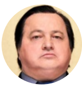
Vasil'chuk 2004 GLACIERS, ICEBERGS AND GROUND ICE



Yurij Vasil'chuk
Lomonosov Moscow State University

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GLACIERS, ICEBERGS AND GROUND ICE

Yu. K. Vasil'chuk *and* **A.C. Vasil'chuk**
Departments of Geography and Geology, Lomonosov's Moscow State University, Moscow, Russia

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Contents

Summary

Ice is the most abundant solid substance on Earth. It can exist in a great range of temperatures and pressures. The total ice mass contained in glaciers, icebergs, ground ice, snow cover and the atmosphere is 2423 x 10^{22} tons. As a consequence of climate, more than 96.6% of ice area and 90% of ice volume is concentrated in the Greenland and Antarctic ice sheets. Outside the ice sheet area there are several isolated ice domes. Large slabs of floating ice, called ice shelves surround much of the Antarctic continent. The ice shelf is considered in relation to the ice sheet that feeds it. There are several types of icebergs: tabular, domelike, pyramidal and destroyed. There are 28 types of fresh ice, which can be divided into three groups: congelation ice (all ice of fresh water reservoirs and streams), ground ice (sedimentation ice) and metamorphic (glacier) ice. Floating ice can be distinguished by its origins into marine, river, lake and land varieties, as well as by dynamics (immobile or fast ice and mobile, also called drift or pack ice).

Ground ice is a general term use to refer to all types of ice formed in freezing and frozen ground. Sediments that contain excess ice are often referred as "ice rich" or "icy" sediments. Various types of buried ice (glacier ice, sea, lake or river ice) fall under this classification, which results in ten mutually exclusive ground ice forms. The terms massive ice or massive ice bodies are usually reserved for relatively pure ground ice where the ice content averages at least 250% for a thickness of several meters. Segregated ice is a broad term for soil with a high ice content. Major developments in the field of glaciology have been due to the wealth of information obtained from analyses of cores from polar ice sheets.

1. Ice in Space

Ice can exist in a great range of temperatures and pressures. There are eleven varieties of ice, which are differentiated from each other in a manner similar to rocks. Under terrestrial conditions, it is possible to form only one variety of ice, which is called ice–I. The other ten varieties are only stable under specific conditions (definite ratios of pressure, density and temperature). However, almost all of them may be found in the Solar System.

Water content in the planets increases with distance from the Sun, and ice becomes the predominant water state. On Mercury there is no water, while on Venus the atmosphere contains 0.02% water, occurring as water vapor. On Earth, water exists as vapor, liquid and solid. Both poles of Mars are covered by ice sheets, which oscillate seasonally. In winter they range up to 45^{o} latitudes. In sub–polar areas, the ice sheets mainly consist of frozen CO_2 with some water ice. In the middle part of the ice sheets there are gas hydrate zones, while in circumferential areas the ice sheets consist of ordinary ice. There is no surface water on Jupiter and Saturn, but their nuclei partially consist of ice. Jupiter's moons consist of as much as 70–90% ice. Small moons (with diameters less than 730 km) are composed of common ice (ice–I), while larger ones are built up from various ice modifications (ice I, II, V, VI, VII.). Jupiter's moon Europa is completely covered by ice. Uranus' moons are possibly covered by layers of icy particles.

Ice is the most abundant rock on Earth. The total ice mass contained in glaciers, icebergs, ground ice, snow cover and the atmosphere is 2423 x 10^{22}tons (see Table1**)**. This is nearly thirty times more than the total mass of land surface water. The high purity of glacier water is very important when considering the worsening water deficiency on Earth.

Table 1. Mass and distribution area of ice on Earth.

Source: Shumsky, Krenke, 1969

2. Glaciers

Ice covers more than 16.3 x 10^6 km^2, i.e. 11% of the Earth surface. The total ice volume of modern glaciers ranges from 26.8 x 10^6 km^3 to 30.3 x 10^6km^3. If this ice layer were to cover the Earth uniformly, its thickness would be approximately 55–60 m). However, ice is distributed unevenly throughout the Earth. Due to climatic features, more than 96.6% of ice area and 90% of ice volume is concentrated in the Greenland and Antarctic ice sheets (see Table 2).

Table 2. Modern surface glaciation of Earth

Source: Dolgushin, 2000

Glaciers are divided into several types according to their morphology and location:

- Summit glaciers (cone summit, flat summit and caldera glaciers).
- Slope glaciers (slope glaciers, hanging glaciers, corrie glaciers and corrie-valley glaciers).
- Valley glaciers (valley glaciers, dendrite glaciers, expanded–foot or bulb glaciers, piedmont glaciers and basin glaciers).
- Plateau glaciers, diffluent glaciers, recemented glaciers and others.

3. Ice Sheets

The area of the Antarctic Continent is 14 x 10^6 km^2; the mean diameter of the Antarctic ice sheet is 4000 km, the least is 2900 km, and the maximum is 5500 km. Mean surface level is 2040 m, the maximum is more than 4000 m above sea level. The Antarctic Continent is covered by thick ice sheets broken in several places by nunataks and mountains.

The coastline is about 30 000 km, of which 28 000 km is ice cliff. Exposed rock occupies only 0.3% of the total area, resulting in the area of the Antarctic ice sheet being roughly equal to the continental area. Most of the central Antarctic continental bedrock is lower than sea level. According to geologic structure, the continent is divided into the Western and Eastern Antarctic Continent. The surface of the Eastern Antarctic Continent is high and flat, with several slightly pronounced domes. The ice sheet surface rises sharply from the coast to the inland areas where the sub-horizontal ice surface is more than 3000 m above sea level.

In the Western Antarctic Continent, the ice surface is lower, and there are three distinctive ice domes, as follows: the Middle Dome (2000 m above sea level), another in Marie Byrd Land (2000 m above sea level), and in the south of the Antarctic Peninsula (2150 m above sea level).

The Antarctic ice sheet consists of (1) stagnate ice slightly sloping toward the sea with several domes, (2) active glaciers and (3) shelf glaciers. The thickness of the shelf glaciers ranges from ten meters to 1300 m, with a mean thickness of about 400 m. The thickness of active glaciers is ten meters to 1000 meters. Mean thickness of the Antarctic ice cover is 1800 m, and the total volume is 24.9 x 10^6 km, i.e. this is 22.4 x 10^6 km^3 or 90% of the ice on Earth.

Greenland is the largest island on Earth. Its area is 2 186 000 km^2; the length is about 2900 km, and the maximum diameter is 1200 km. Ice sheet covers 80% (1726.4 km^2) of the island's area, mainly inland areas. Wide ice streams come down to the ocean and fiords.

There are many flat hills, ridges, terraces and cavities in the elliptical surface of the Greenland Ice Sheet. They are formed due to local differences in snow accumulation, deflation, ablation, and ice movement. The surface is made rugged by summer melt water streams at heights lower than 1800 m in marginal areas. The mean height of the ice surface is 2135 m. The summit area of the ice sheet is a convex crown stretching from the North to the South. There are two top points of the crown, the South Dome (64^o N) at 2850 m and the North Dome (72^o N) at 3300 m.

An ice divide is located along 37^o W, close to the eastern margin of Greenland. In the North, it is close to the center of the island. Outside the ice sheet area there are several isolated ice domes. The largest is Succertoppen with an area of 2330 km^2 and a height of 1200 – 1700 m, located on the southwest coast. Fleid-Iceblink on the northeast coast of Greenland is another. The mean thickness of the ice sheet is 1790 m. The maximum thickness is 3416 m in the center part of the island, which is considerably less than that in the Antarctic continent (4350 m).

4. Ice caps of Arctic and Antarctic Islands

The majority of glaciers in the Eurasian and Canadian Arctic are classified as ice caps by their morphology and as diffluent glaciers by their movement features. The ice caps of many islands come down to the sea, forming ice cliffs as a result. The largest Eurasian ice caps are located in Iceland, Franz Josef Land, Spitsbergen, Novaya Zemlya Island, Severnaya Zemlya Island, Bennett Island, Henriette Island, Jeannette Island, Victoria Island, Ushakov Island and Schmidt Island. The largest ice caps of the Canadian Arctic Archipelago are on Baffin Land, Ellesmere Island, Devon Island, Axel-Heiberg, Melville Island, and Meighen Island. In the vicinity of Ellesmere Island is located a small shelf glacier named Word Hunt, that produces table icebergs.

The largest glacier in Iceland is Vatnajцkull, with a maximum thickness of about 1000 m. The thickness of most arctic glaciers (both Eurasian and Canadian) ranges from 300 to 500 m. Therefore, the total volume of water contained in arctic glaciers is near 70 000 km^3 (see Table 3).

Table 3. The largest ice caps of Arctic islands.

Source: Dolgushin, 2000

Due to the very cold and humid climate around the Antarctic Continent, glaciers form on the many islands. The closer to the Antarctic Continent, the thicker the glaciers become, and mountain glaciers become mountain ice sheets.

5. Mountain glaciers

Mountain glaciers in many dry areas of the Earth provide a considerable part of the water used for irrigation in those places. Snow line location is responsible for glacier formation in otherwise arid geography. Above the snow line the accumulation of solid precipitation is greater than the loss of mass to thawing, evaporation, or run off. The snow line level widely oscillates, depending on moisture and heat balance and local climatic conditions from sea level in the Antarctic Continent to 6000 – 6500 m above sea level on the Tibetan Plateau.

In Europe, modern glaciation is concentrated in Scandinavia, the Alps, Caucasus and Ural Mountains. There are small glaciers in the Khibiny and Perinea Mountains. According to Dolgushin (2000), there are 9529 glaciers in Europe, with a total area of 7395 km^2. On the Scandinavian Peninsula, mountain ice sheets are predominant while in the Caucasus and Alps there are complete and simple Alpine glaciers. In the Ural, Khibiny and Perinea Mountains, corrie and hanging glaciers are found. Westerly

Moisture Transport causes changes in snow line elevation from west to east. Its height in the French Alps is 2500 m, while in the Central Caucasus it is 3700 – 3800 m.

In Asia, ice covers the high mountain areas of Tien Shan, the Pamirs, Karakorum and the Himalayas where many very large dendrite-type and complex mountain valley glaciers come down considerably lower than the snow line. The majority of the mountain areas in Asia are located far from sea coasts in continental climatic conditions with poor precipitation. For this reason, the snow line is located very high (for example in Tibet at 5000 –6000 m). In the north of Asia, in spite of low temperatures and topography, glaciation is only moderate because of dryness and weak precipitation. The glaciers of Kamchatka, the Far East and South-East Asia aliment mainly from the Pacific. The glaciers in Karakorum and the Himalayas receive alimentation from the Indian Ocean, while in the case of the glaciers of western and central Asia the moisture comes from the Atlantic.

The largest centers of modern glaciation in North America are positioned in the North American Cordillera. As the relatively warm Pacific Ocean is very close to the western slopes of the Cordillera, they accumulate a great quantity of precipitation (more than 3000 mm per year). On the opposite slopes the precipitation is only 2000 mm per year. The snow line on the western and southwestern slopes is at 500 to 1000 m while on the eastern and northeastern slopes it is between 2500 and 3000 m. The sublatitudinal Alaska Ridge is also a large center of glaciation that aliments from the Pacific. Glaciation of the Rocky Mountains is not very great because no moisture source is near them.

