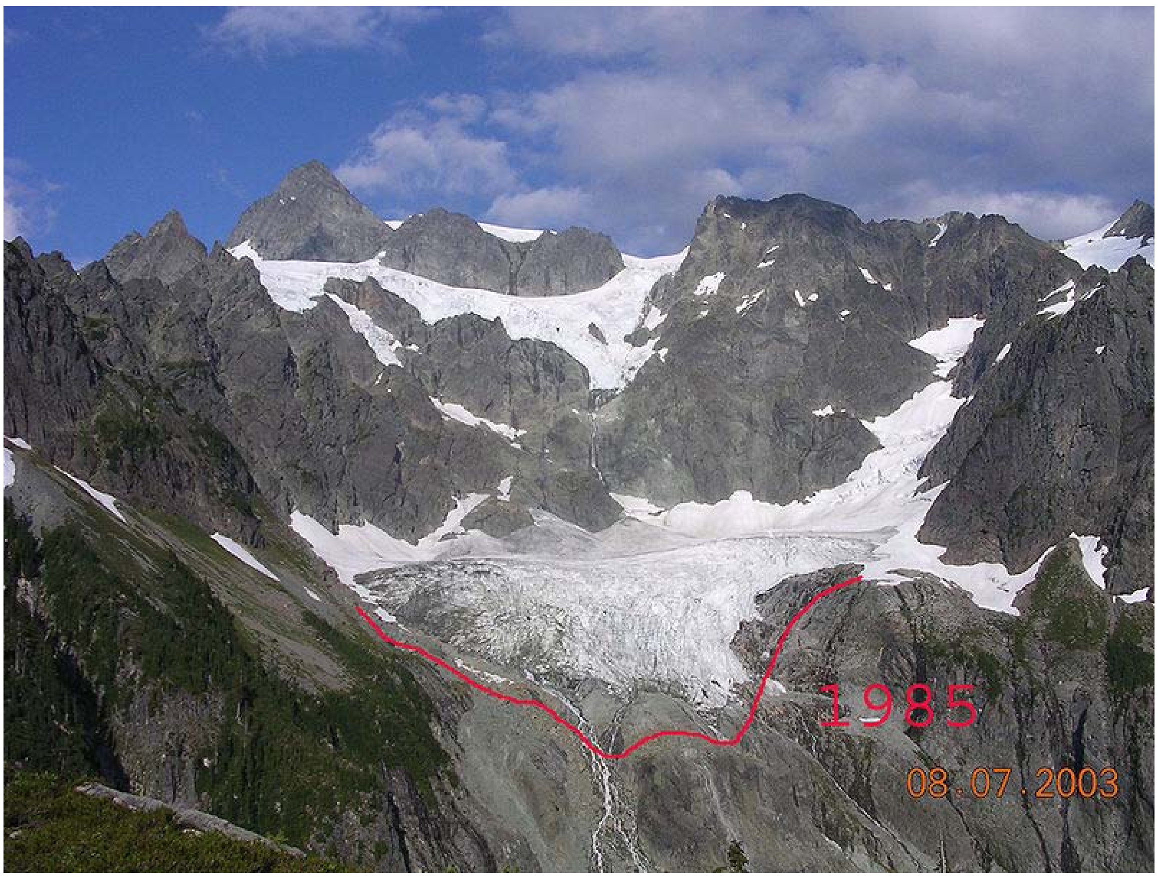
Outlet glaciers

Outlet glaciers are channels of ice that flow out of an ice sheet, but are constrained on the sides with exposed bedrock.

Valley glaciers

Valley glaciers, which provide drainage for icefields, are also constrained by underlying topography. Ice-free exposed bedrock and slopes often surround valley glaciers, providing snow and ice from above to accumulate on the glacier via avalanches. True fjords are formed when valley glaciers retreat and water fills the void (due to land upheaval).

Cirque glaciers



Lower Curtis Glacier is a corrie glacier in the North Cascades in the State of Washington

A **cirque glacier** is formed in a cirque, bowl-shaped depressions on the side of mountains. Snow and ice accumulation in corries often occurs as the result of avalanching from higher surrounding slopes.

In these depressions, snow persists through summer months, and becomes glacier ice. Snow may be situated on the leeward slope of a mountain, where it is sheltered from wind. Rock fall from above slopes also plays an important role in sheltering the snow and ice from sunlight.

Randklufts may form beneath corrie glaciers as open space between the ice and the bedrock, where meltwater can play a role in deposition of the rock.

Chapter-2

Motion of Glaciers

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 *bergshrund*. 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 Greenlands'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

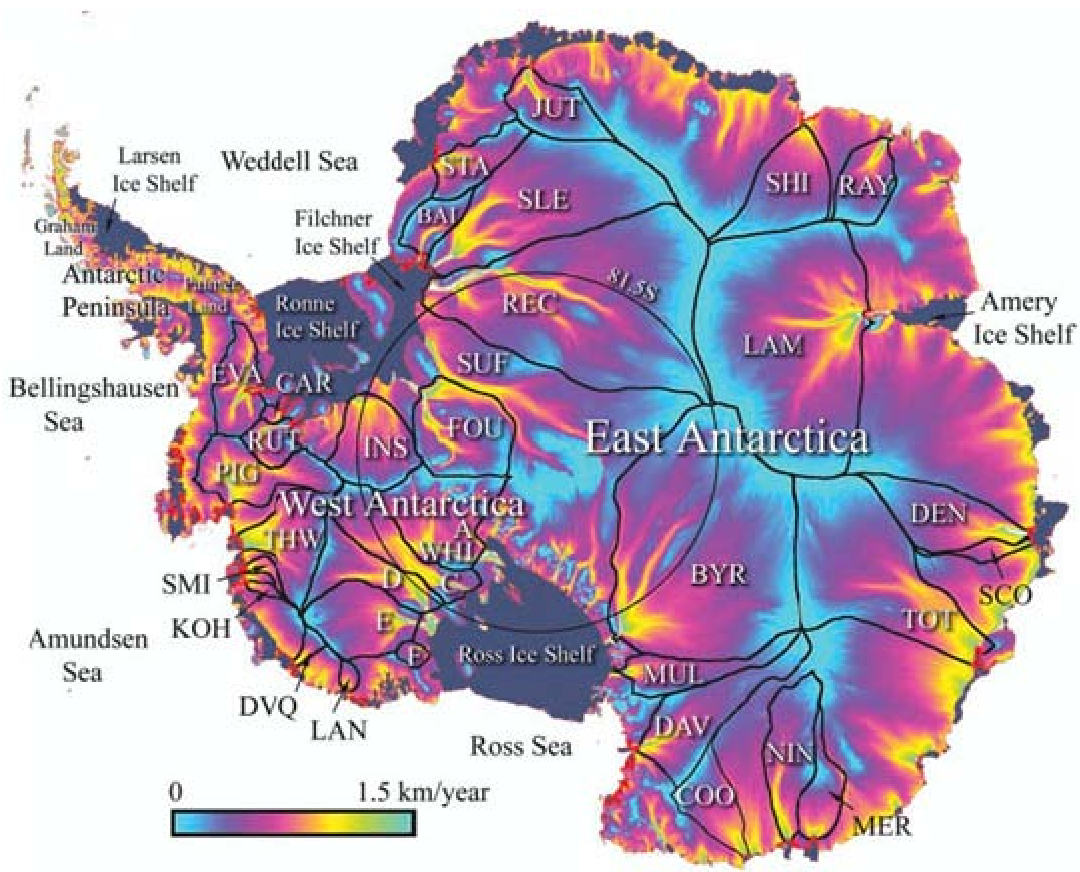
An **ogive** is a geographic feature of certain types of glaciers. They are identified as light and dark bands of ridges and valleys found at the base of glacial icefalls. They are usually found on 'advancing' or 'neutral' glaciers.

Ogives are created by icefalls. The icefalls produce a pattern of alternating ridges and swales that gradually flatten out downglacier. They also bend throughout the glacier, because the velocity of the glacier is greater near the center.

As ice passes through an icefall it is usually very broken-up, which greatly increases the ablation surface area in the summer, and provides space for enhanced entrapment of snowfall during the winter. Winter ice that passes through the icefall becomes bulgy, irregular and hummocky at the base of the icefall, while summer ice turns into a swale.

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.

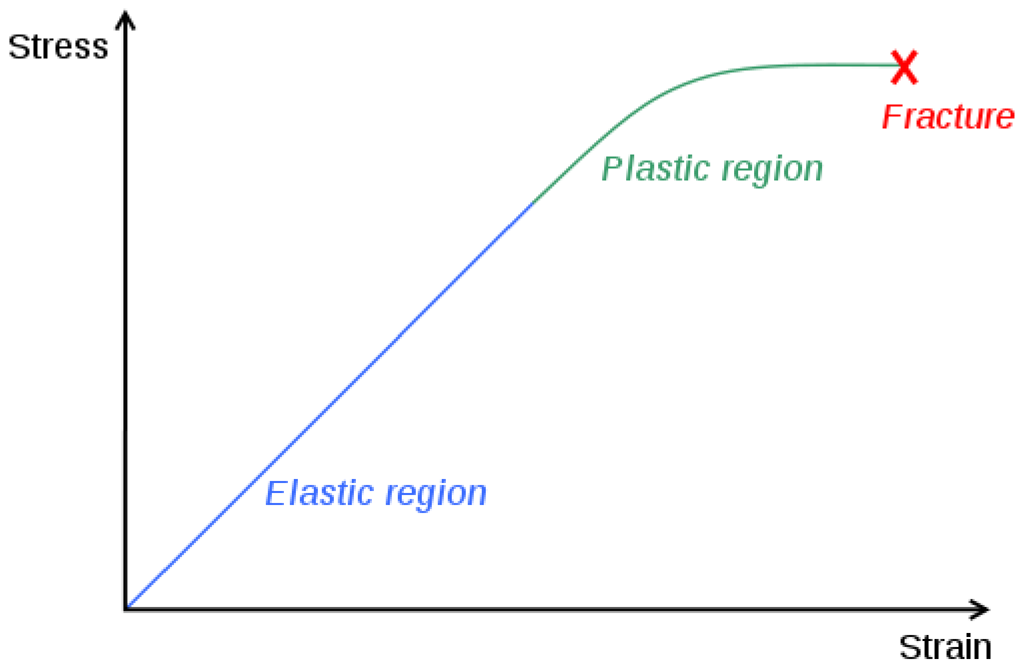
Ice sheet dynamics



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

Σ = shear strain (flow) rate
τ = 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 τ_D is the driving stress, and α the angle of repose.
τ_B is the basal shear stress, a function of bed temperature and softness.

τ_F, the shear stress, is the lower of τ_B and τ_D. It controls the rate of plastic flow, as per the figure (inset, right).

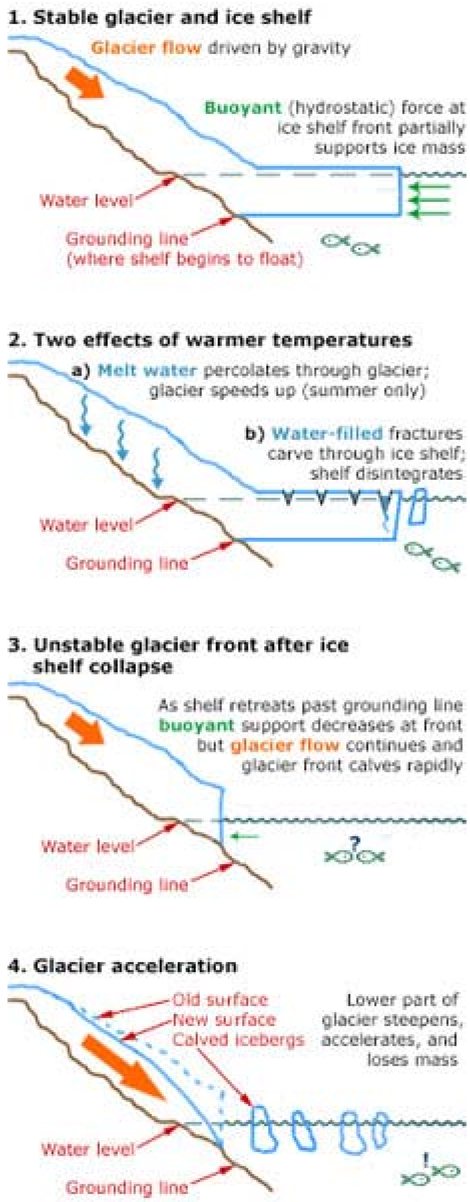
For a given glacier, the two variables are τ_D, which varies with h, the depth of the glacier, and τ_B, the basal shear stress.

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 τ_B). If the sediment strength falls far below τ_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 ρgh. 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, τ_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 ~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.

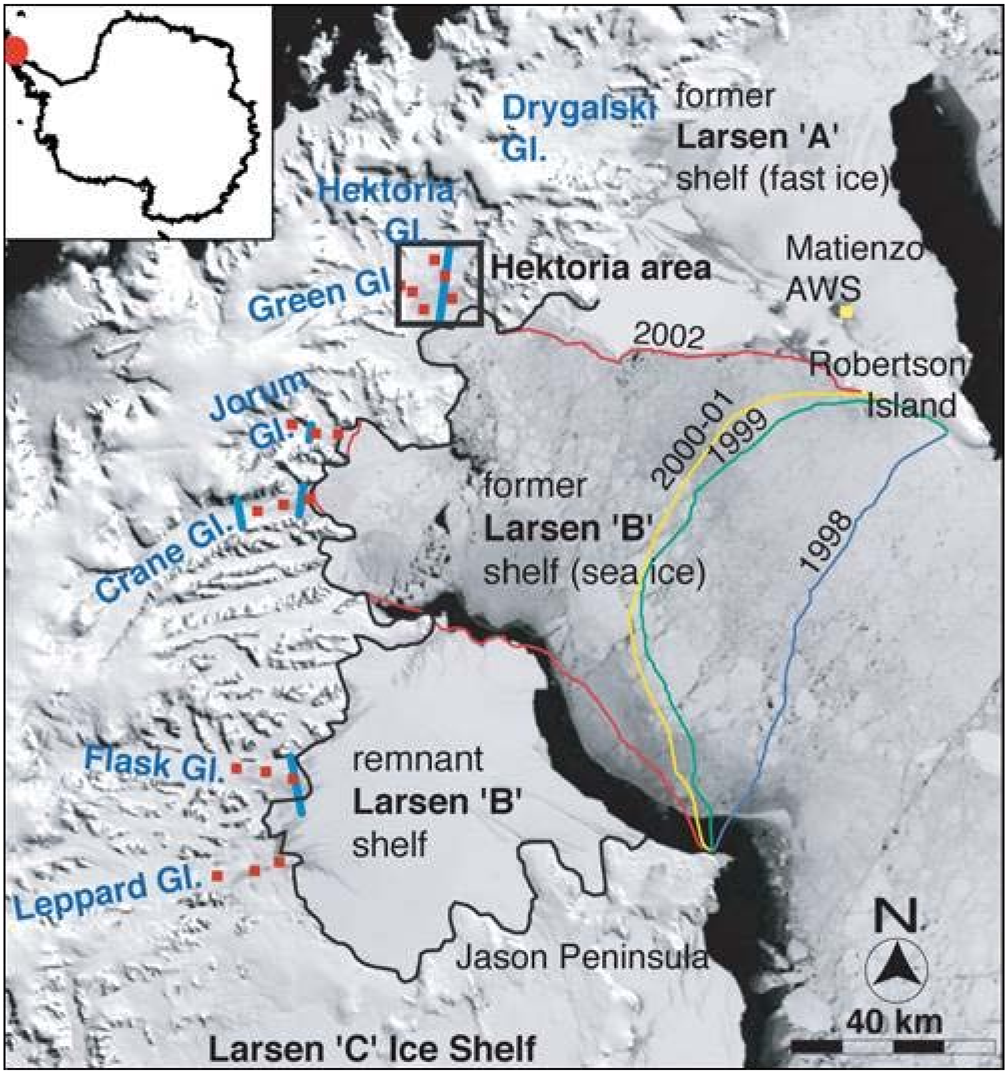
Pipe and sheet flow

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 (Crane, Jorum, Green, Hektoria - see image) 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^{-1} variability in ice shelf thickness.

Long term changes

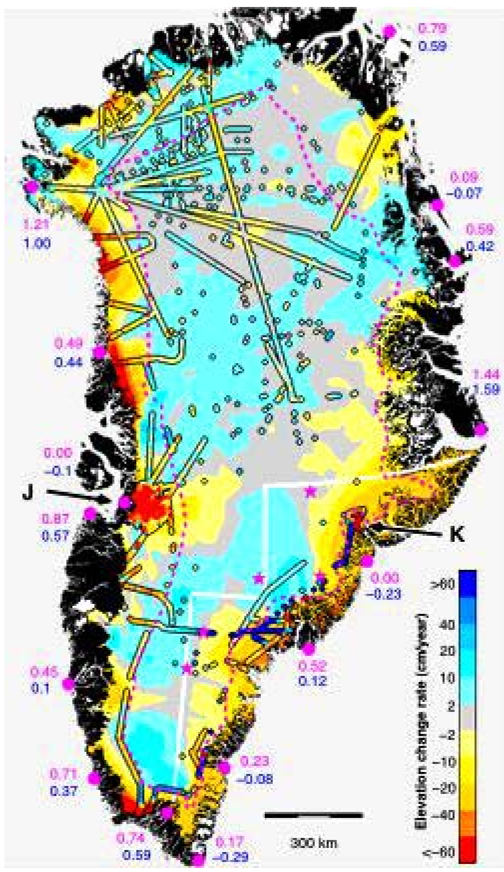
Over long time scales, ice sheet mass balance is governed by the amount of sunlight reaching the earth. This variation in sunlight reaching the earth, or insolation, over geologic time is in turn determined by the angle of the earth to the sun and shape of the Earth's orbit, as it is pulled upon by neighboring 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 peridoicity, 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.

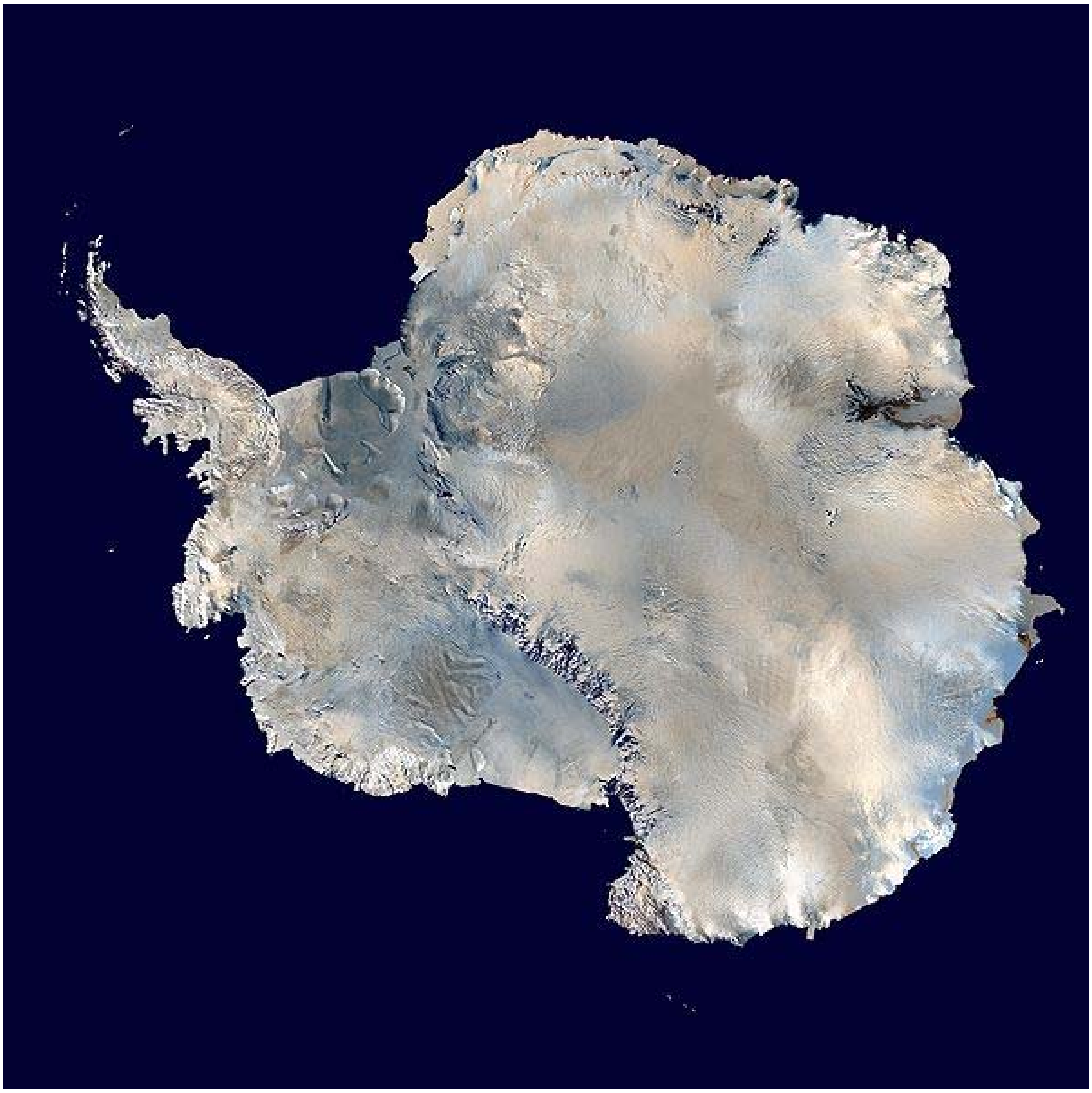
Increasing global temperatures take around 10,000 years to directly propagate through the ice before they influence bed temperatures, but may have an effect through increased surfacal melting, producing more supraglacial lakes, which may feed warm water to glacial bases and facilitate glacial motion.

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.

