2 The polar regions as components of the global climate system

> The Arctic and Antarctic are the cooling chambers of our planet. Having a very limited supply of solar radiation, they attract warm air and ocean currents from the tropics, cool them down and send them back towards the equator as floating ice. In this way, the polar regions regulate the distribution of heat on the Earth. This mechanism will only continue to function smoothly, however, if the interactions between sea ice, glaciers, ocean and atmosphere do not change.

Why it is so cold in the polar regions

> The climate in the polar regions is the result of a self-reinforcing process. Because so little solar energy is received, the water freezes to ice, which then, like a mirror, reflects the small amount of radiation that does arrive. A multi-layered, complex wind system, which plays a decisive part in the weather and climate on our planet, is driven by differences in temperature and pressure between the warm and icy regions.

It doesn't get any colder

According to the World Meteorological Organization (WMO), the coldest place in the world is the Russian Antarctic research station Vostok. It was established in 1957 in the middle of the East Antarctic Ice Sheet, where it lies at an elevation of 3488 metres above sea level. From the station building it is about 1300 kilometres to the geographic South Pole. On 21 July 1983, at the standard measurement height of two metres above the ice, the station meteorologist measured a low temperature of minus 89.2 degrees Celsius – officially the coldest temperature ever directly measured on the Earth.

But at a height of just a few centimetres above the surface of the East Antarctic Ice Sheet the air temperature drops even further. According to satellite data obtained between 2004 and 2016, in a region of the ice sheet further to the south with a higher elevation, near-surface air temperatures can fall to minus 98 degrees Celsius.

2.1 > Elongated ridges of snow on the East Antarctic Ice Sheet. These are called sastrugi, and they are formed when the wind sweeps over the surface in places with somewhat harder snow, carrying loose snow crystals with it and carving aerodynamic elevations or furrows in the previously higher snow cover.

The thermal engine of the Earth's climate

The singular interplay between sun, ice, humidity and wind is the key to the extremely cold climate in the polar regions. The sun is the primary driving force of weather and climate on the planet. Its radiation warms the continents, the oceans and the atmosphere. The intensity with which the sun's rays impinge upon the outer boundary of the Earth's atmosphere has remained fairly constant since satellite measurements began in 2000. But because of the spherical shape of the Earth, not all locations on its surface receive the same amount of solar radiation. Where the rays intersect with the atmosphere at right angles, the light energy has a strength of 1361 watts per square metre (solar constant). Where the solar radiation strikes the Earth's atmosphere at a much lower angle, as in the polar regions, the incoming solar energy per unit of area is substantially reduced. Moreover, the radiation always falls only on the side of the Earth that is facing towards the sun. Accordingly, the global average solar energy arriving at the upper margin of the atmosphere can be calculated as approximately 340 watts per square metre. The much smaller amount of heat that reaches the polar regions can be illustrated by a simple example: If sunlight falls on the Antarctic continent at an angle of 30 degrees on a cloudless summer day, only half as much energy will arrive there as will fall on the surface near the equator at an angle of 90 degrees.

The major reason for the differences in heat input during the year is the fact that the Earth is spinning like a top in space, and its axis of rotation is not exactly perpen-

The levels of the atmosphere

The Earth is surrounded by a blanket of gas that extends from the surface of the planet to an altitude of about 500 kilometres, and which is held in place by the Earth's gravity. Its name "atmosphere" derives from the Greek words *atmós* and *sfaira* (vapour and sphere). The Earth's atmosphere consists mainly of the gases nitrogen (78.1 per cent), oxygen (20.9 per cent) and argon (0.93 per cent). However, trace gases, also known as greenhouse gases, such as water vapour, carbon dioxide, methane and ozone, with a combined proportion of well below one per cent, are of crucial importance for a climate on Earth that is capable of supporting life. These absorb a portion of the incoming solar radiation as well as a large part of the outgoing heat radiation from the Earth, thus contributing significantly to the warming of the atmosphere. Without the trace gases, the Earth would have an average temperature of around minus 18 degrees Celsius, and the blue planet would more closely resemble a snowball.

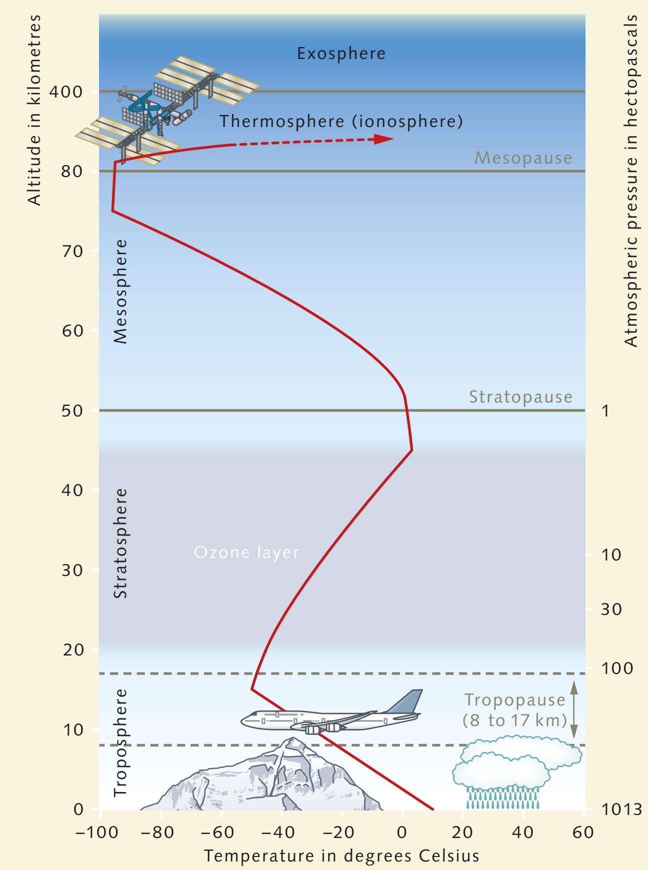
The Earth's atmosphere is composed of several layers, which can be distinguished by their physical and chemical properties. From the bottom upwards it is divided into the troposphere, stratosphere, mesosphere, thermosphere and exosphere.

However, only the two lower layers are important for the weather and climate on the Earth. Weather events occur here, especially in the troposphere, where the temperature decreases with increasing altitude by an average of about 6.5 degrees Celsius per 1000 metres. Above the equator, the troposphere extends to an altitude of about 17 kilometres. In the polar regions, by contrast, it is only half as high, namely eight kilometres.