In South America, modern glaciation occurs in the Andes, which extend for 7000 km along the Pacific coast. Throughout most of the Andes the snow line in located above 5000 m, and large centers of mountain valley glaciation occur in the Northern and Southern Patagonia Plateau. In the northern part of the range from 32^{o}S the glaciers receive moisture from the Atlantic and Amazon Lowland and less from the Pacific, while to the south it comes from the Pacific only. The snow line height ranges from 5000 m in the Atlantic province to 1000 – 2000 m in the Pacific one. Charles Darwin described a glacier in southern Chile that is located very close to subtropical forest.There are only three mounts which are higher than the snow line height (i.e. as high as 5000 m) in Africa. These are Mt. Kilimanjaro, Mt. Kenya and the Ruwenzoris. Their glaciers receive alimentation from the Indian Ocean.

In New Zealand, glaciation takes place only in the Southern Alps on South Island (between 43 and 44^{o}30'S). Moisture comes here from the warm Tasman Sea. The snow line is as low as 300 –400 m above sea level and glaciers meet subtropical forest with *Podocarpus, Dacridium, Phyllocladus*and other species.

On the New Guinea Island glaciers and firn occur in the Carstence Mountains at a height of more than 5000 m. The most interesting feature of the glaciers here is the absence of seasonal temperature and moisture oscillations, while daily ones are noticeable. In the morning and bright sunshine the glaciers and firn thaw, but in the afternoon the convective clouds cover the mountains and snow falls. In the evening the temperature is below zero and snow is preserved until the morning.

6. Hydrology of glaciers

Flow records of glacier-fed rivers provide some information about the movement of water through ice. The discharge (volume of flow per unit of time) has a marked diurnal variation superimposed on a base flow volume that changes more slowly. The maximum discharge may be roughly twice the minimum. The peak daily discharge comes a few hours after the peak in melt but, as summer advances, the daily rise and fall in discharge becomes more rapid and the time lag decreases. Total daily discharge usually reaches its maximum in late July or early August. When melting is stopped by summer snowfalls, the diurnal variation in discharge ceases and the base flow declines. When melting begins again, the base flow takes several days to reach its former level. Water flows throughout the winter; there is no diurnal variation and the base flow appears to be comparable with the volume of water melted from the underside of the glacier by geothermal and frictional heat.

During the melt season, a temperate glacier acts like a reservoir that is drained continuously by the streams emerging at its terminus, that are supplied daily by melt water. The glacier contains a considerable amount of free water. While much of this water is in the snow and firn, the ice also contains appreciable quantities in crevasses, moulins, and cavities. Some of these are isolated, at least temporally, while others are connected by a system of channels. As long as channels and cavities contain enough water, ice flow cannot close them. They close during the winter and are gradually reopened by melt water at the beginning of the following summer. During this period, some water is stored because the channels are too small to carry off all the melt water. The channel system develops during the summer as increasing amounts of melt water enlarge old channels and open up new ones. Thus the time that water spends in the glacier, and the lag between maximum melt and maximum run-off are reduced. During this period also, cavities that were initially isolated become connected to this main drainage system, so that, for a time, run-off exceeds melting. Floods recorded in stream flow records probably resulted from the draining of a series of a previously filled cavities, e.g. when barriers between a series of water filled crevasses are breached successively. Water pressure is determined by the water supply and the resistance of the channels to flow. Diurnal variations in pressure, and the high pressures observed after rain or heavy melt, must be caused by changes in the amount of water because the channels do not have time to enlarge. High pressure results from the restricted size of the channels; ice flow starts to close them in winter, when surface melting stops.

Increases in velocity during summer, especially after periods of heavy melting, have been observed in ablation areas of glaciers in polar regions. Variations in sliding velocity appear to offer the only explanation. This in turn implies that the basal ice is at the melting point and that surface melt water can penetrate to the bed, even though the bulk of the ice is well below the melting point. In polar ice sheets, the basal ice may be at the melting point in some places and not in others. Melt water formed by geothermal and frictional heat is contained by surrounding cold ice to form sub-glacial lakes. Such lakes have been detected in Antarctica. When ice flows from a region where the basal ice is melting to a place where it is not, melt water will refreeze on the ice and in this way may be incorporated into the ice. The largest known Antarctic sub-glacial lake is located at, and extends 230 km north of, Vostok station, East Antarctica. This sub-glacial lake is approximately 50 km wide and several hundred meters deep. Large sub-glacial lakes manifest themselves as flat regions on the ice surface. The mean length of sub-glacial lakes that have expression on the ice-sheet surface is about 8.3 km, while those that do not exhibit a surface morphological manifestation have a mean length about 3.3 km.

7. Surges

A surge is a period of exceptionally rapid sliding during which a large volume of ice is transferred from a reservoir area to the terminal part of the glacier. Surges are confined to the lower parts of many glaciers. In others the whole glacier is affected. During quiescent periods, flow with exceptionally high values of velocity thickens the upper part of the area affected by the surge while the lower part stagnates and is thinned by ablation, and stagnant ice advances down the glacier. The surge starts when the glacier attains a critical profile. Rapid sliding is probably preceded by an accumulation of water at the bed. There is no evidence of a surge in any glacier known to be permanently frozen to its bed. Surges occur at more or less regular intervals in the same glacier. The interval is determined by the time needed to build up the critical profile. This depends on the area of the glacier and accumulation and ablation rates. Parts of at least some surging glaciers slide during quiescent periods. The velocities increase from year to year, at least in the area of rapid thickening. The velocity and the accumulation are much greater in summer than in winter. This suggests that the sliding velocity is predominantly controlled by the profile while surface melt water has an important secondary effect. An improved understanding of glacier sliding is a prerequisite to a full understanding of surges. The triggering mechanism and the reasons for the very high sliding velocities will remain obscure until we know how much water is present at the glacier bed, where it is stored, and how it drains. The best–studied surging glacier, and the only one to have been observed over a whole period between surges, is Medvezhiy Glacier in the Pamirs Mountains of Central Asia (Dolgushin, 2000). This glacier surged in 1963 and 1973, and each surge lasted about three months. Medvezhiy Glacier has an area of 25 km^2 and spans an elevation range from 2850 to 3500 m. Icefall separates the accumulation area from the tongue, which is confined in a narrow valley. Surging glaciers are found only in certain regions, but these cover a wide range of climates and geological features. A total of 204 surging glaciers have been identified in western North America, most of them near the Alaska – Yukon border. Most of the surging glaciers in the Alaska Range lie along parts of the Denali Fault and in the St. Elias Mountains. Most the glaciers in the valley of Steel Creek surge. About 40 surging glaciers have been identified in the Pamirs, 21 in Tian–Shan, 7 in the Caucasus, and one in the Kamchatka Peninsula. They have also been reported from the Karakoram, the Chilean Andes, Iceland, East Greenland, and Arctic Canada. Several glaciers and ice caps in Spitsbergen have surged, including one that advanced 21 km, the greatest movement so far recorded in a surge. Glaciers in the same area do not necessarily surge at the same time.

8. Jokulhlaup

Jokulhlaup is the term for sudden and rapid draining of a glacier-dammed lake. Such events, which cause extensive flooding and have often resulted in loss of life, have been reported from Iceland (where the name *jokulhlaup* originates), Norway, Russia, Alaska, Canada, South America, and New Zealand. There are historical records of such catastrophes in the European Alps, but large floods attributable to this phenomenon do not occur there at present.

Glacier-dammed lakes form in various situations. Probably the most common is when a glacier blocks the stream draining a side valley. Another case, and a particularly dangerous one because of the large volume of water impounded, is when a surging glacier blocks the drainage from a major valley. Jokulhlaups also result from the sudden release of water stored in cavities within a glacier or in sub-glacial lakes. The surface topography over a sub-glacial lake has an important effect upon the critical condition for the triggering of jokulhlaup from a sub-glacial lake. The critical lake level may be determined by the presence of an annulus of thicker ices bordering a surface depression. A jokulhlaup event lasting 6 months occurred near Casey Station, Low Dome, Antarctica in late March 1985. The event first recorded was an outburst of water from beneath a cold ice-cap terminus on Low Dome. From the results of oxygen isotope and solute analyses, the water was found to have originated as basal melt water, which had been derived from the melting ice from before the last glacial maximum. It contained a high total solute load with a dominant enrichment in alkalis Na^+ and K^+. This alkali dominance indicated that the water had been squeezed through fine grained sub-glacial sediments with a significant clay-sized fraction. Evidence from sub-glacial topography, basal ice exposures, and the sedimentology of nearby supra-glacial moraines support the presence of an ice-marginal sub-glacial reservoir as the storage source of the jokulhlaup water.

In the great majority of jokulhlaups, the lake drained under the ice before the water level has reached the ice surface. The floods are sometimes extremely large. The peak flow during the 1934 outburst at Grimsvotn in Iceland was about one-quarter the flow of the Amazon. An area of 100 km^{2}was flooded.

9. Icebergs

The extent and thickness of Arctic and Antarctic sea–ice cover controls the gas and energy balance between the atmosphere and the ocean, affecting ocean circulation and the global climate system.

There are several types of icebergs: tabular, domelike, pyramidal and destroyed. Tabular icebergs have a flat surface because they usually break off from a shelf glaciers. They may be several kilometers in length and width. They are often found around the Antarctic Continent, but sometimes they form near Greenland and near the ice caps of Arctic Islands. Domelike icebergs mainly come from diffluent glaciers or from ice bluffs; their height is about 70 –100 m. Pyramidal icebergs have a pyramid shape, while destroyed icebergs have irregular shapes and often have several summits. The lower part of the iceberg is under the water.

There are different mechanisms of iceberg destruction. A fast moving diffluent glacier can split off due to external stress. Sometimes icebergs split off as a result of tides. Large icebergs may traverse a long distance, more than 4000– 6000 km from their place of origin. The Circumpolar Stream around the Antarctic Continent is a cause of ice drift moving from east to west. In the Ross Sea and Weddell Sea local circulation occurs up to 60^{o} S. Ice discharge in the Antarctic ice sheet takes place due to iceberg calving. There are tens of thousands of icebergs around the Antarctic Continent, sometimes with a volume of nearly 1000 km^3. Such an iceberg can contain half of the annual Antarctic ice sheet discharge. Some icebergs are released from the Antarctic Circumpolar Stream and advance far to the north in the Pacific, Indian and especially in the Atlantic Ocean, where they reach the latitude of Buenos Aires. Single icebergs may be met as far as 24^{o}S. In the summer of 1978, a giant fragment of an Antarctic shelf glacier with a length of 50 km reached the coast of Southern Africa and interfered with navigation.

The largest iceberg on record was encountered in 1966 by the American icebreaker "Glacier". Its volume was near 5000 km^3. It calved from the Ross shelf glacier and slowly drifted to the northeast to Drake Strait where the iceberg was found.

While drift ice moves under the action of winds, icebergs move with ocean streams. One may see an iceberg act like an icebreaker, cutting across ice fields and pack ice.

The nature of icebergs is related to glacier dynamics and morphology. Both iceberg dimensions and rates of relative iceberg production are affected. Various type of icebergs form in Svalbard (Spitsbergen) because its ice masses exhibit considerable variations in their scale and dynamics, and contrasting topographic, oceanographic and sea ice conditions are found on its east and west coasts. A comparison of the length of tidewater ice–mass margins in the Canadian Arctic islands illustrates the importance of the iceberg–producing coastline in Svalbard. Ellesmere and Devon Islands are the only two islands with over 100 and 150 km of ice–cliff coast. Tide–water ice masses fringe significant parts of the major islands within the Svalbard archipelago. Icebergs are calved from the tidewater margins of Svalbard glaciers and ice caps, which extend for over 1000 km around the archipelago (see Table 4).

