GEOS 120, Part B: GLACIERS

Lectures TR 2:00-3:30 pm, 10/11/01 through 11/13/01

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Required text:

Post & LaChapelle (2000): Glacier ice, University of Washington Press, 145pp.

Grading criteria:

Quizzes (Thursdays, starting Oct. 18):	15 % (total)
Final exam (Nov. 13):	50 %
Labs:	35 % (total)*

* Attendance of labs is mandatory. Lab exercises are to be completed and handed in by the end of each lab session.

Course outline

1. Introduction and overview (Oct 11)

- Glaciers: What are they? Where do they occur? Why do we study them?

2. Ice, glaciers and landscapes (Oct 16, 18, 23, 25)

- The building blocks of a glacier: Snow, firn and ice
- Anatomy and flow of a glacier
- Glacier types & Glacier hydrology
- Ice sheets, ice streams and ice shelves
- Periglacial phenomena and the glacial landscape

3. Ice ages and climate (Oct. 30, Nov. 1, 6)

- Glaciation and the ice ages
- Glaciers, ice sheets and global climate change
- Ice cores as an archive of past climate

4. Sea ice (Nov. 8)

Lecture notes (Acrobat pdf file, requires free Adobe Acrobat Software to read):
http://www.gi.alaska.edu/~eicken/he_teach/G120Notes.pdf

1. Introduction to Glaciers: What are they? Where do they occur? Why do we study them? (reading: Post & LaChapelle, pp. 3-13; information on the web: http://www.glacier.rice.edu/)

1.1. What is a glacier?

- Large **mass of ice** (and firn and snow) **flowing** and **sliding** down slope or valley as a result of the **pull of gravity**; individual glaciers or ice masses may link up into one large ice cap or ice sheet (Greenland or Antarctic ice sheet)
- Glaciers can in some ways be thought of as **rivers of ice**, with the ice flowing down an inclined surface (but there are still considerable differences between the way water flows down a river and ice flows down a glacier!); the deformation behaviour of a glacier is in some ways similar to deformation in the earth's crust (there are even icequakes) and in some ways similar to the way lava flows down the side of a volcano, yet, the timescales are different with glacier movement occurring at speeds of several meters to several hundreds of meters per year (and sometimes even more)

1.2. Where do glaciers occur?

- throughout the **solar system** one can find different types of larger ice bodies, not only water ice but also ice made up of ammonia, carbon dioxide or other substances that are gases or liquids at conditions typical for the earth's surface but may freeze solid at much lower temperatures; while the planets between sun and earth (Venus, Mercury) are too warm for ice to build up, **Mars** has a big polar icecap composed of water and carbon dioxide ice; this ice caps grows and shrinks with the seasons, similar to the seasons on earth with different amounts of solar radiation received at the earth's surface in winter and summer; the surface layers of Jupiter and Saturn are composed of different types of ices and one of Jupiter's moons, **Europa**, actually has an ice crust of several tens to possibly hundreds of kilometers thickness which is in many ways similar to ice features visible on earth; even our moon may have some ice present within the upper soil layers
- on earth, we find glaciers and icesheets on (almost) **every continent** (in fact, if one counts New Zealand as part of the Australian continent, then all continents have glaciers); roughly 10 % of the total land surface are covered by glaciers or ice sheets; sea ice covers roughly 10 % of the world's ocean surface
- if we take into account the distribution of ice sheets during previous glaciations, an even larger fraction of the earth's surface has at some point in time been covered by glaciers; as a result, a significant fraction of the **landscapes** on earth are influenced by **glacial action**

Table 1.1:
The Distribution of Water on the Earth

	Volume (km^3)	%
Oceans	1 348 000 000	97.39
Polar ice caps, icebergs, glaciers	227 820 000	2.01
Ground water, soil moisture	8 062 000	0.58
Lakes and rivers	225 000	0.02
Atmosphere	13 000	0.001
Sum	1 384 120 000	100.00
Fresh water	36 020 000	= 2.60%
Fresh water as a percent of its total		
Polar ice caps, icebergs, glaciers		77.23%
Ground water to 800 m depth		9.86%
Ground water from 800 to 4000 m depth		12.35%
Soil moisture		0.17%
Lakes (fresh water)		0.35%
Rivers		0.003%
Hydrated earth minerals		0.001%
Plants, animals, humans		0.003%
Atmosphere		0.04%
Sum		100.00%

After Baumgartner and Reichel, *The World Water Balance*, Elsevier, 1975.

1.3. Why do we study glaciers?

- from the earliest of times, people have been in close interaction with glaciers and icesheets; thus, a significant part of our ancestors lived on an earth with vast ice sheets covering a large part of the northern hemisphere; **changes in sealevel** associated with the buildup of the great **Laurentide Ice Sheet** in North America and the Fennoscandian Ice Sheet in northern Europe were crucial in creating the **Bering landbridge** and making Alaska the stepping stone into the Americas

- today, as mankind's impact on global climate becomes increasingly apparent, it is the impact of glacier and ice sheet melt on global sealevel that causes concerns; moreover, snow and ice play an important role in the **global climate system** because of their ability to reflect a significant fraction of the incoming sunlight (snow has an albedo of around 0.7 to 0.95, such that 70 to 95 % of the shortwave radiation coming from the sun is reflected back from the surface); changes in the extent of glaciers and ice sheets may not only impact the development of climate but are also something of an early **indicator of impending climate change**

- apart from issues of climate change and sea-level rise, glaciers are of immediate importance in areas where **water supply** and **hydroelectric power** depend on glacier-fed streams; in fact, it has even been discussed to tow icebergs from the Antarctic to arid regions to alleviate the problems of water shortage; given the large area of industrialized regions settled on glacial or peri-glacial (i.e. in the vicinity of glaciers or ice sheets), glaciers also have an indirect impact on questions of water supply, foundation of structures etc.

- apart from these and other practical considerations, glaciers (and ice in general) are **aesthetically and culturally important**, our planet would be considerably less appealing in the absence of ice

1.4. A note on units

- to describe quantities (such as the density of a piece of ice or the length of a glacier) scientific researchers use units that conform with the so-called SI International System of Units; this is based on the kilogram (kg), the meter (m), the second (s) and a number of other units; the table shown below provides information on how to transform a quantity measured in a given unit into that of the imperial system of units

A. Units of length

1 cm = 0.39 in; 1 in = 2.54 cm

100 cm = 1 m = 1.09 yd = 3.3 ft; 1 ft = 0.31 m

1 km = 0.62 mi; 1 mi = 1.61 km

B. Units of volume

1 l = 1000 cm^3 = 1000 ml = 0.264 gal; 1 gal = 3.79 gal

1 m^3 = 1000 l = 264 gal

C. Units of weight

1 kg = 1000 g = 2.205 lbs = 35.3 oz

1 t = 1000 kg = 2205 lbs

D. Units of temperature

Relative: 1 ˚C = 1 K (Kelvin) = 1.8 ˚F

Absolute: T in ˚C = 5/9 x (˚F – 32); T in ˚F = 9/5 x ˚C + 32

T in K = ˚C + 273.15

2. The building blocks of a glacier: snow, firn and ice (information on the web: http://snowflakebentley.com/index.htm)

2.1. Ice: not just like any average rock

- **water molecules** are composed of one oxygen and two hydrogen atoms (**H_2O**); in solid ice, the individual molecules are linked with one another through so-called hydrogen bonding, which is based on electrostatic forces; with each water molecule having two bonds available for sharing with neighbouring water molecules, the molecules form a crystal lattice that is composed of layers of molecules; because these **layers** are easily separated and may **glide** against one another, ice offers little resistance to deformation forces, which is important for the way a glacier flows; within these layers molecules are arranged in a six-fold (**hexagonal**) symmetry which is also reflected in the appearance of individual snow flake crystals; this six-fold symmetry and the characteristic shape of snow flakes with six branches (which may sprout further branches) has fascinated people in different cultures for a considerable time; Johannes Kepler, the famous astronomer, wrote a small booklet in 1611 about the nature of the snow flake and explained this six-fold symmetry through the layered arrangement of individual little building blocks (which he likened to peas), thereby coming quite close to our modern understanding of ice being composed of water molecules