Above the troposphere lies the stratosphere, which reaches an altitude of around 50 kilometres. In this layer, the temperature gradually increases upwards again because of the heat that is generated when the ultraviolet radiation in sunlight is absorbed in the ozone layer, located 20 to 45 kilometres above the mid-latitudes. Although the stratosphere, unlike the troposphere, contains almost no water vapour, stratospheric clouds with a pearlescent lustre can form under extremely cold conditions, especially in the polar regions.

Overlying the stratosphere is the coldest layer of the Earth's atmosphere – the mesosphere. It extends to an altitude of around 85 kilometres. With increasing altitude, the temperature and air pressure drop significantly, so that the average temperature at the upper margin of the layer is minus 90 degrees Celsius.

In the subsequent layer, the thermosphere, the density of the air is so low that the distance between individual gas molecules can be as much as several thousand metres, so that collisions and the associated exchange of energy seldom take place. The orbit of the International Space Station (ISS) is located in the thermosphere at an altitude of around 400 kilometres. At an altitude of 500 kilometres, the thermosphere transitions into interplanetary space. This transitional zone is called the exosphere. Within this realm the US American satellite SORCE orbits at an altitude of about 640 kilometres. Since 2003 it has been measuring the amount of solar radiation arriving at the outer edge of the atmosphere.



2.2 > The Earth's atmosphere is divided into several layers, of which only the lower two are important for weather processes. Clouds and precipitation form in the troposphere. In the stratosphere ozone absorbs incoming ultraviolet light as well as heat energy radiated from the Earth.

2.3 > Ice and snow surfaces in the polar regions reflect up to 90 per cent of the incoming solar radiation back into space, which results in a cooling of the Earth.

Weather and climate
For meteorologists, "weather" means the current conditions within the lower atmosphere (troposphere) as well as short-term changes at a particular time and place. They use measurements of the temperature, precipitation, wind direction and other parameters to describe these conditions. The term "climate", on the other hand, refers to the statistics of weather. It uses statistical mean and extreme values to describe the weather patterns at some location over a period of 30 years.

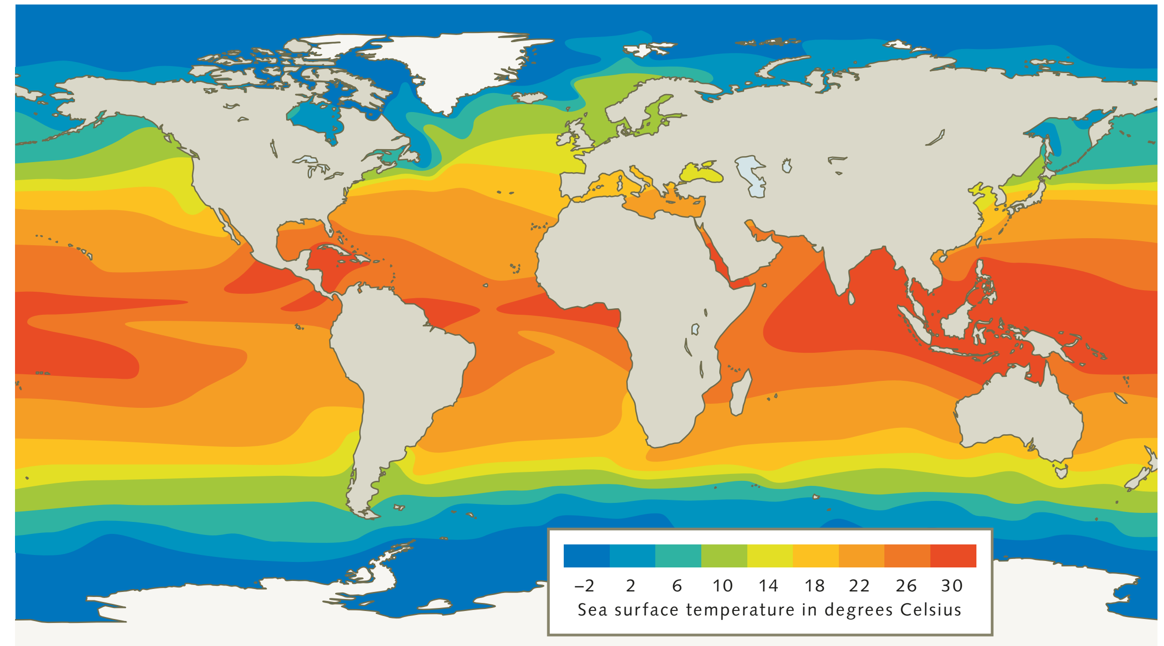
dicular to its plane of orbit around the Sun. Instead, it is presently tilted at an angle of 23.4 degrees. If the Earth's axis were at right angles to its orbit there would be no seasons. Due to its inclination, however, the northern hemisphere faces toward the sun during the northern summer and receives more sunlight, but in the northern winter it is tilted away from the sun and thus receives less radiation. In the southern hemisphere the exact opposite is the case.

The seasons are more pronounced the further one moves away from the equator, but these are not the only phenomena that can be attributed to the planet's inclination. It is also the reason why the sun does not set at all in the polar regions at certain times of the year (polar day) and why it remains hidden below the horizon (polar night) at others. At the geographic North and South Poles, the polar night lasts for almost a half year. If one moves from the Pole towards the Polar Circle, the length of time during which the sun does not rise above the horizon decreases steadily until it is only 24 hours precisely at the Polar Circle.

During its polar night, each polar region is therefore completely cut off from the solar source of heat energy. But even during the polar day, the period of continuous sunlight, only a relatively small amount of solar energy reaches the Arctic or Antarctic regions due to the low angle of incoming rays. These two phenomena thus form the basis for sustained cold conditions in the northern and southern polar regions. There are, however, two other important factors: albedo and water vapour.