Table 4. The length of ice cliffs providing a source for icebergs on the main islands of Svalbard (after Dowdeswell, 1989)

Iceberg parent glaciers include tide–water valley glaciers, which generally flow within steep sided fjords, and larger ice cap and outlet glaciers, which terminate in long sections of vertical ice cliffs that are more open to the ocean. In addition, ice masses of both types have been observed to undergo periodic surges and the dynamic difference between surging and either quiescent or non–surging Svalbard glaciers requires a third category.

Surge-type glaciers provide the clearest demonstration of the way in which ice–mass dynamics affect the nature of iceberg calving. Surges are characterized by velocity increases of several orders of magnitude, usually accompanied by rapid glacier advance and the transfer of mass from an upper, accumulation area to a lower receiving zone. This behavior is a result of enhanced lubrication and high water pressures at the glacier bed, including rapid sliding. It persists for only a short time and is followed by a much longer quiescent period, during which the ice–mass terminus stagnates and retreats, while snow accumulation provides mass input and renewed build–up in the upper basin. A large number of surges are punctuated by longer quiescent periods, and many of these glaciers end in tidewater.

Valley glaciers ending in tidewater are most common in the fjords of western Spitsbergen. The termini of these glaciers can be subdivided into two major types. First, there are relatively active glaciers in which heavy crevassing is present. Sometimes icebergs calve from ice caps and large outlet glaciers. The calving of tabular icebergs also produces bergy bits (less than 10 m across) in abundance, and icebergs of irregular shape also result from the collapse of smaller lengths of ice cliff (Dowdeswell, 1989).

Among the principal influences of Antarctic ice sheet stability are rifting and tabular iceberg calving along the seaward margins of the ice shelves. These processes determine the location of an ice shelf's seaward terminus which, in turn, influences the extent of ocean/atmosphere heat exchange surrounding Antarctica, and the degree to which inland ice is buttressed by the ice shelf. Many of the large rifts that appear to form the boundaries where tabular icebergs may eventually detach from the ice shelf are filled with a melange of sea ice, ice–shelf debris and wind–blown snow. This melange tends to deform coherently in response to the ice–shelf flow and has sufficient strength to trap large tabular ice-shelf fragments for several decades before the fragments eventually become icebergs. In many instances, the motion of the tabular fragments is a rigid body rotation about a vertical axis that is driven by velocity shear within the melange. The mechanical role of the rift filling melange may be to bind tabular ice fragments to the main ice shelf before they calve. This suggests two possible mechanisms by which climate could influence tabular iceberg calving. First, spatial gradients in the oceanic and atmospheric temperature may determine where the melange melts, and, thus, the location of the iceberg–calving margin. Second, melting or weakening of ice melange as a consequence of climate change could trigger a sudden or widespread release of tabular icebergs and lead to rapid ice shelf disintegration.

There is evidence in deep-sea cores from the North Atlantic for the occurrence of short-lived episodes of sudden intensified iceberg discharge into the North Atlantic, known as Heinrich (H) events, during the last glacial period. During these episodes, very large numbers of icebergs are believed to have drifted from the Labrador Sea into the North Atlantic and than eastward along a 40-50^{o}N latitude band. H events also appear to take place at the ends of cooling trends of the surface ocean and they are accompanied by a contemporaneous change in deep-water formation in the North Atlantic. Although they clearly relate to abrupt climate changes, the triggering mechanism of H events remains elusive. Insolation changes, sea-level rise and global temperature changes are general mechanisms proposed to explain their occurrence. In contrast, it was proposed that basal melting due to trapping of geothermal heat by part of the Laurentide Ice Sheet located over Hudson Bay was responsible for an ice sheet surge associated with H events. Heat accumulated at the glacial bed thaws the sub-glacial till that, once water-soaked, acts as a lubricant, triggering a sudden ice surge. The phase of fast ice flow finishes when the geothermal and frictional heat are dissipated as the ice thins. There was agreement that the events are controlled by the factors that regulate basal melting, but proposed instead that internal friction in the ice and downward advection of atmospheric temperature conditions to the base of the ice sheet rather than geothermal heating are critical factors. Under these circumstances a change in atmosphere conditions could have been the driving force of past H events.

Sediment cores from the continental rise west of the Antarctic Peninsula and the northern Weddell and Scotia Seas were investigated for their ice-rafted debris content by lithofacies logging and counting of particles >0.2 cm from core x-radiographs. Cofaigh et al. (2001) investigated whether there are iceberg-rafted units similar to the Heinrich layers of the North Atlantic that might record periodic, widespread catastrophic collapse of basins within the Antarctic Ice Sheet during the Quaternary. Cores from the Antarctic Peninsula margin contain prominent ice-rafted debris-rich units, with maximum ice-rafted debris concentrations in oxygen isotope stages 1, 5, and 7. However, the greater concentration of ice-rafted debris in interglacial stages is the result of low sedimentation rates and current winnowing, rather than regional-scale episodes of increased iceberg rafting. This is also supported by markedly lower mass accumulation rates during interglacial periods versus glacial periods. Furthermore, thinner ice-rafted debris layers within isotope stages 2-4 and 6 cannot be correlated between individual cores along the margin. This implies that the ice sheet over the Antarctic Peninsula did not undergo widespread catastrophic collapse along its western margin during the Late Quaternary (isotope stages 1-7). Sediment cores from the Weddell and Scotia Seas are characterized by low ice-rafted debris concentrations throughout, and the ice-rafted debris signal generally appears to be of limited regional significance with few strong peaks that can be correlated between cores. Tentatively, this argues against pervasive, rapid ice-sheet collapse around the Weddell embayment over the last few glacial cycles.

10. Iceberg interaction with sea ice

The pattern of sea–ice distribution in space and time has several implications for iceberg movement. The presence of fast ice may have two main effects. First, it is one of several processes restricting winter iceberg calving. Secondly, it curtails the movement of any icebergs that are calved during the winter months. The presence of winter fast ice protects the ice cliffs at the margins of tidewater ice masses. Undercutting and impact stress from wave action is effectively suppressed. The winter calving rate of such glaciers is therefore likely to be significantly reduced relative to summer iceberg production, particularly when the effects of reduction of ice surface velocity during winter and the likelihood of melt water freezing in crevasses during winter are also considered. Winter observations at the tidewater margin indicated a lack of calved ice blocks, but the presence of buckled and over-thrust sea ice suggested a winter re–advance of the glacier terminus. Icebergs calved in winter have been observed nonetheless, punctuating the relatively smooth fast ice surface immediately beyond the tidewater termini of a number of Spitsbergen ice masses. These icebergs are held in this location until break-up in spring. Once summer calving or release from winter fast ice has occurred, icebergs are susceptible to overturning and fragmentation when exposed to wave and wind forces. The presence of ice floes acts to damp wave energy, and any icebergs moving within a pack–ice field will experience considerable protection from wave perturbation. The degree of protection increases with distance into the pack ice from the sea–ice margin and with increasing floe density. Thus, icebergs in pack ice will be less likely to experience overturning stimulated by wave forces than icebergs of comparable dimensions in the open ocean. Sedimentation from icebergs into the marine environment may increase during glacier surges due to the high rate of production of small icebergs and a possible increase in their sediment content. Large, tabular icebergs will transport included debris the farthest distances from parent tidewater ice masses.

The difference between the mean distribution of icebergs in spring and summer is slight. The primary difference is that there are more icebergs in the summer season, but they are still distributed over the same region. So, iceberg density distribution starts with low iceberg density in early spring, increases to high densities in late spring and early summer and then tapers off again in late summer.

11. Ice Shelves

Large slabs of floating ice, called ice shelves surround much of the Antarctic continent. They are nourished by flow from the inland ice sheet and by snow accumulation. Calving of icebergs and melting at the base are the normal forms of ablation. Basal accumulation by freezing of seawater may, however, also contribute near landward margins. The ice spreads under its own weight as it moves out to sea. Most ice shelves are confined in bays, the shores of which produce a drag on the moving ice. Movement is also restricted by islands and grounding on shoals. Where an ice shelf is grounded over an appreciable area, its surface is dome–shaped. Such a region is called an ice–rise. It has its own radial flow pattern independent of flow in the ice shelf. Its margins are marked by extensive crevasses formed by shear as the shelf ice flows around the ice rise.

The Ross Ice Shelf is the largest with an area of 525 x 1000 km^2. Its thickness varies from over 1000 m where the ice starts to float to about 250 m at the seaward margin, which consists of an ice cliff about 30 m high. (It is limited in its height by the plastic properties of the ice). The accumulation rate ranges from about 0.07 to 0.25 m per year water equivalent. The bottom melting rate is difficult to measure. It is believed to approach 1 m per year at the ice front and decrease to zero at about 150 km from it. Typical velocities are a few hundred meters per year increasing to about 1 km per year at the ice front. The thickness gradient is inversely proportional to the width of the ice shelf. Ice rises and other areas where the ice is grounded provide an upstream retaining force in addition to the shear along the sides and the seawater pressure at the ice front. This can have an appreciable effect on the profile of an ice shelf. Moreover, shoals that ground the outer margin are important to the stability of ice shelves whether or not they are confined in bays. Ice shelves are unlikely to extend beyond the outermost shoals surrounding Antarctica. Real ice shelves are more complex than their models. Thickness and velocity do not vary uniformly over the Ross Ice Shelf. Valley glaciers draining ice through the Transantarctic Mountains persist as streams of thick, fast moving ice for some distance into the shelf. Most of the ice from Byrd Land, on the opposite side of the shelf, drains through a few fast–moving ice streams that occupy bedrock channels. These streams also persist in the shelf. Such ice streams in the ice shelf are retarded by the drag of the surrounding ice. They decrease rapidly in thickness towards their margins and, as they more further into the shelf, the shear and the transverse variations in thickness and velocity are reduced. An ice shelf will grow thicker if snowfall increases or if the bottom–melting rate is reduced. It will also grow thicker if its spreading rate is reduced by emergence of new areas of grounding. This could result from a fall in sea level or from rebound of the seabed as a delayed response to the removal of ground ice at the end of the last glaciation. Moreover, the ice shelf must be considered in relation to the ice sheet that feeds it.

Increased flow from the ice sheet will make the ice shelf grow thicker. Thickening of the ice shelf will change the position of the grounding line, the junction between the ice sheet and the ice shelf. This, in turn, will affect flow in the ice sheet.

12. Stability of the Antarctic Ice Sheet

Glaciers and ice sheets are normally regarded as relatively stable features, advancing or retreating slowly in response to climatic changes. However, glacier surges show that this is not always the case. It is suggested that the Antarctic Ice Sheet may have inherent instabilities that could cause appreciable parts of it to disintegrate. It has been proposed that the surges in Antarctica are the cause of the ice ages. Such a surge would greatly increase the ice cover of the Southern Ocean. This would increase the amount of solar radiation reflected back to space and cool the atmosphere sufficiently to start the growth of ice sheets in the Northern Hemisphere. Irrespective of whether or not it could cause an ice age, such a surge would certainly change the climate and the rise in sea level resulting from the surge would submerge all the world's ports. The rapid decay of the ice sheets at the end of the last glaciation has also been attributed to instability.