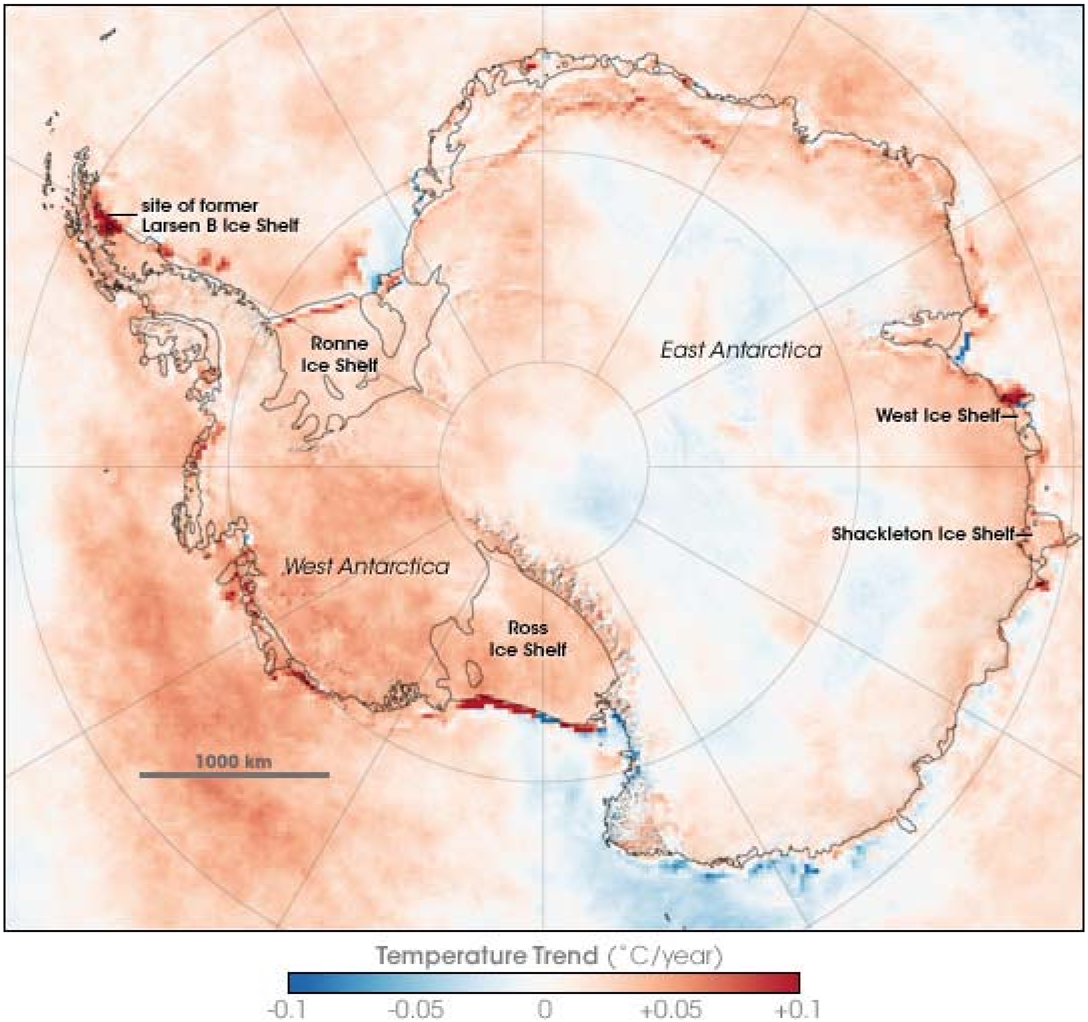
Observations, while currently limited in scope, do agree with these predictions of an increasing rate of ice loss from both Greenland and Antarctica.

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_2 release, further climate change.

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_2 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_2 levels to have been more important.

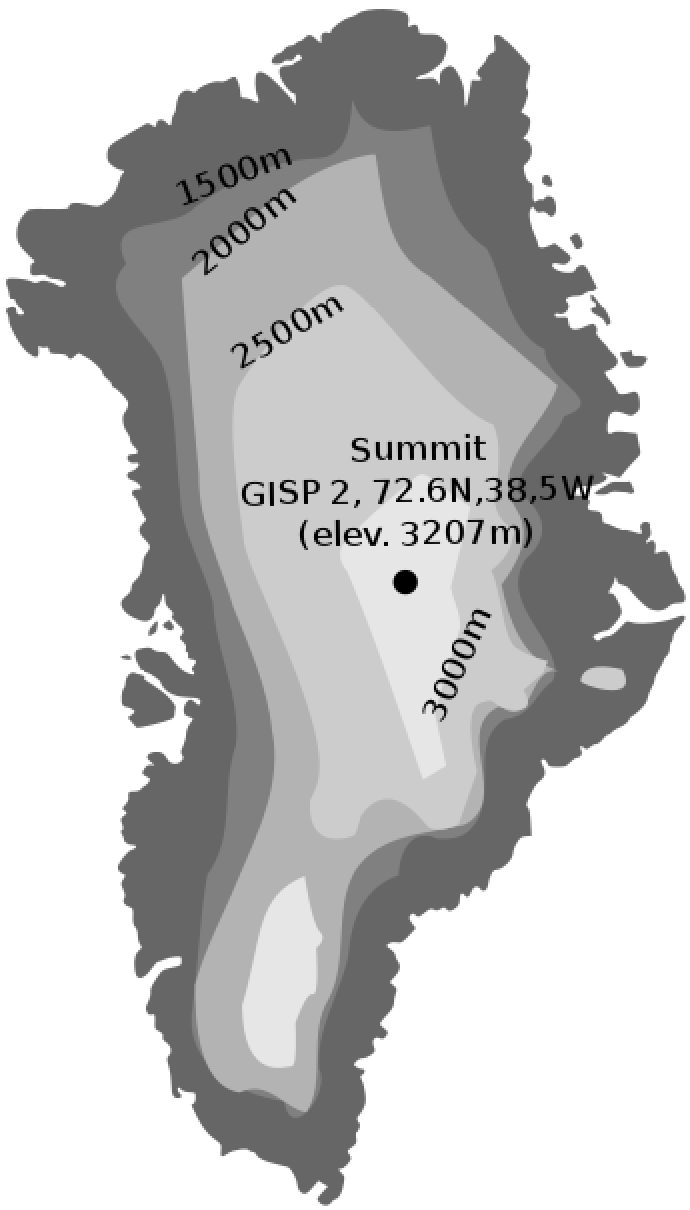
Ice enters the sheet through precipitation as snow. This snow is then compacted to form glacier ice which moves under gravity towards the coast. Most of it is carried to the coast by fast moving ice streams. The ice then passes into the ocean, often forming vast floating ice shelves. These shelves then 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°C/decade since 1957. West Antarctica has warmed by more than 0.1°C/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



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) (see picture) 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 (Kalaallisut: *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 °C (−23.8 °F), occur on the north-central part of the north dome, and temperatures at the crest of the south dome are about −20 °C (−4 °F).

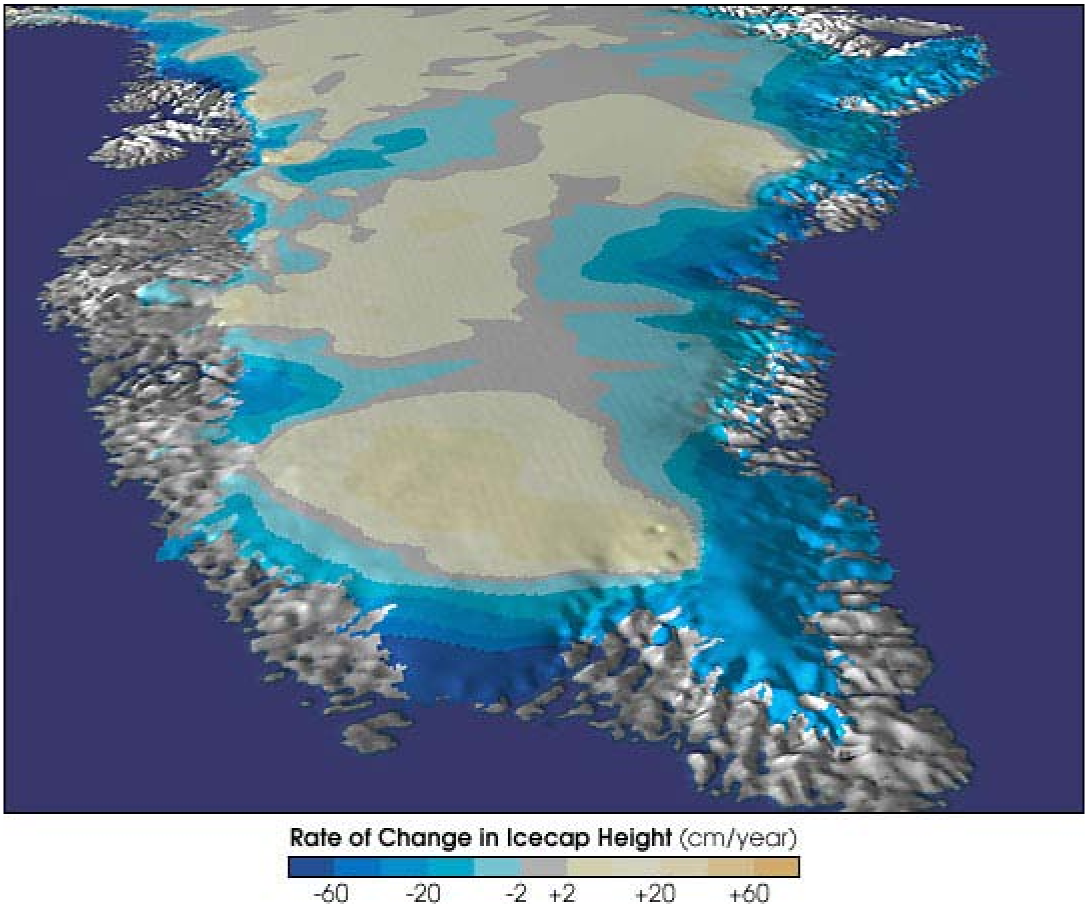
During winter, the ice sheet takes on a clear blue/green color. During summer, the top layer of ice melts leaving pockets of air in the ice that makes it look white.

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.

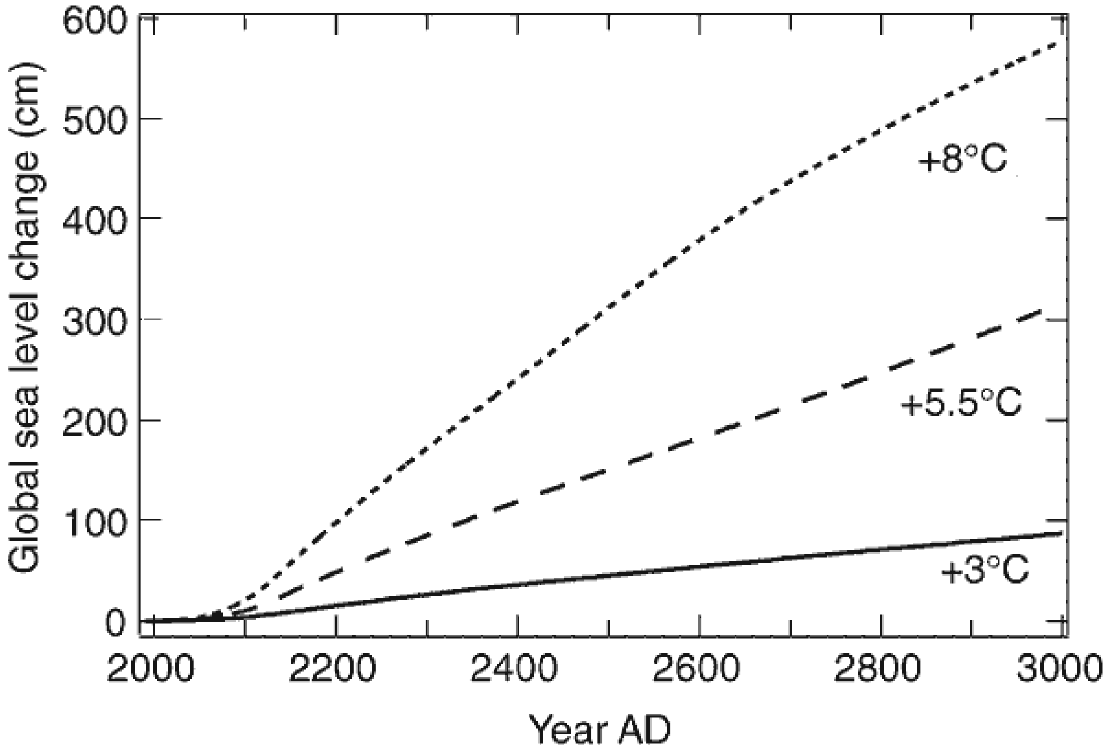
The melting ice sheet



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^3 (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^2) of floating ice between 2000 and 2001.
- Between 2001 and 2005, a breakup of Sermeq Kujalleq erased 36 square miles (93 km^2) 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^2) 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.

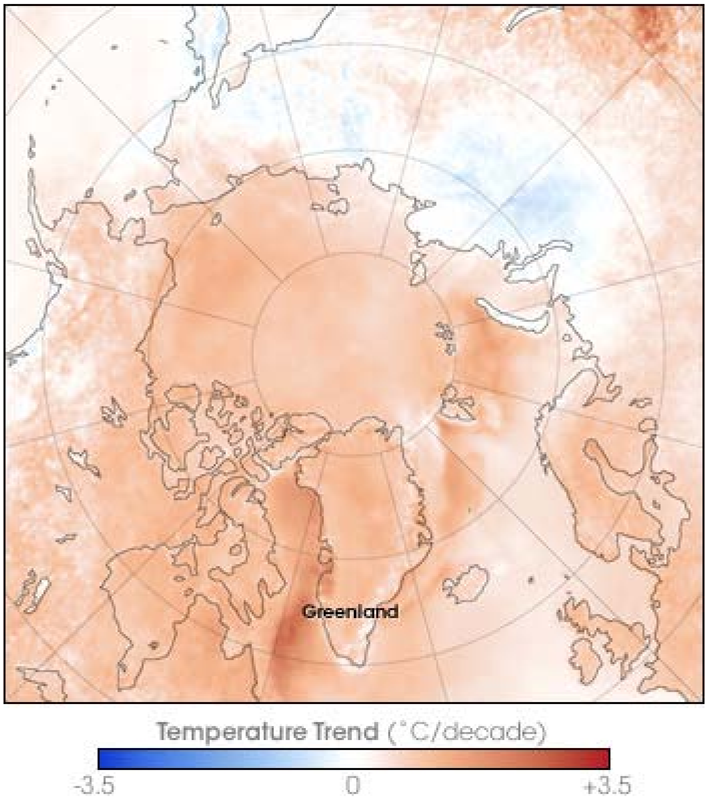
An examination of 32 outlet glaciers in southeast Greenland indicates that the acceleration is significant only for marine terminating outlet glaciers. That is glaciers that calve into the ocean. Further, noted that the thinning of the ice sheet is most pronounced for marine terminating outlet glaciers. 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-1 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 ± 26 Gigatonnes of ice per year, runoff and bottom melting to 297±32 Gt/yr and 32±3 Gt/yr, respectively, and iceberg production to 235±33 Gt/yr. On balance, the IPCC estimates -44 ± 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 km^3 or 23.0 cu mi per year in mass from its ice sheet. In 2005, this had increased to about 220 km^3 or 52.8 cu mi a year due to rapid thinning near its coasts, while in 2006 it was estimated at 239 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 km^3 (142.0 cu mi). Also snowfall was unusually low, which lead to unprecedented negative -65 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 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.

Chapter-3

Retreat of Glaciers Since 1850

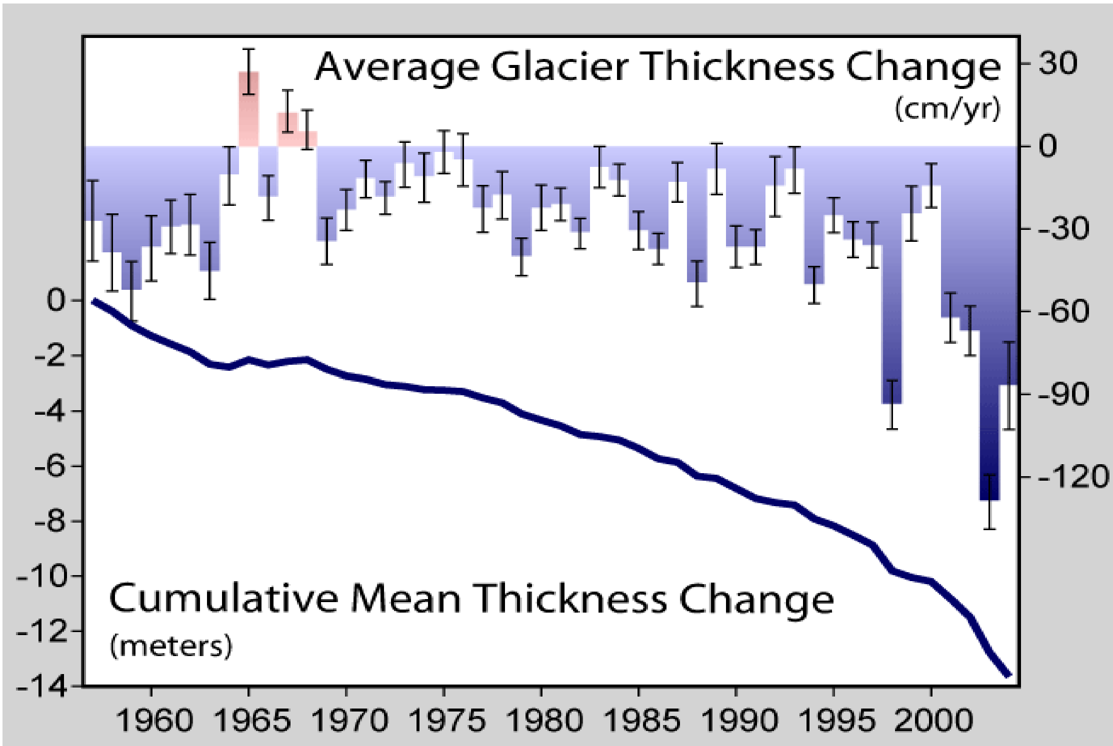
A view down the Whitechuck Glacier in Glacier Peak Wilderness in 1973

In a similar view as seen in 2006, where this branch of glacier retreated 1.9 kilometres (1.2 mi)

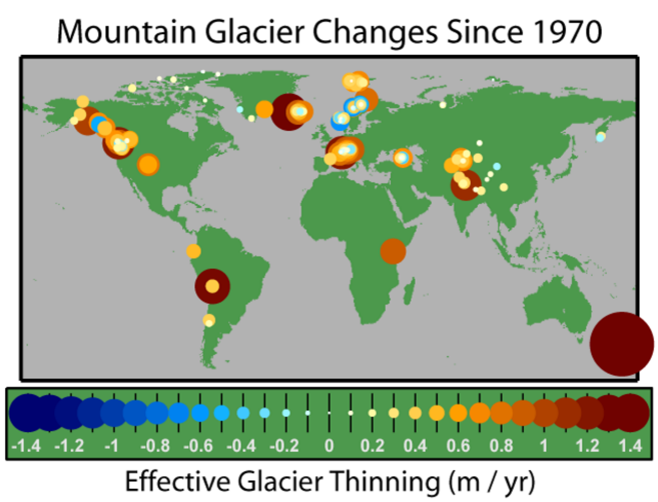
The **retreat of glaciers since 1850** affects the availability of fresh water for irrigation and domestic use, mountain recreation, animals and plants that depend on glacier-melt, and in the longer term, the level of the oceans. Studied by glaciologists, the temporal coincidence of *glacier retreat* with the measured increase of atmospheric greenhouse gases is often cited as an evidentiary underpinning of global warming. Mid-latitude mountain ranges such as the Himalayas, Alps, Rocky Mountains, Cascade Range, and the southern Andes, as well as isolated tropical summits such as Mount Kilimanjaro in Africa, are showing some of the largest proportionate glacial loss.

The Little Ice Age was a period from about 1550 to 1850 when the world experienced relatively cooler temperatures compared to the present. Subsequently, until about 1940, glaciers around the world retreated as the climate warmed substantially. Glacial retreat slowed and even reversed temporarily, in many cases, between 1950 and 1980 as a slight global cooling occurred. Since 1980, a significant global warming has led to glacier retreat becoming increasingly rapid and ubiquitous, so much so that some glaciers have disappeared altogether, and the existence of a great number of the remaining glaciers of the world is threatened. In locations such as the Andes of South America and Himalayas in Asia, the demise of glaciers in these regions will have potential impact on water supplies. The retreat of mountain glaciers, notably in western North America, Asia, the Alps, Indonesia and Africa, and tropical and subtropical regions of South America, has been used to provide qualitative evidence for the rise in global temperatures since the late 19th century. The recent substantial retreat and an acceleration of the rate of retreat since 1995 of a number of key outlet glaciers of the Greenland and West Antarctic ice sheets, may foreshadow a rise in sea level, having a potentially dramatic effect on coastal regions worldwide.

Glacier mass balance



Global glacial mass balance in the last fifty years, reported to the WGMS and NSIDC. The downward trend in the late 1980s is symptomatic of the increased rate and number of retreating glaciers.



Map of mountain glacier mass balance changes since 1970. Thinning in yellow and red, thickening in blue. The 1970s were a decade of more positive mass balance than the 1980-2004 period as seen above.

Crucial to the survival of a glacier is its **mass balance**, the difference between accumulation and ablation (melting and sublimation). Climate change may cause variations in both temperature and snowfall, causing changes in mass balance. Changes in mass balance control a glacier's long term behavior and is the most sensitive climate indicator on a glacier. From 1980-2008 the mean cumulative mass loss of glaciers reporting mass balance to the World Glacier Monitoring Service is -12 m. This includes 19 consecutive years of negative mass balances.

A glacier with a sustained negative balance is out of equilibrium and will retreat, while one with a sustained positive balance is out of equilibrium and will advance. Glacier retreat results in the loss of the low elevation region of the glacier. Since higher elevations are cooler than lower ones, the disappearance of the lowest portion of the glacier reduces overall ablation, thereby increasing mass balance and potentially reestablishing equilibrium. However, if the mass balance of a significant portion of the accumulation zone of the glacier is negative, it is in disequilibrium with the local climate. Such a glacier will melt away with a continuation of this local climate. The key symptom of a glacier in disequilibrium is thinning along the entire length of the glacier. For example, Easton Glacier (pictured below) will likely shrink to half its size, but at a slowing rate of reduction, and stabilize at that size, despite the warmer temperature, over a few decades. However, the Grinnell Glacier (pictured below) will shrink at an increasing rate until it disappears. The difference is that the upper section of Easton Glacier remains healthy and snow-covered, while even the upper section of the Grinnell Glacier is bare, melting and has thinned. Small glaciers with shallow slopes such as Grinnell Glacier are most likely to fall into disequilibrium if there is a change in the local climate.