White reflects

One of these reinforcing or amplifying factors is the reflective capacity of the Earth's surface, called the albedo effect. It determines how much of the incident solar radiation is reflected by the Earth's surface. As a basic rule of thumb, the darker or rougher a surface is, the less radiation it reflects. A freshly ploughed field may reflect about ten per cent of the sun's energy; green meadows and pastures can account for about 25 per cent. Light-coloured surfaces such as desert sands have an albedo of about 40 per cent, but still do not begin to approach the reflection values of snow and ice. Freshly fallen snow, for example, will reflect up to 90 per cent of the incident solar energy. Depending on its age and surface structure, sea ice



2.4 > Ocean surface temperatures reflect the strong contrast between the warm equatorial regions and cold polar regions. Where the sea and air are warm, greater amounts of water evaporate, and clouds are formed. In colder areas, however, evaporation rates are low.

can have an albedo of 50 to 70 per cent. This rather large range is partially due to the deposition of dust and soot particles on the ice surface over time, which change its colour, particularly in the Arctic. Ice floes from which the wind has blown away the snow layer also have a different surface texture than floes with a hardened snow surface. Snow-free glacier ice, for example, reflects up to 60 per cent of the radiant energy, but with a fresh snow cover the albedo is greater.

Thus, the general situation for the Arctic and Antarctic is as follows: In both regions a relatively small amount of solar energy reaches the Earth's surface for long periods of time because of the spherical shape of the Earth and the tilt of its axis. A large proportion of the energy that does arrive impinges on white ice or snow surfaces and is mostly reflected away. As a result, it is not stored as thermal energy in the ground or ocean, and therefore does not contribute to the warming of the air layers near the land surface or sea. In this way, the high reflectivity of the snow and ice surfaces reinforces cooling in the polar regions. This means that more sea ice is formed in response to the increasing cold, which in turn increases the total albedo levels. This then results in even more solar radiation being reflected. Climate researchers refer to such self-amplifying processes as positive feedback.

Water vapour – invisible regulator of heat

A third factor relating to the origin of cold climate at the poles is water vapour. Water is an extremely versatile element of our climate system. It can evaporate, condense and freeze, and it occurs in nature in three physical states: as a liquid (water), frozen (ice), and as a gas (water vapour).

This odourless and invisible gas is formed when liquid water evaporates. The Earth's atmosphere contains around 13 trillion cubic metres of water. This amount represents about 0.001 per cent of the accessible water on the Earth, whereby the largest proportion of water in the atmosphere is in the gaseous state. If all of the water vapour in the atmosphere were to condense and fall to the surface as rain, it would cover the entire globe with a layer of water about 25 millimetres thick. Still, the proportion of water vapour in the air by mass is on average only 0.25 per cent.

The Earth's heat and radiation balance

The Earth obtains its energy almost exclusively from the sun. With a surface temperature of around 5500 degrees Celsius, it emits a hundred thousand times more energy than the Earth, whose average temperature is around 15 degrees Celsius. The sun's energy impinges on the Earth's atmosphere as extra-terrestrial radiation. This radiation transmits energy in the form of electromagnetic waves. To understand the radiation balance of the Earth, one has to be aware of three physical laws: For one, every body, whether solid, liquid or gaseous, emits electromagnetic radiation as a function of its surface temperature. This is true for the glowing star that is our sun as well as for the Earth. Our home planet has a temperature of 288 degrees Kelvin (14.58 degrees Celsius), and thus also radiates independently without the help of the sun.

Secondly, the wavelengths of radiation emitted depend upon the temperature of the body. The hotter it is, the shorter the waves of the radiation released. The filament in an incandescent bulb, for example, becomes so hot that it glows and emits white light, i.e. visible radiation. When the light is turned off, the filament cools down and continues to glow slightly reddish for a moment. This indicates that its radiation maximum has shifted from the short-wave to the long-wave scale. But it would still be unwise to touch the bulb at this time, because even when the metal filament stops glowing and is no longer giving off visible radiation, it continues to emit invisible heat radiation in the infrared range. It is still hot enough to burn one's fingers. As a relatively cool body, our Earth emits only long-wave heat radiation in the infrared range.

Thirdly, the radiation emitted by one body can be reflected or absorbed by other bodies. This is also the case for the Earth-sun system. The global average solar radiation arriving at the outer edge of the Earth's atmosphere is 340 watts per square metre. About seven per cent of the incoming radiation is UV radiation, 46 per cent is in the range of visible light, and the remaining 47 per cent is in the infrared spectrum.

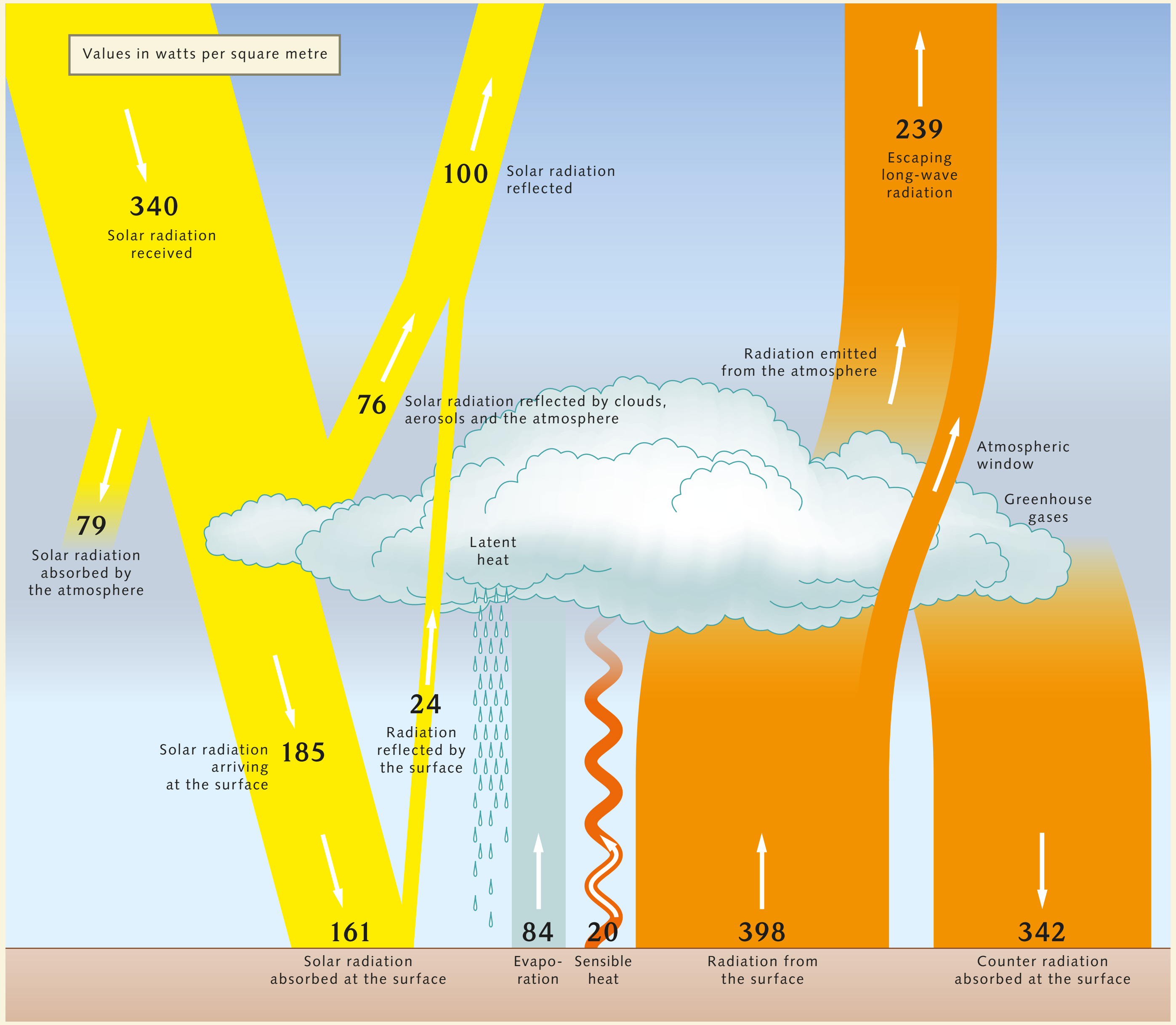
About 30 per cent of the 340 watts per square metre of incoming radiation is reflected directly back into space by the atmosphere and the Earth's surface. This amount, around 100 watts per square metre, is called the planetary albedo. Only 240 watts per square metre remain in the climate system and are available to warm the planet. About one-third of this amount (79 watts per square metre) is absorbed by the clouds, water vapour, dust particles and ozone in the atmosphere. This portion thus warms the atmosphere above us. The gases, in particular, filter within very specific wave ranges. Ozone, for example, absorbs almost all of the incoming ultraviolet radiation and is therefore a very effective UV filter.