The Transantarctic Mountains divide the Antarctic Ice Sheet into two parts that must be considered separately as regards stability because they have different characteristics. Most of the base of the West Antarctic Ice Sheet lies well below sea level and would remain so after rebound following removal of all ice. An ice sheet in this situation is called a marine ice sheet. Most of the ice in West Antarctica drains to the sea through the Ross and Ronne Ice Shelves. The East Antarctic Ice Sheet, which contains over 80% of all the ice, is a broad dome with a maximum elevation of over 4000 m. Except for two relatively small areas, the underlying bedrock is above sea level. Melting of the East Antarctic Ice Shelf would raise the world sea level by about 60 m. The value for West Antarctica is about 5 m, not only because the ice sheet is smaller but also because much of it is at present displacing ocean water.

Whether an ice sheet – ice shelf system is stable depends on the slope of the seabed and on whether or not the ice shelf is confined in a bay or pinned by ice rises. If the seabed slopes down in a seaward direction, a small change in ice thickness will cause the grounding line to advance or retreat to a new stable position. The response is different if the seabed slopes down towards the center of the ice sheet. Any increase in thickness will cause the ice to advance to the margin of the continental shelf before it starts to float, whereas a decrease in thickness will cause retreat until the whole ice sheet has disintegrated into a floating ice shelf.

13. Ice Cores Studies

One of the major developments in glaciology has been the wealth of information obtained from analyses of cores from polar ice sheets. Ice cores gives access to palaeoclimate series that includes local temperature and precipitation rate, moisture source conditions, wind strength and aerosol fluxes of marine, volcanic, terrestrial, cosmogenic and anthropogenic origin. They are also unique, with their entrapped air inclusions, in providing direct records of past changes of atmospheric trace-gas composition.

The GRIP (Greenland Ice Core Project) core is a deep high resolution core from the Summit in Central Greenland (72^{o}N, 37^{o}W). The complete core is 3028 m long and extends back 250,000 yr. BP. The GISP2 (Greenland Ice Sheet Project 2) site (72^{o}N, 38^{o}W), 28 km west of the Summit site, was completed shortly after drilling at the GRIP site. The complete GISP2 core is 3053 m long and it also extends back into the penultimate glacial period.

The GRIP chronology is constructed by counting annual layers back to 14,500 yr. BP and by using an ice flow model for the earlier part of the record.

Comparative studies of the GRIP and GISP2 cores demonstrate that 90% of the records are in good agreement, but that major discrepancies occur on the lower 10%. From 87,000 yr. to the present (i.e., from 2700 m to the surface in the GISP2 core and from 2670 m depth to the surface in the GRIP core) the ECM records are well correlated, both for major events and for abrupt low amplitude changes spanning a few years. This increases confidence in the climate inferences for the Holocene and most of the glacial period made from the top 90% of both cores. The lower 10% of both cores contains the records where many events cannot be correlated as shown by $\delta^{18}O$ data and the ECM record (Dansgaard et al., 1993). The extent of deformation in the lower parts of the GISP2 and GRIP cores is not known at present. The GRIP is located at the present day ice divide where there is no shear flow to generate ice folds similar to those which appear to have deformed the GISP2 core. The GISP2 site is considered likely to suffer from shear flow at the present day. This would explain the discrepancies between the two records. However, it is equally possible that, due to climate change , the ice divide has changed position, meaning that the GRIP core could also have been subject to shear flow and ice folding at sometime in the past. It has been suggested that there is the likelihood of shifts from divide to flank flow in the Summit region. Marshall and Caffey (2000) applied a three dimensional thermo-mechanical ice sheet model to examine the evolution of the ice divide location and divide residence time over the last 160 ka and find that the shifts correspond to glacial interglacial geometric modes, with the most substantial dynamical variations in the divide region occurring during transitions between glacial and interglacial states. The Greenland Summit position has moved significantly through time in the last 100 ka, but never (excepting the modern summit) enough to cause a significant increase in shear strain rates along flow paths leading to present ice core sites. The summit position was likely quite stable through the last 60 ka of the glacial climate, but then shifted abruptly during deglaciation Changes in flow direction accompanying the summit shift, and the shift itself, are plausible causes for the absence of a Raymond bump beneath the Summit, and for the stratigraphic disruption of Eemian ice in the GRIP and GISP2 cores (Marshall, Caffey, 2000). Comparison with high-resolution Antarctic ice cores may help to resolve some uncertainties.

The Vostok ice-drilling project undertaken in the framework of a long-term collaboration between Russia, the USA and France at the Russian Vostok station in the East Antarctic (78^{o}S, 106^{o}E, elevation 3488 m, mean temperature –55^{o}C) has already provided a wealth of such information for the past two glacial–interglacial cycles. Much colder temperatures and reduced precipitation characterize glacial periods in Antarctica along with more vigorous large-scale atmospheric circulation. There is a close correlation between Antarctic temperature and atmospheric concentrations of CO_2 and CH_4. This discovery suggests that greenhouse gases are important as amplifiers of the initial orbital forcing and may have significantly contributed to the glacial-interglacial changes. The Vostok ice cores were also used to infer an empirical estimate of the sensitivity of global climate to future anthropogenic increases of greenhouse gas concentrations.

The recent completion of ice core drilling at Vostok has allowed researchers to considerably extend the ice core records of climate properties at this site. In January 1998, the Vostok project yielded the deepest ice core ever recovered, reaching a depth 3623 m. Drilling than stopped approximately 120 m above the surface of Vostok Lake, a deep sub-glacial lake which extends below the ice sheet over a large area, in order to avoid any risk that drilling fluids would contaminate the lake water. The Vostok ice-core record extended through four climate cycles, with ice slightly older than 400,000 yr. BP at the depth of 3310 m, thus spanning a period comparable to that covered by numerous oceanic records.

The Vostok data include deuterium content of the ice (δD_{ice}, proxy of local temperature changes), the dust content (desert aerosol), the concentration of sodium (marine aerosol), and from entrapped air the greenhouse gases CO_2 and CH_4, and $\delta^{18}O$ of O_2 (hereafter $\delta^{18}O_{atm}$). The detailed record of δD_{ice} confirms the main features of the third and fourth cycles. Sudden decrease from interglacial-like to glacial-like values rapidly followed by an abrupt return to interglacial-like values occurs between 3320 and 3330 m. A transition from low to high CO_2 and CH_4 values occurs at exactly the same depth. Also, three volcanic ash layers just a few centimeters apart but inclined in opposite directions, have been observed, 10 m above this δD excursion. Similar inclined layers were observed in the deepest parts of the GRIP and GISP2 ice cores from Central Greenland, where they are believed to be associated with ice flow disturbances. Vostok climatic records are probably disturbed below these ash layers, whereas none of the six records show any indications of disturbance above this level.

14. Lake, river and sea ice

At present, the total mass of natural ice is as much as $2.456\ 10^{22}$ kg, and the total area of consolidated ice is as much as $72.4\ 10^{6}\ km^{2}$ (seasonal oscillations are $53.6\ 10^{6}$ to $91.2\ 10^{6}\ km^{2}$). This is 14.2% of the Earth's surface or seasonally 10.5 to 17.9% (Lisitzin, 1994). If iceberg and open pack ice area are taken into account, the total area is as much as $100\ 10^{6}\ km^{2}$ or 19.6% of the Earth's surface (seasonally oscillations from $81\ 10^{6}\ km^{2}$ to $119\ 10^{6}\ km^{2}$).

According to Shumski classification, there are 28 types of fresh ice, which can be divided into three groups: congelation ice (all ice of fresh water reservoirs and streams), ground ice (sedimentation ice) and metamorphic (glacier) ice.

Floating ice can be distinguished first by its origin into marine, river, lacustrine and land, and second by its dynamics into immobile (fast ice) and mobile (drift or pack ice). Number scales are used for the evaluation of pack ice cover. Pack ice is divided according to dimension and shape as follows: an ice field is more than 5000 m wide, floe ice is more than 1000 m, small floe is more than 200 m, cake ice is smaller than small floe, bit ice is less than 5 m and brush ice is less than 2 m.

Some varieties of ice form on the surfaces of lakes. White ice—often called snow ice—usually forms on the ice sheet when sufficient snow accumulates on the surface to depress it below water level. Water seeping through cracks forms slush, which rapidly freezes into ice that is milky white in section in contrast to normal black ice, which grows relatively slowly by freezing at the lower ice/water interface.

The structure of Antarctic sea ice is quite dissimilar from that of Arctic ice, with Antarctic ice being predominantly granular and Arctic ice being predominantly columnar. Compared to columnar ice, granular ice is typically more disorganized and thus has a larger density of ice/brine interfaces. This can result in enhanced scattering and higher albedo.

There are large differences in the properties of pure ice single crystals and sea ice single crystals. It has been long recognized that polycrystalline pure ice and sea ice have different mechanical properties. This has been attributed to such factors as porosity differences and brine pockets. However, it has been shown that sea ice crystals may have low porosity and density approximately equal to that of pure ice. The sea ice was also thoroughly drained before testing, so the amount of liquid brine was as low as could be obtained by draining. The salinity of all sea ice was between 7.0 and 9.0, which is somewhat higher than frequently found in actual polycrystalline sea ice. Microstructures in the thin sections of the sea ice single crystals show the classic substructure of sea ice. The ice had a cellular or platelet substructure with a fairly uniform cell thickness of 0.5 to 0.8 mm in which the cells are separated by brine/salt layers. These brine layers separating the pure ice cells or plates lie parallel to the basal plane and hence should add a fluid or visco–elastic deformation to the dislocation glide on the basal plane. The additional result is viscous–like flow of matter from the brine channels onto lateral surfaces of the ice when a critical stress is reached. This occurred only when the c–axis was perpendicular to the load direction so that the maximum principal stress exceeded 2.0 Mpa. This was not observed where the c–axis was oriented at 45^o, 60^o, or 66^o, regardless of the strain rate. The morphologies observed during the solidification of seawater and other aqueous systems have been found to be consistent with those observed in metallic systems and were excluded, when taking into consideration the general understanding of growth from the melt. Further, it was predicted that a fraction of second phase, amounting to an appreciable volume, form a network during the normal solidification of sea ice. Since in certain temperature ranges this network becomes liquid, deteriorating the strength of the material, a challenge for controlled solidification would be dispersal of this second phase as unconnected regions in the relatively pure ice matrix.

Understanding radiative transfer in sea ice is considerably complicated by the large spatial and temporal variability in the physical, and therefore the optical properties of the medium. Over distances of tens of meters, ice thickness can range from open water to thick pressure rides, while surface conditions can include bare ice, ponded ice and snow–covered ice. On a smaller scale, sea ice forms an intricate structure consisting of ice platelets, brine pockets, brine channels and air bubbles, all of which can vary in size and orientation. In addition, when sea ice is always at or near its salinity–determined freezing point, small changes in ice properties are observed. For example, as sea ice warms the brine volume increases and internal platelet boundaries become smoother, brine pockets enlarge and connect and air bubbles in the ice are released and coalesce into larger bubbles. All changes act to reduce light scattering within the ice, causing a decrease in the albedo. The most pronounced temporal change in the optical properties occurs during the seasonal transition from cold winter conditions to the warm summer melt period. Another seasonal transition occurs when summer ends and winter returns. As the air temperature drops, the ice cools internally, causing a reduction in the brine volume, while melt ponds freeze and new ice forms in areas of open water and the ice become snow–covered.