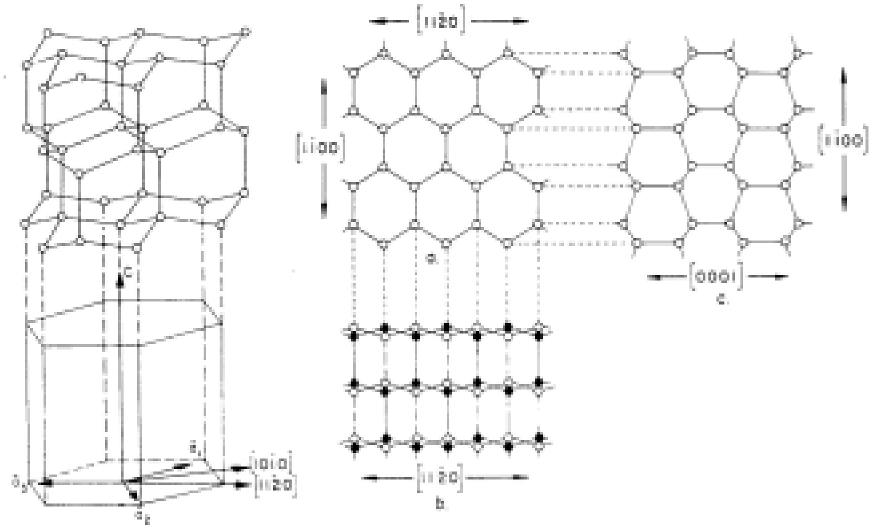

Fig. 2.1: Crystal structure of ordinary ice. The water molecules are shown as white/black circles. Note the hexagonal (six-sided) arrangements of molecules in the horizontal plane and the way these planes are arranged parallel to one another.

Table 2.1: Important properties of water and ice

Melting point of ice:	0 ˚C*
Freezing point of water:	0 ˚C*
Density of water:	1.0 g/cm^3
Density of ice:	0.92 g/cm^3
Crystal sizes:	0.1 mm to >100 mm
Regelation:	because of ice density being lower than water density, ice melts under pressure and refreezes (regeals) when pressure is released
Viscosity:	when a force is applied ice may flow (like wax or other viscous materials); the effective viscosity of ice is 10^{14} times higher than that of water and 10^{7} times lower than that of typical rocks

* for pressures above atmospheric pressure the melting point drops

- technically speaking, ice is a rock composed of just one mineral: ice; however, water ice is very much different from other rocks and minerals insofar as it has a **lower density** in the **solid** than in the liquid state; this is because some of the hydrogen bonds linking all the individual water molecules in the solid break up when the ice starts to melt, such that the liquid is composed of small **clusters of water molecules** still held together by hydrogen bonds but packed more densely than in the ice lattice; the fact that the density of ice is smaller than that of water has at least two important consequences: (1) **ice fishing** is possible: if ice would not have a smaller density than water, lakes would freeze from the bottom up, so that by the time you could put your hut out on the ice and drill some holes, the fish would be dead and frozen solid; (2) **glaciers** may **melt** at the **bottom** even under temperatures below 0˚C (or 32˚F): the pressure of the overlying ice may cause ice to melt at the base of glaciers and ice sheets even if the temperature is below 0˚C; a particularly impressive demonstration of this fact is the existence of a lake the size of Lake Ontario underneath the Antarctic ice sheet, buried below more than 3 km of ice with liquid water at a temperature of -3 ˚C; this melting at the base is of particular importance for the **sliding** of glaciers on their bed

2.2. From snow to firn to ice

- ice is continuously moving down a glacier, with new material (i.e. snow) deposited in the **accumulation** area and ice being removed through melting (or calving) in the **ablation** zone; thus, every glacier or ice sheet is fed from snow flakes deposited year after year in its upper reaches

- once **snow** crystals have been deposited on the ground, they are subject to a number of important changes; first, the delicate, **highly branched dendritic crystals** may be blown around by the wind such that breakage of crystals results in much smaller particles which can be compacted to a much larger degree; second, even if not redistributed, the highly branched crystals with a very high surface area in relation to their volume tend to change their shape (in a process called **metamorphism**) to a more compact spherical shape, which corresponds to a decrease in the surface area relative to the volume of ice; this process is driven by the tendency of substances to **minimize** their **surface energy** similar to the way small droplets of water tend to join into larger droplets, e.g. with rain flowing down a window pane; further ageing and summer melt result in the transformation of snow into more compact, higher density **firn** (see Table 2.2); as layer upon layer of snow is deposited on a glacier or ice sheet, the firn is compressed under the weight of the overlying snow and ice (with the density increasing, see Table 2.2) until eventually the **air space** inbetween the grains of ice is compressed to individual small **bubbles**; this transition from an open pore space, which is connected to the surface of the glacier (and hence the atmosphere), to isolated pores which are completely enclosed corresponds to the transition from **firn** (with an open pore space) to **ice** (with a closed pore space); the firn-ice transition usually occurs at an ice density of approximately 0.83 g/cm^3 or 830 kg/m^3 (see Table 2.2 and Fig. 2.3)

Table 2.2: Typical densities of snow, firn and ice at different stages of development (Paterson, 1994)

Type	Density, g/cm^3
New snow (immediately after falling under calm conditions)	0.05 - 0.07
Damp new snow	0.1 - 0.2
Settled snow	0.2 - 0.3
Depth hoar	0.1 - 0.3
Wind-packed snow	0.35 - 0.4
Firn	0.4 - 0.83
Very wet snow and firn	0.7 - 0.8
Glacier ice	0.83 - 0.92

- as a result of the different **atmospheric conditions** prevailing during snow formation and deposition as well as the differences in **metamorphism of the snowpack** on the ground, there are distinct differences between the type of snow one finds in different areas of the world at different times of year; Fig. 2.2 shows a classification scheme of snow types that is based on the air temperature, the amount of precipitation (i.e. how much snow falls within a given period of time) and the wind speed
- thus, high temperatures, higher wind speeds and high amounts of precipitation would give rise to a **maritime snow pack** with higher densities, ice layers and lenses, high liquid water content (this is typical for instance for snowfall in the **coastal eastern U.S.** and in some parts of coastal Alaska; the high density of this snow and the sometimes very large precipitation rates can result in a substantial risk of **avalanche** release in some areas)
- cold, dry and less windy conditions such as typical of the forested areas surrounding Fairbanks yield the typical Taiga snowpack (which you will be studying in more detail in one of the labs)

- snow provides very good thermal insulation because of its high air content which is one of the reasons why Inuit and other northern peoples rely on certain types of snow as a building material
- air, however, does still circulate in the snow pack and because of temperature gradients in the snow (as a result of thermal insulation) this air carries moisture which in turn promotes metamorphism of the snow grains; an important type of snow that forms as a result of moisture deposition in very steep temperature gradients (more than 10 to 25 ˚C per m) is so-called **depth hoar**; these are big crystals in the shape of cups, prisms or columns with sharp, stepped edges that typically form at the base of the snow cover where the ground provides both heat and moisture for the crystals to grow; depth hoar can also form just below the surface of a snow pack when solar heating provides for a steep temperature gradient (the latter is typical of the snow pack on the large polar ice sheets, where the formation of **summer depth hoar layers** is important for dating of ice layers based on the visual appearance of the snow and ice in a vertical section - the so-called snow or **ice stratigraphy**)
- because of their shape and size, depth hoar crystals are not as tightly linked with one another as the spherical snow grains of ordinary windpacked or metamorphosed snow; consequently depth hoar layers in a snow pack have very little strength and are easily sheared off; on a hill slope, shearing along depth hoar layers may result in an entire, thick slab of snow (i.e. an **avalanche**) sliding down the hill along a low-strength depth hoar layer; in particular in interior Alaska such slab avalanches are quite common and of considerable potential danger

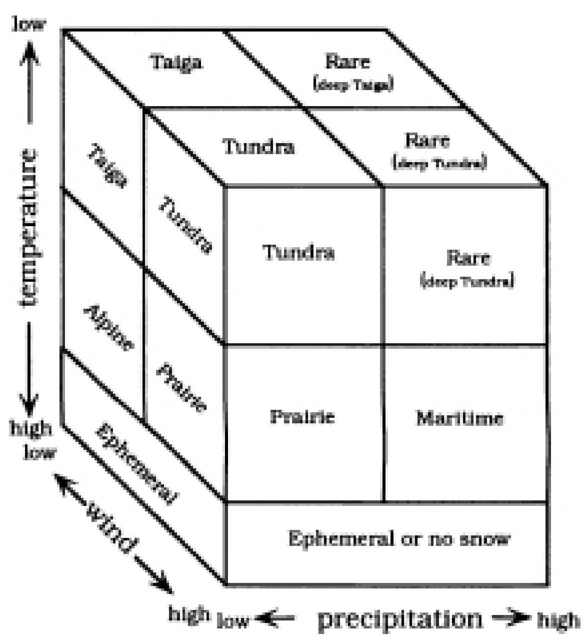

Fig. 2.2: Classification of snow pack based on the amount of precipitation, the wind speed and the air temperature (Sturm et al., 1995).