It is important at this point to note that the air molecules, water droplets, dust particles and ice crystals present in the atmosphere reflect the incoming radiation in all directions. The portions deflected upwards go back into space, the rays deflected downwards reach the Earth as diffuse sky radiation. Two forms of radiation therefore arrive at the Earth's surface: the direct solar radiation, which only reaches the surface through a cloudless sky, and the diffuse sky radiation.

Almost 60 per cent of the incident solar radiation (185 watts per square metre) actually reaches the Earth's surface. Of this around 13 per cent (24 watts per square metre) is reflected directly back into space. This is a portion of the planetary albedo. The remaining 161 watts per square metre are absorbed by the Earth's surface and provide it with warmth.

The Earth's surface releases its heat in three ways: firstly, in the form of evaporation – called latent heat (84 watts per square metre), secondly through the rising of warm air masses – called sensible heat (20 watts per square metre), and thirdly by radiating long-wave heat rays (398 watts per square metre). However, only a very small amount of the heat radiation is actually lost directly into space. On its way through the atmosphere it collides with the same obstacles that previously hindered the incoming short-wave solar radiation. This time, however, it is primarily the molecules of the trace or greenhouse gases that absorb the long-wave radiation and ultimately emit it again in all directions as heat radiation. They thus trap part of the heat in the lower atmosphere and generate what is called counter radiation (342 watts per square metre). For this reason, the Earth receives a large portion of its emitted radiation back again.

This process is often referred to as the greenhouse effect. But at the same time, the atmosphere does radiate some heat back into space. Comparing the incident solar radiation with the total long-wave radiation emitted at the outer edge of the atmosphere clearly shows that the Earth absorbs slightly more energy than it releases. This fact is of crucial importance, as will be explained later.



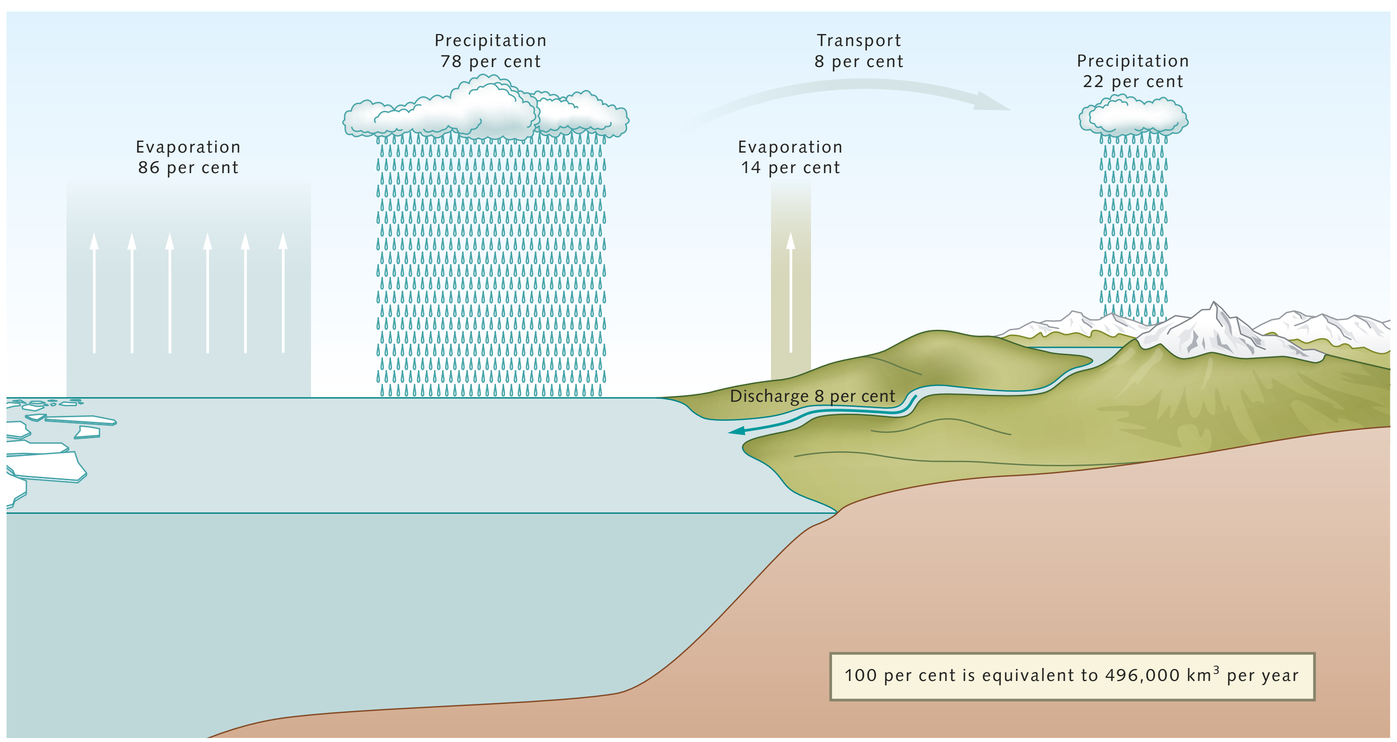
However, this average value is misleading because water vapour is distributed very unevenly throughout the atmosphere. Its concentration decreases rapidly with increasing elevation, due in part to the fact that warm air can hold more water vapour than cold air. Accordingly, large amounts of water can be converted into water vapour in warm regions and less in colder regions. In the polar regions, because of the low temperatures, evaporation and water-vapour content in the atmosphere are very low in winter. The water-vapour capacity of the atmosphere increases with every degree Celsius of air temperature. As an example, one cubic metre of air at a temperature of minus 20 degrees Celsius can hold at most 1.1 grams of water vapour. However, if this volume is heated to plus 20 degrees Celsius, it can contain a maximum of 17.2 grams of water vapour.