Large accumulations of very close pack ice, called ice massifs, have been revealed in the ice-covered water areas of the World Ocean. Thicker and more compacted ice massifs present a troublesome obstruction for navigation and have caused the loss of many vessels. The ridges of the ice massifs of the Arctic basin are distinguished from the local massifs. The main peculiarity of the inter-annual and seasonal variability of the areas and location of massifs is explained mostly by the differing influences of anemobaric conditions.

A detailed profile in a 5.54 m multi-year sea ice core from the rift area in the southern part of the George VI ice shelf has been studied by Tison et al. (1991) in order to discriminate different parent water sources for the ice growth. The core consists mainly of granular ice, which is essentially frazil ice. The core presents a wide range of salinity and isotopic values, with Na ranging from 0.00019 to 1.6 (global salinity ranges from 0.00062 to 5.2), δD from –162.5 to –8.2, $\delta^{18}O$ from –20.8 to 1.15. These ranges are comparable to those observed for salinity and $\delta^{18}O$, in land fast multi-year sea ice from the Arctic. Half of the core consists of sea ice formed in the winter. Although during this period the parent water is close to normal sea water, the Na and $\delta^{18}O$ curves in the core show a few sympathetic large-scale fluctuations indicating dilution events reaching 20% of fresh water input that might reflect small variations in the basal melting of the ice shelf. A sharp transition from sea ice to brackish ice (at 450 cm depth), which must result from mixing with melt water from basal shelf ice, indicates the importance of this process during the summer period, when maximum dilution coefficients reach up to 80% fresh water. Winter accretion consists almost exclusively of frazil-ice production, while summer accretion alternates between frazil-ice formed at depth close to the freshwater input location (low salinity and low-isotopic signal) and congelation ice formed by direct progression of a freezing front under the sea-ice cover into a less diluted parent water. Instability of thermal fluxes through the ice cover leads to variability of vertical temperature distribution. There are three main types of temperature vs. depth curves. The first type has a positive gradient and is common to early winter. The second type is characterized by minimum temperatures in ice thickness and is typical to spring. The third type is intermediate between the first and the third and may occur through the winter.

In North America the term "icing" is often replaced by a variety of terms. Aufeis (German), flood ice, flood-plain icing, ice field, naled (Russian) and overflow ice usually indicate icings formed on river ice and floodplains. Chrystocrene (or crystocrene), ground icing, groundwater icing and spring icing usually indicate icings formed by freezing of ground-water discharge.

Aufeis systems are divided into systems of underground water aufeis and those of river water aufeis. Despite the discrete nature of aufeis phenomena, each of them is characterized by typical features and relatively stable spatial and temporal regularities. Ignoring the peculiarities in the development of aufeis, the constituent systems of elements can be placed within the limits of the regularities with a 30% accuracy. The formation of underground water aufeis is mainly determined by the properties of geological cryogenic systems, while river aufeis is greatly influenced by climate and weather. This is why the dynamics of the former are more stable over time and their spreading over territory is generally independent of the environment. Inverse regularity is typical of river aufeis systems.

15. Ground ice

Ground ice is a general term used to refer to all types of ice formed in freezing or frozen ground. Ground ice occurs in the pores, cavities, voids, or other openings in soil or rock and includes massive ice. Buried glaciers, lakes, rivers and snowbank ice are considered as one type of ground ice. Traditionally permafrost is defined on the basis of temperature: i.e. soil or rock that remains at or below 0 ^{o}C for at least two consecutive years. However, permafrost may not necessarily be frozen since the freezing point of included water may be depressed several degrees below 0 ^{o}C.

The ice content in permafrost is probably the most important feature relevant to human life in the North. Ice in perennially frozen ground exists in various sizes and shapes, with definite distribution characteristics grouped into five main types: pore ice, segregated or Taber ice, foliated ice or ice wedges, pingo ice and buried ice. About 10% by volume of the upper 3.5 m of permafrost of the Alaskan coastal plain is composed of ice wedges. Taber ice is the most extensive type—in places representing 75% of the ground ice by volume. The pore and Taber ice content in the depth between 0.5 and 3.5 m (surface to 0.5 m is seasonally thawed) is 61%, between 3.5 and 8.5 m is 41% by volume. The total amount of pingo ice is less than 0.1% of the permafrost. The total amount of perennial ice in the permafrost of the Arctic Coastal plain of Alaska is estimated to be 1500 km^3 and below 8.5 m, most of that is present as pore ice.

Two quantitative parameters are used to describe ground ice conditions. First, the ice content of a soil is defined as the weight of ice to dry soil, and is expressed as a percentage. Low ice content soils are generally regarded as those having ice contents less than 40 to 50%. Soils with high ice contents are usually fine grained and have ice content values that commonly range between 50 and 150%. The amount of "excess ice" is a second parameter commonly used in the description of ground ice. Excess of ice refers to the volume of supernatant water present if a vertical column of frozen sediment were thawed. In this case, the sample is allowed to thaw and relative volumes of supersaturated sediments and standing water (i.e. excess ice) are noted. The volume of supernatant water is expressed as a percentage of the total volume of sediment and water. Sediments that contain excess ice are often referred as "ice rich" or "icy" sediments. Frozen sediments containing excess ice are thaw sensitive and may be contrasted with thaw stable materials. The latter contain no excess ice and are not subject to thaw settlement and retain much of their mechanical strength when thawed.

A commonly used classification developed by Mackay (1990) is based upon (a) the source of water immediately prior to freezing, and (b) the principal transfer process which moves water to the freezing plane. Various types of buried ice (glacier ice, sea, lake or river ice and snow banks) are from this classification, which results in ten mutually exclusive ground ice forms. The term massive ice or massive ice bodies are usually reserved for relatively pure ground ice, the ice content of which averages at least 250% for a thickness of several meters.

In addition to its clarity, an advantage of this classification is that it emphasizes the variety and complexity of transfer process, recognizing at least six basic mechanisms.

Moreover, although the theoretical principles behind these transfer processes are beyond the scope of this article, the classification has the added advantage that it focuses attention upon three broad types of ground ice: wedge ice, segregated ice and intrusive ice. These three types of ground ice are particularly important from a geomorphologic viewpoint since their localized occurrence is distinctive of periglacial landforms and terrain, such as ice-wedge polygons, frost mounds and various types of ice-cored topography. In addition, pore ice and segregated ice are the major determinants of cryostructures and frost heave. Ice within perennially frozen sediments imparts structures distinct from those found in other sedimentary environments. These are termed cryostructures. They are determined by the amount and distribution of ice within pores (often termed pore and cement ice), and by lenses of segregated ice. Cryostructures must be distinguished from cryotextures. The latter refers to the grains and/or ice crystal size and shape, and nature of the contacts between grains and ice crystals in frozen earth materials. Both cryotextures and cryostructures are useful in determining the nature of the freezing process and the conditions under which the sediments have accumulated.

The more important types of ice and their characteristics require a brief description. Pore ice is the bonding cement that holds soil grains together. The distinction between pore ice and segregated ice is related to the water content of the soil. It is best determined by thawing the soil and noting the presence or absence of excess ice or supernatant water. If supernatant water is present, this indicates that the frozen soil was supersaturated and that segregated ice was present.

Segregated ice is a broad term for soil with a high ice content. Usually, the ice lenses are visible to the naked eye, but in certain soils, particularly those that are fine-grained, lensing may be minimal. Segregated ice lenses vary in thickness from layers a few centimeters thick to massive ice bodies, sometimes tens of metres thick. In theory, segregated ice may be distinguished from intrusive ice by the relative purity of the latter and the stratification and presence of soil particles and air bubbles oriented normal to the freezing plane in the former. However, when dealing with massive icy bodies, this distinction is often difficult to make.

In general terms ground ice may be either "epigenetic" (i.e. develops inside the enclosing rock and after the latter has formed), or "syngenetic" (i.e. forms at, or almost at, the same time as the enclosing sediments are deposited and usually associated, therefore, with surface aggradation). The distinction between epigenetic and syngenetic becomes especially important when understanding ice wedges.

Vein ice is formed by the penetration of water into open fissures developed at the ground surface. In contrast to segregated ice, therefore, the origin of the water prior to its solidifying is of a surface nature, usually snowmelt water and summer rain. Vein ice can be distinguished from segregated ice by its vertical foliation and structures. Single vein ice develops in small fissures usually formed in the upper 60 cm of permafrost, probably by thermal contraction. The majority is very thin, less than 0.2 cm in thickness. In silty/clayey material, reticulate ice veins are common. Repeated vein ice is a specific variety of single vein ices and is the result of successive ice formation in frost fissures periodically forming in the same place for many years. The result is the formation of vertical or near vertical sheets of ice, which because of their form are termed ice wedges. These are important features of permafrost environments.

Intrusive ice is formed by water intrusions, usually under pressure, into seasonally or perennially frozen zones. Sill and pingo ices are the two types of intrusive ice that are usually identified. The former grows when water is intruded into confining material and freezes in a tabular mass along the base of the active layer and parallel to the permafrost surface. In many instances, small frost mounds can result.

The amount of ground ice present within permafrost can vary from negligible, as in certain igneous and metamorphic rocks, to considerable, as in the case of unconsolidated, fine-grained Quaternary–age rocks. The total volumetric ice content varies between 35 and 60%, of which by far the majority (66–80%) is either segregated or pore ice. Clearly, in such regions ground ice is an extremely important component of permafrost. Elsewhere, typical ground ice content varies from 1–2% to 10–15%.

16. Ice wedges

One of the most common cryostructures observed in silty and /or clayey permafrost is the reticulate ice vein network. The most conspicuous and controversial types of ground ice in permafrost are large ice wedges or masses characterized by parallel or sub–parallel foliation structures. Most foliated ice masses occur as wedge-shaped, vertical, or inclined sheets or dikes 1 cm to 3 m wide and 1 to 10 m high when seen in transverse cross-section. Some masses seen on the faces of frozen cliffs may appear as horizontal bodies a few centimeters to 3 m in thickness and 0.5 to 15 m in length. The true shape of these ice wedges can be seen only in three dimensions. Ice wedges are parts of a polygonal network of ice enclosing cells of frozen ground 3 to 30 m or more in diameter. They reflect a polygonal micro-relief pattern on the surface of the ground that is a characteristic feature of permafrost terrain. Snow, hoar frost or freezing water partly fill winter contraction cracks outlining polygons, commonly 5-20 m in diameter, or on the surface of the ground. Moisture comes from the atmosphere. Increments of ice 0.1 to 0.2 mm are added annually to the wedges, which squeeze enclosing permafrost.

Late Pleistocene syngenetic ice wedges are widespread in the arctic coastal plains, river valleys and intermontane depressions of Northern Eurasia. The Arctic coastal plains occur in the north of Western Siberia, Yakutia and Chukotka. They include both Holocene and Pleistocene river valleys and deltas. The plains are characterized by lakes, lagoons, deltas and alluvial fans. These vast areas have not been glaciated during the last 40 000 years.

Intermontane depressions are located at a height of 500 m above sea level in interfluves of Yana and Omoloj (North Yakutia), in northeastern Chukotka (Lake Koolen') in the Upper Kolyma River and in the Chara River valley (Trans-Baikal Region).

Parts of the mountainous areas were glaciated in the Late Pleistocene by glaciers that did not cover the whole area of the depressions. Consequently, syngenetic permafrost sediments with large ice wedges accumulated here.