2.3. Ice stratigraphy and fundamentals of ice-core analysis

- with more and more snow layers deposited in the accumulation area of a glacier or an ice sheet (see Section 3 for details), the transformation of snow to firn to ice takes place, with the ice preserving important information about the **depositional history** and the **environmental conditions** prevailing at the time the snow fell and was deposited (Fig. 2.3)

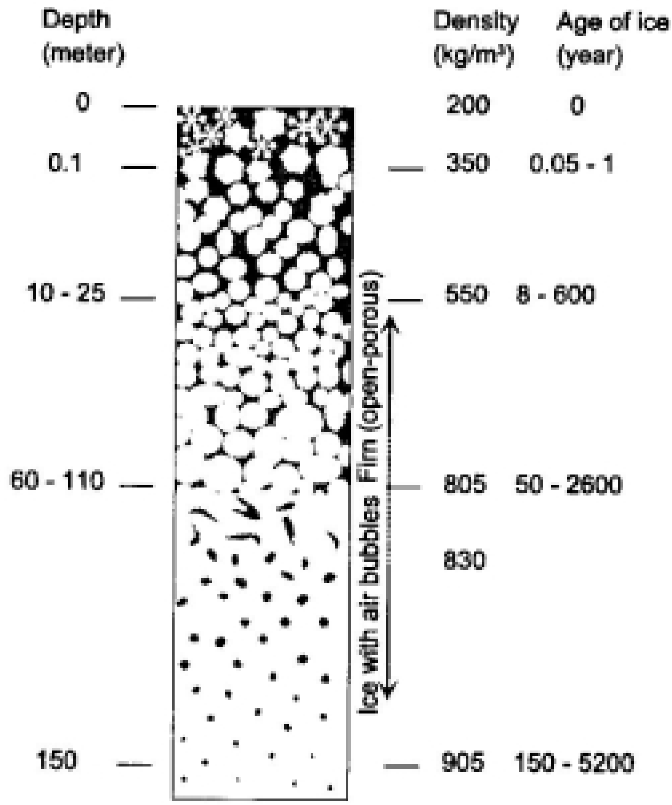

Fig. 2.3: Stratigraphy of the upper snow and ice layers in an ice sheet, showing the increase in ice density and age with depth (Schwander, 1996).

- in particular the central parts of ice sheets, where the lateral motion of the ice is very small and undisturbed deposition of snow preserves an environmental record of thousands or tens of thousands of years, are of great interest as **archives of past climates**; as changes in the air temperature are of particular interest in the study of climate change, glaciologists (i.e., ice researchers) have developed a technique to deduce past temperatures from the chemical composition of the ice; in order to do this, the concentration of two **isotopes** of oxygen in the water molecules contained in an ice sample are measured; as shown in Fig. 2.4, the two common isotopes Oxygen-16 and Oxygen-18 (^{16}O or ^{18}O) differ in atomic mass; as the heavier atom Oxygen-18 has a tendency to **sublimate at higher temperatures** (the process of turning solid ice into vapour or vice versa is called sublimation), we find higher concentrations of Oxygen-18 in snow or ice formed at warmer temperatures; thus, by measuring the concentration of Oxygen-18 in an ice sample one can calculate the temperature at which the snow was formed and deposited (a more detailed discussion of this method and examples of such ice core temperature records will follow later)

- apart from such isotope chemistry measurements, the crystal microstructure of ice cores is also commonly analysed to learn more about the deposition and metamorphism of the ice; for this type of microstructural analysis, **thin sections** (usually as thin as 0.2 mm) are cut from the ice; studying these thin sections in **polarized light** produced by orienting two polarizer sheets at 90 degrees, one above and one below the sample; light where all the light rays correspond to waves oscillating in only one direction is polarized; the polarizer acts like a row of fence posts which only allow passage of objects parallel to the poles; upon entering an ice crystal each ray of light is split up into two rays travelling at different speeds; when recombining these two rays in the second polarizer, they interfere and as a result crystals appear in different colours; it is these different colours that allow us to **distinguish** different **ice crystals** and learn more about the **formation and history of the ice**

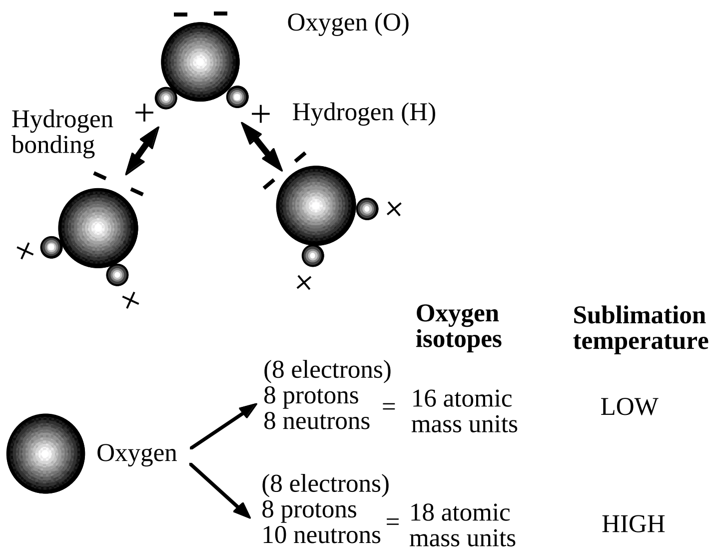

Fig. 2.4: Structure of water molecule, hydrogen bonding and oxygen isotopes.

3. The anatomy of a glacier (reading: Post & LaChapelle, pp. 3-13; information on the web: http://www-water-ak.usgs.gov/glaciology/gulkana/ - Gulkana glacier mass balance monitoring by the USGS)

- glaciers or ice sheets are characterized by an **accumulation zone**, where the surface is **gaining mass** averaged over the course of an entire year and an **ablation zone** where the surface is **losing mass** (mostly through melting) over the course of a year (Fig. 3.1); glacier motion thus transports snow and ice from the accumulation to the ablation area

3.1. Accumulation zone

- that part of the accumulation that experiences no or only minute amounts of melting is termed the **dry snow zone**, basically the entire Antarctic ice sheet and large parts of the Greenland ice sheet are a dry snow zone; the **wet snow zone** and **percolation zone** are those parts of a glacier where the snow collects meltwater and where the meltwater may subsequentially run down into the glacier, either refreezing or **circulating through channels** within and at the bottom of a glacier; the water circulating within and at the bottom of a glacier can carve out impressive features from the underlying bedrock (discussed at a later stage as part of glacial erosion)

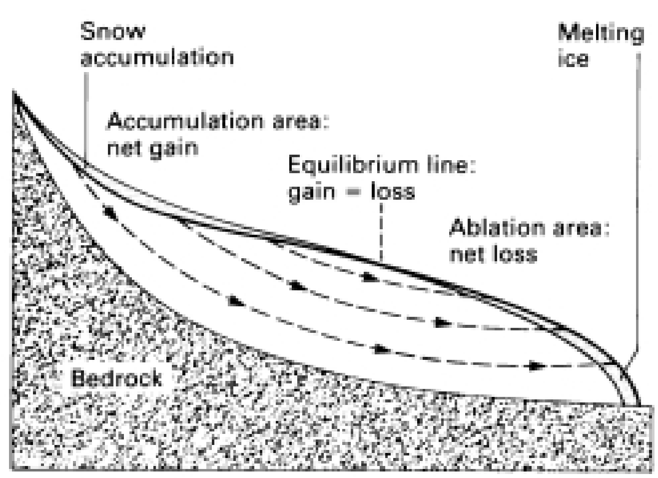

Fig. 3.1: Anatomy of a glacier (Hambrey and Alean, 1992).