The amount of water vapour present in the atmosphere at a given time is commonly referred to as "humidity". When meteorologists report a condition of high humidity, this means that the air contains a large amount of water vapour. The most common measure used is relative humidity in per cent. Because a given volume of air at a given temperature and pressure can only hold a certain maximum amount of water vapour, we refer to a relative humidity of 100 per cent when this maximum amount is reached.

As a general rule, when water evaporates over the sea or on land, no more than ten days will pass before the water vapour leaves the atmosphere again in the form of precipitation. In contrast to carbon dioxide, which may be retained for several centuries, water vapour leaves the atmosphere rather quickly and it is thus referred to as short-lived. Nevertheless, water vapour is regarded as the most important natural greenhouse gas. Firstly, this is because it occurs in higher concentrations in the atmosphere than carbon dioxide, methane or nitrous oxide (laughing gas). Secondly, it contributes two to three times more to the natural greenhouse effect than does carbon dioxide.

The Earth's climate, and particularly the climate of the polar regions, is strongly influenced by the presence or absence of water vapour. The atmosphere has to contain water vapour before fog or clouds can form. However, the water vapour only **condenses** when the air is supersaturated with the gas, i.e. when it contains more water vapour than it can physically retain. This supersaturation occurs when warm humid air masses rise and are cooled, and thus lose their capacity to absorb more water vapour. The gas condenses into small droplets or, in certain circumstances, into small ice particles that waft freely in the air and commonly become visible from the ground as clouds or fog.

There are two ways in which clouds are important for the global climate. The billions of water droplets they contain refract sunlight from above, preventing these rays from striking the Earth's surface directly. Instead, they are deflected in many different directions. A certain portion even escapes back into space. Ultimately, therefore, less solar radiation reaches the ground than it would if there were no cloud cover. As a consequence, the cloud cover effectively cools the Earth. On the other hand, however, clouds also block the long-wave heat radiation rising from the Earth. They absorb a large portion of this heat radiation and release the heat again in all different directions. In this way clouds can also contribute to warming in the atmosphere. Which of the two features is dominant depends upon the type of cloud. Clouds are most commonly differentiated based on their altitude and form. Visibly thick, low-hanging clouds primarily reflect the incoming sunlight and cool the Earth. High thin clouds, on the other hand, let the solar radiation through. They subsequently block the outgoing heat radiation from the Earth and absorb a large portion of the thermal energy. The day-night effect also plays a role. Obviously, a cloudless sky usually means warmer temperatures during the day because the sun's rays are unobstructed. But it becomes cooler at night with no clouds because the Earth's absorbed heat energy can be radiated outward again unhindered.



2.5 > Over the oceans large amounts of water are constantly evaporating, especially in the warmer ocean regions. The moisture does not remain in the atmosphere for long, however. Within ten days it falls to the Earth again as precipitation.

Freeze-dried air

The Arctic and Antarctic are fundamentally different with regard to the influence of clouds. While dense fog and cloud cover are phenomena often observed during the summer in the Arctic – much to the dismay of polar explorers who usually plan their expeditions for the summer – in Antarctica they normally only form in coastal areas. The air above central Antarctica is simply too cold due to the limited amount of solar radiation, and therefore contains too little water vapour for condensation to form a thick cloud cover. Instead, with increasing cold, all of the residual moisture condenses into ice crystals and falls to the ground as a form called diamond dust. The air is thus essentially freeze-dried, which is why Antarctica is considered to be the world's driest continent.

For comparison: In Germany around 700 litres of precipitation per square metre fall each year. The same amount is also recorded at the weather station on the Antarctic Peninsula. In the coastal area of the Weddell Sea, i.e. near the German Antarctic research station Neumayer III, there are only 300 litres of precipitation per square metre, which is equivalent to a layer of snow about one metre thick. In central Antarctica, on the other hand, annual precipitation rates are less than 50 litres per square metre over vast areas because of the extremely dry air. Only under exceptional conditions have meteorologists reported a thin veil of clouds over the Antarctic Ice Sheet. However, these are not substantial enough to prevent the ice surface from radiating the small amount of incident heat back into space, which leads to further cooling of the air above Antarctica.

In the Arctic, on the other hand, water vapour, clouds and fog can promote warming, especially in summer. One reason for this is the shrinking of the sea-ice cover in the Arctic Ocean during the summer. The white ice floes, drifting in the winter and spring and reflecting a large portion of the sun's radiation, are partially replaced in summer by the much darker sea surface. This absorbs up to 90 per cent of the sun's energy, which causes a rise in the sea-surface temperature. Because this is accompanied by a corresponding warming of the air, the atmosphere can absorb more moisture. The humidity increases, so that only small soot, dust or salt particles in the air are required for the water vapour to condense and form clouds or fog.

In addition to the fact that clouds can be formed from it, water vapour possesses another property that is significant for the heat balance and weather patterns: It stores heat energy. This heat cannot be detected by a thermometer or felt by humans. Meteorologists therefore refer to it as latent heat. It is sometimes referred to as evaporation heat because its value corresponds precisely to the energy originally required to evaporate the water. What is special about the heat storage of water vapour, however, is that as soon as the vapour condenses into water droplets in the atmosphere, the stored heat from evaporation is released again as condensation energy and warms the surrounding air. In regions with high atmospheric water-vapour content, this effect causes additional warming. In areas with low humidity or little water vapour in the atmosphere, this effect is much less significant.