In the case of positive mass balance, the glacier will continue to advance expanding its low elevation area, resulting in more melting. If this still does not create an equilibrium balance the glacier will continue to advance. If a glacier is near a large body of water, especially an ocean, the glacier may advance until iceberg calving losses bring about equilibrium.

Measurement methods

The Easton Glacier which retreated 255 m from 1990 to 2005 is expected to achieve equilibrium.

Grinnell Glacier in Glacier National Park (U.S.) showing recession since 1850 of 1.1 km
USGS

Mass balance

Mass balance is measured by determining the amount of snow accumulated during winter, and later measuring the amount of snow and ice removed by melting in the summer. The difference between these two parameters is the mass balance. If the amount of snow accumulated during the winter is larger than the amount of melted snow and ice during the summer, the mass balance is positive and the glacier has increased in volume. On the other hand, if the melting of snow and ice during the summer is larger than the supply of snow in the winter, the mass balance is negative and the glacier volume decreases. Mass balance is reported in meters of water equivalent. This represents the average thickness gained (positive balance) or lost (negative balance) from the glacier during that particular year.

To determine mass balance in the accumulation zone, snowpack depth is measured using probing, snowpits or crevasse stratigraphy. Crevasse stratigraphy makes use of annual layers revealed on the wall of a crevasse. Akin to tree rings, these layers are due to summer dust deposition and other seasonal effects. The advantage of crevasse stratigraphy is that is provides a two-dimensional measurement of the snowpack layer, not a point measurement. It is also usable in depths where probing or snowpits are not feasible. In temperate glaciers, the insertion resistance of a probe increases abruptly when its tip reaches ice that was formed the previous year. The probe depth is a measure of the net accumulation above that layer. Snowpits dug through the past winters residual snowpack are used to determine the snowpack depth and density. The snowpack's mass balance is the product of density and depth. Regardless of depth measurement technique the observed depth is multiplied by the snowpack density to determine the accumulation in water equivalent. It is necessary to measure the density in the spring as snowpack density varies. Measurement of snowpack density completed at the end of the ablation season yield consistent values for a particular area on temperate alpine glaciers and need not be measured every year. In the ablation zone, ablation measurements are made using stakes inserted vertically into the glacier either at the end of the previous melt season or the beginning of the current one. The length of stake exposed by melting ice is measured at the end of the melt (ablation) season. Most stakes must be replaced each year or even mid-way through the summer.

Measuring snowpack in a crevasse on the Easton Glacier, North Cascades, USA, the two dimensional nature of the annual layers is apparent

Measuring snowpack on the Taku Glacier in Alaska, this is a slow and inefficient process, but is very accurate

Net balance

Net balance is the mass balance determined between successive mass balance minimums. This is the stratigraphic method focusing on the minima representing a stratigraphic horizon. In the northern mid-latitudes, a glacier's year follows the hydrologic year, starting and ending near the beginning of October. The mass balance minimum is the end of the melt season. The net balance is then the sum of the observed winter balance (bw) normally measured in April or May and summer balance (bs) measured in September or early October.

Measuring snowpack on the Easton Glacier by probing to the previous impenetrable surface, this provides a quick accurate point measurement of snowpack

Annual balance

Annual balance is the mass balance measured between specific dates. The mass balance is measured on the fixed date each year, again sometime near the start of October in the mid northern latitudes.

Geodetic methods

Geodetic methods are an indirect method for the determination of mass balance of glacier. Maps of a glacier made at two different points in time can be compared and the difference in glacier thickness observed used to determine the mass balance over a span of years. This is best accomplished today using Differential Global Positioning System. Sometimes the earliest data for the glacier surface profiles is from images that are used to make topographical maps and digital elevation models. Aerial mapping or photogrammetry is now used to cover larger glaciers and icecaps such found in Antarctica and Greenland, however, because of the problems of establishing accurate ground control points in mountainous terrain, and correlating features in snow and where shading is common, elevation errors are typically not less than 10 m (32 ft). Laser altimetry provides a measurement of the elevation of a glacier along a specific path, e.g., the glacier centerline. The difference of two such measurements is the change in thickness, which provides mass balance over the time interval between the measurements. Again a good method over a span of time but not for annual change detection. The value of geodetic programs is providing an independent check of traditional mass balance work, by comparing the cumulative changes over ten or more years.

Mass balance research worldwide

Mass balance studies have been carried out in various countries worldwide, but have mostly conducted in the Northern Hemisphere due to there being more mid-latitude glaciers in that hemisphere. The World Glacier Monitoring Service annually compiles the mass balance measurements from around the world. From 2002-2006, continuous data is available for only 7 glaciers in the southern hemisphere and 76 glaciers in the Northern Hemisphere. The mean balance of these glaciers was its most negative in any year for 2005/06. The similarity of response of glaciers in western North America indicates the large scale nature of the driving climate change.

Alaska

The Taku Glacier near Juneau, Alaska has been studied by the Juneau Icefield Research Program since 1946, and is the longest continuous mass balance study of any glacier in North America. Taku is the world's thickest known temperate alpine glacier, and experienced positive mass balance between the years 1946 and 1988, resulting in a huge advance. The glacier has since been in a negative mass balance state, which may result in a retreat if the current trends continue. The Juneau Icefield Research Program also has studied the mass balance of the Lemon Creek Glacier since 1953. The glacier has had an average annual balance of -0.44 m per year from 1953–2006, resulting in a mean loss of over 27 m of ice thickness. This loss has been confirmed by laser altimetry.

Austrian Glacier Mass Balance

The mass balance of Hintereisferner and Kesselwandferner glaciers in Austria have been continuously monitored since 1952 and 1965 respectively. Having been continuously measured for 55 years, Hintereisferner has one of the longest periods of continuous study of any glacier in the world, based on measured data and a consistent method of evaluation. Currently this measurement network comprises about 10 snow pits and about 50 ablation stakes distributed across the glacier. In terms of the cumulative specific balances, Hintereisferner experienced a net loss of mass between 1952 and 1964, followed by a period of recovery till 1968. Hintereisferner reached an intermittent minimum in 1976, briefly recovered in 1977 and 1978 and has continuously lost mass in the 30 years since then. Total mass loss has been 26 m since 1952 Sonnblickkees Glacier has been measured since 1957 and the glacier has lost 12 m of mass, an average annual loss of -0.23 m per year.

New Zealand

Glacier mass balance studies have been ongoing in New Zealand since 1957. Tasman Glacier has been studied since then by the New Zealand Geological Survey and later by the Ministry of Works, measuring the ice stratigraphy and overall movement. However, even earlier fluctuation patterns were documented on Franz Josef and Fox Glaciers in 1950. Other glaciers on the South Island studied include Ivory Glacier since 1968, while on the North Island, glacier retreat and mass balance research has been conducted on the glaciers on Mount Ruapehu since 1955. On Mount Ruapehu, permanent photographic stations allow repeat photography to be used to provide photographic evidence of changes to the glaciers on the mountain over time.

An aerial photograpic survey of 50 glaciers in the South Island has been carried out for most years since since 1977. The data was used to show that between 1976 and 2005 there was a 10% loss in glacier volume.

North Cascade glacier mass balance program

The North Cascade Glacier Climate Project measures the annual balance of 10 glaciers, more than any other program in North America. These records extend from 1984–2008 and represent the only set of records documenting the mass balance changes of an entire glacier clad range. To monitor an entire glaciated mountain range in North America, which was listed as a high priority of the National Academy of Sciences in 1983. North Cascade glaciers annual balance has averaged −0.48 m/a from 1984–2008, a cumulative thickness loss of over 13 m or 20–40% of their total volume since 1984 due to negative mass balances. The trend in mass balance is becoming more negative which is fueling more glacier retreat and thinning.

Norway mass balance program

Norway maintains the most extensive mass balance program in the world and is largely funded by the hydropower industry. Mass balance measurements are currently performed on twelve glaciers in Norway. In southern Norway six of the glaciers have been measured for 42 consecutive years or more, and they constitute a west-east profile reaching from the very maritime Ålfotbreen Glacier, close to the western coast, to the very continental Gråsubreen Glacier, in the eastern part of Jotunheimen. Storbreen Glacier in Jotunheimen has been measured for a longer period of time than any other glacier in Norway, a total of over 55 years, while Engabreen Glacier has the longest series (35 years) in northern Norway. The Norwegian program is where the traditional methods of mass balance measurement were largely derived.

Sweden Storglaciären

The Tarfala Research Station in the Kebnekaise region of northern Sweden is operated by Stockholm University. It was here that the first mass balance program was initiated immediately after World War II, and continues to the present day. This survey was the initiation of the mass balance record of Storglaciären Glacier, and constitutes the longest continuous study of this type in the world. Storglaciären has had a cumulative negative mass balance from 1946-2006 of -17 m. The program began monitoring the Rabots Glaciär in 1982, Riukojietna in 1985, and Mårmaglaciären in 1988. All three of these glaciers have had a strong negative mass balance since initiation.

Iceland Glacier mass balance

Glacier mass balance is measured once or twice annually on numerous stakes on the several ice caps in Iceland by the National Energy Authority. Regular pit and stake mass-balance measurements have been carried out on the northern side of Hofsjökull since 1988 and likewise on the Þrándarjökull since 1991. Profiles of mass balance (pit and stake) have been established on the eastern and south-western side of Hofsjökull since 1989. Similar profiles have been assessed on the Tungnaárjökull, Dyngjujökull, Köldukvíslarjökull and Brúarjökull outlet glaciers of Vatnajökull since 1992 and the Eyjabakkajökull outlet glacier since 1991.

Swiss mass balance program

Temporal changes in the spatial distribution of the mass balance result primarily from changes in accumulation and melt along the surface. As a consequence, variations in the mass of glaciers reflect changes in climate and the energy fluxes at the Earth's surface. The Swiss glaciers Gries in the central Alps and Silvretta in the eastern Alps, have been measured for many years. The distribution of seasonal accumulation and ablation rates are measured in-situ. Traditional field methods are combined with remote sensing techniques to track changes in mass, geometry and the flow behaviour of the two glaciers. These investigations contribute to the Swiss Glacier Monitoring Network and the International network of the World Glacier Monitoring Service (WGMS).

United States Geological Survey (USGS)

The USGS operates a long-term "benchmark" glacier monitoring program which is used to examine climate change, glacier mass balance, glacier motion, and stream runoff. This program has been ongoing since 1965 and has been examining three glaciers in particular. Gulkana Glacier in the Alaska Range and Wolverine Glacier in the Coast Ranges of Alaska have both been monitored since 1965, while the South Cascade Glacier in Washington State has been continuously monitored since the International Geophysical Year of 1957. This program monitors one glacier in each of these mountain ranges, collecting detailed data to understand glacier hydrology and glacier climate interactions.

Geological Survey of Canada-Glaciology Section (GSC)

The GSC operates Canada's Glacier-Climate Observing System as part of its Climate Change Geoscience Program. With its University partners, it conducts monitoring and research on glacier-climate changes, water resources and sea level change using a network of reference observing sites located in the Cordillera and the Canadian Arctic Archipelago. This network is augmented with remote sensing assessments of regional glacier changes. Sites in the Cordillera include the Helm, Place, Andrei, Kaskakwulsh, Haig, Peyto, Ram River, Castle Creek, Kwadacha and Bologna Creek Glaciers; in the Arctic Archipelago include the White, Baby and Grise Glaciers and the Devon, Meighen, Melville and Agassiz Ice Caps. GSC reference sites are monitored using the standard stake based glaciological method (stratigraphic) and periodic geodetic assessments using airborne lidar. Detailed information, contact information and database available here: Helm Glacier (-33 m) and Place Glacier (-27 m) have lost more than 20% of their entire volume, since 1980, Peyto Glacier (-20 m) is close to this amount. The Canadian Arctic White Glacier has not been as negative at (-6 m) since 1980.

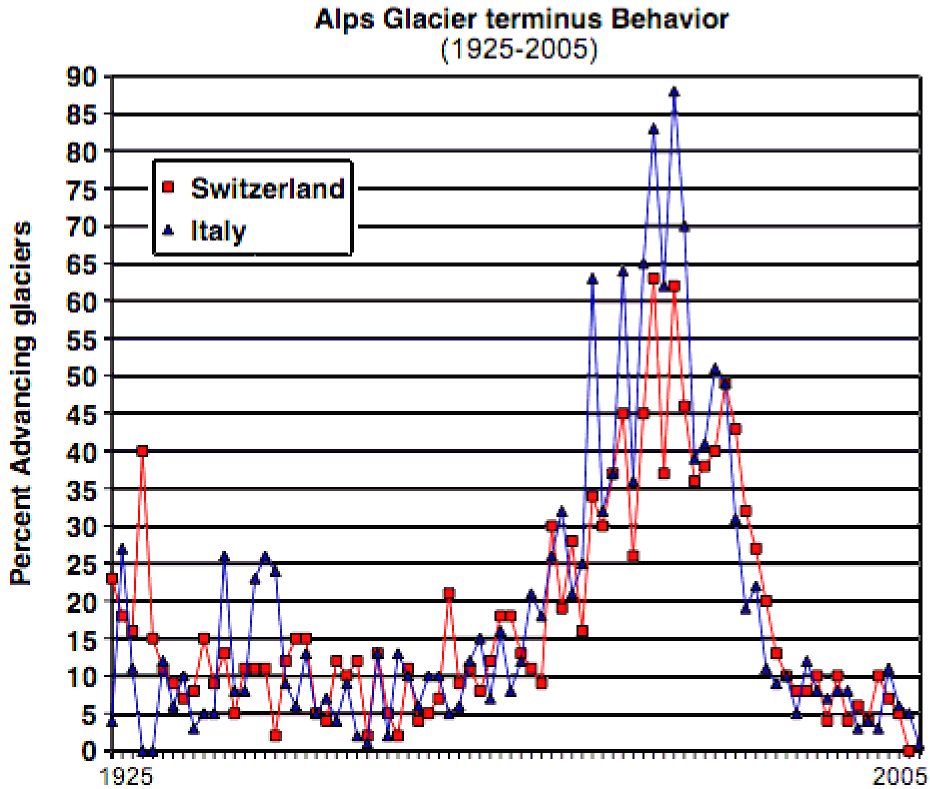
Bolivia mass balance network

The glacier monitoring network in Bolivia, a branch of the glacio-hydrological system of observation installed throughout the tropical Andes mountains by IRD and partners since 1991, has monitored mass balance on Zongo (6000 m asl), Chacaltaya (5400 m asl) and Charquini glaciers (5380 m asl). A system of stakes has been used, with frequent field observations, as often as monthly. These measurements have been made in concert with energy balance to identify the cause of the rapid retreat and mass balance loss of these tropical glaciers.

Mid-latitude glaciers

Mid-latitude glaciers are located either between the Tropic of Cancer and the Arctic Circle, or between the Tropic of Capricorn and the Antarctic Circle. These two regions support glacier ice from mountain glaciers, valley glaciers and even smaller icecaps, which are usually located in higher mountainous regions. All of these glaciers are located in mountain ranges, notably the Himalayas; the Alps; the Pyrenees; Rocky Mountains and Pacific Coast Ranges of North America; the Patagonian Andes in South America; and mountain ranges in New Zealand. Glaciers in these latitudes are more widespread and tend to be more massive the closer they are located to the polar regions. These glaciers are the most widely studied over the past 150 years. As is true with the glaciers located in the tropical zone, virtually all the glaciers in the mid-latitudes are in a state of negative mass balance and are retreating.

Eastern hemisphere



This map from the annual Glacier Commission surveys in Italy and Switzerland shows the percentage of advancing glaciers in the Alps. Mid-20th century saw strong retreating trends, but not as extreme as the present; current retreats represent additional reductions of already smaller glaciers.

Europe

The World Glacier Monitoring Service reports on changes in the terminus, or lower elevation end, of glaciers from around the world every five years. In their 2000-2005 edition, they noted the terminal point variations of glaciers across the Alps. Over the five-year period from 2000–2005, 115 of 115 glaciers examined in Switzerland retreated, 115 of 115 glaciers in Austria reatreated, in Italy during 2005 50 glaciers were retreating and 3 stationary , and all 7 glaciers observed in France were in retreat. French glaciers experienced a sharp retreat in the years 1942–53 followed by advances up to 1980, and then further retreat beginning in 1982. As an example, since 1870 the Argentière Glacier and Mont Blanc Glacier have receded by 1,150 m (3,770 ft) and 1,400 m (4,600 ft), respectively. The largest glacier in France, the Mer de Glace, which is 11 km (6.8 mi) long and 400 m (1,300 ft) thick, has lost 8.3% of its length, or 1 km (0.62 mi), in 130 years, and thinned by 27%, or 150 m (490 ft), in the midsection of the glacier since 1907. The Bossons Glacier in Chamonix, France, has retreated 1,200 m (3,900 ft) from extents observed in the early 20th century. In 2005, of 91 Swiss glaciers studied, 84 retreated from where their terminal points had been in 2004 while the remaining 7 showed no change.

Other researchers have found that glaciers across the Alps appear to be retreating at a faster rate than a few decades ago. In 2008, the Swiss Glacier survey of 85 glaciers found 78 retreating, 2 stationary and 5 advancing. The Trift Glacier had retreated over 500 m (1,600 ft) just in the three years of 2003 to 2005, which is 10% of its total length. The Grosser Aletsch Glacier, the largest glacier in Switzerland, has retreated 2,600 m (8,500 ft) since 1880. This rate of retreat has also increased since 1980, with 30%, or 800 m (2,600 ft), of the total retreat occurring in the last 20% of the time period. Similarly, of the glaciers in the Italian Alps, only about a third were in retreat in 1980, while by 1999, 89% of these glaciers were retreating. In 2005, the Italian Glacier Commission found that 123 glaciers were retreating, 1 advancing and 6 stationary. Repeat photography of glaciers in the Alps provides clear evidence that glaciers in this region have retreated significantly in the past several decades. Morteratsch Glacier, Switzerland is one key example. The yearly measurements of the length changes started in 1878. The overall retreat from 1878 to 1998 has been 2 km (1.2 mi) with a mean annual retreat rate of approximately 17 m (56 ft) per year. This long-term average was markedly surpassed in recent years with the glacier receding 30 m (98 ft) per year during the period between 1999–2005. One major concern which has in the past had great impact on lives and property is the death and destruction from a Glacial Lake Outburst Flood (GLOF). Glaciers stockpile rock and soil that has been carved from mountainsides at their terminal end. These debris piles often form dams that impound water behind them and form glacial lakes as the glaciers melt and retreat from their maximum extents. These terminal moraines are frequently unstable and have been known to burst if overfilled or displaced by earthquakes, landslides or avalanches. If a glacier has a rapid melting cycle during warmer months, the terminal moraine may not be strong enough to hold the rising water behind it, leading to a massive localized flood. This is an increasing risk due to the creation and expansion of glacial lakes resulting from glacier retreat. Past floods have been deadly and have resulted in enormous property damage. Towns and villages in steep, narrow valleys that are downstream from glacial lakes are at the greatest risk. In 1892 a GLOF released some 200,000 km^3 (2.6×10^{14} cu yd) of water from the lake of the Glacier de Tête Rousse, resulting in the deaths of 200 people in the French town of Saint Gervais. GLOFs have been known to occur in every region of the world where glaciers are located. Continued glacier retreat is expected to create and expand glacial lakes, increasing the danger of future GLOFs.

Though the glaciers of the Alps have received more attention from glaciologists than in other areas of Europe, research indicates that throughout most of Europe, glaciers are rapidly retreating. In the Kebnekaise Mountains of northern Sweden, a study of 16 glaciers between 1990 and 2001 found that 14 glaciers were retreating, one was advancing and one was stable. During the 20th century, glaciers in Norway retreated overall with brief periods of advance around 1910, 1925 and in the 1990s. In the 1990s, 11 of 25 Norwegian glaciers observed had advanced due to several consecutive winters with above normal precipitation. However, following several consecutive years of little winter precipitation since 2000, and record warmth during the summers of 2002 and

2003, Norwegian glaciers have decreased significantly since the 1990s. By 2005 only 1 of the 25 glaciers monitored in Norway was advancing, two were stationary and 22 were retreating. In 2009 18 glaciers reatreated, three were stationary (less than 2 meters of change) and two advanced. In 2006 glacier mass balances were very negative in Norway and of the 26 glaciers examined, 24 were retreating with one stationary and one advancing. The Norwegian Engabreen Glacier has retreated 185 m (607 ft) since 1999, while the Brenndalsbreen and Rembesdalsskåka glaciers have retreated 276 m (906 ft) and 250 m (820 ft), respectively, since 2000. The Briksdalsbreen glacier retreated 96 m (315 ft) in 2004 alone—the largest annual retreat recorded for this glacier since monitoring began in 1900. This figure was exceeded in 2006 with five glaciers retreating over 100 m (330 ft) from the fall of 2005 to the fall of 2006. Four outlets from the Jostedalsbreen ice cap, Kjenndalsbreen, Brenndalsbreen, Briksdalsbreen and Bergsetbreen had a frontal retreat of more than 100 metres. Gråfjellsbrea, an outlet from Folgefonna, had a retreat of almost 100 m (330 ft). Overall, from 1999 to 2005, Briksdalsbreen retreated 336 metres (1,102 ft).

In the Spanish Pyrenees, recent studies have shown important losses in extent and volume of the glaciers of the Maladeta massif during the period 1981-2005. These include a reduction in area of 35.7%, from 2.41 km^2 (600 acres) to .627 km^2 (155 acres), a loss in total ice volume of .0137 km^3 (0.0033 cu mi) and an increase in the mean altitude of the glacial termini of 43.5 m (143 ft). For the Pyrenees as a whole 50-60% of the glaciated area has been lost since 1991. At least three glaciers Balaitus, Perdigurero and La Munia have disappeared in this period. Monte Perdido Glacier has shrank from 90 hectares to 40 hectares.

Siberia

Siberia and the Russian Far East, although typically classified as polar regions, owing to the dryness of the winter climate have glaciers only in the high Altai Mountains, Verkhoyansk Range and Cherskiy Range. Kamchatka, exposed to moisture from the Sea of Okhotsk, has much more extensive glaciation totaling around 2,500 square kilometres (970 square miles).