Northern Eurasia comprises two permafrost zones: a) continuous in the north and b) discontinuous in the south. In the continuous zone, permafrost exists everywhere beneath the land surface and varies in thickness from about 100 m at the southern limit to over 350–500 m in Yakutia and Northern Siberia and over 1000 m in the Anabar Plateau. Permafrost thickness in the discontinuous zone varies from a few meters at the southern limit to about 100 m at the boundary with the continuous zone.

The distribution of permafrost is broadly related to air temperature. The mean annual ground temperature of permafrost varies between slightly below 0 ^{o}C (in the discontinuous zone) and –10 to –13 ^{o}C (in northern areas of the continuous zone). The boundary between the continuous and discontinuous permafrost zones corresponds to a mean annual air temperature of –5.5 ^{o}C in Eastern Europe (below the –3000 to –3100 degree-days for total winter temperatures and below the +600 to +800^{o} degree-days for total summer temperatures), –7.0 ^{o}C in Western Siberia (below the 3500 to –3700 degree-days for total winter temperatures and below the +1000 to +1200 degree-days for total summer temperatures), –7.5 ^{o}C in Yakutia (below the –4200 to –4400 degree-days for total winter temperatures and below the +1300 to +1500 degree-days for total summer temperatures), and–3.5 ^{o}C in Chukotka (below the –2500 degree-days for total winter temperatures and below the +600 to +800 degree-days for total summer temperatures).

Very large syngenetic ice wedges are widespread in the Northern Eurasia permafrost zone. There are several well known exposures in different regions, as follows: Seyaha exposure (the height of the ice wedge system is more than 30 m) in the north of Western Siberia; near the Zelyony Mys settlement (height about 40 m), Duvanny (height about 45 m), Vorontsovski (height about 50 m), Stanchikovsky (height about 35 m) and Oiyagossky Yar's (height about 40 m) in Yakutia. In North America similar thick ice wedges were found in the valley of the Titaluk River (height about 30 m) in northern Alaska. The width of large syngenetic ice wedges varies from 1.5 to 3.0 m.

Syngenetic ice wedges are a direct indicator of the existence of a "very cold" permafrost zone, with mean annual ground temperatures not exceeding–2 to –3 ^{o}C. Syngenetic ice wedges are formed in such a way that ice becomes vertically stratified.

Ice wedges are usually divided into two major types: epigenetic and syngenetic, although Mackay (1990) distinguishes a third type, anti–syngenetic.
Epigenetic ice wedges grow in pre–existing permafrost on a stable surface and are younger than the host material. As epigenetic ice wedges grow they increase in width but not in height. Few epigenetic ice wedges exceed a height of 3 to 4 m. In theory, the vertical dimension of an epigenetic fissure is limited by the depth of cracking.
Syngenetic ice wedges may grow both in thickness and height as the upper permafrost surface rises in response to the addition of material on the ground surface. The added material may include alluvium, lacustrine sediments and peat.
Syngenetic icewedges are often vertically nested in chevron patterns. In Northern Eurasia, large syngenetic ice wedges are a dominant form of ice. It is widely thought that syngenetic ice wedges form only under stable conditions of slow, continuous sedimentation accompanied by repeated frost cracking. However, we believe that this situation occurs rarely and that sedimentation during the last 20 to 30 thousand years has been episodic, with big pulses of subaqueous deposition alternating with subaerial conditions of ice wedge growth i.e. a macrocyclic mechanism takes place (see Vasil'chuk and Vasil'chuk, 1996). Syngenetic ice-wedge growth proceeds subaerially during the accumulation of peat or peaty sediments. Periodically, when gravel, sand, sandy loam, loam, silt, and clay are deposited under subaqueous conditions, ice wedge growth decreases or stops. When the subaerial regime returns, ice wedge growth recommences. If the thickness of subaqueous strata is thin enough (<3–4m), the toes of younger and stratigraphically higher ice wedges penetrate into buried ice wedges of the previous phase. Conversely, if the thickness of subaqueous sediment exceeds 4–5 m, the stratigraphically higher ice wedges do not penetrate the lower wedges. This model of syngenetic ice wedge growth is supported by the distribution ofice wedges in higher and lower levels of sediment aggradation. For example, the polygonal network on high floodplains of northern rivers tends to be widespread, whereas that on low floodplains is rare. This confirms that ice wedge growth occurs preferentially in subaerial conditions. When the tail of a new ice vein is incorporated into the underlying ice wedge, a single multistage (cryocyclite) ice wedge forms. The multistage process may be due to climatic changes or to changes in sedimentation regime. Sometimes, buried ice wedges can be plastically uplifted (extruded) because of lateral compression, although the process is not fully understood. As a rule, the uplift of ice wedge ice can vary from 0.5 to about 1.5 m. However, Black reported in 1975 that buried ice wedges in McLeod Point in northern Alaska have flowed upward 1–3 m. The scale of uplift has been observed in the lower part of ice wedge cross-sections at the Tambey River valley. Here the sand extruded both wide and narrow ice wedges and formed an anticline with ice wedges in the central part. The ice wedges were uplifted together with laminated sand, forming overturned (almost to 90 degrees) bedding (Vasil'chuk, 1992).

The macrocyclic model of ice wedge formation in permafrost zones is similar to the cyclical syngenetic ice wedge casts in Late Pleistocene deposits near Grouw in the northern Netherlands (radiocarbon dated to 43 to 35 ka BP). It was proposed that the mechanism of cyclic ice wedge formation is characterized by substantial sedimentation between phases of ice wedge development and by the partial melting of the wedge ice from a previous growth phase. The term 'cyclic syngenetic cast formations' describes a step-like formation of ice wedges by partial melting of the wedge ice from a previous phase on floodplains. By contrast, Vasil'chuk's macrocyclic model has no growth stages and melting of ice wedges, but the almost continuous development of ice wedge systems. It seems that this model describes conditions in southern areas of ice wedge formation, where the icewedge growth stages are more dependent on climatic oscillations and slight warming can result in the melting of ice wedges. Thawing of ice wedgescan also take place in northern areas if surface water is more than 1.5 to 2 m.

It must be emphasized that short cycles which are caused by annual-scale changes of accumulation also occur in syngenetic permafrost sedimentation at a scale of about 1 to 10 cm. These differ significantly from macro-layering (3–5 m thickness), in which minerogenic sediments with peat or peaty sediments reflect long-term changes of sedimentation regime.

17. Water source of ice-wedge ice

The main source of water for syngenetic ice wedges is snowmelt. This water is generally from spring snowmelt. Minor sources include hoarfrost and the melting of active-layer ice (<5–10%).

Within high terraces and divides, the syngenetic ice wedges are formed exclusively of atmospheric water that freezes within the frost cracks (those of the epigenetic type). On flood plains and coastal plains, small ice wedges also form from atmospheric water flowing into the frost cracks (if the crack is open to the surface) or from water from the seasonally thawed layer (in the case of intra-stratal frost cracks). Highly mineralized water may penetrate into cracks when there is a salty lake nearby or as a result of an extremely active tide or a surge. Such tides and storm surges usually occur only in summer when the sea or bay surface is free from ice. By this time, the majority of the frost cracks are already closed and thus, in only a few instances will this water freeze within the ice wedge.

18. Massive ice bodies

Thick, often bedded and sometimes deformed layers of massive ground ice and icy sediments, often tens of meters thick, are the most spectacular of ground ice forms. They are important not only because of their origin and the light this may throw on permafrost histories, but also because of the thaw settlement properties of terrain underlying such material. Massive ground ice and massive icy bodies are known to exist in parts of Western Siberia, Chukotka, Northern Alaska and the Western Canadian Arctic, and in China. Massive ices may be divided into two main groups according to their origin, namely: (1) autochthonous, which is segregated ice that transforms into intrusive ice due to the water injection process, and (2) allochthonous or buried ice, without making a clear distinction being made between glacier ice derived from snow and subglacier regelation ice which, in essence, is segregated ice. Autochthonous ices may be classified as segregated, infiltration-segregated, injection-segregated, and injection. Allochthonous ices may be divided into buried pack and floated ices, bottom ices, glacier ices, and aufeis ices. Natural exposures of massive ground ice are rare and usually short-lived. Data from Canada and Siberia indicate two salient facts about the occurrence of these massive ice bodies. First, in the vast majority of instances where massive ice is encountered, the ice was overlain by clay-grade sediment and underlain by sand-grade sediment. Second, massive ice is usually located at depths in excess of 30 m, sometimes at depths of 100 to 200 m. Any field differentiation between massive segregated and buried glacier ice is complicated by the fact that the two may appear very similar. For that reason, it is necessary to study their biological, chemical, and especially isotopic properties. Most massive ground ice bodies in the Western Arctic occur near the outer limits of known glaciation. This suggests some relationship between glacial history, permafrost evolution, and ground ice formation. The ice bodies are of different ages and of several different origins. Some ice bodies of non-glacial nature have been strongly deformed.

In a massive segregated ice body, the contact between the ice and the overlying/ underlying material should be gradational. In places, soil fragments might be suspended in clear ice as far as 10–20 cm beneath the ice–soil contact. Coversely, buried glacier ice must be older than the overlying sediments, and therefore the upper ice–soil contact of such an ice body must represent an erosional surface (i.e. a thaw unconformity). If the ice had initially been exposed to thaw and radiation melting, one possible sign of this event may be that the bubbles immediately at the ice contact are truncated and filled in with sediment from above. Thus, if glacier ice is buried by unfrozen sediments, the nature of bubbles at the ice-soil contact might indicate such a history.

In a massive segregated ice body, there may be a progressive change in water quality with depth, the trend continuing upwards and downwards into the enclosing sediments. If it were buried glacier ice, no such trends would be discernible and, in fact there might be an abrupt discontinuity between the hydrochemistry of the massive ice and that of any ice found within the overlying and underlying sediments. If freezing occurred in a closed system, isotopic fractionation might also result, as indicated by $\delta^{18}O$ and δD values. There might be a straight-line relationship between these two isotope values, with a regression line slope of approximately 6.0. This is significantly lower than the slope of the meteoric water line (8.0), which is what one might expect if the ice were buried glacier ice (i.e. derived from snow). In a massive segregated ice body, bubble trains will extend downwards from the upper soil–ice contact, indicating that freezing progressed downwards. If sediment bands are present, soil fragments may be matched and an ice coating may be found on the undersides of stones within soil pods or sediment layers. By contrast, buried glacier ice may be highly variable in terms of air bubbles and mineral inclusions. These characteristics will depend among other things upon the position of the ice within the glacier at the time it became "dead" and the nature of the glacier itself. For example, referring specifically to glaciers on Bylot Island, R. Klassen differentiates between "Dirty Ice", "Banded Ice" and "Foliated Glacier Ice". The ice at the base of a glacier may contain significant layers of debris and probably will have experienced numerous thawing and re-freezing episodes, as sub-ice pressures fluctuate. To all intents and purposes buried basal glacier ice may be very similar to massive segregated ice: large quantities of sediments may be present or, alternatively, the ice may be sediment-poor if it represents ice stratigraphically high within the glacier ice body. The three-dimensional nature of the ice body, and its original location within the glacier ice mass at the time of the detachment, assume critical importance in recognizing buried glacier ice.