In some small depressions on the southern slope of the East Antarctic Ice Sheet, the paucity of water vapour is one of the reasons that it can get even colder than it does at the Vostok Research Station. In July and August, the air layer directly above the ice sheet becomes so cold that, according to scientific reckoning, it cannot become any colder. Minus 98 degrees Celsius seems to be the coldest temperature possible on the Earth under natural conditions.

For the air in the depressions to become this cold, a number of conditions must be met. Incoming solar radiation has to be absent for several weeks, which can only occur during the polar nights. Furthermore, the air above the snow-covered ice sheet may not contain any water vapour that could give off heat in the case of condensation, or could absorb radiation energy reflected from the snow and then be held in the atmosphere. According to researchers, the air in the region contains so little water vapour in winter that, considered as a water column, its height would be just 0.04 to 0.2 millimetres. Ideally, the water-vapour content has to be less than 0.1 millimetres. Additionally, the wind must be extremely weak and the sky free of clouds for several days.

Under these conditions, the layer of air directly above the snow cools down stepwise. It becomes denser and heavier, slowly flows downslope, and collects in the depressions where researchers have been able to detect it from satellites.

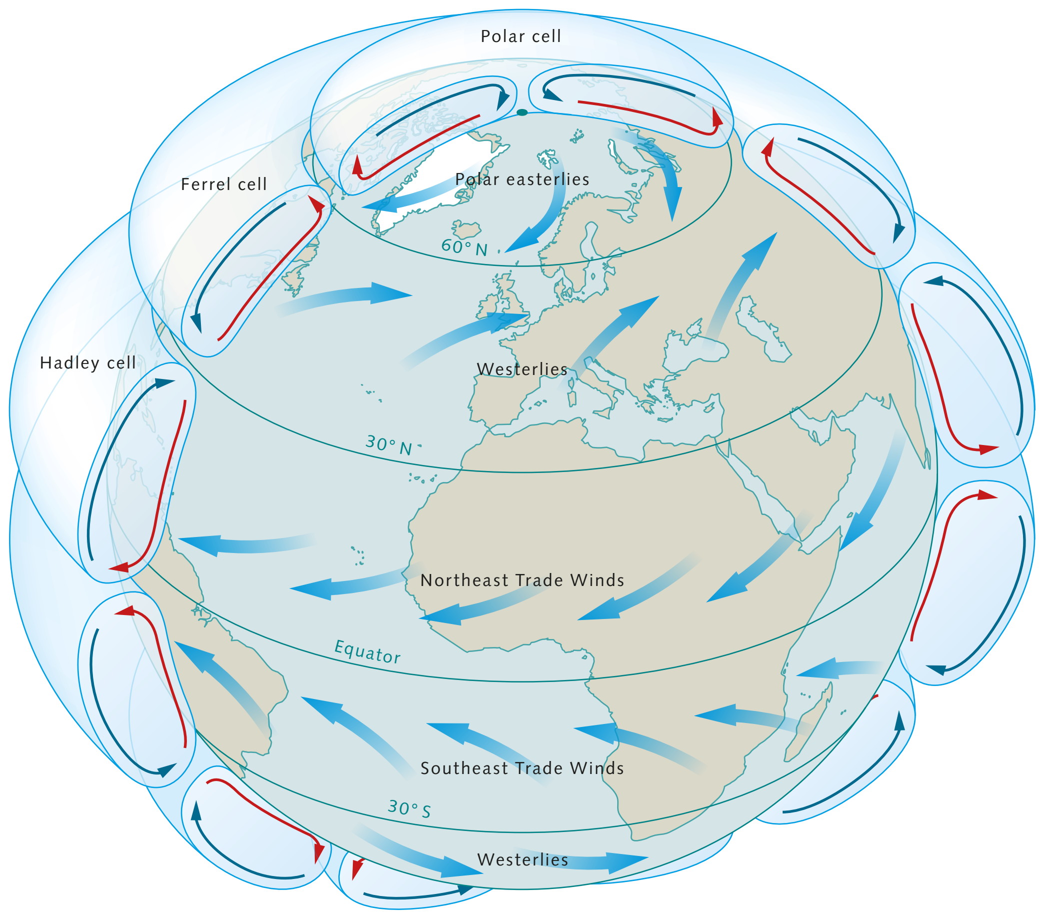
Winds – the driving forces of weather

Looking at the polar regions through the eyes of a physicist, the Arctic and Antarctic are regions where the lack of solar radiation and the high proportion of heat reflection due to the albedo effect result in temperatures that are lower by far than in other regions of the world. Temperature differences are accompanied by density differences; cold air masses are denser and thus heavier than warmer ones. Cold air sinks and warm air rises. These density-differences and resulting air motions are generated by differences in the atmospheric pressure at different locations. Where air cools down and sinks, a high-pressure area develops near the ground, a phenomenon known in both the central Arctic and the Antarctic as a polar or cold high. In low-pressure areas like the tropics, by contrast, warm air rises.

These atmospheric temperature and pressure differences between the warm tropical and the cold polar regions are the true "weather generators" of the Earth. They drive the large wind and current systems of the Earth and thereby also global air circulation. All of the processes in the atmosphere are geared toward equalizing these temperature differences and pressure contrasts. This means that the warm air masses from the tropics migrate poleward at high altitudes, while the cold air masses from the polar regions flow towards the equator closer to the ground.

If the Earth were not rotating on its axis, the paths of the different air masses on a map might be seen as straight lines both near the ground and at higher altitudes. But because the Earth is turning, every air current travelling from a high-pressure to low-pressure area is diverted to the right in the northern hemisphere and to the left in the southern hemisphere. This effect is caused by the Coriolis force – an apparent force arising from the rotation of the Earth. It affects both air and ocean currents, increases with latitude, and is the reason why, for example, the trade winds in the northern hemisphere do not travel in a straight line directly southward towards the equator from the high-pressure area at 30 degrees North. Instead, they are deflected to the right with respect to their flow direc-

2.6 > In late Antarctic spring, the sun rises above the horizon again and marks the end of polar night in Antarctica. In most of the coastal regions of the southern continent this lasts about two months. The nearer one moves to the South Pole, however, the longer the period of darkness lasts.