Because the collapse of Communism has caused a large reduction in the number of monitoring stations, details on the retreat of Siberian glaciers is much poorer than in most other regions of the world. Nonetheless, available records do indicate a general retreat of all glaciers in the Altai Mountains and (with the exception of volcanic glaciers) in Kamchatka. Sakha's glaciers, totaling seventy square kilometers, have shrunk by around 28% since 1945, whilst in moister regions of Siberia and on the Pacific coast, the shrinkage is considerably larger, reaching several percent annually in some places.

Asia

This NASA image shows the formation of numerous glacial lakes at the termini of receding glaciers in Bhutan-Himalaya.

The Himalayas and other mountain chains of central Asia support large regions that are glaciated. These glaciers provide critical water supplies to arid countries such as Mongolia, western China, Pakistan, Afghanistan and India. As is true with other glaciers worldwide, the glaciers of Asia are experiencing a rapid decline in mass. The loss of these glaciers would have a tremendous impact on the ecosystem of the region.

In the Wakhan Corridor of Afghanistan 28 of 30 glaciers examined retreated significantly during the 1976-2003 period, the average retreat was 11 meters per year. One of these glaciers, the Zemestan Glacier, has retreated 460 m during this period, not quite 10% of its 5.2 km length. In examining 612 glaciers in China between 1950 and 1970, 53% of the glaciers studied were retreating. After 1990, 95% of these glaciers were measured to be retreating, indicating that retreat of these glaciers was becoming more widespread. Glaciers in the Mount Everest region of the Himalayas are all in a state of retreat. The Rongbuk Glacier, draining the north side of Mount Everest into Tibet, has been retreating 20 m (66 ft) per year. In the Khumbu region of Nepal along the front of the main Himalaya of 15 glaciers examined from 1976-2007 all retreated significantly, average retreat was 28 m per year. The most famous of these Khumbu Glacier retreated at a rate of 18 m per year from 1976-2007. In India the Gangotri Glacier, retreated 34 m (112 ft) per year between 1970 and 1996, and has averaged a loss of 30 m (98 ft) per year since 2000. However, the glacier is still over 30 km (19 mi) long. In 2005 the Tehri Dam was finished on the Bhagirathi River, it is a 2400 mW facility that began producing hydropower in 2006. The headwaters of the Bhagirathi River is the Gangotri and Khatling Glacier, Garhwal Himalaya. Gangotri Glacier has retreated 1 km in the last 30 years, and with an area of 286 km2 provides up to 190 m3/second (Singh et al., 2006). For the Indian Himalaya retreat ranged from -19 meters per year for 17 glaciers all retreating. In Sikkim 26 glaciers examined were retreating at an average rate of 13.02 m per year from 1976 to 2005. For the 51 glaciers in the main Himalayan Range of India, Nepal and Sikkim 51 are retreating, at an average rate of 23 m per year. In the Karokoram Range of the Himalaya there is a mix of advancing and retreating glaciers with 18 advancing and 22 retreating during the 1980-2003 period. Many of the advancing Karakoram glaciers are surging.

With the retreat of glaciers in the Himalayas, a number of glacial lakes have been created. A growing concern is the potential for Glacial Lake Outburst Floods—researchers estimate 20 glacial lakes in Nepal and 24 in Bhutan pose hazards to human populations should their terminal moraine dams fail. One glacial lake identified as potentially hazardous is Bhutan's Raphstreng Tsho, which measured 1.6 km (0.99 mi) long, .96 m (0.00096 km) wide and was 80 m (260 ft) deep in 1986. By 1995 the lake had swollen to a length of 1.94 km (1.21 mi), 1.13 km (0.70 mi) in width and a depth of 107 m (351 ft). In 1994 a GLOF from Luggye Tsho, a glacial lake adjacent to Raphstreng Tsho, killed 23 people downstream.

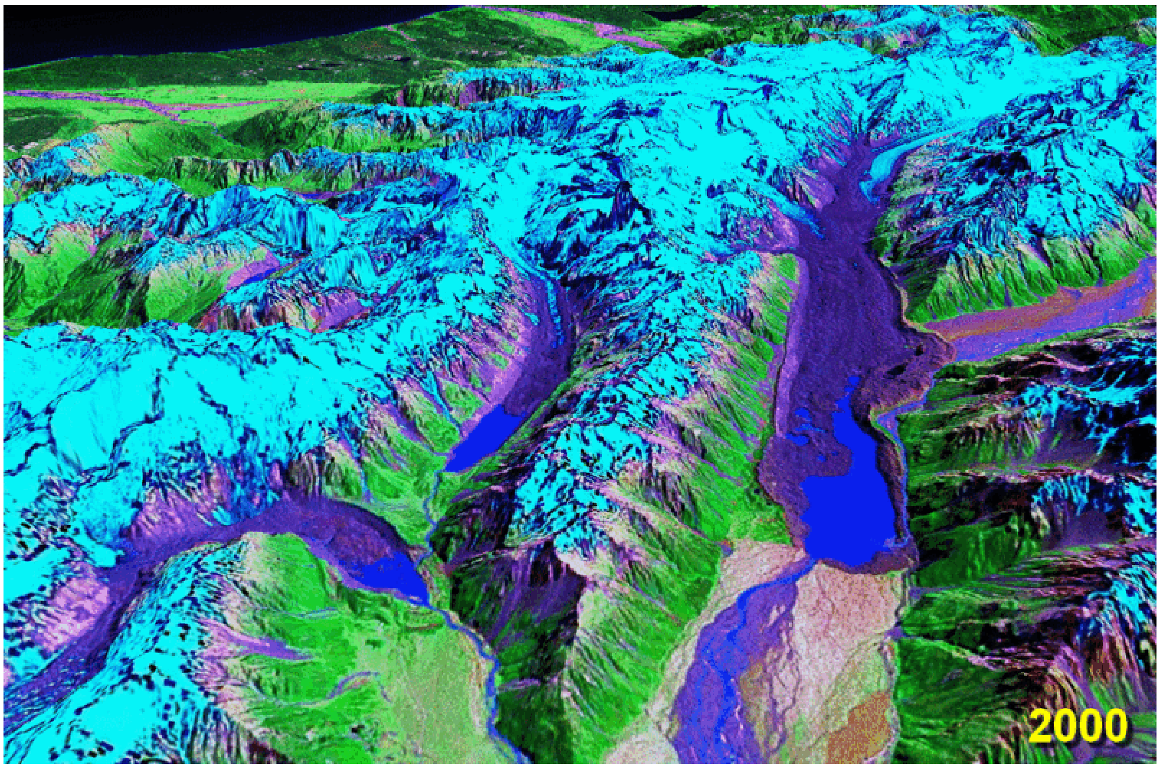
Glaciers in the Ak-shirak Range in Kyrgyzstan experienced a slight loss between 1943 and 1977 and an accelerated loss of 20% of their remaining mass between 1977 and 2001. In the Tien Shan mountains, which Kyrgyzstan shares with China and Kazakhstan, studies in the northern areas of that mountain range show that the glaciers that help supply water to this arid region have been losing nearly 2 km^3 (0.48 cu mi) of ice per year between 1955 and 2000. The University of Oxford study also reported that an average of 1.28% of the volume of these glaciers had been lost per year between 1974 and 1990.

To the south of the Tien Shan, the Pamirs mountain range located primarily in Tajikistan has many thousands of glaciers, all of which are in a general state of retreat. During the 20th century, the glaciers of Tajikistan lost 20 km^3 (4.8 cu mi) of ice. The 70 km (43 mi) long Fedchenko Glacier, which is the largest in Tajikistan and the largest non-polar glacier on Earth, lost 1.4% of its length, or 1 km (0.62 mi), 2 km^3 (0.48 cu mi) of its mass, and the glaciated area was reduced by 11 km^2 (4.2 sq mi) during the 20th century. Similarly, the neighboring Skogatch Glacier lost 8% of its total mass between 1969 and 1986. The country of Tajikistan and neighboring countries of the Pamir Range are highly dependent upon glacial runoff to ensure river flow during droughts and the dry seasons experienced every year. The continued demise of glacier ice will result in a short-term increase, followed by a long-term decrease in glacial melt water flowing into rivers and streams.

The Tibetan Plateau contains the world's third-largest store of ice. Qin Dahe, the former head of the China Meteorological Administration, said that the recent fast pace of melting and warmer temperatures will be good for agriculture and tourism in the short term; but issued a strong warning:

"Temperatures are rising four times faster than elsewhere in China, and the Tibetan glaciers are retreating at a higher speed than in any other part of the world.... In the short term, this will cause lakes to expand and bring floods and mudflows. . . . In the long run, the glaciers are vital lifelines for Asian rivers, including the Indus and the Ganges. Once they vanish, water supplies in those regions will be in peril."

Oceania



These glaciers in New Zealand have continued to retreat rapidly in recent years. Notice the larger terminal lakes, the retreat of the white ice (ice free of moraine cover), and the higher moraine walls due to ice thinning. Photo.

In New Zealand the mountain glaciers have been in general retreat since 1890, with an acceleration of this retreat since 1920. Most of the glaciers have thinned measurably and have reduced in size, and the snow accumulation zones have risen in elevation as the 20th century progressed. During the period 1971–75, Ivory Glacier receded 30 m (98 ft) from the glacial terminus, and about 26% of the surface area of the glacier was lost over the same period. Since 1980 numerous small glacial lakes were created behind the new terminal moraines of several of these glaciers. Glaciers such as Classen, Godley and Douglas now all have new glacial lakes below their terminal locations due to the glacial retreat over the past 20 years. Satellite imagery indicates that these lakes are continuing to expand. There has been significant and ongoing ice volume losses on the largest New Zealand glaciers, including the Tasman, Ivory, Classen, Mueller, Maud, Hooker, Grey, Godley, Ramsay, Murchison, Therma, Volta and Douglas Glaciers. The retreat of these glaciers has been marked by expanding proglacial lakes and terminus region thinning. The loss in volume from 1975-2005 is 11% of the total.

Several glaciers, notably the much-visited Fox and Franz Josef Glaciers on New Zealand's West Coast, have periodically advanced, especially during the 1990s, but the scale of these advances is small when compared to 20th-century retreat. Both glaciers are currently more than 2.5 km (1.6 mi) shorter than a century ago. These large, rapidly flowing glaciers situated on steep slopes have been very reactive to small mass-balance changes. A few years of conditions favorable to glacier advance, such as more westerly winds and a resulting increase in snowfall, are rapidly echoed in a corresponding advance, followed by equally rapid retreat when those favorable conditions end. The glaciers that have been advancing in a few locations in New Zealand have been doing so due to a temporary weather change associated with El Niño, which has brought more precipitation and cloudier, cooler summers since 2002.

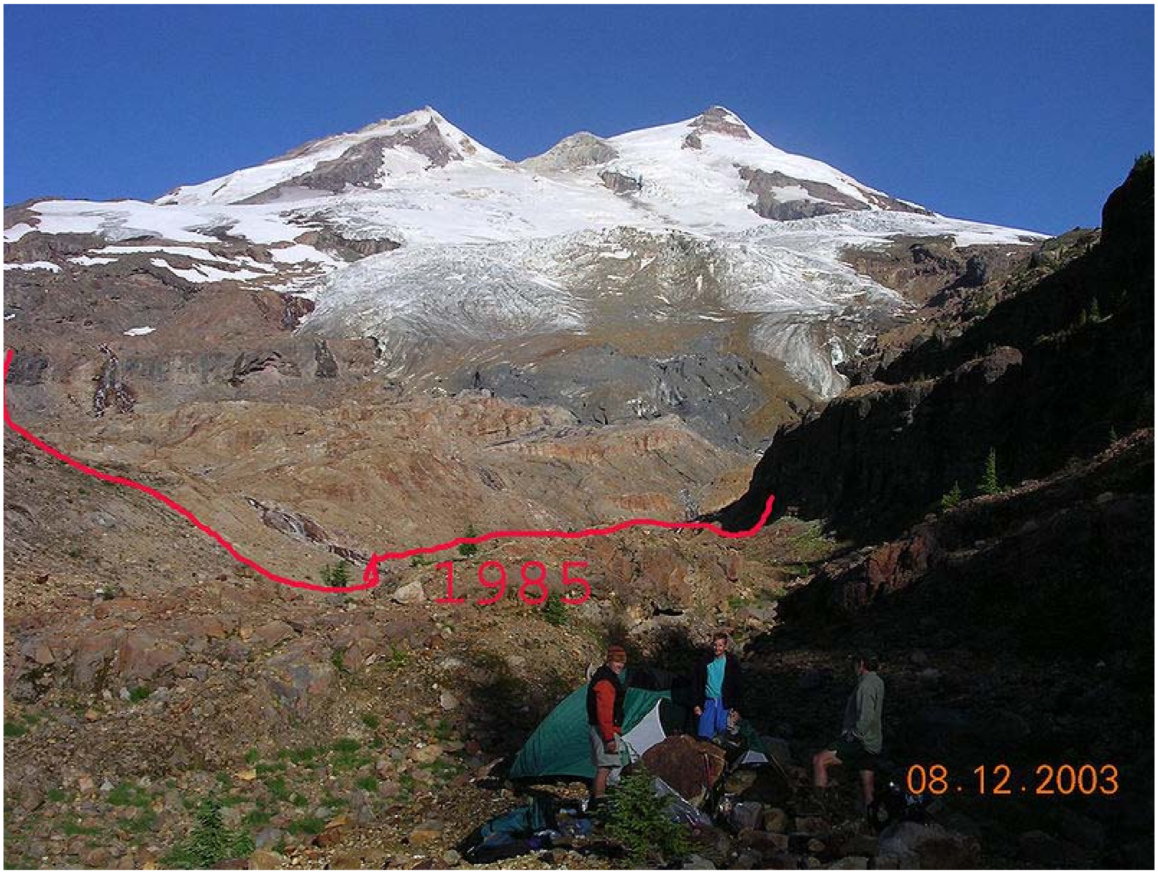
Western hemisphere

The Lewis Glacier, North Cascades National Park after melting away in 1990

North American glaciers are primarily located along the spine of the Rocky Mountains in the United States and Canada, and the Pacific Coast Ranges extending from northern California to Alaska. While Greenland is geologically associated with North America, it is also a part of the Arctic region. Apart from the few tidewater glaciers such as Taku Glacier, that are in the advance stage of their tidewater glacier cycle prevalent along the coast of Alaska, virtually all the glaciers of North America are in a state of retreat. The observed retreat rate has increased rapidly since approximately 1980, and overall each decade since has seen greater rates of retreat than the preceding one. There are also small remnant glaciers scattered throughout the Sierra Nevada mountains of California and Nevada.

Cascades

The Cascade Range of western North America extends from southern British Columbia in Canada to northern California. Excepting Alaska, about half of the glacial area in the U.S. is contained in the more than 700 glaciers of the North Cascades, a portion of the range between the Canadian border and I-90 in central Washington. These glaciers store as much water as that contained in all the lakes and reservoirs in the rest of the state, and provide much of the stream and river flow in the dry summer months, approximating some 870,000 m^3 (1,140,000 cu yd).



The Boulder Glacier retreated 450 m (1,480 ft) from 1987 to 2005.

The Easton Glacier retreated 255 m (837 ft) from 1990 to 2005.

As recently as 1975, many North Cascade glaciers were advancing due to cooler weather and increased precipitation that occurred from 1944 to 1976. However, by 1987 all the North Cascade glaciers were retreating, and the pace of the glacier retreat has increased each decade since the mid-1970s. Between 1984 and 2005, the North Cascade glaciers lost an average of more than 12.5 m in thickness and between 20% and 40% of their volume.

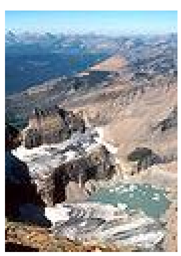
Glaciologists researching the North Cascades glaciers have found that all 47 monitored glaciers are receding and that four glaciers—Spider Glacier, Lewis Glacier (pictured), Milk Lake Glacier, and David Glacier—have disappeared completely since 1985. The White Chuck Glacier (near Glacier Peak) is a particularly dramatic example. The glacier area shrank from 3.1 km^2 (1.2 sq mi) in 1958 to .9 km^2 (0.35 sq mi) by 2002. Between 1850 and 1950, the Boulder Glacier on the southeast flank of Mount Baker retreated 8,700 feet (2,650 m). William Long of the United States Forest Service observed the glacier beginning to advance due to cooler/wetter weather in 1953. This was followed by a 2,438 feet (743 m) advance by 1979. The glacier again retreated 450 m (1,480 ft) from 1987 to 2005, leaving barren terrain behind. This retreat has occurred during a period of reduced winter snowfall and higher summer temperatures. In this region of the Cascades, winter snowpack has declined 25% since 1946, and summer temperatures have risen

0.7 °C (1.2 °F) during the same period. The reduced snowpack has occurred despite a small increase in winter precipitation; thus, it reflects warmer winter temperatures leading to rainfall and melting on glaciers even during the winter. As of 2005, 67% of the North Cascade glaciers observed are in disequilibrium and will not survive the continuation of the present climate. These glaciers will eventually disappear unless temperatures fall and frozen precipitation increases. The remaining glaciers are expected to stabilize, unless the climate continues to warm, but will be much reduced in size.

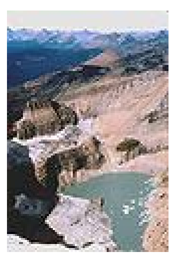
U.S. Rocky Mountains

On the sheltered slopes of the highest peaks of Glacier National Park in Montana, its eponymous glaciers are diminishing rapidly. The area of each glacier has been mapped by the National Park Service and the U.S. Geological Survey for decades. Comparing photographs taken in the mid-19th century with contemporary images provides ample evidence that the glaciers in the park have retreated notably since 1850. Repeat photography over the decades since clearly show that glaciers throughout the park such as Grinnell Glacier are all retreating. The larger glaciers are now approximately a third of their former size when first studied in 1850, and numerous smaller glaciers have disappeared completely. Only 27% of the 99 km^2 (38 sq mi) area of Glacier National Park covered by glaciers in 1850 remained covered by 1993. Researchers believe that by the year 2030, the vast majority of glacial ice in Glacier National Park will be gone unless current climate patterns reverse their course. Grinnell Glacier is just one of many glaciers in Glacier National Park that have been well documented by photographs for many decades. The photographs below clearly demonstrate the retreat of this glacier since 1938.

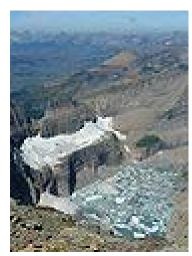
1938 *T.J. Hileman GNP*



1981 *Carl Key (USGS)*



1998 *Dan Fagre (USGS)*



2009 *Lindsey Bengtson (USGS)*

The semiarid climate of Wyoming still manages to support about a dozen small glaciers within Grand Teton National Park, which all show evidence of retreat over the past 50 years. Schoolroom Glacier, located slightly southwest of Grand Teton, one of the more easily reached glaciers in the park, is expected to disappear by 2025. Research between 1950 and 1999 demonstrated that the glaciers in Bridger-Teton National Forest and Shoshone National Forest in the Wind River Range shrank by over a third of their size during that period. Photographs indicate that the glaciers today are only half the size as when first photographed in the late 1890s. Research also indicates that the glacial retreat was proportionately greater in the 1990s than in any other decade over the last 100 years. Gannett Glacier on the northeast slope of Gannett Peak is the largest single glacier in the

Rocky Mountains south of Canada. It has reportedly lost over 50% of its volume since 1920, with almost half of that loss occurring since 1980. Glaciologists believe the remaining glaciers in Wyoming will disappear by the middle of the 21st century if the current climate patterns continue.

Canadian Rockies and British Columbia Coast Range

Fast-melting toe of the Athabasca Glacier, 2005

The Athabasca Glacier in the Columbia Icefield of the Canadian Rockies, has retreated 1,500 m in the last century.

Valdez Glacier has thinned 90 m (300 ft) over the last century and the barren ground near the glacial margins have been exposed due to the glacier thinning and retreating over the last two decades of the 20th century.

In the Canadian Rockies, the glaciers are generally larger and more widespread than they are to the south in the United States Rocky Mountains. One of the more accessible glaciers in the Canadian Rockies is the Athabasca Glacier, which is an outlet glacier of the 325 km^2 (125 sq mi) Columbia Icefield. The Athabasca Glacier has retreated 1,500 m (4,900 ft) since the late 19th century. The rate of retreat for this glacier has increased since 1980, following a period of slow retreat from 1950 to 1980. The Peyto Glacier in Alberta covers an area of about 12 km^2 (4.6 sq mi), and retreated rapidly during the first half of the 20th century, stabilized by 1966, and resumed shrinking in 1976. Illecillewaet Glacier in British Columbia's Glacier National Park (Canada) has retreated 2 km (1.2 mi) since first photographed in 1887.

In Garibaldi Provincial park in SW British Columbia over 505 km2, or 26%, of the park, was covered by glacier ice at the beginning of the 18th century. Ice cover decreased to 297 km2 by 1987–1988 and to 245 km2 by 2005, 50% of the 1850 area. The 50 km2 loss in the last 20 years coincides with negative mass balance in the region. During this period all nine glaciers examined have retreated significantly.