- the metamorphism of the snow is typically characterized by the following four stages: (1) **rounding** of crystal faces or pointed crystals through **melting/sublimation**, (2) **settling** of the snow, (3) **sintering** whereby individual crystals link up with one another through necks, eventually melding into single larger grains, (4) **recrystallization**, with ice crystals changing size and shape as a result of deformation and ageing; extensive melting during the summer months enhances the speed of the transformation from snow to ice

- for typical **temperate** glaciers (with ice reaching melting point at some point during the season) the transformation from snow/firn to ice (see Table 2.2 for the corresponding density values) may be completed within **3-10 years** time corresponding to a depth in the glacier of **10 to 35 m**; in **polar** ice sheets (where the temperature is mostly below the melting point throughout the upper layers) this transformation requires **100-500 years** and occurs at depths of **60-100 m** (see Fig. 2.3)

3.2. Ablation zone

- the accumulation zone is separated from the ablation zone by the **equilibrium line**, where the loss of ice through ablation equals the gain of ice through accumulation; the **equilibrium line altitude** (ELA) is an important climatological variable that depends on the **temperature** and the **precipitation** rate and may change considerably with time

- while most of the ablation in glaciers typically takes place through **melting**, ice may also be lost through **sublimation** (turning ice into water vapour, particularly in dry regions); for some glaciers and in particular the Antarctic ice sheet, mass is also lost through **iceberg calving**; in most glaciers, summertime melt rates range between 5 and 10 cm of snow or firn and 1 and 5 cm of ice per day; the melt rate increases at lower altitudes or with the deposition of thin (<1-2 mm) **dirt layers** on the ice (because they increase the amount of sunlight absorbed by the ice); thicker dirt layers reduce the melt rate because they thermally insulate the ice from the atmosphere

4. Glacier motion and crevassing (reading: Post & LaChapelle, pp. 15-32; information on the web: http://www.asf.alaska.edu:2222/intro_begin.html - Glacier Power at UAF with nice crevasse pictures)

4.1. Glacier motion

- glaciers move by **sliding on the bed** (or **deformation of the bed**) and **internal deformation** of the ice; the relative importance of these processes can vary considerably depending on the type of glacier, but typically sliding at the base contributes between **25 and 75 %** to the total flow of a glacier

- **internal deformation** of ice is favoured by the **ice crystal lattice** (Fig. 2.1) which allows ice layers to **glide** past one another along the basal plane of the lattice

- the sliding on the bed depends on the extent of "lubrication" provided by water present at the base of a glacier; the higher the water pressure at the bed (providing some "float" to the ice), the faster the glacier can slide at the bed; glaciers resting on a bed of till (muddy, clayey sediments accumulating from erosion by the glacier and subglacial water movement) may also move through **internal deformation** of the **till**

- in horizontal cross-section the velocity profile through a glacier can range between **plug flow**, where the velocity is constant throughout the glacier and drops off sharply to 0 at the margin (the glacier thus flows like a plug sliding through a channel), or **parabolic** where the velocity is at a maximum in the center of the glacier and drops off steadily towards the sides; in longitudinal cross-section, the flow lines (i.e. the lines along which a particle travels after being deposited on the glacier surface) **dip into the glacier** in the **accumulation** zone and **emerge from out of the glacier** in the **ablation** zone (Fig. 3.1); the velocity is at a **maximum** in the area of the equilibrium line

- as with a mountain stream, the velocity of a glacier depends on the **slope** of the glacier and its bed (due to internal deformation), the steeper the slope (the larger the slope angle α) the faster the glacier moves; for glacier flow as a result of internal deformation, the ice velocity V also depends on the **thickness** H (and to some extent the temperature) of the ice such that

$$V = k (\sin \alpha)^3 H^4$$

where k is a temperature-dependent constant; thus, if one doubles the slope, the velocity increases by a factor of $2^3 = 2 \times 2 \times 2 = 8$, a doubling of the ice thickness increases the velocity 16-fold; similar to a river, the ice flows in the direction of the maximum slope

- if the slope of the underlying ground is 0, i.e. the ice is resting on a horizontal surface such as in the case of a large ice sheet, the ice velocity depends only on the ice thickness and the surface slope of the ice sheet (and the conditions at the base of the ice sheet); in such a case the shape of the ice sheet takes on a parabolic form

4.2. Crevassing

- changes in slope or changes in the sliding conditions at the bed of a glacier thus result in **changes in ice velocity**; the extensional or compressional forces associated with these changes can become quite large and once the strength of the ice is exceeded, they induce **crevassing** such that the **ice fails** along a plane perpendicular to the direction of maximum stretching and a crack opens up; these cracks are typically **limited** to the upper **30 to 40 m** of a glacier or ice sheet, because the pressure of the overlying ice causes cracks at greater depth to be **sealed** quickly by plastic flow of the ice; similarly, plastic behaviour of the uppermost snow layers can also conceal crevasses (this is typically observed in Antarctic ice streams and ice shelves, where some crevasses are only visible by using radar that penetrates through the top snow layers and shows the reflections from the walls of the crevasses below); also, on glaciers crevasses are often bridged by snow accumulating on the glacier which increases the hazard of travelling on a glacier in winter (or the accumulation area during summer)

- depending on the type of forces acting on a glacier, three main types of crevasses can be distinguished (Fig. 4.1):
(1) **marginal or shear crevasses** occurring along the margin of a glacier where it slides past solid bedrock, a moraine or a slower or stagnant body of ice; these crevasses open up as a result of shear forces and are oriented at roughly 45˚ to the margin, pointing upglacier; (2) **transverse crevasses** occurring in a zone of extension where the glacier accelerates, e.g., at the top of an ice fall, where the ice “plunges” down a slope that is much steeper than the upstream slope of the glacier; (3) **longitudinal (splaying) crevasses** through a zone of compression where the glacier slows down, such as in the ablation zone near the terminus

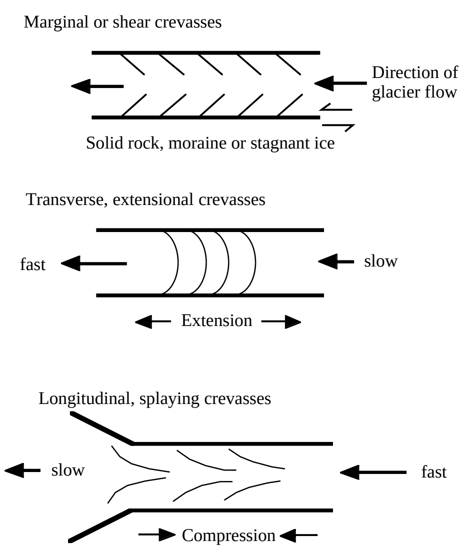

Fig. 4.1: Major types of crevasses occurring in a glacier.

- the irregular crevasse appearing in the uppermost reaches of a glacier where it detaches itself from stagnant firn or ice or a rock wall is called the **bergschrund**

4.3. Ogives

- Ogives appear as **alternating bands** of **dark** and **light** ice, that are convex in the direction of flow (because the ice in the center of the glacier moves faster than at the sides)

- ogives are produced in **ice falls**, where the ice is **stretched** by longitudinal forces as it moves through an icefall, enlarging the surface area which in turn increases the amount of surface ablation and the amount of dust collecting on the surface during summer (**dark bands** correspond to ice moving through ice fall in **summer months**, they also correspond to a slight **trough** in the glacier surface; **light bands** move through ice fall in **winter**); thus, ogives can be used to measure the velocity of a glacier since every pair of dark and light bands corresponds to one year of motion; by counting the number of band pairs one can estimate how far a glacier has travelled from one point to another along a valley

5. Types of glaciers (reading: Post & LaChapelle, pp. 125-137; information on the web: http://nsidc.org/glaciers/information.html - National Snow and Ice Data Center Glacier pages)

5.1. Temperate and cold (polar) glaciers

- an important distinction in glacier types relies on the temperature in the body of the glacier; thus, glaciers which are **at or close to the melting point** except for an upper layer that may cool out in winter are known as **temperate glaciers**; most glaciers in the world except for the polar ice sheets and some high-altitude glaciers are temperate (even most glaciers in Alaska); because of the presence of meltwater, temperate glaciers have a much stronger impact on bedrock or the underlying topography than **cold or polar glaciers** which are mostly below the freezing point and often frozen to their bed; as a result of the ground heat flow at the base of thick ice sheets and because of the decrease in the melting point of ice with an increase in pressure even the cold, polar ice sheets show an increase in temperature towards the base (Fig. 5.1) which is of importance for ice flow (warmer ice deforms more easily) and may also result in the formation of meltwater pools at the base of the ice sheet in warmer spots (see under-ice lake mentioned in Section 2.3)