2.7 > Atmospheric circulation of the air masses surrounding the globe is so complex that researchers sometimes use this highly simplified model. It shows the six circulation cells and wind systems that are formed by deflected air-mass currents, and that are almost identical in the two hemispheres.

tion and therefore sweep across Africa and the Atlantic as northeast winds.

The Coriolis force is also the reason why air high above the equator moving towards the poles is more strongly deflected to the right with every metre that it travels. At about 30 degrees of latitude this deflection is so great that the air current flows parallel to the lines of latitude as high westerly winds, or the subtropical jet stream, and it can no longer perform its actual task of heat balance between the equator and the pole. Instead the air, which has become cooler by this time, descends and flows back toward the equator near the surface as trade winds. This is how the trade-wind circulation, also called Hadley circulation after the British hobby meteorologist George Hadley (1685–1768), is generated between the equator and 30 degrees latitude, both in the northern and southern hemispheres. In simple, idealized descriptions it is referred to as a closed cell.

A very similar circulation pattern, called the polar cell, forms above the northern and southern polar regions. Here, due to cooling, the cold heavy air in the centre descends (high-pressure area) and then flows near the surface toward the polar circle (low-pressure area). Again, the Coriolis force acts so that the air currents become polar easterly winds. On its way towards the Arctic Circle, however, the air warms up enough to rise and return to the pole at high altitude as a counter-current.

Located between the two homologous systems of the Hadley and polar cells, there is room for a third system, called the Ferrel cell, named after the American meteorologist William Ferrel (1817–1891). In this, the air masses circulate in the opposite direction. This means that the near-surface air here is transported towards the pole and is deflected to the right (in the northern hemisphere), so that the winds blow from the west. This zone is therefore referred to as the prevailing westerlies. Unlike the polar cell or the Hadley cell, however, turbulence in the air masses of this zone produce low-pressure cells, which wander back and forth as waves and cause some degree of instability in the circulation. This instability is due to the large temperature contrast between the tropics and the polar regions, which cannot be balanced directly because of the strong Coriolis force. Instead, nature makes use of high- and low-pressure areas, which, like a paddle wheel, shovel warm air to the north and polar air to the south on the opposite side of the depression. For this reason, what meteorologists typically refer to as weather only occurs in the area of the Ferrel cell. In the other cells, the seasons rather than the weather tend to determine meteorological events.

In laboratory experiments scientists have tested how the atmospheric circulation would be different if the planet rotated faster on its axis. The results indicate that the increased Coriolis force would cause five cells to form on each hemisphere. If the Earth were to rotate much more slowly, there would be only one cell between the pole and equator, in which the air masses would then flow directly from the tropics to the poles and back. This kind of single-cell circulation is found, for example, on Venus.

The protective vortices

The wind and current patterns of atmospheric circulation are of great importance for the polar regions. So far, they have been very reliable in preventing warm air masses from reaching the centres of the Arctic or Antarctic regions. In order to understand how the winds protect the polar regions, we have to take a somewhat closer look at the atmosphere in the polar areas.

The air above the polar regions cools down drastically in the autumn and winter (polar night) and descends to the Earth's surface. Whereas a high-pressure area, the polar high, forms near the surface, higher up at an altitude of eight to ten kilometres an area of depression is created. This can extend to an altitude of up to 50 kilometres, and is known as the stratospheric polar vortex. The air masses of this vortex are surrounded and held together by a strong westerly wind called the polar night jet. This develops because the same principles of current flow apply in the stratosphere as in the underlying troposphere.

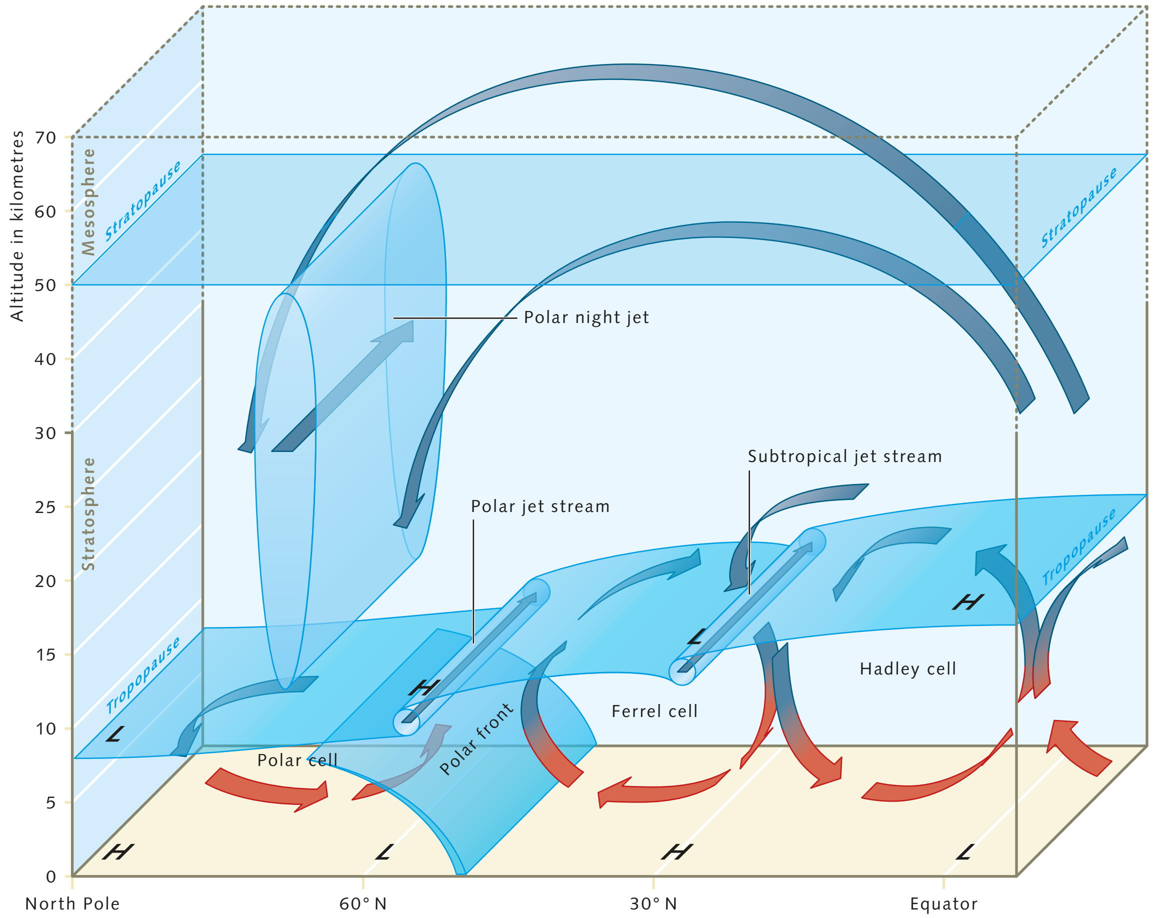
That is to say, air always flows from a high-pressure area, in this case the upper-level high-pressure area above the equator, to a low-pressure area, here the upper-level low-pressure area above the polar regions. However, the air current moving toward the pole is again deflected to the right because of the Earth's rotation, which for the northern hemisphere results in the creation of a westerly wind. The polar night jet is thus located at an altitude above ten kilometres and blows from the west towards the east completely encircling the North Pole. The wind attains its highest velocity at a latitude of about 60 degrees. Here it forms a kind of barrier that isolates the upper-level polar low-pressure area from the air masses coming from the equatorial region, thus preventing the high-altitude warm air masses from advancing further toward the pole.