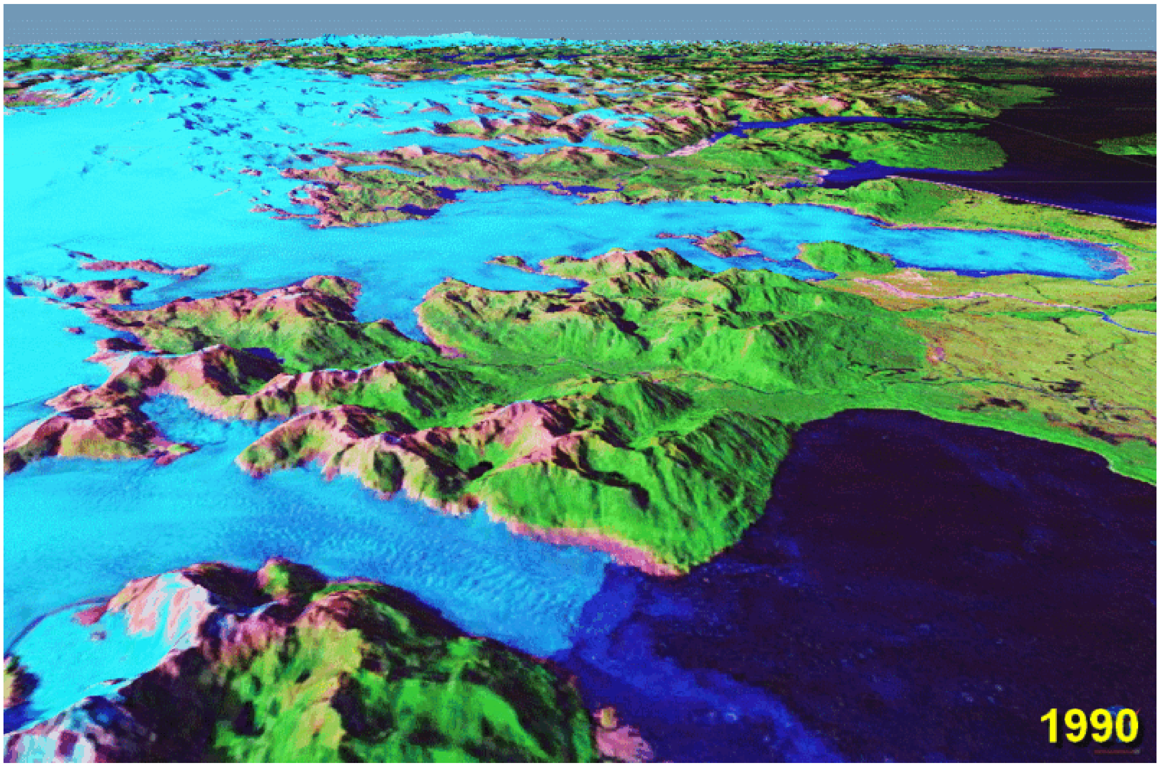
Alaska

There are thousands of glaciers in Alaska, though only a relative few of them have been named. The Columbia Glacier near Valdez in Prince William Sound has retreated 15 km (9.3 mi) in the last 25 years. Icebergs calved off this glacier were a partial cause of the Exxon Valdez oil spill, as the oil tanker had changed course to avoid the icebergs. The Valdez Glacier is in the same area, and though it does not calve, it has also retreated significantly. "A 2005 aerial survey of Alaskan coastal glaciers identified more than a dozen glaciers, many former tidewater and calving glaciers, including Grand Plateau, Alsek, Bear, and Excelsior Glaciers that are rapidly retreating. Of 2,000 glaciers observed, 99% are retreating." Icy Bay in Alaska is fed by three large glaciers—Guyot, Yahtse, and Tyndall Glaciers—all of which have experienced a loss in length and thickness and, consequently, a loss in area. Tyndall Glacier became separated from the retreating Guyot Glacier in the 1960s and has retreated 24 km (15 mi) since, averaging more than 500 m (1,600 ft) per year.

The Juneau Icefield Research Program has monitored the outlet glaciers of the Juneau Icefield since 1946. On the west side of the ice field, the terminus of the Mendenhall Glacier, which flows into suburban Juneau, Alaska, has retreated 580 m (1,900 ft). Of the nineteen glaciers of the Juneau Icefield, eighteen are retreating, and one, the Taku Glacier, is advancing. Eleven of the glaciers have retreated more than 1 km (0.62 mi) since 1948 — Antler Glacier, 5.4 km (3.4 mi); Gilkey Glacier, 3.5 km (2.2 mi); Norris Glacier, 1.1 km (0.68 mi) and Lemon Creek Glacier, 1.5 km (0.93 mi). Taku Glacier has been advancing since at least 1890, when naturalist John Muir observed a large iceberg calving front. By 1948 the adjacent fjord had filled in, and the glacier no longer calved and was able to continue its advance. By 2005 the glacier was only 1.5 km (0.93 mi) from reaching Taku Point and blocking Taku Inlet. The advance of Taku Glacier averaged 17 m (56 ft) per year between 1988 and 2005. The mass balance was very positive for the 1946–88 period fueling the advance; however, since 1988 the mass balance has been slightly negative, which should in the future slow the advance of this mighty glacier.

Long-term mass balance records from Lemon Creek Glacier in Alaska show slightly declining mass balance with time. The mean annual balance for this glacier was −0.23 m (0.75 ft) each year during the period of 1957 to 1976. Mean annual balance has been increasingly negatively averaging −1.04 m (3.4 ft) per year from 1990 to 2005. Repeat glacier altimetry, or altitude measuring, for 67 Alaska glaciers find rates of thinning have increased by more than a factor of two when comparing the periods from 1950 to 1995 (0.7 m (2.3 ft) per year) and 1995 to 2001 (1.8 m (5.9 ft) per year). This is a systemic trend with loss in mass equating to loss in thickness, which leads to increasing retreat—the glaciers are not only retreating, but they are also becoming much thinner. In Denali National Park, all glaciers monitored are retreating, with an average retreat of 20 m (66 ft) per year. The terminus of the Toklat Glacier has been retreating 26 m (85 ft) per year and the Muldrow Glacier has thinned 20 m (66 ft) since 1979. Well documented in Alaska are surging glaciers that have been known to rapidly advance, even as much as 100 m (330 ft) per day. Variegated, Black Rapids, Muldrow, Susitna and Yanert are examples of surging glaciers in Alaska that have made rapid advances in the past. These glaciers are all retreating overall, punctuated by short periods of advance.

Andes and Tierra del Fuego



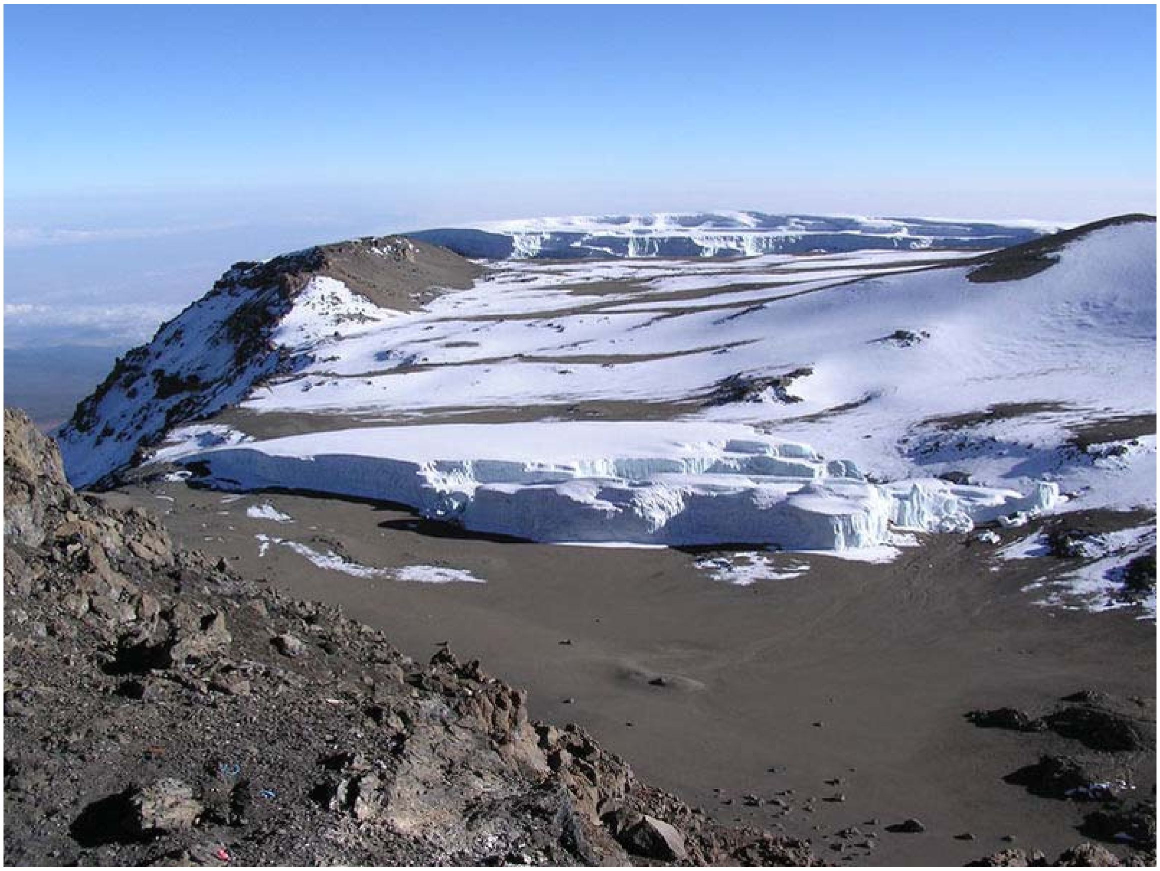
Retreat of San Rafael Glacier from 1990 to 2000. San Quintín Glacier is shown in the background

A large region of population surrounding the central and southern Andes of Argentina and Chile reside in arid areas that are dependent on water supplies from melting glaciers. The water from the glaciers also supplies rivers that have in some cases been dammed for hydroelectric power. Some researchers believe that by 2030, many of the large ice caps on the highest Andes will be gone if current climate trends continue. In Patagonia on the southern tip of the continent, the large ice caps have retreated a 1 km (0.62 mi) since the early 1990s and 10 km (6.2 mi) since the late 19th century. It has also been observed that Patagonian glaciers are receding at a faster rate than in any other world region. The Northern Patagonian Ice Field lost 93 km^2 (36 sq mi) of glacier area during the years between 1945 and 1975, and 174 km^2 (67 sq mi) from 1975 to 1996, which indicates that the rate of retreat is increasing. This represents a loss of 8% of the ice field, with all glaciers experiencing significant retreat. The Southern Patagonian Ice Field has exhibited a general trend of retreat on 42 glaciers, while four glaciers were in equilibrium and two advanced during the years between 1944 and 1986. The largest retreat was on O'Higgins Glacier, which during the period 1896–1995 retreated 14.6 km (9.1 mi). The Perito Moreno Glacier is 30 km (19 mi) long and is a major outflow glacier of the Patagonian ice sheet, as well as the most visited glacier in Patagonia. Perito Moreno Glacier is presently in equilibrium, but has undergone frequent oscillations in the period 1947–96, with a net gain of 4.1 km (2.5 mi). This glacier has advanced since 1947, and has been essentially stable since 1992. Perito Moreno Glacier is one of three glaciers in Patagonia known to have advanced, compared to several hundred others in retreat. The two major glaciers of the Southern Patagonia Icefield to the north of Moreno, Upsala and Videma Glacier have retreated 4.6 km (2.9 mi) in 21 years and 1 km (0.62 mi) in 13 years respectively. In the Aconcagua River Basin, glacier retreat has resulted in a 20% loss in glacier area, declining from 151 km^2 (58 sq mi) to 121 km^2 (47 sq mi). The Marinelli Glacier in Tierra del Fuego has been in retreat since at least 1960 through 2008.

Tropical glaciers

Tropical glaciers are located between the Tropic of Cancer and the Tropic of Capricorn, in the region that lies 23° 26′ 22″ north or south of the equator. Tropical glaciers are the most uncommon of all glaciers for a variety of reasons. Firstly, the tropics are the warmest part of the planet. Secondly, the seasonal change is minimal with temperatures warm year round, resulting in a lack of a colder winter season in which snow and ice can accumulate. Thirdly, few taller mountains exist in these regions upon which enough cold air exists for the establishment of glaciers. All of the glaciers located in the tropics are on isolated high mountain peaks. Overall, tropical glaciers are smaller than those found elsewhere and are the most likely glaciers to show rapid response to changing climate patterns. A small temperature increase of only a few degrees can have almost immediate and adverse impact on tropical glaciers.

Africa



Furtwängler Glacier atop Kilimanjaro in the foreground and snowfields and the Northern Icefields beyond.

With almost the entire continent of Africa located in the tropical and subtropical climate zones, glaciers are restricted to two isolated peaks and the Ruwenzori Range. Kilimanjaro, at 5,895 m (19,341 ft), is the highest peak on the continent. Since 1912 the glacier cover on the summit of Kilimanjaro has apparently retreated 75%, and the volume of glacial ice is now 80% less than it was a century ago due to both retreat and thinning. In the 14-year period from 1984 to 1998, one section of the glacier atop the mountain receded 300 m (980 ft). A 2002 study determined that if current conditions continue, the glaciers atop Kilimanjaro will disappear sometime between 2015 and 2020. A March 2005 report indicated that there is almost no remaining glacial ice on the mountain, and it is the first time in 11,000 years that barren ground has been exposed on portions of the summit. Researchers reported Kilimanjaro's glacier retreat was due to a combination of increased sublimation and decreased snow fall.

The Furtwängler Glacier is located near the summit of Kilimanjaro. Between 1976 and 2000, the area of Furtwängler Glacier was cut almost in half, from 113,000 m^2 (1,220,000 sq ft) to 60,000 m^2 (650,000 sq ft). During fieldwork conducted early in 2006, scientists discovered a large hole near the center of the glacier. This hole, extending through the

6 m (20 ft) remaining thickness of the glacier to the underlying rock, is expected to grow and split the glacier in two by 2007.

To the north of Kilimanjaro lies Mount Kenya, which at 5,199 m (17,057 ft) is the second tallest mountain on the African continent. Mount Kenya has a number of small glaciers that have lost at least 45% of their mass since the middle of the 20th century. According to research compiled by the U.S. Geological Survey (USGS), there were eighteen glaciers atop Mount Kenya in 1900, and by 1986 only eleven remained. The total area covered by glaciers was 1.6 km^2 (0.62 sq mi) in 1900, however by the year 2000 only about 25%, or 0.4 km^2 (0.15 sq mi) remained. To the west of Mounts Kilimanjaro and Kenya, the Ruwenzori Range rises to 5,109 m (16,762 ft). Photographic evidence of this mountain range indicates a marked reduction in glacially covered areas over the past century. In the 35-year period between 1955 and 1990, glaciers on the Rwenzori Mountains receded about 40%. It is expected that due to their proximity to the heavy moisture of the Congo region, the glaciers in the Ruwenzori Range may recede at a slower rate than those on Kilimanjaro or in Kenya.

South America

A study by glaciologists of two small glaciers in South America reveals another retreat. More than 80% of all glacial ice in the northern Andes is concentrated on the highest peaks in small glaciers of approximately 1 km^2 (0.39 sq mi) in size. A 1992 to 1998 observation of the Chacaltaya Glacier in Bolivia and Antizana Glacier in Ecuador indicated that between 0.6 m (2.0 ft) and 1.9 m (6.2 ft) of ice was lost per year on each glacier. Figures for Chacaltaya Glacier show a loss of 67% of its volume and 40% of its thickness over the same period. Chacaltaya Glacier has lost 90% of its mass since 1940 and is expected to disappear altogether sometime between 2010 and 2015. Research also indicates that since the mid-1980s, the rate of retreat for both of these glaciers has been increasing. In Colombia, the glaciers atop Nevado del Ruiz have lost more than half their area in the last 40 years. Further south in Peru, the Andes are at a higher altitude overall, and there are approximately 722 glaciers covering an area of 723 km^2 (279 sq mi). Research in this region of the Andes is less extensive but indicates an overall glacial retreat of 7% between 1977 and 1983. The Quelccaya Ice Cap is the largest tropical icecap in the world, and all of the outlet glaciers from the icecap are retreating. In the case of Qori Kalis Glacier, which is Quelccayas' main outlet glacier, the rate of retreat had reached 155 m (509 ft) per year during the three year period of 1995 to 1998. The melting ice has formed a large lake at the front of the glacier since 1983, and bare ground has been exposed for the first time in thousands of years.

Oceania

Puncak Jaya icecap 1936 USGS

Puncak Jaya glaciers 1972. Left to right: Northwall Firn, Meren Glacier, and Carstensz Glacier. USGS.

On the large island of New Guinea, there is photographic evidence of massive glacial retreat since the region was first extensively explored by airplane in the early 1930s. Due to the location of the island within the tropical zone, there is little to no seasonal variation in temperature. The tropical location has a predictably steady level of rain and snowfall, as well as cloud cover year round, and there has been no noticeable change in the amount of moisture which has fallen during the 20th century. The 7 km^2 (2.7 sq mi) ice cap on Puncak Jaya is the largest on the island, and has retreated from one larger mass into several smaller glacial bodies since 1936. Of these smaller glaciers, research between 1973 and 1976 showed glacier retreat for the Meren Glacier of 200 m (660 ft) while the Carstensz Glacier lost 50 m (160 ft). The Northwall Firn, another large remnant of the icecap that once was atop Puncak Jaya, has itself split into several separate glaciers since 1936. Research presented in 2004 of IKONOS satellite imagery of the New Guinean glaciers provided a dramatic update. The imagery indicated that in the two years from 2000 to 2002, the East Northwall Firn had lost 4.5%, the West Northwall Firn 19.4% and the Carstensz 6.8% of their glacial mass. Researchers also discovered that, sometime between 1994 and 2000, the Meren Glacier disappeared altogether. An expedition to the remaining glaciers on Puncak Jaya in 2010 discovered that the ice on the glaciers there is about 32 metres (105 ft) thick and thinning at a rate of 7 metres (23 ft) annually. At that rate, the remaining glaciers are expected to last another 5 years or 2015. Separate from the glaciers of Puncak Jaya, another small icecap known to have existed on the summit of Puncak Trikora completely disappeared sometime between 1939 and 1962.

Polar regions

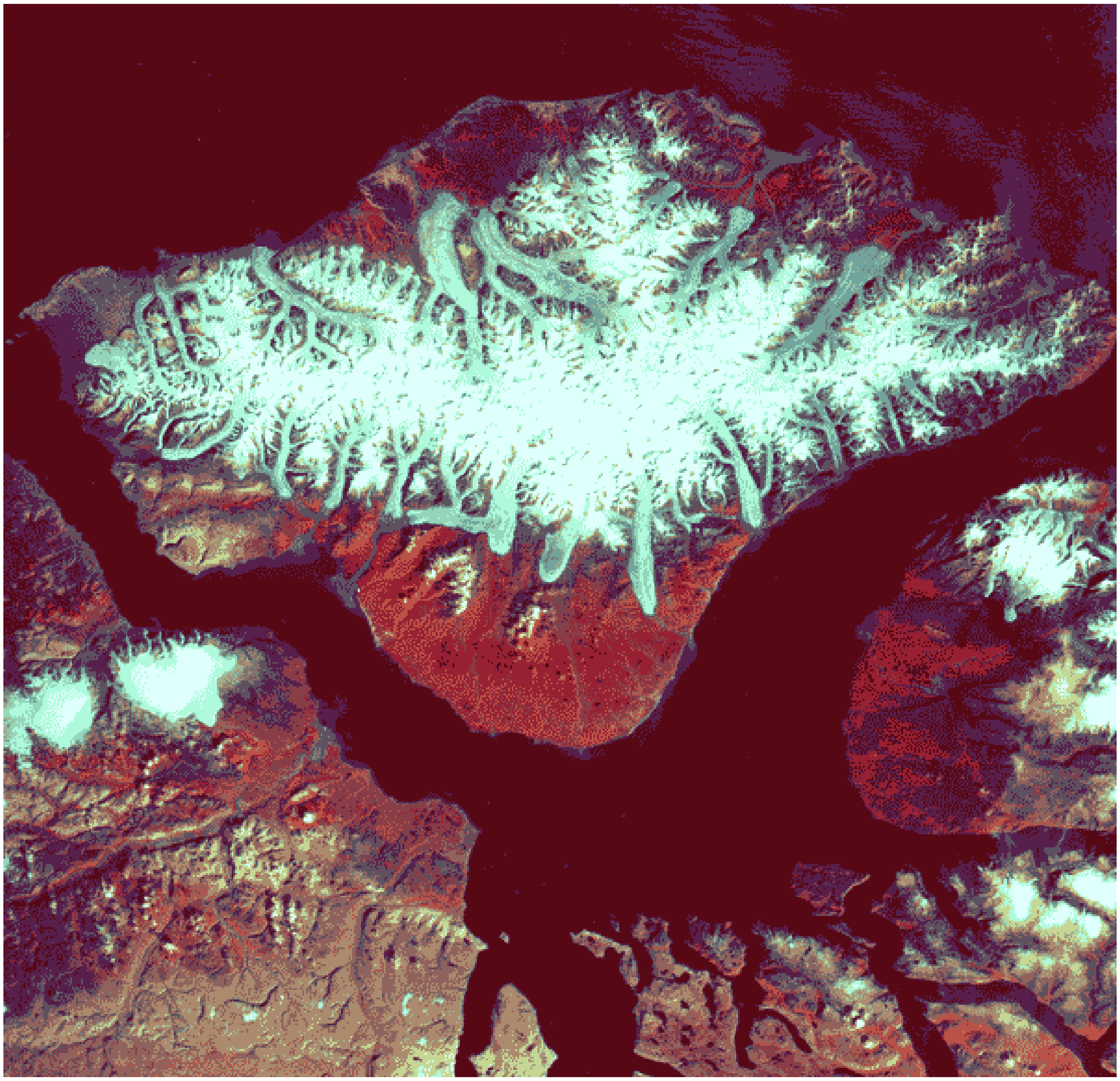
Despite their proximity and importance to human populations, the mountain and valley glaciers of tropical and mid-latitude glaciers amount to only a small fraction of glacial ice on the Earth. About 99% of all freshwater ice is in the great ice sheets of polar and subpolar Antarctica and Greenland. These continuous continental-scale ice sheets, 3 km (1.9 mi) or more in thickness, cap much of the polar and subpolar land masses. Like rivers flowing from an enormous lake, numerous outlet glaciers transport ice from the margins of the ice sheet to the ocean.

Iceland

The northern Atlantic island nation of Iceland is home to the Vatnajökull, which is the largest ice cap in Europe. The Breiðamerkurjökull Glacier is one of the Vatnajökull outlet glaciers, and had receded by as much as 2 km (1.2 mi) between 1973 and 2004. In the early 20th century, Breiðamerkurjökull extended to within 250 m (820 ft) of the ocean, but by 2004 Breiðamerkurjökull's terminus had retreated 3 km (1.9 mi) further inland. This glacier retreat exposed a rapidly expanding lagoon that is filled with icebergs calved from its front. The lagoon is 110 m (360 ft) deep and nearly doubled its size between 1994 and 2004. Mass-balance measurements of Iceland's glaciers show alternating positive and negative mass balance of glaciers during the period 1987–95, but the mass balance has been predominantly negative since. On Hofsjökull ice cap, mass balance has been negative each year from 1995-2005.