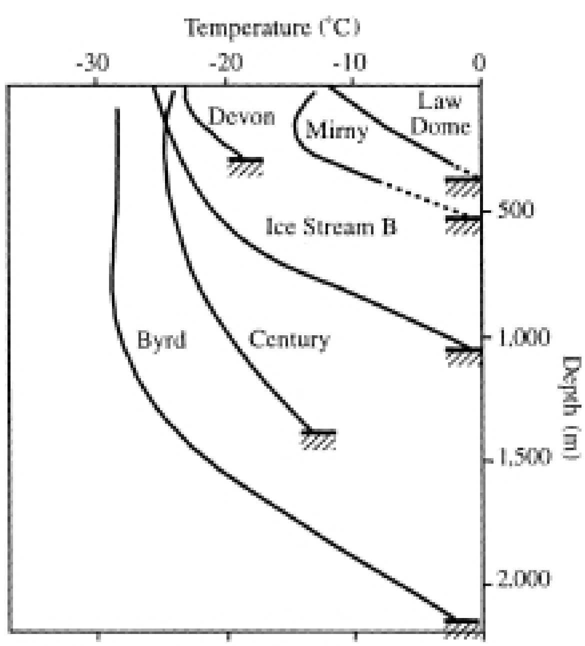

Fig. 5.1: Temperature profiles through polar ice caps and ice sheets. Note the increase with temperature at depth (Paterson, 1994).

5.2. Glacier types based on shape, occurrence and flow

- **ice sheets** and **ice caps** are large masses of ice, mostly confined to the polar regions, that cover a larger area (in the case of Antarctica an entire continent), with snow accumulating in the central, higher parts of the ice sheet and ice flowing down to its margins where ablation takes place through melting or iceberg calving (a more detailed discussion of ice sheets can be found in the subsequent Section along with discussion of **ice shelves** and **ice streams**)

- most of the other types of glaciers can in some be classified as **mountain glaciers**:

(1) **cirque glaciers**: resting in a bowl-shaped depression carved from a mountain side by glacial action

(2) **valley glaciers**: the “prototype” of a typical glacier with the ice flowing down a valley, carving out a U-shaped valley in the process, possibly with several **tributary** glaciers that feed into a main valley

(3) **hanging glaciers**: clinging to a steep slope or rock wall (with ablation taking place also through ice tumbling down below)

(4) **Piedmont glaciers**: spilling out from a valley into a flat plain with small slope, as a result the glacier spreads out into a large lobe

(5) **Rock glaciers**: mixtures of rock and ice (slow moving or even near-stagnant) that are the result of massive accumulations of debris on an ordinary glacier or result out of ground ice and permafrost processes

(6) **Tidewater glaciers**: terminating in tidewater bays or inlets, with the ice grounded but almost afloat; the high velocities favoured by enhanced sliding at the base result in heavily crevassed termini, producing numerous **icebergs** (cragged shape and hence likely to capsize or disintegrate); tidewater glaciers rely on a terminal moraine (formed from debris accumulating at the terminus and being continuously pushed forward by the slowly advancing glacier) as a toehold; once that toehold is lost, e.g. when the glacier enters into a wider bay or deeper water, the glacier tongue may break up and retreat at substantial speeds (up to a km a year); while **advance** may thus take around **1000 years**, **retreat** by the same distance can be completed within **less than a century**; a prominent Alaskan example of such tidewater glaciers can be found in **Glacier Bay** where a substantial retreat has been observed in this century (see p. 51 in Hambrey & Alean)

- **surging glaciers** are characterized by periods of **rapid advance** with velocities increasing from several tens to few hundred meters per year to several tens of meters per day at maximum; surging behaviour is quite common (also in Alaskan glaciers) and not yet completely understood; however, it is clear that the buildup of pressure in **water-vein systems** in the interior and the base of the glacier play an important role in triggering a surge; as a result of a surge, the glacier bulges out in the lower reaches and thins in the upper parts with ice moved rapidly down-glacier (see Fig. 5.2)

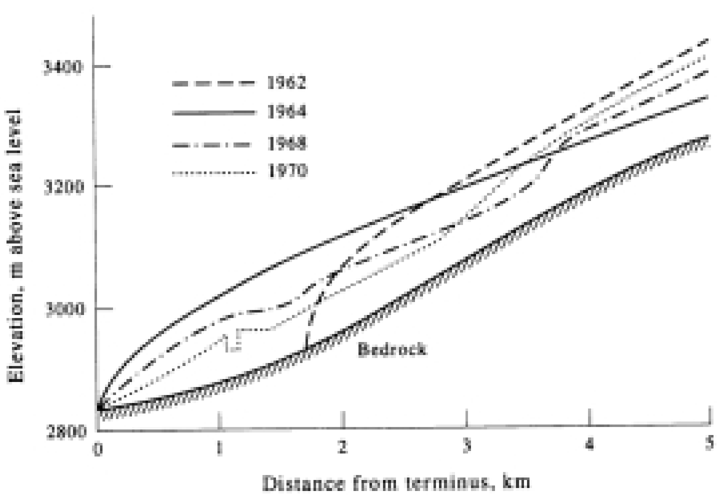

Fig. 5.2: Cross-section of Medvezhiy Glacier (Ural Mountains) before during and after surge. Note how ice bulges out at front during and immediately after surge (Paterson, 1994).

6. Ice sheets, ice shelves and ice streams (Reading: Post & LaChapelle, pp. 93-102; information on the web: http://www.space.gc.ca/csa_sectors/earth_environment/radarsat/radarsat_info/amm/antarc.asp - tracing Amundsen's and Scott's steps to the Pole by satellite; http://www.natice.noaa.gov - giant icebergs at the National Ice Center)

- in regions with **sufficient snowfall** and **cold temperatures** that allow snow to stay on the ground the entire year (i.e. where the equilibrium line occurs at very low altitudes, typically this is only the case in the polar regions), larger **ice sheets** may start to build up and continue to grow until they cover vast regions or even an entire continent as is the case in Antarctica; thus, ice sheets are not confined to mountainous regions (although that is often where they start to form) but may cover flat, low areas as well

- presently, two large ice sheets are found on earth:
(1) the **Antarctic ice sheet** covering the entire continent of Antarctica with an area of approximately 12 million km^2, an average thickness of slightly less than 2.5 km and a volume of roughly 25 million km^3; in the vicinity of the **South Pole** the ice sheet is roughly 4000 m thick at an altitude of around 4000 m above sealevel; the Antarctic ice sheet has a long history and has been in existence for at least **20 million years**
(2) the **Greenland ice sheet** covering the island of Greenland with an area of approximately 2 million km^2, an average thickness of 1500 m and a volume of roughly 2.5 million km^3; the Greenland ice sheet was built up mostly during the previous (or present?) ice age and is probably not older than about **2 million years** (in fact, model results show that if one were to remove the Greenland ice sheet today, it would not form again in its present size but only as a number of smaller ice caps separated by non-glaciated lands)
- combined, both of these ice sheets account for roughly two-thirds of the worlds freshwater reservoirs; if they were to melt completely they could raise global sealevel by as much as 60 m

- in the North American and Eurasian Arctic one can find miniature ice sheets, so-called **ice caps** which may reach up to 1 km in thickness and several tens to at most few hundreds kilometers in size

- just as in a glacier, ice sheets are fed by snow being deposited in the **accumulation zone** (which usually occupies almost the entire area of the ice sheet, see Fig. 6.1) and with ice flowing down towards sealevel; ablation typically occurs through **iceberg calving** and to a much lesser extent than in ordinary glaciers through surface melting (at least in Greenland; in the Antarctic there is essentially no surface melt but underneath ice shelves – see below – bottom melt is observed); in those spots where the ice flows downslope to all sides (so-called **domes** or **ice-divides**, see the one marked at the center in Fig. 6.1) one can obtain long, undisturbed records of snow and ice accumulation by drilling and extracting a core from the entire thickness of the ice sheet