Numerous ground ice slumps testify to the existence of massive ground ice and icy sediments within the moraine in the Late Wisconsinan lateral moraine complex called the Sandhills Moraine on southwest Banks Island (North-West Territories, Canada). The massive ice is overlain by compact gray-blue pebbly clay and alternating silty clay and fine-to-medium sand. While the contact between the ice and the overlying clay is distinct, the others are lithostratigraphic boundaries. The compact pebbly clay is interpreted as an ablation till resulting from the melt-out of debris within the underlying ice (i.e. the contact is the thaw unconformity). The main body of ice-rich pebbly clay is regarded as a till that is overlain by ice-contact (i.e. outwash) sands. Because the massive ice in Sandhills moraine has an irregular upper contact with overlying glacigenic materials—pebbly clay and sandy-gravelly sediments—the ice is almost certainly not of segregated nature. Petrofabric analysis indicates a strongly preferred orientation to the ice crystals explained by glacial origin (French, Harry, 1990).

Buried glacier ice is also found in Eskimo Lakes (North-West Territories, Canada). The strongest evidence in support of a glacial origin for these ice bodies is again provided by cryostratigraphic observation. The total thickness of the massive ice is approximately 6 m. The ice was overlain by as much as 10 m of cross-stratified, fine, medium and coarse sand containing ripped-up clasts near the base. Small bubbles within the ice were truncated at the junction with the overlying soil indicating that contact had been exposed to thaw and radiation melting.

Some tabular ice bodies such as those described by French (1996) from the Klondike District, Yukon Territory, and from Northern Alaska, are clearly of a non-glacial origin, since they occur in never-glaciated terrain. Other ice bodies, which are frequently exposed in the Mackenzie Delta coastal lowlands, are best explained in terms of segregation–intrusive origin. These ice bodies occur within the limits of mapped Early Wisconsinan Toker point and Franklin Bay stades. At Peninsula Point near Tuktoyaktuk, the presence of ice dykes in association with massive ice suggests that this ice body is a massive tabular body of segregation–injection ice (French, 1996). The ice dykes are cryogenically analogous to the peripheral dykes commonly with the laccoliths and other intrusive bodies. The horizontal banding within the ice and the clear evidence of under-loading structures are further evidence in favor of a segregation origin at some depth. At Nicholson Points, approximately 100 km northeast of Tuktoyaktuk, a 6-8 m thick ice body of highly deformed and stratified ice has been examined. The ice rests conformably beneath and upon silty sandy sediments of a non-glacial nature. Such a stratigraphic position argues strongly for a segregation origin with subsequent deformation either by glacial-tectonics or by creep/unloading.

Massive ice bodies of differing origins are found in Northern Eurasia. In the North of Western Siberia, massive ice bodies are found at heights +75 to–150 m ASL. Large massive tabular bodies of segregation–injection ice are described in the Yamal and Gydan Peninsulas and in Chukotka. Most of them are located in the upper 50 m, in contact with sandy and clay sediments. Sand underlies the massive ice body and clay or loam overlies it. Within the Yamal Peninsula, segregation–injection ice bodies were found in the Bovanenkovo area. The total thickness of this massive ice is approximately 5-18 m, the area is 4 380 000 m^2 and the ice volume is 21 900 000 to 78 560 000 m^3. Surrounding sediments are marine sand, loam and sandy loam. The ice is stratified with horizontal bands of icy ground, and rests conformably beneath and upon silty sandy sediments. The existence of the ice dykes supports a segregation–intrusive origin. On the west coast of Neito Lake, at a depth of 5 to 20 m, a very large massive ice body has been found. Its area is 20 200 000 m^2 and the ice volume here is 400 000 000 m^3. Large lakes were formed after thawing similar massive ice bodies in Yamal and Gydan Peninsulas in the Early Holocene.

A massive ice ground body more than 20 m thick has been found in the continuous permafrost of northeastern China at Huola Basin ($52^{o}57'$ to $53^{o}03'$N, $121^{o}03'$ to $122^{o}04'$E). Permafrost varies between 98 and 28 m in thickness. The massive ice lies in mudstone at a depth of 46 to 68 m below the ground surface. The base of the massive ice coincides with the base of the permafrost. The ice is relatively pure, containing only small quantities of fine mineral material, apparently suspended within the ice. Small bubbles were unusually elongated in form and arranged in near vertical lines. The ice is of the sodium–calcium bicarbonate–chlorate type, and the degree of mineralization is 0.05 g/l. This is less than that of the currently circulating ground water. This body of massive ground ice may have been formed by the freezing of bulk water in an existing aquifer, i.e. a cavity filled in by ground water under artesian pressure started to freeze as permafrost aggraded or *in situ* intruded under pressure into permafrost.

19. Mountain permafrost

The lower limit of global permafrost distribution roughly corresponds to sea level at about 60^{o}N and shows an overall trend to rise about 110 m per degree latitude to reach a maximum altitude of about 5000 m ASL in the subtropical zones. It is estimated to occupy some 5 million km^2 in mountain ranges at middle and low latitudes, excluding the uplands of eastern Siberia and the Far East. It is mainly concentrated in the mountains of Central Asia (Himalayas, Tibet, Karakoram, Kunlun, Hindu Kush, Pamirs and Tien Shan) and in southern Siberia and Mongolia (Altai-Sayang, Khangai and Khentai). Less extensive areas underlain by permafrost are found in the Andes, the Alps, the Scandinavian mountains, the Rocky Mountains, The New Zealand Alps, the mountains of Japan and the Caucasus, as well as on individual summits of the Pyrenees, the Tatra, the Carpathians and mountain ranges in Iran and Turkey (see Table 5).

Table 5. Estimated area underlain by non-polar mountain permafrost (km^2).

After Haeberli et. al., 1993

The main factors of permafrost distribution in mountains are: (1) mean annual temperature (altitude effect, regional scale); (2) direct solar radiation (relief effect, local topographic scale); (3) snow cover characteristics (effects of wind drift and avalanche activity, meso- and micro-scale of surface roughness. Continuous permafrost in the form of negative temperature bedrock with numerous ice filled cracks is known to commonly exist above the equilibrium line on glaciers. Under extremely dry conditions, ice-rich continuous permafrost also occurs in Quaternary sediments below glacier equilibrium lines in addition to the frozen bedrock. However, where annual precipitation exceeds about 2000 mm, the equilibrium line on the glacier approaches the lower boundary of the permafrost occurrence and even the timberline. Maritime climate mountains such as the Cascades in North America or Patagonia in South America mainly have seasonally frozen ground below the glacier equilibrium lines. In localities of extreme shade, permafrost occurrences are known at altitudes where the mean annual temperature is far above 0 ^{o}C. Special conditions also exist in mountainous areas of extreme aridity, such as the eastern Pamirs, where the soil moisture tends to be very low and frozen ground mainly occurs in lacustrine sediments and glacial moraines.

The area of perennially frozen Quaternary sediments between the timberline and the equilibrium line on glaciers is usually characterized by air temperatures between 0 and –8 ^{o}C, due to the insulating effect of the winter snow cover. Mountain permafrost is typically a few tens of meters up to considerably more than 100 m thick. Supersaturation of fine-grained, frost-susceptible sediments with ice contents between 40 and 100% by volume seems to be common, whereas coarse material is more likely to be saturated with about 30 to 50% ice by volume. Considerable volumes of massive ice can exist within creeping rock glacier permafrost and probably also within perennially frozen moraines and debris cones. Buried sedimentary ice from avalanche cones or glaciates can also occur in alpine permafrost but may be less widespread than often assumed.

In mountainous areas (e.g. in Magadan Region, Siberia) large (more than 20 m high) Late Pleistocene syngenetic ice wedges can be found. These findings of large syngenetic ice wedges in slope sediments of the Utinoe and Phoenix ice-wedge complexes are important in the study of ice-wedgecomplex origin. Attachment to a slope and content of rock debris are undoubted evidence of the main role of gravitation in sediment formation. It is gravitation processes that are responsible for the removal of material along the slope. The intermountain depressions commonly provide little information, which is why it is very important to find an appropriate object for Late Pleistocene palaeo-reconstruction. The study of ice-wedgecomplexes has allowed the authors to reconstruct palaeo-facial and palaeo-climatic features reliably and in detail. The primary results of these investigations is the first description of syngenetic ice-wedge complexes in intermountain depressions, slope sediments containing rock debris, and also the obtaining of the first radiocarbon and oxygen isotope characteristics. The $\delta\ ^{18}O$ values of recent ice wedges are 25 to 27 (modern mean January temperature is –39 ^{o}C). The same in relict Late Pleistocene ice wedges are 30 to 32. Such values are typical for ice wedges from Phoenix and Utinoe cross-sections, which could be dated about 30 (25?) to 10 ka BP. Hence the mean winter temperatures of that period were lower than modern ones by 5–6 ^{o}C, i.e. –30 to –32 ^{o}C, and mean January temperatures were lower by 7–9 ^{o}C, i.e. –46 to –48 ^{o}C.

20. Global climate change and changes in glaciers and ground ice

Changes in the extent of the Greenland and Antarctic ice sheets may lead to a worldwide sea level increase of as much as 10 m. Of special concern to periglacial environments will be inundation of coastal tundra lowlands of the western Northern American Arctic and Siberian lowlands. A climate–induced rise in sea level will increase coastal erosion, flooding and inshore marine sedimentation. There is a strong possibility that the Arctic Ocean's ice cover, currently about 1.5 m thick on average and with a mean annual area of about 4 000 000 km^2, will disappear. The change in net albedo, the increased source of atmospheric moisture from an ice–free ocean, and the consequent increase in cloudiness and precipitation will alter the environment of the high Arctic islands, many of which are technically deserts. Since about 1840, when reliable observations on sea ice began in the Northern Hemisphere, its maximum extent has fluctuated by about 50%. General experience suggests that rapid climatic warming would produce an almost instantaneous response of lighter sea–ice conditions in the marginal areas of the Arctic Ocean. Depending on storm patterns, cloudiness and precipitation, sea ice in the Central Arctic Ocean might take a decade or more to respond; but warming of the amounts postulated would lead to instability and eventual disappearance of the ice. Perhaps some of the most dramatic changes will be associated with permafrost.

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Glossary

Aufeis (icing) : stratified ice mass on the surface of the earth or ice as a result of periodic eruption and freezing of water.

Epigenetic ice : ground ice developing inside enclosing rock and after the latter has formed.

Glacier discharge : volume flowing in a unit of time.

Equilibrium line : a line on a glacier surface at which the ablation (thawing) is equal to accumulation.

Ice cap : a diffluent glacier coming down to the sea; ice cliffs are formed as a result. The largest Eurasian ice caps are located in Iceland, Franz Josef Land, Spitsbergen, Novaya Zemlya Island, Severnaya Zemlya Island, Bennett Island, Henriette Island, Jeannette Island, Victoria Island, Ushakov Island and Schmidt Island. The largest ice caps of the Canadian Arctic Archipelago are on Baffin Land, Ellesmere Island, Devon Island, Axel–Heiberg, Melville Island, and Meighen Island.

Ice sheet : a vast area covered with a thickness of ice. More than 96.6% of ice area and 90% of ice volume is concentrated in the Greenland and Antarctic ice sheets, due to climatic features.

Ice shelves : large slabs of floating ice surrounding much of the Antarctic continent.

Intrusive ice : formed by water intrusions, usually under pressure, into a seasonally or perennially frozen zone.

Jokulhlaup : the sudden and rapid draining of a glacier dammed lake.

Pingo ice : grows when water is intruded into confining material and freezes in a tabular mass along the base of the active layer and parallel to the permafrost surface.