Most of the Icelandic glaciers retreated rapidly during the warm decades from 1930 to 1960, slowing down as the climate cooled during the following decade, and started to advance after 1970. The rate of advance peaked in the 1980s, after which it slowed down as a consequence of rapid warming of the climate that has taken place since the mid-1980s. Most glaciers in Iceland began to retreat after 1990, and by 2000 all monitored non-surge type glaciers in Iceland were retreating. An average of 45 non-surging termini were monitored each year by the Icelandic Glaciological Society from 2000–2005.

Canada



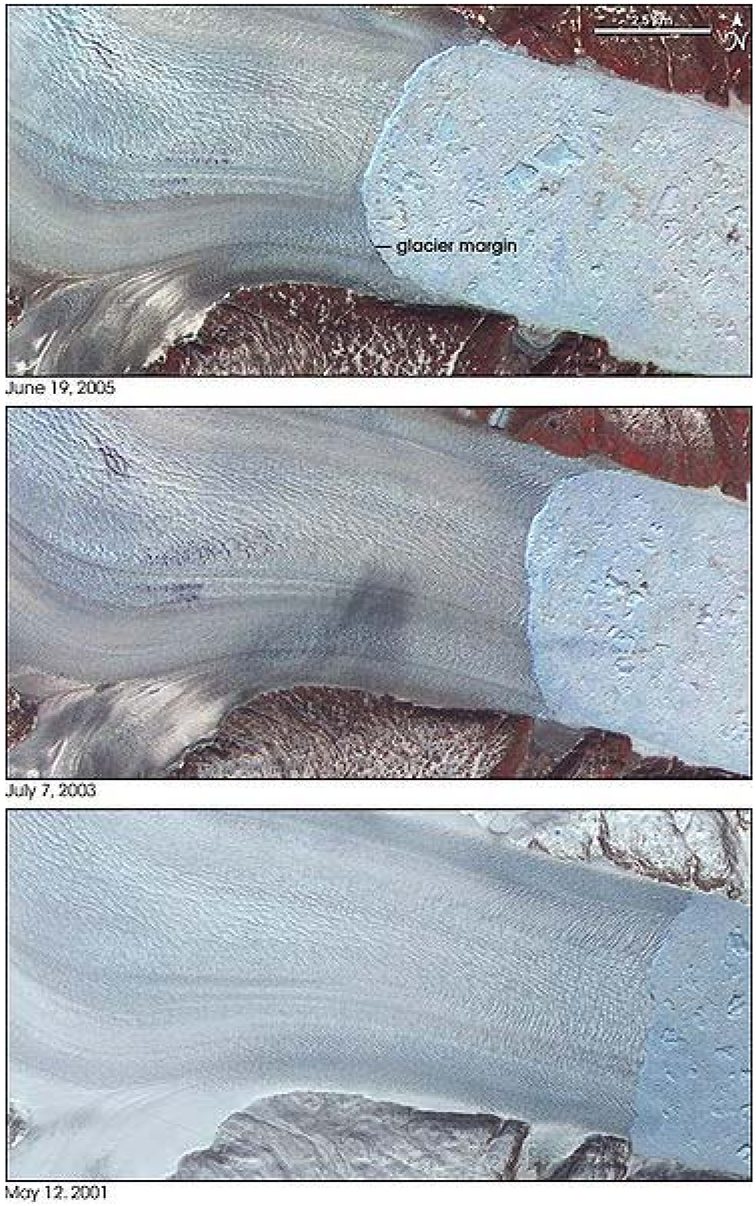
Bylot Ice Cap on Bylot Island, one of the Canadian Arctic islands, August 14, 1975 (USGS)

The Canadian Arctic islands have a number of substantial ice caps, including Penny and Barnes ice caps on Baffin Island, Bylot Ice Cap on Bylot Island, and Devon Ice Cap on Devon Island. All of these ice caps have been thinning and receding slowly. The Barnes and Penny ice caps on Baffin Island have been thinning at over 1 m (3.3 ft) per year in the lower elevations from 1995 to 2000. Overall, between 1995 and 2000, ice caps in the Canadian Arctic lost 25 km^2 (9.7 sq mi) of ice per year. Between 1960 and 1999, the Devon Ice Cap lost 67 km^3 (16 cu mi) of ice, mainly through thinning. All major outlet glaciers along the eastern Devon Ice Cap margin have retreated from 1 km (0.62 mi) to 3 km (1.9 mi) since 1960. On the Hazen Plateau of Ellesmere Island, the Simmon Ice Cap has lost 47% of its area since 1959. If the current climatic conditions continue, the remaining glacial ice on the Hazen Plateau will be gone around 2050. On August 13, 2005 the Ayles Ice Shelf broke free from the north coast of Ellesmere Island, the 66 km^2 (25 sq mi) ice shelf drifted into the Arctic Ocean. This followed the splitting of the Ward Hunt Ice Shelf in 2002. The Ward Hunt has lost 90% of its area in the last century.

Northern Europe

Arctic islands north of Norway, Finland and Russia have all shown evidence of glacier retreat. In the Svalbard archipelago, the island of Spitsbergen has numerous glaciers. Research indicates that Hansbreen (Hans Glacier) on Spitsbergen retreated 1.4 km (0.87 mi) from 1936 to 1982 and another 400 m (1,300 ft) during the 16-year period from 1982 to 1998. Blomstrandbreen, a glacier in the King's Bay area of Spitsbergen, has retreated approximately 2 km (1.2 mi) in the past 80 years. Since 1960 the average retreat of Blomstrandbreen has been about 35 m (115 ft) a year, and this average was enhanced due to an accelerated rate of retreat since 1995. Similarly, Midre Lovenbreen retreated 200 m (656 ft) between 1977 and 1995. In the Novaya Zemlya archipelago north of Russia, research indicates that in 1952 there was 208 km (129 mi) of glacier ice along the coast. By 1993 this had been reduced by 8% to 198 km (123 mi) of glacier coastline.

Greenland



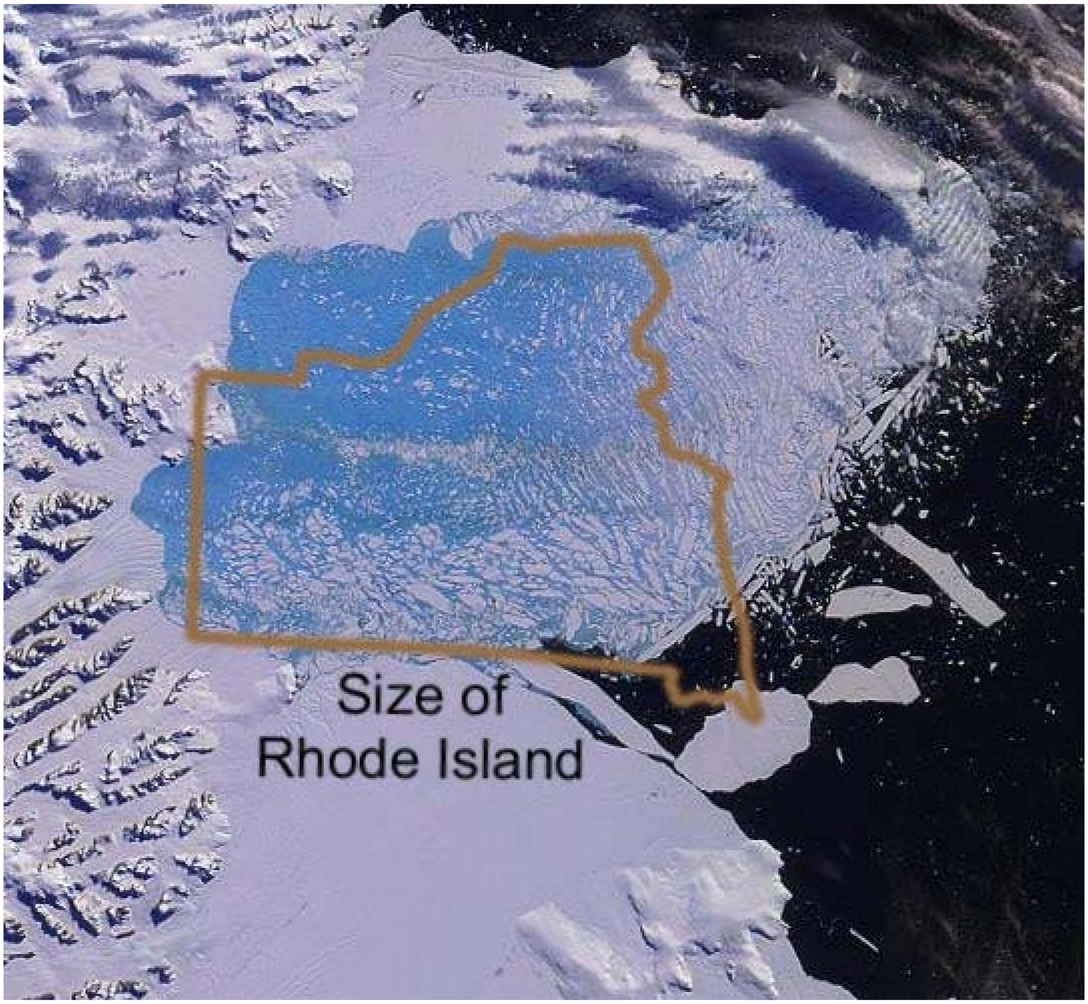
Retreat of the Helheim Glacier, Greenland

In Greenland, glacier retreat has been observed in outlet glaciers, resulting in an increase of the ice flow rate and destabilization of the mass balance of the ice sheet that is their source. The net loss in volume and hence sea level contribution of the Greenland Ice Sheet (GIS) has doubled in recent years from 90 km^3 (22 cu mi) to 220 km^3 (53 cu mi) per year. Researchers also noted that the acceleration was widespread affecting almost all glaciers south of 70 N by 2005. The period since 2000 has brought retreat to several very large glaciers that had long been stable. Three glaciers that have been researched—Helheim Glacier, Kangerdlugssuaq Glacier, and Jakobshavn Isbræ—jointly drain more than 16% of the Greenland Ice Sheet. In the case of Helheim Glacier, researchers used satellite images to determine the movement and retreat of the glacier. Satellite images and aerial photographs from the 1950s and 1970s show that the front of the glacier had remained in the same place for decades. In 2001 the glacier began retreating rapidly, and by 2005 the glacier had retreated a total of 7.2 km (4.5 mi), accelerating from 20 m (66 ft) per day to 35 m (115 ft) per day during that period.

Jakobshavn Isbræ in west Greenland, a major outlet glacier of the Greenland Ice Sheet, has been the fastest moving glacier in the world over the past half century. It had been moving continuously at speeds of over 24 m (79 ft) per day with a stable terminus since at least 1950. In 2002 the 12 km (7.5 mi) long floating terminus of the glacier entered a phase of rapid retreat, with the ice front breaking up and the floating terminus disintegrating and accelerating to a retreat rate of over 30 m (98 ft) per day. On a shorter timescale, portions of the main trunk of Kangerdlugssuaq Glacier that were flowing at 15 m (49 ft) per day from 1988 to 2001 were measured to be flowing at 40 m (130 ft) per day in the summer of 2005. Not only has Kangerdlugssuaq retreated, it has also thinned by more than 100 m (330 ft).

The rapid thinning, acceleration and retreat of Helheim, Jakobshavns and Kangerdlugssuaq glaciers in Greenland, all in close association with one another, suggests a common triggering mechanism, such as enhanced surface melting due to regional climate warming or a change in forces at the glacier front. The enhanced melting leading to lubrication of the glacier base has been observed to cause a small seasonal velocity increase and the release of meltwater lakes has also led to only small short term accelerations. The significant accelerations noted on the three largest glaciers began at the calving in front and propagated inland and are not seasonal in nature. Thus, the primary source of outlet glacier acceleration widely observed on small and large calving glaciers in Greenland is driven by changes in dynamic forces at the glacier front, not enhanced meltwater lubrication. This was termed the *Jakobshavns Effect* by Terence Hughes at the University of Maine in 1986.

Antarctica



The collapsing Larsen B Ice Shelf in Antarctica is similar in area to the U.S. state of Rhode Island.

The climate of Antarctica is one of intense cold and great aridity. Most of the world's freshwater ice is contained in the great ice sheets that cover the continent of Antarctica. The most dramatic example of glacier retreat on the continent is the loss of large sections of the Larsen Ice Shelf on the Antarctic Peninsula. Ice shelves are not stable when surface melting occurs, and the collapse of Larsen Ice Shelf has been caused by warmer melt season temperatures that have led to surface melting and the formation of shallow ponds of water on the ice shelf. The Larsen Ice Shelf lost 2,500 km^2 (970 sq mi) of its area from 1995 to 2001. In a 35-day period beginning on January 31, 2002, about 3,250 km^2 (1,250 sq mi) of shelf area disintegrated. The ice shelf is now 40% the size of its previous minimum stable extent The recent collapse of Wordie Ice Shelf, Prince Gustav Ice Shelf, Mueller Ice Shelf, Jones Ice Shelf, Larsen-A and Larsen-B Ice Shelf on the Antarctic Peninsula has raised awareness of how dynamic ice shelf systems are. Jones Ice Shelf had an area of 35 km^2 (14 sq mi) in the 1970s but by 2008 it had disappeared. Wordie Ice Shelf has gone from an area of 1500 square kilometers in 1950 to 140 km2 in 2000.

Prince Gustav Ice Shelf has gone from an area of 1600 km2 to 11 km2 in 2008. After their loss the reduced buttressing of feeder glaciers has allowed the expected speed-up of inland ice masses after shelf ice break-up. The Wilkins Ice Shelf is another ice shelf that has suffered substantial retreat. The ice shelf had an area of 16,000 km^2 (6,200 sq mi) in 1998 when 1,000 km^2 (390 sq mi) was lost. In 2007 and 2008 significant rifting developed and led to the loss of another 1,400 km^2 (540 sq mi) of area. Some of the calving occurred in the Austral winter. The calving seemed to have resulted from preconditioning such as thinning, possibly due to basal melt, as surface melt was not as evident, leading to a reduction in the strength of the pinning point connections. The thinner ice than experienced spreading rifts and breakup. This period culminated in the collapse of an ice bridge connecting the main ice shelf to Charcot Island leading to the loss of an additional 700 km^2 (270 sq mi) in February–June 2009.

Pine Island Glacier, an Antarctic outflow glacier that flows into the Amundsen Sea, thinned 3.5 m (11 ft)± 0.9 m (3.0 ft) per year and retreated a total of 5 km (3.1 mi) in 3.8 years. The terminus of the Pine Island Glacier is a floating ice shelf, and the point at which it starts to float retreated 1.2 km (0.75 mi) per year from 1992 to 1996. This glacier drains a substantial portion of the West Antarctic Ice Sheet and along with the neighboring Thwaites Glacier, which has also shown evidence of thinning, has been referred to as the weak underbelly of this ice sheet. Additionally, the Dakshin Gangotri Glacier, a small outlet glacier of the Antarctic ice sheet, receded at an average rate of 0.7 m (2.3 ft) per year from 1983 to 2002. On the Antarctic Peninsula, which is the only section of Antarctica that extends well north of the Antarctic Circle, there are hundreds of retreating glaciers. In one study of 244 glaciers on the peninsula, 212 have retreated an average of 600 m (2,000 ft) from where they were when first measured in 1953. The greatest retreat was seen in Sjogren Glacier, which is now 13 km (8.1 mi) further inland than where it was in 1953. There are 32 glaciers that were measured to have advanced; however, these glaciers showed only a modest advance averaging 300 m (980 ft) per glacier, which is significantly smaller than the massive retreat observed.

Impacts of glacier retreat

The continued retreat of glaciers will have a number of different quantitative impacts. In areas that are heavily dependent on water runoff from glaciers that melt during the warmer summer months, a continuation of the current retreat will eventually deplete the glacial ice and substantially reduce or eliminate runoff. A reduction in runoff will affect the ability to irrigate crops and will reduce summer stream flows necessary to keep dams and reservoirs replenished. This situation is particularly acute for irrigation in South America, where numerous artificial lakes are filled almost exclusively by glacial melt. Central Asian countries have also been historically dependent on the seasonal glacier melt water for irrigation and drinking supplies. In Norway, the Alps, and the Pacific Northwest of North America, glacier runoff is important for hydropower.

Some of this retreat has resulted in efforts to slow down the loss of glaciers in the Alps. To retard melting of the glaciers used by certain Austrian ski resorts, portions of the Stubai and Pitztal Glaciers were partially covered with plastic. In Switzerland plastic sheeting is also used to reduce the melt of glacial ice used as ski slopes. While covering glaciers with plastic sheeting may prove advantageous to ski resorts on a small scale, this practice is not expected to be economically practical on a much larger scale.

Many species of freshwater and saltwater plants and animals are dependent on glacier-fed waters to ensure the cold water habitat to which they have adapted. Some species of freshwater fish need cold water to survive and to reproduce, and this is especially true with salmon and cutthroat trout. Reduced glacial runoff can lead to insufficient stream flow to allow these species to thrive. Alterations to the ocean currents, due to increased freshwater inputs from glacier melt, and the potential alterations to thermohaline circulation of the world's oceans, may impact existing fisheries upon which humans depend as well.

The potential for major sea level rise depends mostly on a significant melting of the polar ice caps of Greenland and Antarctica, as this is where the vast majority of glacial ice is located. If all the ice on the polar ice caps were to melt away, the oceans of the world would rise an estimated 70 m (230 ft). Although previously it was thought that the polar ice caps were not contributing heavily to sea level rise (IPCC 2007), recent studies have confirmed that both Antarctica and Greenland are contributing 0.5 mm a year each to global sea level rise (Cazenave et al. 2009, Velicogna 2009). The fact that the IPCC estimates did not include rapid ice sheet decay into their sea level predictions makes it difficult to ascertain a plausible estimate for sea level rise but recent studies find that the minimum sea level rise will be around 0.8 m (Pfeffer et al. 2008).

Chapter-4

Glacial Geology

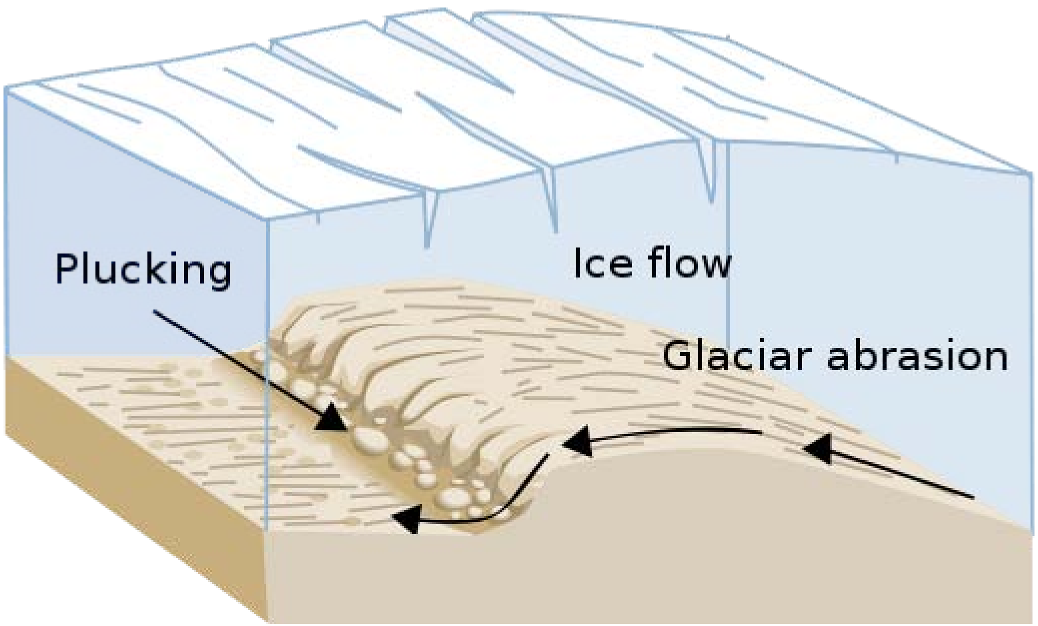


Diagram of glacial plucking and abrasion

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

Moraines can be classified either by their origin or shape. The first aproach is suitable for moraines associated to contemporary glaciers but more difficult to apply to old moraines whose glaciers has disapeared long ago. Moraines types like rogen and veiki moraines are defined by their particular morphology since their origin has been a matter of dispute. Some moraine types are only known from ancient glaciers, like the two former ones, while medial moraines of valley glaciers are poorly preserved and difficult to distinguish after the retreat or melting of the glacier.

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

Lake Rogen, Sweden as seen from the north. The forested ridges in the lake are 'Rogen moraines' of which this is the type location

A **Rogen moraine** (also called **ribbed moraine**) is a subglacially (*i.e.* under a glacier or ice sheet) formed type of moraine landform, that mainly occur in Fennoscandia, Scotland, Ireland and Canada. They cover large areas that have been covered by ice, and occur mostly in what is believed to be the central areas of the ice sheets. Rogen moraines are named after Lake Rogen in Härjedalen, Sweden, the landform's type locality.

The landform occurs in groups that are often closely and regularly spaced. They consist of glacial drift, with till being the most common constituent. The individual moraines are large, wavy ridges orientated transverse to ice flow. Drumlins are often found in close proximity of Rogen moraines, and are often interpreted to be formed at the same time as the Rogen moraines. Although Rogen moraines can span a large range of sizes, the most common distribution seems to be 10-30 metres high, 150-300 metres wide and 300-1,200 metres long.

The exact mechanics of Rogen moraine formation are not known, but since the 1970's, four main theories on the formation have been proposed:

- Megaripples eroded in the basal ice fill during subglacial outburst floods.

- Already existing landforms, such as drumlins and flutes or marginal moraines are reshaped due to a ~90° change in the direction of the ice flow.
- Debris-rich basal ice or pre-existing sediments are sheared and stacked, or folded during compressive ice flow.
- Sediment sheets become fractured and extended during a transition of the overlying glacier from being cold based ice to warm based.

However, it has been suggested that, due to the diversity of morphological characteristics displayed by Rogen moraine, different processes might be able to create the landform. This means that all four of the processes mentioned above might be correct. The different theories that proposed a formation near or at the glacial margin have largely been abandoned. Some of these theories proposed that Rogen moraines had an origin as a series of end moraines, that they formed in association with calving ice termini in glacial lakes, or that Rogen moraines formed in dead-ice, where supraglacial material fell down into crevasses in the ice.