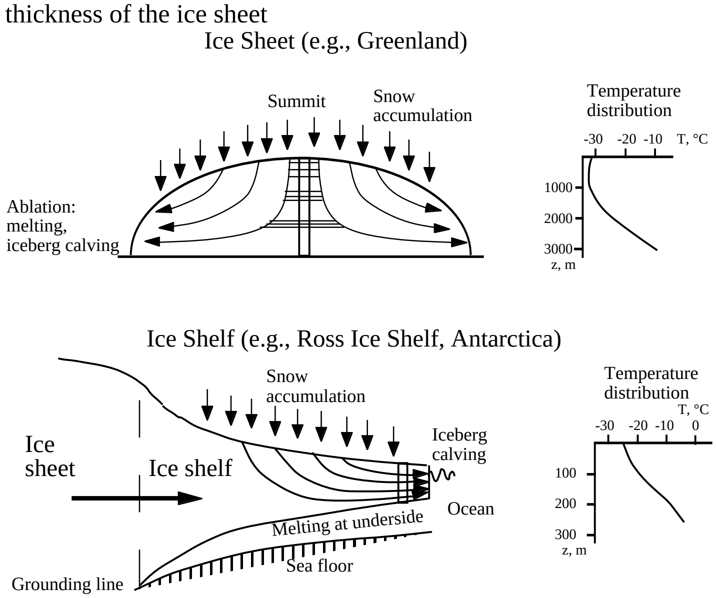

Fig. 6.1: Schematic cross-section through an ice sheet (top) and an ice shelf (bottom).

- the Antarctic ice sheet can be divided into the large **East Antarctic ice sheet** which rests on rock mostly **above sealevel** (a terrestrial-based ice sheet) and the **West Antarctic ice sheet** which has a base that is **below sealevel** in vast areas (a **marine-based** ice sheet); thus, changing, e.g., the thickness of the latter may allow water to penetrate at the base of the ice sheet and partially or completely float those parts that are resting on a base below sealevel; while we presently do not know very much about how likely such a scenario is or what it could be triggered by, we do know that a **collapse** of the West Antarctic ice sheet (which, even though termed a "collapse" would occur only on timescales of decades to centuries) would **raise sealevel by several meters** and thus put coastal populations at risk

- significant parts of the Antarctic ice sheet float up on the ocean as the ice reaches the coast; these **floating** extensions of the ice cap are termed **ice shelves** (see Fig. 6.1); they account for little under half of the total coastline of Antarctica and total more than 1.5 million km^2 in area; ice shelves are important sources for icebergs (**tabular icebergs** which can reach more than 50 km in diameter, such as some of those tracked by the National Ice Center through satellite remote sensing, see web pages indicated above); ice shelves are important ablation sites not only because of iceberg calving but also because the ocean circulating underneath may melt substantial amounts of ice, an ablation process that is of great importance in the Antarctic but almost insignificant in other areas of the world

- ice shelves are fed by **ice streams**, which can be thought of as glaciers moving several hundred meters per year within a field of stagnant or slow moving ice surrounding them; while it is presently not fully understood what makes the ice move fast in an ice stream, the sliding conditions at the base of an ice stream (with lubrication provided by highly deformable till beds) appear to play a major role

7. Glacial deposits and the glacial landscape (Reading: Post & LaChapelle, pp. 45-52, 111-124; information on the web: http://daac.gsfc.nasa.gov/DAAC_DOCS/geomorphology/GEO_9/ GEO_CHAPTER_9_TABLE.HTML - satellite remote sensing of glacial landforms)

7.1. Bedrock features

- glaciers have a strong **impact on the landscape**, in particular as a result of **erosive action** (i.e. removal of solid and loose rocks and sediments); apart from **meltwater** circulating underneath or emerging from a glacier, erosion mostly takes the form of **rock inclusions** in the basal ice layers interacting with the underlying rocks or sediments or ice **plucking** solid rocks (which have been loosened through freeze-thaw cycles at the glacier bed) or entraining loose sediments and carrying them away;

- these abrasive and erosive processes may already start in firn fields or small cirque glaciers: by enlarging the cirques through glacial action, **arête ridges** or – if the several cirques cut into a single peak – **horns** can be carved out (see Fig. 7.1)

- as the glacier seeks the path of least resistance down a given slope, it usually follows pre-existing valleys; however, while rivers cut valleys with a caracteristic V-shape in cross-section, **glacier valleys** are **U-shaped**; if the glacial valley extends all the way out to the sea it may form a **fiord**

- if a glacier slides over solid bedrock, various types of **scratchmarks** (**striation**, **chattermarks**) are left behind on the surface; the direction and shape of these marks allows conclusions about the direction of ice movement during past glaciations (because such marks may be preserved for hundreds of millions of years – such as in Australia or South America – they are important evidence in the study of past periods of glaciation)

- on a larger scale, glaciers can carve large knobs called **roches moutonnées** (sheep rocks, because they look like herds of grazing sheep) out of the solid bedrock; these knobs can be several tens to hundreds of meters in diameter and are asymmetric with a gentle slope on their up-glacier side and a very steep face on the down-glacier side where material is continually being plucked from the bedrock

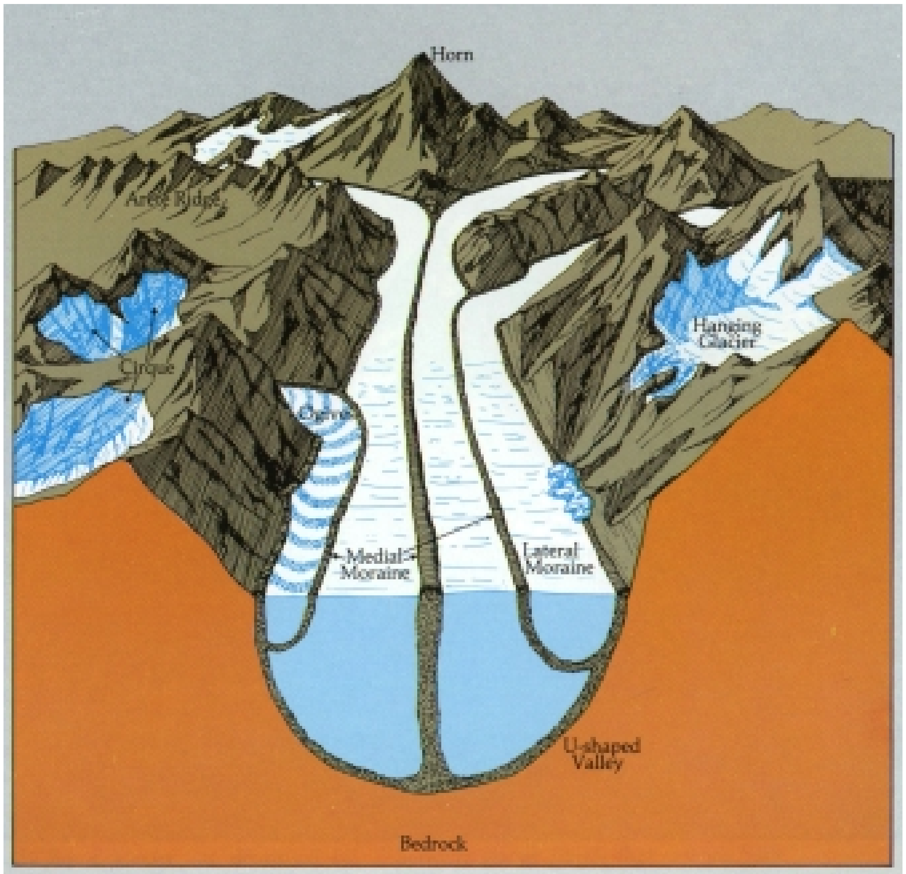

Fig. 7.1: Erosional rock landforms and different types of moraines (Molnia, 1983).