Surge : a period of exceptionally rapid sliding during which a large volume of ice is transferred from a reservoir area to the terminal part of a glacier.

Syngenetic ice : ground ice forming at, or almost at, the same time as the enclosing sediments are deposited and usually associated with surface aggradation.

Taber ice : Segregated ground ice that can range in thickness from hairline to more than 10 m. It commonly occurs in alternating layers of ice and soil (also *ice lens*).

Vein ice : formed by the penetration of water into open fissures developed at the ground surface.

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Biographical Sketches

Vasil'chuk Yurij Kirillovich, Doctor of Sciences, Professor, Academician of the Russian Academy of Natural Sciences (2004, Corresponding Member - 2000), Professor of Cryolithology and Glaciology Department, Head of Regional Engineering Laboratory, Lomonosov's Moscow State University 119992, Moscow, Lenin Hills, Lomonosov's Moscow State University, Geology Faculty, Engineering and Ecological Geology Department, Main Building, Zone A, room No 111

Prof. Vasil'chuk Yurij Kirillovich was born in 1954 in Lazo, (Moldova). He graduated Lomonosov's Moscow State University with excellent degree in geocryology and glaciology in 1975. He received a PhD in 1982 and Doctor of Sciences degree in 1991. He is Academician of the Russian Academy of Natural Sciences from.2004. Since 1992 he is a head of Glaciology and Geocryology Data Centre of Theoretical Problems Department of Russian Academy of Sciences. Since 1996 he serves as a professor of Cryolithology and Glaciology Department of Geography faculty of Lomonosov's Moscow State University and from 1997 as a Head of Regional Engineering Laboratory of Engineering and Ecological Geology Department of Geology faculty of Lomonosov's Moscow State University. His principal science interests are in area of isotope geochemistry, geochronology, Quaternary Geology, stratigraphy, geocryology, glaciology and geomorphology. He undertook field investigations in nearly all permafrost regions of Eurasia, such as Gydan and Yamal Peninsulas in the North of Western Siberia, Central and Northern Yakutia, Chukotka, Magadan region, Trans-Baikal region and Arctic Islands.

Yu.K.Vasil'chuk is the author of over 200 publications, from them there are 7 monographies, such as: "Oxygen-Isotope Composition of Ground Ice" (Application to paleogeocryological reconstructions) 2-volum issued in 1992 and the textbook "Principles of Isotope Geocryology and Glaciology" (coathered with Academician RAS V.M.Kotlyakov) issued in 2000 et al., about 20 papers he has published in "Transactions of Russian Academy of Sciences" and more than 25 ones in the International Journals, such as *Radiocarbon, Permafrost and Periglacial Processes, Nuclear Instruments and Methods in Physics Research B, Earth* and *Planetary Science Letters* etc. His recent textbook, "Soil Engineering» (2005, Lomonosov' Moscow University Press), was co-authored with V.T.Trofimov et al. This textbook characterized the ground ice as a base for constructions. Currently he prepared the new book "Ice wedge: Heterocyclity, Heterogeneity, Heterochroneity".

Vasil'chuk Alla Constantinovna, PhD, Senior Researcher of the Geography Department, Laboratory of Geoecology of the North, Senior researcher, Moscow State University named M.V.Lomonosov, Vorob'yovy Hills Moscow, Russia, 119899.

Dr.Alla Vasil'chuk was born in 1955 in Groznyi (formerly USSR). In 1973 she has attended a Russian secondary school in Groznyi with gold medal. She graduated a Geography Faculty Moscow State University with excellent degree in Palaeogeography (palynology) in 1979. From 1998 Dr. Alla Vasil'chuk is Official Applicant of Doctor Habilitate Degree in Laboratory of Palaeogeography of Pleistocene, Moscow State University. Her principal interests are focused in Facies and Genetic Palaeogeographic and Palaeotemperature reconstructions based on the pollen data of Late Quaternary permafrost sediments. Dr.Alla Vasil'chuk participate in field investigations in many permafrost regions in Gydan and Yamal Peninsulas in the North of Western Siberia, Central and Northern Yakutia, Chukotka, Magadan Region and Arctic Islands: Ayon, Belyi et al. Alla C.Vasilchuk is the author of 120 publications. Her monograph, "Formation features of pollen spectra in Russia permafrost area" (2005, Lomonosov'Moscow University Press) firstly contain ^{14}C dating of pollen concentrate from ice-wedge ice. It has been shown, that due to good safety of ancient pollen and spores the ^{14}C age of pollen concentrate often is older than the organic micro inclusions. At the moment she is completing her monograph "Palynology and chronology of polygonal ice wedge complexes in Russia permafrost area".

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PROPERTIES OF GLACIAL, ICEBERG AND PERMAFROST WATER

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Biographical Sketch

Prof. Vasil'chuk Yurij Kirillovich was born in 1954 in Lazo, (Moldova). He graduated Lomonosov's Moscow State University with excellent degree in geocryology and glaciology in 1975. He received a PhD in 1982 and Doctor of Sciences degree in 1991. He is an Academician of the Russian Academy of Natural Sciences since 2004. He is the head of Glaciology and Geocryology Data Centre of Theoretical Problems Department of Russian Academy of Sciences since 1992, and of Regional Engineering Laboratory of Engineering and Ecological Geology Department of Geology faculty of Lomonosov's Moscow State University since 1997. He is also a professor of Cryolithology and Glaciology Department of Geography faculty of Lomonosov's Moscow State University since 1996. His principal scientific interests are in area of isotope geochemistry, geochronology, Quaternary Geology, stratigraphy, geocryology, glaciology and geomorphology. He undertook field investigations in nearly all permafrost regions of Eurasia, such as Gydan and Yamal Peninsulas in the North of Western Siberia, Central and Northern Yakutia, Chukotka, Magadan region, Trans-Baikal region and Arctic Islands.
Yu.K.Vasil'chuk is the author of over 200 publications, from them there are 7 monographies, such as: "Oxygen-Isotope Composition of Ground Ice" (Application to paleogeocryological reconstructions) 2-volum issued in 1992 and the textbook "Principles of Isotope Geocryology and Glaciology" (coathered with Academician RAS V.M.Kotlyakov) issued in 2000 et al., about 20 papers he has published in "Transactions of Russian Academy of Sciences" and more than 25 ones in the International Journals, such as *Radiocarbon, Permafrost and Periglacial Processes*, *Nuclear Instruments and Methods in Physics Research B, Earth* and *Planetary Science Letters* etc. His recent textbook, "Soil Engineering» (2005, Lomonosov' Moscow University Press), was co-authored with V.T.Trofimov et al. This textbook characterized the ground ice as a base for constructions. Currently he prepared the new book "Ice wedge: Heterocyclity, Heterogeneity, Heterochroneity".

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Table 1. Mass and distribution area of the ice on Earth.

Ice type	Mass of (tons)	Distribution area

		%	Million Km^2	%
Glaciers	$2{,}398\times 10^{16}$	98.95	16.3	10.9 Land
Icebergs	7.65×101^{2}	0.03	63.5	18.7 Ocean
Snow cover	1.05×10^{13}	0.04	72.4	14.2 Earth
Marine ice	$3{,}483\times 10^{13}$	0.14	26.0	7.2 Ocean
Ground ice	$(2–5)\times 10^{14}$	0.83	21.0	14.1 Land
Atmosphere ice	1.68×10^{12}	0.01	510.0	100 Earth
Total	$2{,}423\times 10^{16}$	100		

Source: Shumsky, Krenke, 1969

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Table 2. Modern surface glaciation of the Earth

Area	**Glaciation area, km^2**	**Water reserves of glaciers, Km^3**
Greenland	1,802,600	2,704,400
Canadian Archipelago	149,990	42,770
Eurasian Arctic Islands and Byrranga mountains	91,130	22,710
Iceland	11,785	3,968
Europe(continent)	7,395	482
Asia (continent)	119,730	11,140
North America (continent)	123,700	24,960
Northern Hemisphere	**2,306,330**	**2,810,430**
South America (continent)	32,300	5,430
Africa (continent)	13	1
Australia (continent)	0	

New Zealand	817	64
New Guinea Island	7	1
Antarctic Islands	25,500	5,700
Antarctica (continent)	13,589,000	27,480,000
Southern Hemisphere	**13,647,637**	**27,491,196**
Earth Land	**15,953,967**	**30,301,196**

Source: Dolgushin, 2000

Table 3. The largest ice caps of Arctic islands (after Dolgushin, 2000)

Eurasian Arctic		**Canadian Arctic archipelago**	
Glaciers	**Area, km^2**	**Glaciers**	**Area, km^2**
Iceland	11,785	Baffin Land	36,830
Vatnajцkull	8,400	Terra Nivea Ice Cap	165
Langjцkull	1,025	Grinell Ice Cap	130
Western Svalbard	21,240	Hall Ice Cap	490
North–East Land	11,135	Penny Ice Cap	5,960
West Ice	2,623	Barnes Ice Cap	5,935
East and Southern Ice	7,985	Ellesmere Island	77,180
Franz Josef Land	13,735	Glaciers of United States Ridge	28,210
King George Land	2,241	GlaciersVictoria and Abert mountainsl	20,400
Vilchek Land	1,892	Glaciers of Prince of Wales mountains	17,300
Gram Bell Island	1,215	Axel Heiberg Island	12,560
Edgeшya Island	1,880	McGill (Akaioa) Ice Cap	7,250
Novaya Zemlya	24,322	Steacie Ice Cap	3,040

North Island	23,802	Bylot Island	4,895
South Island	520	Devon Island	16,575
Severnaya Zemlya	17,472	Koburg Island	230
Komsomoletz Island	5,903	North Kent Island	140
October Revolution Island	7,580	Meighen Island	76
Bolshevik Island	3,318	Melville Island	335
Total	**102,950**	**Total**	**149,990**

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Table 4. The length of ice cliffs providing a source for icebergs on the main islands of Svalbard

Island	**Total coast length, Km**	**Length ice cliffs, Km**	**Ice cliffs as % of coast**
Spitsbergen	3068	484	16
Nordaustlandet	1369	306	22
Edgeoya	429	79	18
Prins Karls Forland	208	17	8
Barentsoya	186	23	12
Kvitoya	115	106	92
Storoya	32	13	41
Totals	5407	1028	19

Source: Dowdeswell, 1989

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Table 5. Estimated area underlain by non-polar mountain permafrost (km^2).
After Haeberli et. al. (1993)

Region	**Area (km^2)**

Northern Eurasia	1,000,000
China, Nepal, India, Pakistan, Afghanistan	2,000,000
Mongolia	1,000,000
Caucasus	20,000
European Alps	50,000
Scandinavian mountains	80,000
Iceland	50,000
North America	400,000
South America	270,000
New Zealand Alps	10,000

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Table 1. ^{36}Cl concentration in the polar snow (after Tuniz et al., 1998)

Object	Concentration	Reference
Snow, Antarctica	10^6 atom/kg	Elmore et al.,1987; Suter et al., 1987
Snow, Greenland	10^7 atom/kg	Nishizumi,1983

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Table 2. ^{10}Be, ^{14}C and ^{36}Cl concentration of (10^3 atoms in 1 g) in ice cores near the surface (after Tuniz et al., 1998)

Region	Station	10Be	14C	36Cl
Greenland	Dye–3	0.5–1	–	1.4
	Cemp Century	0.5–1	–	–
	Milcent	0.5–1.2		2.0
	Summit	–	–	2.7
Antarctic	Byrd	2.5	–	–