End or terminal moraines

The *Salpausselkä* terminal moraines in Southern Finland

The Marseilles terminal moraine located southwest of Chicago

Alpine terminal moraine near St. Moritz, Upper Engadin, Switzerland, view east towards the Muretto pass, the national border with Italy

A **terminal moraine**, also called **end moraine**, is a moraine that forms at the end of the glacier called the snout.

Terminal moraines mark the maximum advance of the glacier. An end moraine is at the present boundary of the glacier.

Terminal moraines are one of the most prominent types of moraines in the Arctic. One famous terminal moraine is the Giant's Wall in Norway which, according to legend, was built by giants to keep intruders out of their realm. It is now known that terminal moraines are created at the edge of the greatest extent of the glacier. At this point, the debris that has been accumulated by plucking and abrasion, that has been pushed by the front edge of the ice is driven no farther, but instead is dumped in a heap. Because the glacier acts very much like a conveyor belt, the longer it stays in one place, the greater the amount of material that will be deposited. The moraine is left as the marking point of the terminal extent of the ice.

In North America, the Outer Lands is a name given to the terminal moraine archipelago of the northeast United States (Cape Cod, Martha's Vineyard, Nantucket, Block Island

and Long Island). Other prominent examples of terminal moraines are the Tinley Moraine and the Valparaiso Moraine, perhaps the best examples of terminal moraines in North America. These moraines are most clearly seen southwest of Chicago.

In Europe, virtually all the terrain in the central Netherlands is made up of an extended terminal moraine. In Switzerland, alpine terminal moraines can be found, one striking example being the moraine at the end of the valley of the glacier Forno in the south-eastern canton of Graubünden near St. Moritz and the Italian border.

In New Zealand the Franz Josef Glacier on the West Coast has created the terminal moraine called the Waiho Loop.

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** (Swedish: Veiki morän) is a particular type of moraine found in northern Sweden and parts of Canada that is characterized by forming a hummocky landskape of moraine plateaus with elevated rims that are intercalated with ponds. The name Veiki derives from the type locality of Veiki in Swedish Lapland. Towards southern Norrland Veikimoraines are gradually replaced by Rogen moraines as the most prominent morainic landforms, and to the north near Finland there is a gradual transition to what is called Pulju moraines. One of the most prominent Veiki moraines is the Lainio Arc (Swedish: Lainio Bågen) in Torne älv, a large scale moraine lobe pointing southeastwards that is made of Veiki moraines. The lobe have been interpreted as the remaines of a large glacial surge.

Although originally thought as being remains of the last glacial period Swedish Veiki moraines are now believed to be of late Saalian age. Veiki moraines are believed to have been formed by supraglacial till that during the late Saale glaciation or the early Eemian interglacial partially protecting dead-ice from melting. This insultation did not hinder parts of the ice to melt creating ponds between the till covered dead-ice. When dead-ice finally melted the ponds remained like plateaus sorrouned by rims of till. The subsequent Weichsel glaciation did not signifantly altered the morphology of the Veiki moraines now exposed.

Glacial moraines are formed by the deposition of material from a glacier and are exposed after the glacier has retreated. These features usually appear as linear mounds of till, a non-sorted mixture of rock, gravel and boulders within a matrix of a fine powdery material. Terminal or end moraines are formed at the foot or terminal end of a glacier. Lateral moraines are formed on the sides of the glacier. Medial moraines are formed when two different glaciers, flowing in the same direction, coalesce and the lateral moraines of each combine to form a moraine in the middle of the merged glacier. Less apparent is the ground moraine, also called *glacial drift*, which often blankets the surface underneath much of the glacier downslope from the equilibrium line. Glacial meltwaters contain rock flour, an extremely fine powder ground from the underlying rock by the glacier's movement. Other features formed by glacial deposition include long snake-like ridges formed by streambeds under glaciers, known as *eskers*, and distinctive streamlined hills, known as *drumlins*.

Stoss-and-lee erosional features are formed by glaciers and show the direction of their movement. Long linear rock scratches (that follow the glacier's direction of movement) are called *glacial striations*, and divots in the rock are called *chatter marks*. Both of these features are left on the surfaces of stationary rock that were once under a glacier and were formed when loose rocks and boulders in the ice were transported over the rock surface. Transport of fine-grained material within a glacier can smooth or polish the surface of rocks, leading to glacial polish. Glacial erratics are rounded boulders that were left by a melting glacier and are often seen perched precariously on exposed rock faces after glacial retreat.

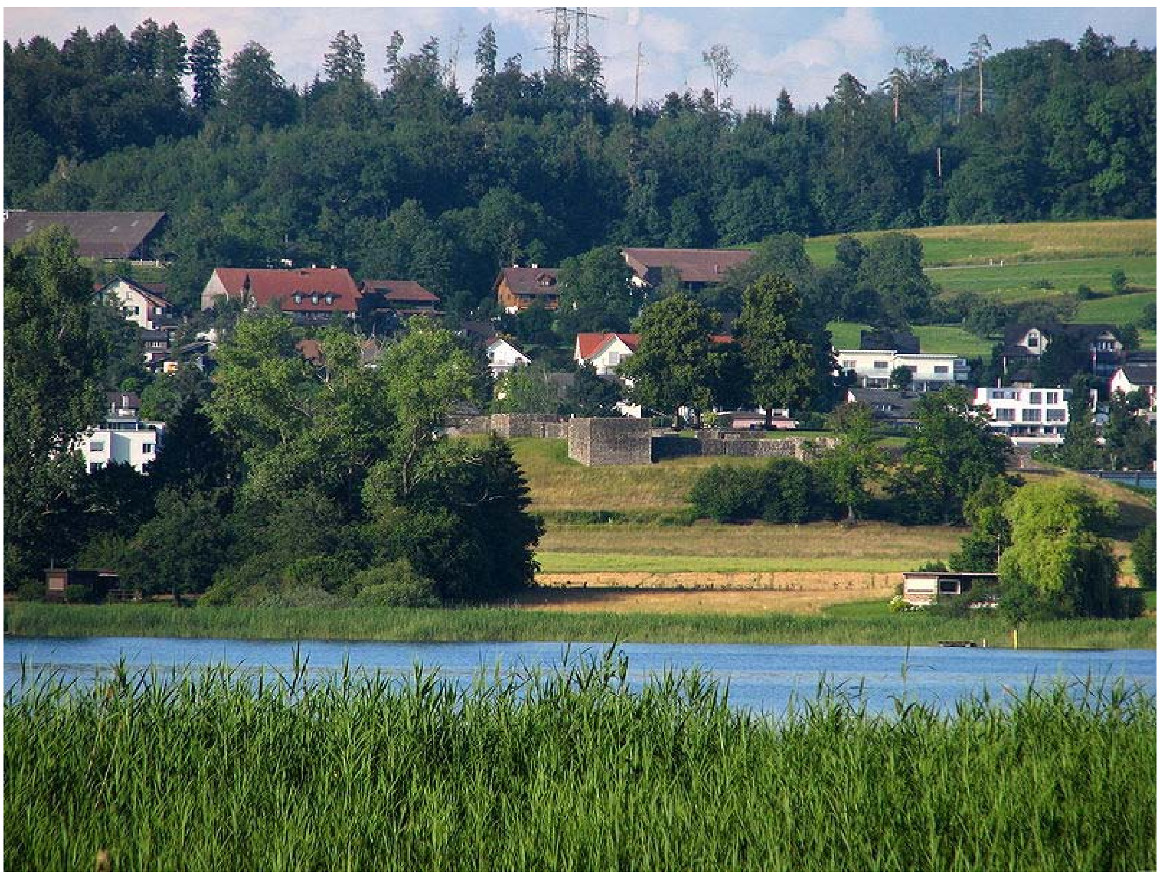
The term *moraine* is of French origin. It was coined by peasants to describe alluvial embankments and rims found near the margins of glaciers in the French Alps. In modern geology, the term is used more broadly, and is applied to a series of formations, all of which are composed of till.

Drumlins

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 composed 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

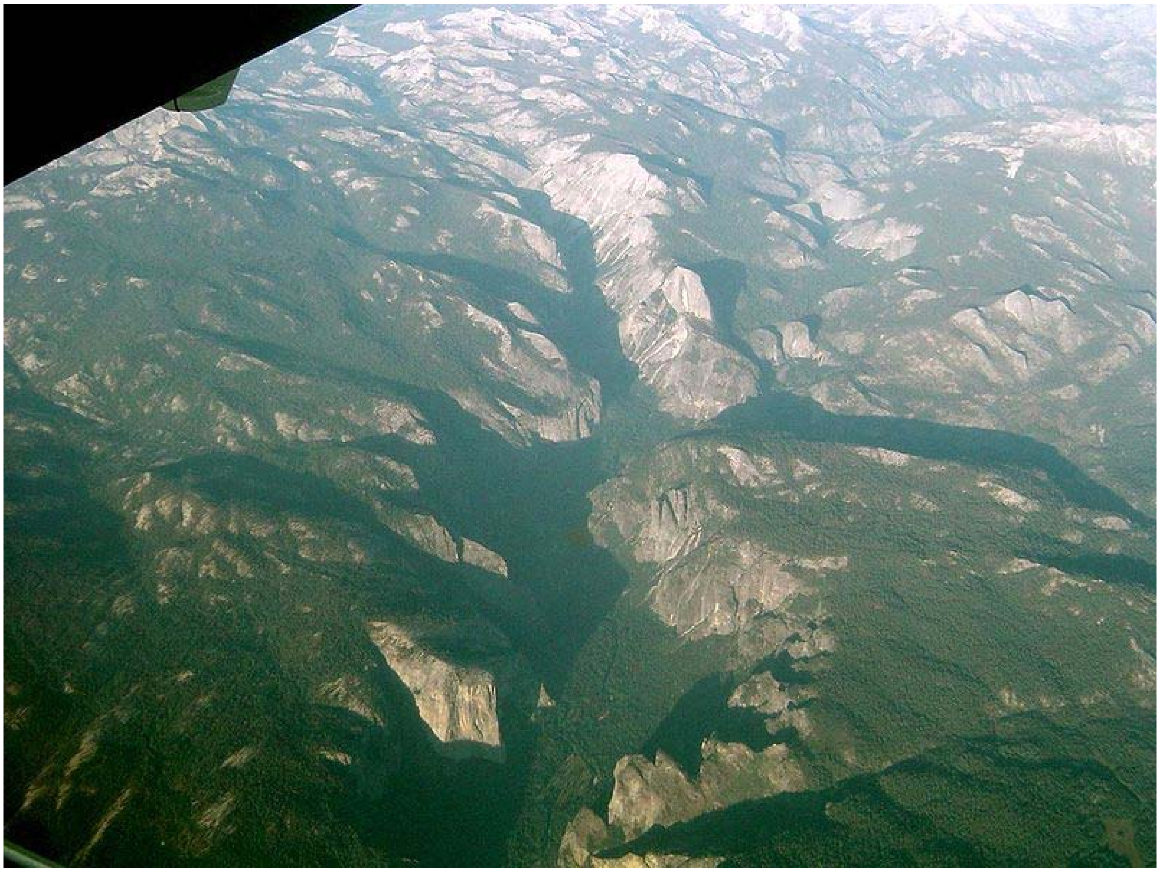
There are many theories as to the exact mode of origin and plenty of controversy among geologists interested in geomorphology. Some consider them a direct formation of the ice, while a theory proposed since the 1980s by John Shaw and others postulates creation by a catastrophic flooding release of highly 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.

Related features

A similar formation, with a more resilient core (generally composed of igneous or metamorphic rock), is a crag. A drumlin composed entirely of rock is known as a rock drumlin.

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.

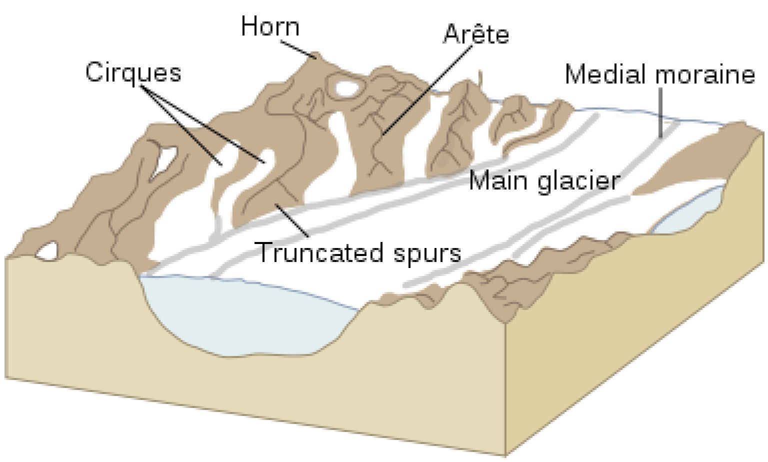
Many glaciers deepen their valleys more than their smaller tributaries. Therefore, when the glaciers recede from the region, the valleys of the tributary glaciers remain above the main glacier's depression, and these are called hanging valleys.

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

Roche moutonnée landform in the Cascade Mountains near Cle Elum, Washington. Direction of glacier movement is from right to left.

In glaciology, a **roche moutonnée** (or **sheepback**) is a rock formation created by the passing of a glacier. When a glacier erodes down to bedrock, it can form tear-drop shaped hills that taper in the up-ice direction.

Name

The 18th-century Alpine explorer Horace-Bénédict de Saussure coined the term 'roches moutonnées' in 1786. He saw in these rocks a resemblance to the wigs that were fashionable amongst French gentry in his era and which were smoothed over with mutton fat (hence 'moutonnée') so as to keep the hair in place.

Nature

Roche moutonnée near Llyn Cau on Cadair Idris, Wales. Direction of glacial movement was from left to right.

The appearance of the erosional stoss and lee feature is very defined on Roche moutonnée as all the sides and edges have been smoothed and eroded in the direction that the glacier that once passed over it. It is often marked with glacial striations.

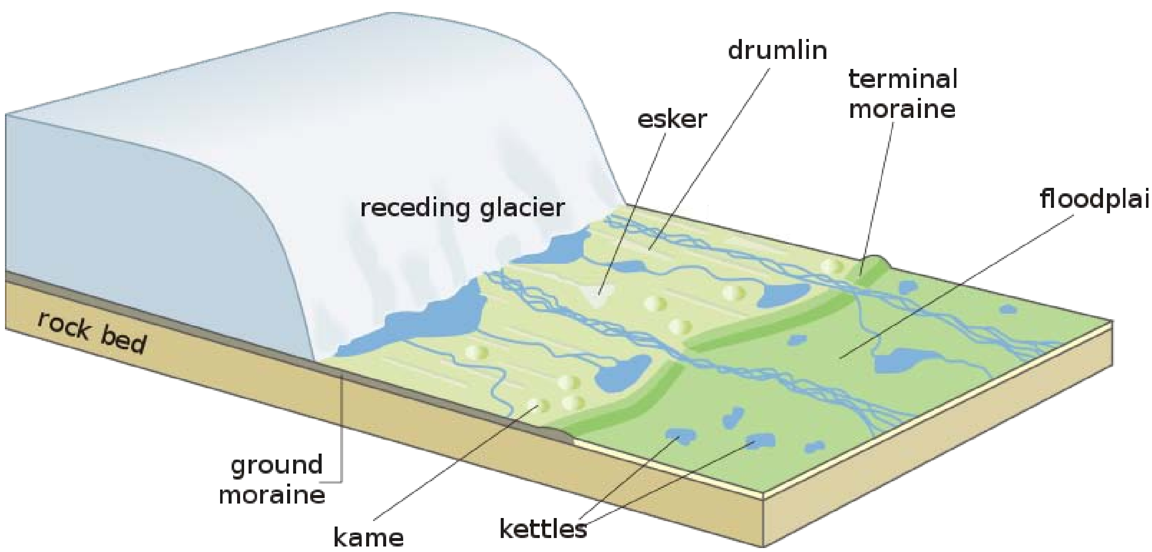
The rough and craggy down-ice side is formed by "plucking", the erosional process in which ice melts slightly by pressure and seeps into cracks in the rock. When the water refreezes, the rock becomes attached to the glacier. But as the glacier continues its forward progress it subjects the stone to frost shattering ripping strips away from the rock formation.

Note that the side profile of a roche moutonnée is opposite to that of a drumlin. In a drumlin, the steep side is *facing* the approaching glacier, rather than *trailing* it.

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.

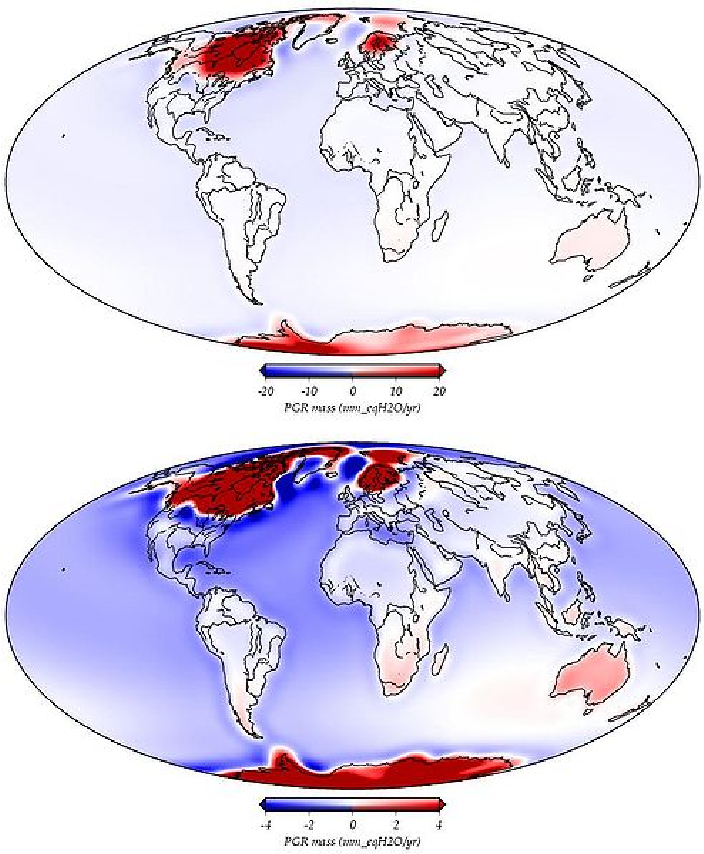
A third type of deposit formed in contact with the ice is characterized by long, narrow sinuous crests, composed fundamentally of sand and gravel deposited by streams of meltwater flowing within, or beneath the glacier. After the ice has melted, these linear ridges or eskers remain as landscape features.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

Chapter-5

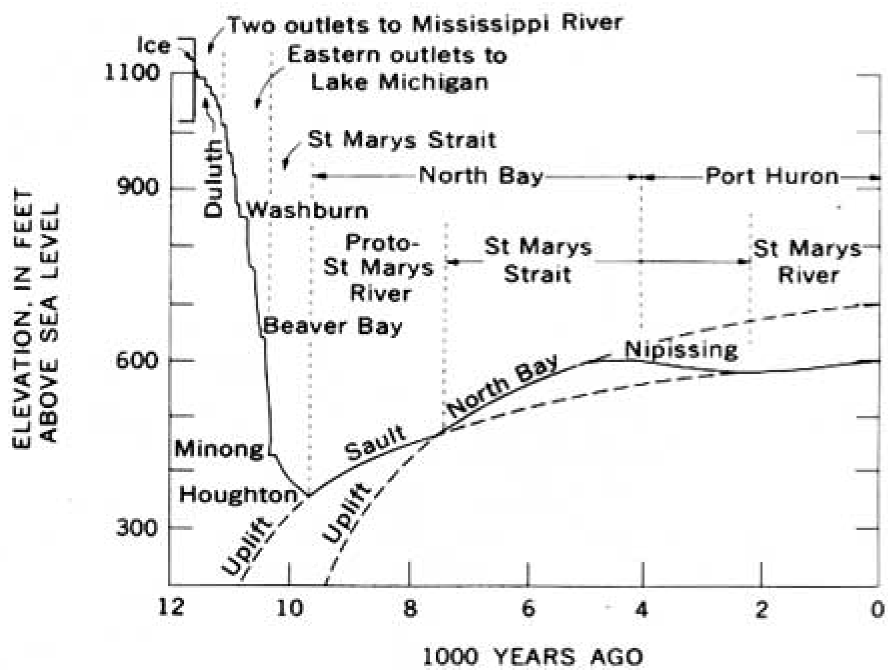
Post-glacial Rebound



A model of present-day surface elevation change due to post-glacial rebound and the reloading of the ocean basins with seawater. Red areas are rising due to the removal of the ice sheets. Blue areas are falling due to the re-filling of the ocean basins when the ice sheets melted and because of the collapse of the forebulges around the ice sheets.

Post-glacial rebound (sometimes called **continental rebound**, **glacial isostatic adjustment**) is the rise of land masses that were depressed by the huge weight of ice sheets during the last glacial period, through a process known as isostasy. It affects northern Europe (especially Scotland, Fennoscandia and northern Denmark), Siberia, Canada, the Great Lakes of Canada and the United States, parts of Patagonia, and Antarctica.