7.2. Glacial deposits

- sediments deposited by a glacier are referred to as a **till**; characteristically these till sediments are not layered and poorly sorted such that boulders of more than a meter in diameter may occur side by side with small pebbles, sand and clay (i.e. submicrometer grain diameters); tills that are actively deposited by the ice are referred to as **lodgment tills**, whereas sediments left behind from a melting glacier are termed **ablation tills**

- the till deposited at the base of a glacier, visible as a blanket of sediments covering the entire area occupied by a glacier once the ice has receded, forms a **ground moraine**; the deposits left behind around the perimeter of a glacier that has been stationary for some time are referred to as **end moraines** or **terminal moraines**; they are important markers of the advances or retreats a glacier has made at different periods of time (see Fig. 7.2)

- **lateral moraines** are found along the sides of a glacier and are composed of material that the glacier has abraded or plucked from the valley walls; in cases where a tributary joins a glacier from a side valley, the lateral moraine displaced towards the center of the glacier turns into a **medial moraine**

- a significant fraction of the vast amounts of material eroded by a glacier are transported by meltwater to the **outwash plain** surrounding the perimeter of the glacier (Fig. 7.2); apart from lakes filling hollows in the ground and streams and rivers cutting through the sediments, ice blocks may be left behind by the glacier or be dumped in front of the terminus, eventually leaving behind **kettles** or kettle holes as the ice melts

- apart from crevasse fillings left behind after glacier melt, one often finds **eskers**, long (up to several kilometers), winding, sandy ridges that were deposited in channels at the glacier bottom

- **kames** deposits occur in places where sediments accumulated in holes or in particular between the glacier and the valley wall over longer periods of time

- **drumlins** are asymmetric, rounded hills (with the gentle slope down-glacier and the steep slope up-glacier, i.e. contrary to the roches moutonnées) that often occur in swarms and are a result of processes occurring at the base of an ice sheet; presently it is not entirely clear how they form, however, since they may be either a leftover from erosional processes or features accreted by active transport of sediments into the drumlin

- while a lot can be learned about past glaciations through the interpretation of glacial landmarks, it should be noted that landscape features indicating an absence of an ice sheet or glaciers may also provide valuable insight into the climate history of a particular region; as an example, **granite tors** – a peculiar weathering pattern observed in granitic rocks with a billowy, stacked appearance - would be eroded away in the presence of a glacier; thus, UAF scientists from the Department of Geology and Geophysics have dated granite tors at locations in the White Mountains north of Fairbanks to establish the lack of glaciers or ice caps in this area during the last glacial maximum

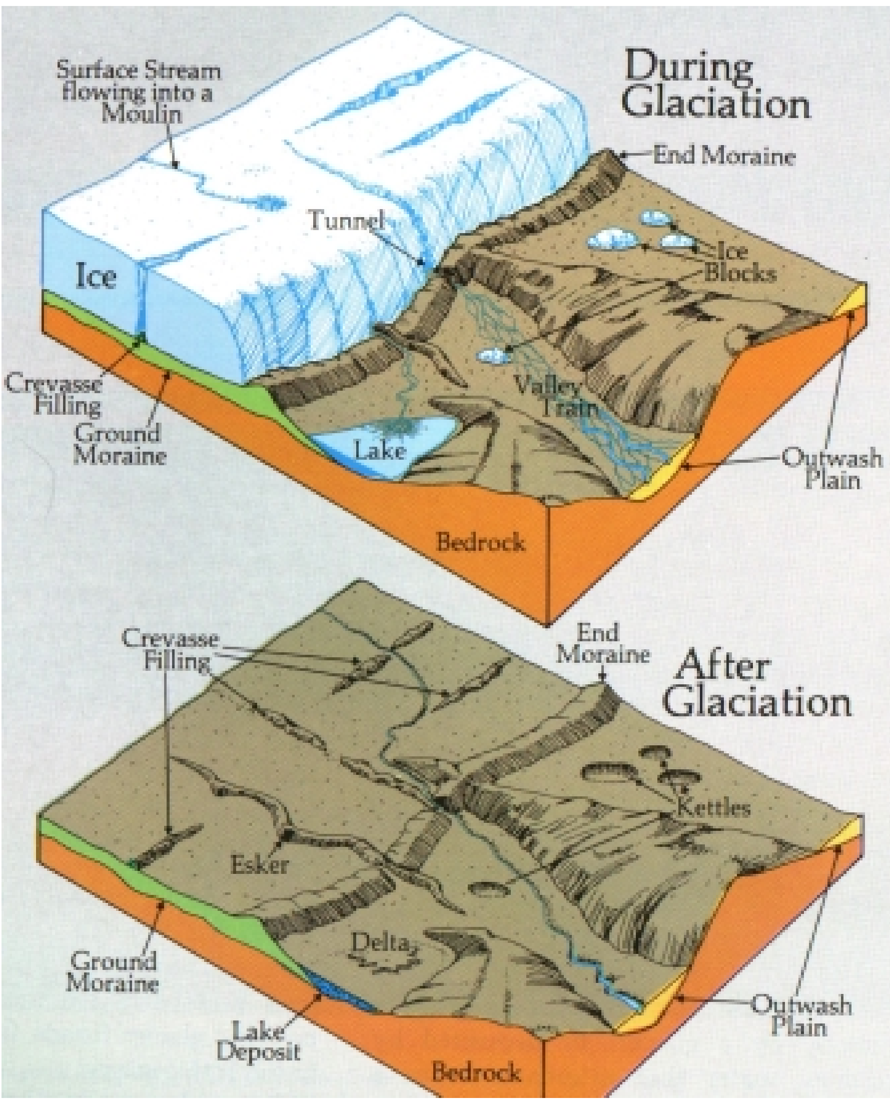

Fig. 7.2: Glacial deposits and landscape features (Molnia, 1983).

8. Ice ages, glaciations and climate (reading: Post & LaChapelle, pp. 45-52, 111-124; information on the web: http://www.ngdc.noaa.gov/paleo/parcs/atlas/beringia/index.html - paleoclimate of Beringia)

8.1 Ice ages and glaciation

- the various landscape features left behind by ice sheets tell a detailed story of ice ages several hundred millions years ago (see above) as well as of glaciations in North America and Europe in the not-so-distant past; periods of cooler climate associated with the presence of larger ice sheets on the earth are termed **ice ages** and may last from several hundred thousand to several million years; such ice ages consist of **colder, glacial periods** (of roughly **100,000 years** duration) during which the ice sheets grow and global climate cools by several degrees C, alternating with **warmer, interglacial periods** (typically **10,000 to 30,000 years** in length) with higher temperatures and reduced ice sheet volumes

- while the earth is presently still (at least for all we know) going through an ice age, we have been enjoying the moderate climate of an interglacial stage (called the **Holocene**) for the past 12,000 years; prior to that the so-called **Wisconsinan glacial** period (named after the southernmost extent of the **Laurentide ice sheet** that covered a large part of North America, centered roughly on Hudson Bay and central Canada) had led to the buildup of the Laurentide ice sheet in North America and the **Fennoscandian ice sheet** in northern Europe during a period of roughly 100,000 years

- the **causes** of ice ages and the factors that determine the length of the glacial and interglacial periods are **only partially understood** at present; it has become increasingly clear, however, that the earth's climate on such long timescales is to a large part **controlled by the sun**; thus, the waxing and waning of the large ice sheets at timescales of 100,000 years corresponds to changes in the distance between the earth and the sun at the point where the earth is farthest away from the sun during its orbit (such changes affect the eccentricity of the earth's orbit around the sun, see Fig. 8.1); similarly, variability on timescales of 41,000 and 23,000 years (see also climate record shown in Fig. 8.2) corresponds quite well to changes in the earth's axial tilt and the orientation of the earth's rotational axis

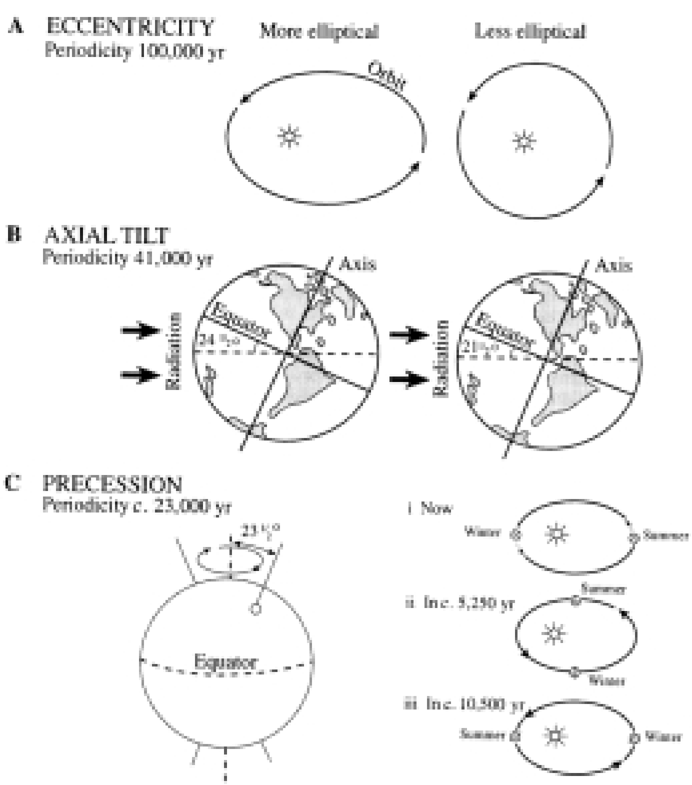

Fig. 8.1: Changes in the earth's orbit and axial orientation and tilt that are of importance for initiation of glacial periods (from Benn & Evans, 1998).

- glaciations, i.e. the build-up of ice sheets during the cooler glacial periods, have a profound impact on the earth's climate:

(1) the larger areas covered by ice and snow **reduce** the amount of **solar heat** absorbed at the earth's surface, because of the high **albedo** (fraction of the incoming sunlight reflected back by a surface), this in turn tends to amplify and increase the amount of cooling

(2) because a significant amount of water is locked up in the icesheets resting on land (i.e., up to 90% of the world's freshwater during a glacial age), **global sealevel** is significantly **lower during the glacial** than the interglacial periods; thus, **presently** sealevel is **120 m higher** than it has been during the last glacial maximum about 20,000 years ago; as a result of lower sealevel during the glacial maxima, landbridges may link islands or continents that are separated by water during the interglacial (e.g., today); 20,000 years ago Australia was connected to the Asian mainland and the Siberian Far East was linked to Alaska through the **Bering Landbridge**, allowing the settling of the Americas

(3) the **weight of the ice** accumulating up to several kilometers thickness is sufficient to **depress** the surface of the earth's **crust**; under the load of the ice the land surface yields and subsides near the center of the ice sheets, while it may actually bulge upward in a ring surrounding the ice sheets; after complete meltback of the ice sheets the earth's crust **rebounds** (by as much as 500 to 700 m in Scandinavia and more than 100 m in Northern Canada)

8.2. Ice cores as archives of climate history

- since written **records of climate** and direct measurements only date back a few centuries at best, we have to turn to other sources or proxies to learn more about the climate of the more distant past and the history of past glaciations; **deep ice cores** (some of them up to 3 km long) drilled to great depths in the ice sheets and ice caps are one of the most important sources of such information, because the ice layers provide us with a continuous **archive** of the temperature, the composition of the air and many other parameters prevailing at the time of ice deposition (see section 2.3 above)

- as outlined in Section 2.3 and discussed in more detail in the lab exercise on ice-core dating, the **isotope chemistry** of the ice (i.e. the concentration of the heavy isotope ^{18}O, written as $\delta^{18}O$) allows conclusions about the temperature at which the snow was deposited; from the analysis of two ice cores drilled in the early 1990's through the entire thickness of the Greenland ice sheet at its summit (see Fig. 6.1), where the ice is more than 3000 m thick, we have learned a lot about past ice age climate and the ups and downs of the interglacial and glacial cycles

- one of the most exciting findings in the temperature record of the Greenland ice cores is the **relative stability** of the **climate of the past 10,000 years** as compared to dramatic swings in the temperature record during the past glacial from about 100,000 to 10,000 years before present (Fig. 8.2)

- while it is not fully understood what these ups and downs in the temperature record prior to the stable period that has prevailed during the past 10,000 years are caused by, one wonders for whether the stable climate is going to hold for future years to come; thus, at some point or another during the next hundreds to thousands or ten-thousands of years the earth's climate **may shift back into a cold, glacial stage** again; at the same time, however, the **emission of greenhouse gases** (in particular CO_2) by mankind is highly likely to result in significant warming (part of which we may already be observing now)

- the ice-core records do in fact show the **increase in CO_2** as a result of the **burning of fossil fuels** (coal, oil, natural gas) starting with the industrial age and thus provide us with a clear picture of our own actions; by learning more from ice cores and other climate archives about the way the climate system works, we may be able to use models to predict what climate has in store for us and our descendants

- apart from their linkage to climate, glaciers and ice sheets are an important factor in controlling **global sealevel**; the shrinking and melting of many mountain glaciers and parts of the Greenland ice sheet has already significantly increased sea level and based on our present understanding will contribute by **another 22 cm** to the rise in sealevel predicted for the period 1990 to 2100 (Fig. 8.3)

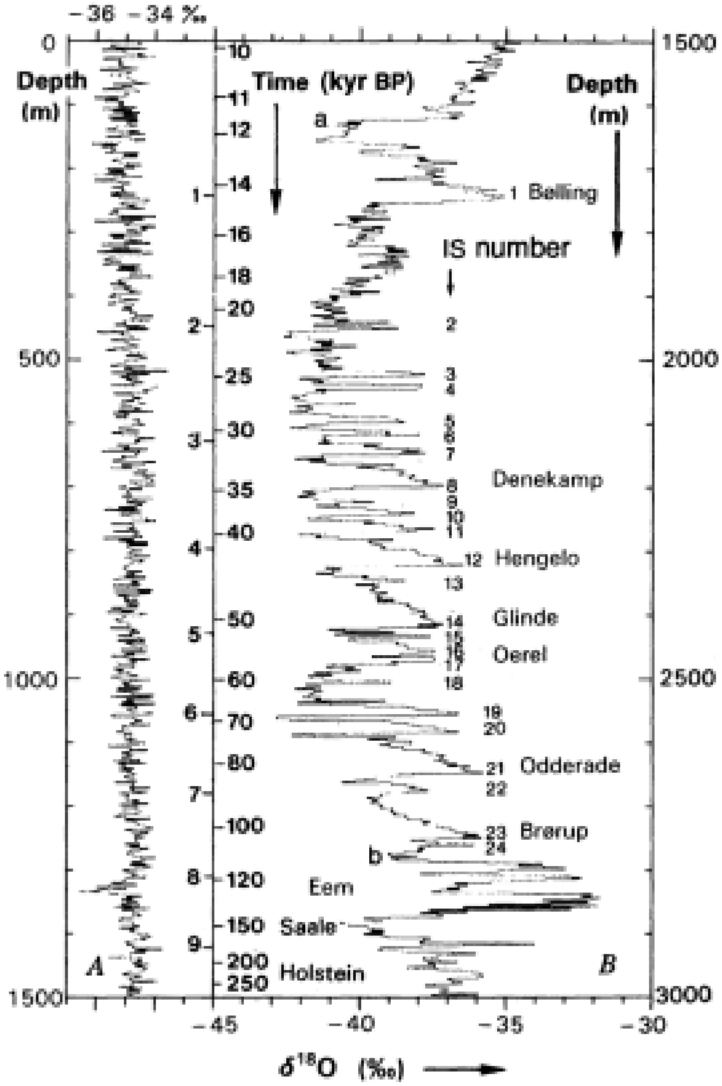

Fig. 8.2: Stable-isotope composition (concentration of heavy water isotope, $\delta^{18}O$) of ice core drilled at summit of Greenland ice sheet as a measure of the temperature during the past 10,000 years (curve in the left half of the diagram) and during the period from 10,000 to about 250,000 before present (curve at right of the diagram). Note how climate has been very stable during the past 10,000 years (the holocene period) as compared to the period prior to that (from Dansgaard et al., 1993).

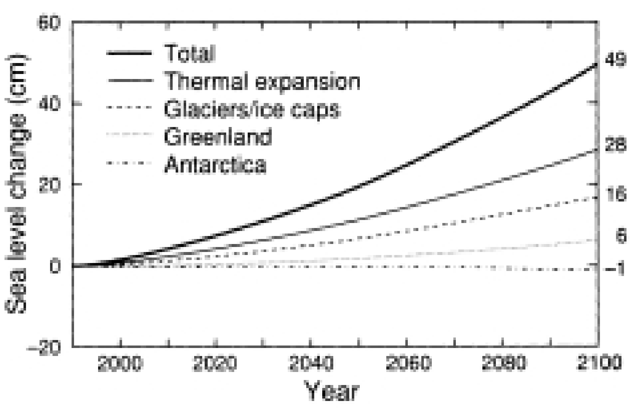

Fig. 8.3: Global sealevel rise predicted for the period 1990 to 2100. Glaciers and ice caps contribute more than one third to the total sealevel change. Because the Antarctic ice sheet may actually grow in the next century, it compensates a small amount of this rise (IPCC, 1995).