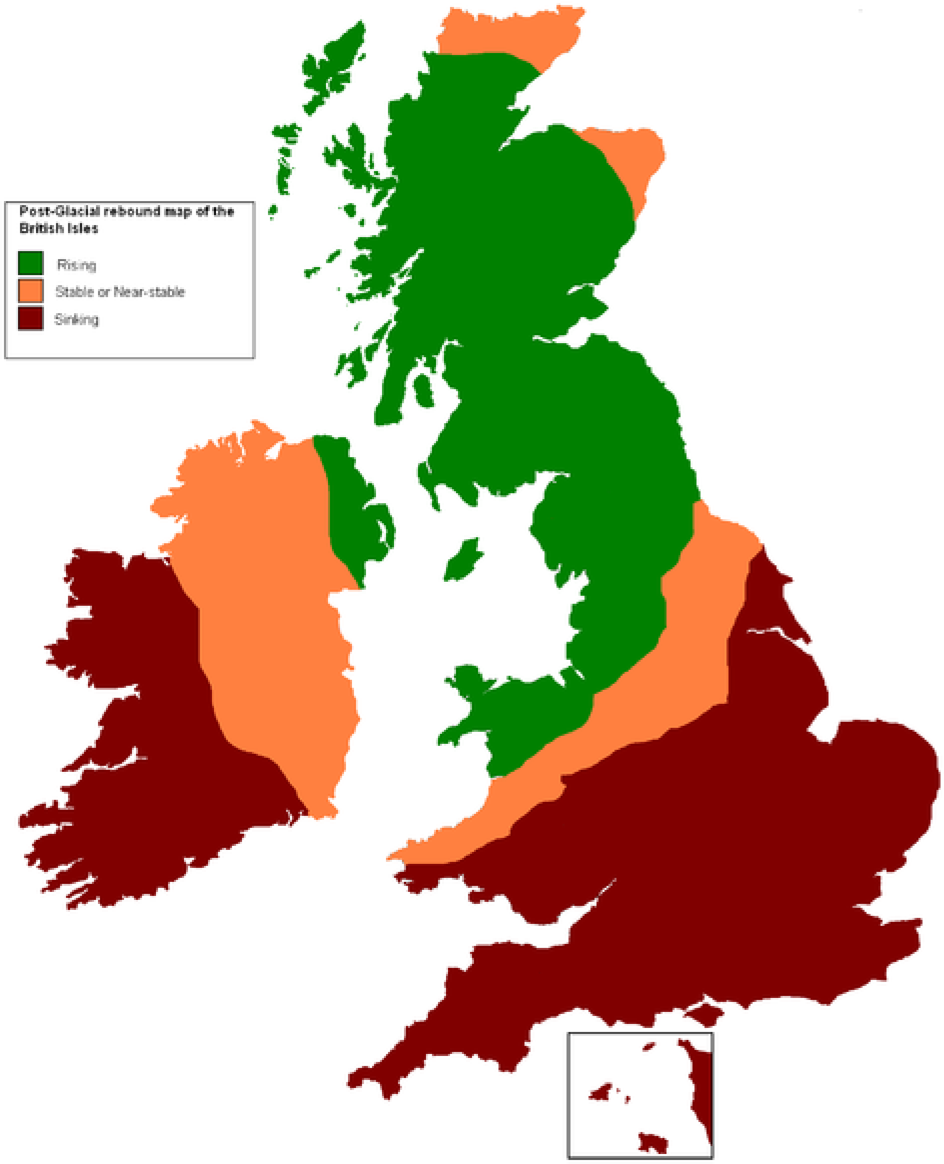
Overview



Changes in the elevation of Lake Superior due to glaciation and post-glacial rebound

During the last glacial period, much of northern Europe, Asia, North America, Greenland and Antarctica were covered by ice sheets. The ice was as thick as three kilometres during the last glacial maximum about 20,000 years ago. The enormous weight of this ice caused the surface of the Earth's crust to deform and warp downward, forcing the fluid mantle material to flow away from the loaded region. At the end of the ice age when the glaciers retreated, the removal of the weight from the depressed land led to slow (and still ongoing) uplift or rebound of the land and the return flow of mantle material back under the deglaciated area. Due to the extreme viscosity of the mantle, it will take many thousands of years for the land to reach an equilibrium level.

Studies have shown that the uplift has taken place in two distinct stages. The initial uplift following deglaciation was rapid (called "elastic"), and took place as the ice was being unloaded. After this "elastic" phase, uplift proceeded by "slow viscous flow" so the rate decreased exponentially after that. Today, typical uplift rates are of the order of 1 cm/year or less. In northern Europe, this is clearly shown by the GPS data obtained by the BIFROST GPS network. Studies suggest that rebound will continue for about at least another 10,000 years. The total uplift from the end of deglaciation depends on the local ice load and could be several hundred metres near the centre of rebound.



Map of Post Glacial Rebound effects upon the land-level of the British Isles.

Recently, the term **post-glacial rebound** is gradually being replaced by the term **glacial isostatic adjustment**. This is in recognition that the response of the Earth to glacial loading and unloading is not limited to the upward rebound movement, but also involves downward land movement, horizontal crustal motion, changes in global sea levels, the Earth's gravity field, induced earthquakes and changes in the rotational motion.

Effects

Post-glacial rebound (or Glacial Isostatic Adjustment) produces measurable effects on: (i) Vertical Crustal Motion, (ii) Global sea levels, (iii) Horizontal Crustal Motion, (iv)

Gravity field, (v) Earth's rotational motion and (vi) State of stress and earthquakes. Studies of Glacial rebound give us information about the flow law of mantle rocks and also past ice sheet history. The former is important to the study of Mantle Convection, Plate Tectonics and the thermal evolution of the Earth. The latter is important to the study of Glaciology, Paleoclimate and changes in Global Sea Level. Understanding postglacial rebound is also important to our ability to monitor recent global change.

Vertical Crustal Motion

This sign in Turku illustrates the sea level in 2,000 BCE.

Much of modern Finland is former seabed or archipelago: illustrated are sea levels immediately after the last ice age.

Erratic boulders, U-shaped valleys, drumlins, eskers, kettle lakes, bedrock striations are just some common signatures of the Ice Age. In addition, post-glacial rebound has caused numerous significant changes to coastlines and landscapes over the last several thousand years, and the effects continue to be significant.

In Sweden, Lake Mälaren was formerly an arm of the Baltic Sea, but uplift eventually cut it off and led to it becoming a freshwater lake in about the 12th century, at the time when Stockholm was founded at its outlet. Marine seashells found in Lake Ontario sediments imply a similar event in prehistoric times. Other pronounced effects can be seen on the island of Öland, Sweden, which has little topographic relief due to the presence of the very level Stora Alvaret. The rising land has caused the Iron Age settlement area to recede from the Baltic Sea, making the present day villages on the west coast set back unexpectedly far from the shore. These effects are quite dramatic at the village of Alby, for example, where the Iron Age inhabitants were known to subsist on substantial coastal fishing.

As a result of post-glacial rebound, the Gulf of Bothnia is predicted to eventually close up at Kvarken. The Kvarken is a UNESCO World Natural Heritage Site, selected as a "type area" illustrating the effects of post-glacial rebound and the holocene glacial retreat.

In several other Nordic ports, like Tornio and Pori (formerly at Ulvila), the harbour had to be relocated several times in the past centuries. Place names in the coastal regions also illustrate the rising land: there are places named 'island', 'skerry', 'rock', 'point' and 'sound' in the inland. For example, Oulunsalo "island of Oulujoki" is a peninsula, with names in the inland such as *Koivukari* "Birch Rock", *Santaniemi* "Sandy Cape", and *Salmioja* "the ditch of the Sound".

In Great Britain, glaciation affected Scotland but not Southern England, and the post-glacial rebound of northern Great Britain is causing a corresponding downward movement of the southern half of the island. This is leading to an increased risk of floods, particularly in the areas surrounding the lower River Thames. Along with rising sea levels caused by global warming, the post-glacial sinking of southern England is likely to seriously compromise the effectiveness of the Thames Barrier, London's most important flood defence, after about 2030.

The combination of horizontal and vertical motion changes the tilt of the surface. That is, locations farther north rise faster. This effect is seen in lakes, where land rises at the northern end and sinks at the southern end. The Great Lakes of North America lie approximately on the 'pivot' line between rising and sinking land. Lake Superior was formerly part of a much larger lake together with Lake Michigan and Lake Huron, but post-glacial rebound raised land dividing the three lakes about 2100 years ago. Today, southern shorelines of the lakes continue to experience rising water levels while northern shorelines see falling levels.

Since the glacial isostatic adjustment process causes the land to move relative to the sea, ancient shorelines are found to lie above present day sea level in areas that were once glaciated. On the other hand, places in the peripheral bulge area which was uplifted during glaciation now begins to subside. Therefore ancient beaches are found below present day sea level in the bulge area. The “relative sea level data”, which consists of height and age measurements of the ancient beaches around the world, tells us that glacial isostatic adjustment proceeded at a higher rate near the end of deglaciation than today.

The present-day uplift motion in northern Europe is also monitored by a GPS network called BIFROST. Results of GPS data shows that a peak rate of about 11 mm/year exist in the north part of the Gulf of Bothnia, but this uplift rate decreases away and become negative outside the former ice margin.

In the near field outside the former ice margin, the land sinks relative to the sea. This is the case along the east coast of the United States, where ancient beaches are found submerged below present day sea level and Florida is expected to be submerged in the future. GPS data in North America also confirms that land uplift becomes subsidence outside the former ice margin .

Global Sea Levels

To form the ice sheets of the last Ice Age, water is taken from the oceans through evaporation, condensation as snow and then deposited as ice in high latitudes. Thus global sea level would fall during glaciation.

The ice sheets at the last Glacial Maximum were so massive that global sea level fell by about 120 metres. Thus continental shelves were exposed and many islands became connected with the continents through dry land. This was the case between the British Isles and Europe, or between Taiwan, the Indonesian islands and Asia. Most important is the existence of a sub-continent between Siberia and Alaska that allowed the migration of people and animals during last glacial maximum.

The fall in sea level also affects the circulation of ocean currents and thus has important impact on climate during the Ice Age.

During deglaciation, the melted ice water returns to the oceans, thus sea level in the ocean increases again. However, geological records of sea level changes show that the redistribution of the melted ice water is not the same everywhere in the oceans. In other words, depending upon the location, the rise in sea level at a certain site may be more than that at another site. This is due to the gravitational attraction between the mass of the melted water and the other masses, such as remaining ice sheets, glaciers, water masses and mantle rocks and the changes in centrifugal potential due to Earth's variable rotation.

Horizontal Crustal Motion

Accompanying vertical motion is the horizontal motion of the crust. The BIFROST GPS network shows that the motion diverges from the centre of rebound. However, the largest horizontal velocity is found near the former ice margin.

The situation in North America is less certain; this is due to the sparse distribution of GPS stations in northern Canada, which is rather inaccessible.

Gravity field

Ice, water and mantle rocks have mass, and as they move around, they exert a gravitational pull of other masses towards them. Thus, the gravity field, which is sensitive to all mass on the surface and within the Earth, will be affected by the redistribution of ice/melted water on the surface of the Earth and the flow of mantle rocks within.

Today, more than 6000 years after the last deglaciation terminated, the flow of mantle material back to the glaciated area causes the overall shape of the Earth to become less oblate. This change in the topography of Earth's surface affects the long wavelength components of the gravity field.

The changing gravity field can be detected by repeated land measurements with **Absolute Gravimeters** and recently by the **GRACE satellite mission**. The changing long wavelength components of Earth's gravity field also perturbs the orbital motion of satellites and has been detected by LAGEOS satellite motion.

Vertical datum

The Vertical datum is a theoretical reference surface for altitude measurement and plays vital roles in many human activities - including land surveying, construction of buildings, bridges etc.. Since postglacial rebound continuously deforms the crustal surface and the gravitational field, the vertical datum needs to be redefined repeatedly through time.

Earth's Rotational Motion

Examination of ancient Chinese and Babylonian eclipse records reveals that the Earth’s rotation rate is not constant. For example, if the rotation rate were constant, then the shadow path of an ancient Babylonian eclipse would lie somewhere across western Europe and the ancient eclipse could not have been observed at the recorded time in Babylon. It is well known that tidal interaction between Earth and the Moon (Tidal Friction or Tidal Dissipation) causes the Earth’s rotation to slow down. But taking into account the tidal interaction alone over-corrects the eclipse path which would lie east of Babylon. In order to have the shadow path pass through Babylon at the recorded time, we need to take into account the effect of Glacial Isostatic Adjustment on Earth’s rotational motion.

To understand how Glacial Isostatic Adjustment affects Earth’s rotation rate, we note that the movement of mass on and beneath the Earth's surface affects the Moment of Inertia of the Earth, and by the Conservation of Angular Momentum, the rotational motion must also change. This is illustrated in the case of a rotating ice skater: as she extends her arms vertically over her head, her moment of inertia decreases and as a consequence, she spins faster. On the other hand, as she extends her arms horizontally, her moment of inertia increases and her spin slows down.

During glaciation, water is taken from the oceans, whose average position is nearer the equator, and deposited as ice over the higher latitudes closer to the poles, which is closer to the rotational axis. This causes the Moment of Inertia of the Earth-ice-water system to decrease and just like the rotating figure skater bringing her arms closer to her body, the earth should spin faster. During deglaciation, the melted ice water returns to the oceans - farther from the rotational axis - and thus causing the Earth’s spin to slow down. Also, the mantle rocks flow in a direction opposite to that of the water, but the rate is much slower. After the end of deglaciation, the dominant mass movement is from the return flow of the mantle rocks back to the glaciated areas at high latititude, making the shape of the Earth less oblate. This process would, in isolation, lead to an increase in the rotation speed of the Earth and therefore to a decrease of the length of day. Lambeck estimated that the isolated effect of post-glacial rebound on the length of the day would be a decrease of about 0.7 milliseconds per century. This process of nontidal acceleration of the rotation of the earth is corroborated by observations of the satellite LAGEOS and is generally attributed to glacial isostatic adjustment.

In addition to the changes in the Earth's rotation rate, the changes in the Moment of Inertia due to Glacial Isostatic Adjustment also cause the rotational axis to move from the current position near the North Pole towards the center of the ice masses at glacial maximum (Polar wander), thus it is moving towards eastern Canada at a rate of about 1 degree per million years.

This drift of the Earth's rotational axis in turn affects the centrifugal potential on the surface of the earth, and thus also affects sea levels.

State of Stress and Intraplate Earthquakes

According to the theory of Plate Tectonics, plate-plate interaction results in earthquakes near plate boundaries. However, large earthquakes are found in intraplate environment like eastern Canada (up to M7) and northern Europe (up to M5) which are far away from present-day plate boundaries. An important intraplate earthquake was the magnitude 8 New Madrid earthquakes that occurred in mid-continental USA in the year 1811.

Glacial loads have provided more than 30 MPa of vertical stress in northern Canada and more than 20 MPa in northern Europe during glacial maximum. This vertical stress is supported by the mantle and the flexure of the lithosphere. Since the mantle and the lithosphere continuously respond to the changing ice and water loads, the state of stress at any location continuously changes in time. The changes in the orientation of the state of stress is recorded in the postglacial faults in southeastern Canada. When the postglacial faults formed at the end of deglaciation 9000 years ago, the horizontal principal stress orientation was almost perpendicular to the former ice margin, but today the orientation is in the northeast-southwest, along the direction of seafloor spreading at the Mid-Atlantic Ridge. This shows that the stress due to postglacial rebound had played an important role at deglacial time, but has gradually relaxed so that tectonic stress has become more dominant today.

According to the Mohr-Coulomb Theory of rock failure, large glacial loads generally suppress earthquakes, but rapid deglaciation promotes earthquakes. According to Wu & Hasagawa, the rebound stress that is available to trigger earthquakes today is of the order of 1 MPa. This stress level is not large enough to rupture intact rocks but is large enough to reactivate pre-existing faults that are close to failure. Thus, both postglacial rebound and past tectonics play important roles in today's intraplate earthquakes in eastern Canada and southeast USA. Generally postglacial rebound stress could have triggered the intraplate earthquakes in eastern Canada and may have played some role in triggering earthquakes in eastern USA including the New Madrid earthquakes of 1811. The situation in northern Europe today is complicated by the current tectonic activities nearby and by coastal loading and weakening.

Recent Global Warming

Recent global warming has caused mountain glaciers and the ice sheets in Greenland and Antarctica to melt and global sea level to rise. Therefore, monitoring sea level rise and the mass balance of ice sheets and glaciers allows us to understand more about global warming.

Recent rise in sea levels has been monitored by tide gauges and Satellite Altimetry (e.g. TOPEX/Poseidon). In addition to the addition of melted ice water from glaciers and ice sheets, recent sea level changes are also affected by the thermal expansion of sea water due to global warming, sea level change due to deglaciation of the last Ice Age (postglacial sea level change), deformation of the land and ocean floor and other factors. Thus, to understand global warming from sea level change, one must be able to separate all these factors, especially postglacial rebound, since it is one of the leading factors.

Mass changes of ice sheets can be monitored by measuring changes in the ice surface height, the deformation of the ground below and the changes in the gravity field over the ice sheet. Thus ICESat, GPS and GRACE satellite mission are useful for such purpose. However, glacial isostatic adjustment of the ice sheets affect ground deformation and the gravity field today. Thus understanding glacial isostatic adjustment is important in monitoring recent global warming.

One of the possible impacts of global warming triggered rebound may be more volcanic activity in previously ice capped areas such as Iceland.

Applications

The speed and amount of postglacial rebound is determined by two factors: the viscosity or rheology (i.e., the flow) of the mantle, and the ice loading and unloading histories on the surface of Earth.

The viscosity of the mantle is important in understanding mantle convection, plate tectonics, dynamical processes in Earth, the thermal state and thermal evolution of Earth. However viscosity is difficult to observe because creep experiments of mantle rocks take thousands of years to observe and the ambient temperature and pressure conditions are not easy to attain for a long enough time. Thus, the observations of postglacial rebound provide a natural experiment to measure mantle rheology. Modelling of glacial isostatic adjustment addresses the question of how viscosity changes in the radial and lateral directions and whether the flow law is linear or nonlinear.

Ice thickness histories are useful in the study of paleoclimatology, glaciology and paleo-oceanography. Ice thickness histories are traditionally deduced from the three types of information: First, the sea level data at stable sites far away from the centers of deglaciation give an eastimate of how much water entered the oceans or equivalently how much ice was locked up at glacial maximum. Secondly, the location and dates of terminal moraines tell us the areal extent and retreat of past ice sheets. Physics of glaciers gives us the theoretical profile of ice sheets at equilibrium, it also says that the thickness and horizontal extent of equilibrium ice sheets are closely related to the basal condition of the ice sheets. Thus the volume of ice locked up is proportional to their instantaneous area. Finally, the heights of ancient beaches in the sea level data and observed land uplift rates (e.g. from GPS or VLBI) can be used to constrain local ice thickness. A popular ice model deduced this way is the ICE5G model. Because the response of the Earth to changes in ice height is slow, it cannot record rapid fluctuation or surges of ice sheets, thus the ice sheet profiles deduced this way only gives the "average height" over a thousand years or so.

Glacial isostatic adjustment also plays an important role in understanding recent global warming and climate change.

Discovery

Before the 18th century, it was thought in Sweden that sea levels were falling. On the initiative of Anders Celsius a number of marks were made in rock on different locations along the Swedish coast. In 1765 it was possible to conclude that it was not a lowering of sea levels but an uneven rise of land. In 1865 Thomas Jamieson came up with a theory that the rise of land was connected with the ice age that had been first discovered in 1837. The theory was accepted after investigations by Gerard De Geer of old shorelines in Scandinavia published in 1890.

Legal status

In areas where the rising of land is seen, it is necessary to define the exact limits of property. In Finland, the "new land" is legally the property of the owner of the water area, not any land owners on the shore. Therefore, if the owner of the land wishes to build a pier over the "new land", he needs the permission of the owner of the (former) water area. The landowner of the shore may redeem the new land at market price.

Chapter-6

List of Glaciers

Canada Glacier in Antarctica

Due to somewhat sparse information, some glaciers, especially those in the tropics, may no longer exist as listed. This is especially true for glaciers in Africa and New Guinea.

Glaciers in India

Jammu and Kashmir

- Siachen Glacier is the second longest glacier outside of the polar regions and largest in the Himalayas-Karakoram region.
- Nubra Glacier

- Chong Kumdan Glacier
- Drang Drung Glacier
- Rimo Glacier

Himachal Pradesh

- Bara Shigri
- Chandra Glacier
- Chandra Nahan Glacier
- Bhadal Glacier
- Bhaga Glacier
- The Lady of Keylong
- Mukkila Glacier

Sikkim

The Zemu Glacier

- Zemu Glacier
- Rathong Glacier
- Lonak Glacier

Uttarakhand

Goumukh, terminus of the Gangotri glacier (lower right in image, behind prayer flag). The Bhagirathi peaks rise in the background.

- Gangotri Glacier
- Kalabaland Glacier
- Meola Glacier
- Milam Glacier
- Namik Glacier
- Panchchuli Glacier

Glaciers in Europe

Iceland

- Barkárdalsjökull
- Eiríksjökull
- Eyjafjallajökull
- Drangajökull
- Gljúfurárjökull
- Langjökull
- Hjaltadalsjökul

Norway

- Austfonna
- Blåmannsisen
- Buarbreen
- Folgefonna
- Frostisen

- Gihtsejiegŋa
- Harbardsbreen

Germany

- Höllentalferner
- Zugspitze

Austria

- Pasterze Glacier
- Schlatenkees Glacier
- Kitzsteinhorn Glacier
- Hintertux Glacier

Switzerland

- Aletsch Glacier
- Allalin Glacier
- Brunegg Glacier
- Corbassière Glacier
- Fee Glacier
- Ferpècle Glacier
- Fiescher Glacier
- Findel Glacier
- Gauli Glacier
- Gorner Glacier
- Haut Glacier d'Arolla
- Bas Glacier d'Arolla
- Hufi Glacier
- Kander Glacier
- Lang Glacier
- Lower Grindelwald Glacier

France

- Mer de Glace
- Glacier d'Argentière
- Glacier de Bellecote
- Glacier des Bossons
- Glacier du Peclet
- Glacier de Saint Sorlin

Italy

- Brenva Glacier
- Calderone glacier (Ghiacciaio del Calderone)
- Canin Glacier

• Careser Glacier
• Crystal Glacier
• Forni Glacier

Russia

- Kolka Glacier
- Severny Island ice cap (Largest by area in Europe)

Glaciers in United States

Glaciers of Alaska

The Homer Spit is believed to be the remains of a glacial moraine

- Aialik Glacier - Kenai Peninsula
- Aurora Glacier - Glacier Bay
- Bacon Glacier
- Barnard Glacier
- Bering Glacier
- Black Rapids
- Brady Glacier
- Carroll Glacier - Glacier Bay
- Chenega Glacier - Prince William Sound
- Clark Glacier - Glacier Bay
- Columbia Glacier - Prince William Sound

- Fairweather Glacier
- Grand Pacific Glacier - Glacier Bay

Gulkana glacier in the Alaska Range

- Grewingk Glacier - Kenai Peninsula
- Guyot Glacier
- Harding Icefield -

Washington

Boulder Glacier, Mount Baker.

There are hundreds of named glaciers in Washington. This list contains most of the glaciers on the volcanoes, but is very incomplete otherwise.

Olympic Mountains

Mount Olympus

- Blue Glacier
- Hoh Glacier
- Hubert Glacier
- Humes Glacier
- Jeffers Glacier
- White Glacier