During the course of the winter, the polar night jet gains strength because, with increased cooling in the stratosphere, more of the air masses in the low-pressure area continue to descend enabling more air to flow in, which boosts the wind strength. However, as soon as the first rays of the sun reach the polar region in spring, the air in the low-pressure area warms up. The differences in density and pressure equilibrate and the wind weakens again.

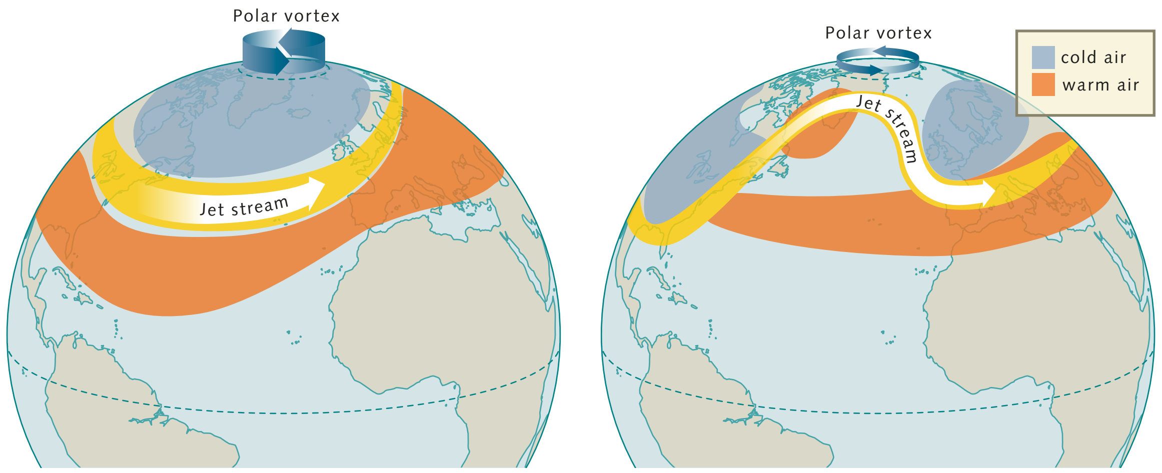
Comparing the stratospheric polar vortex in the Arctic with the one in the Antarctic, it is notable that the wind in the south maintains a more circular path and is significantly stronger than in the high north. On a normal winter day, the winds of the Antarctic polar night jet can achieve velocities of up to 80 metres per second. This is equal to 288 kilometres per hour. By contrast, in the northern hemisphere they blow with an average velocity of only 180 kilometres per hour. In addition, the stratospheric polar vortex over the Antarctic is significantly larger and the temperatures in its interior are lower than those in the north.

The fact that the two stratospheric polar vortices are so different in character is partly due to Rossby waves, which are also referred to in the special literature as planetary waves. These are large masses of air in the troposphere that meander around the globe with the westerlies. This kind of air mass possesses a certain vorticity (spin strength) due to the Coriolis force, which is related to its own rotation. The rotational speed of the air mass depends on the geographical latitude along which it is moving, because the Coriolis force becomes stronger with increasing distance from the equator. Air masses forming at higher latitudes basically rotate faster than those at lower latitudes.

If such a rotating mass of air encounters a mountain or high plateau during its meandering, such as the Rocky Mountains in the USA, or the Urals in Russia, or the high ice sheet of Greenland, it is deflected upwards by the obstacle. When this happens the air mass rises and also forces the overlying air masses to higher altitudes. With this ascent, the vorticity of the air mass changes and it is deflected towards the equator. Here the distance to the Earth's axis is greater than on the original meander path of the air mass. The vorticity of the air mass no longer matches the latitude-dependent vorticity at its new location. As a result, the direction of movement of the air mass changes back toward the pole. It overshoots its original geographical latitude in the opposite direction, then turns back again due to the opposing effect. It thus establishes an oscillating pattern.



2.8 > The polar night jet is a band of wind in the stratosphere that feeds off air masses flowing at high altitudes from the equatorial region towards the high northern latitudes. The polar jet stream, on the other hand, meanders through the troposphere, one atmospheric level lower.



2.9 > The two faces of the polar jet stream: When the polar vortex rotates at full strength in the stratosphere, the winds in the troposphere blow parallel to the equator and effectively block warm air masses from advancing into the Arctic. But if the polar vortex is weaker, the course of the jet stream begins to meander. As a result, cold polar air penetrates southward over North America and Siberia and moist, mild air over the North Atlantic migrates into the Arctic region.

A mass of air that was originally located at 50 degrees north latitude may fluctuate back and forth within the troposphere between 40 and 60 degrees, its path defining a waving line that snakes around the entire globe. This is the Rossby wave, named after the US American meteorologist Carl-Gustaf Arvid Rossby (1898–1957).

Because the Rossby wave also spreads upwards, under certain conditions its effect may extend into the stratosphere and disrupt the polar vortex to such an extent that it is weakened or even collapses completely. If this natural barrier is eliminated, warm air from the mid-latitudes can flow in and lead to a rapid warming of the stratospheric polar region. In the Arctic, scientists observe such a surge and the associated sudden rise of temperature in the stratosphere about every two years. The presence of mountains as well as the stark temperature differences between land and sea surfaces in the northern hemisphere are conducive to the formation of strong planetary waves.

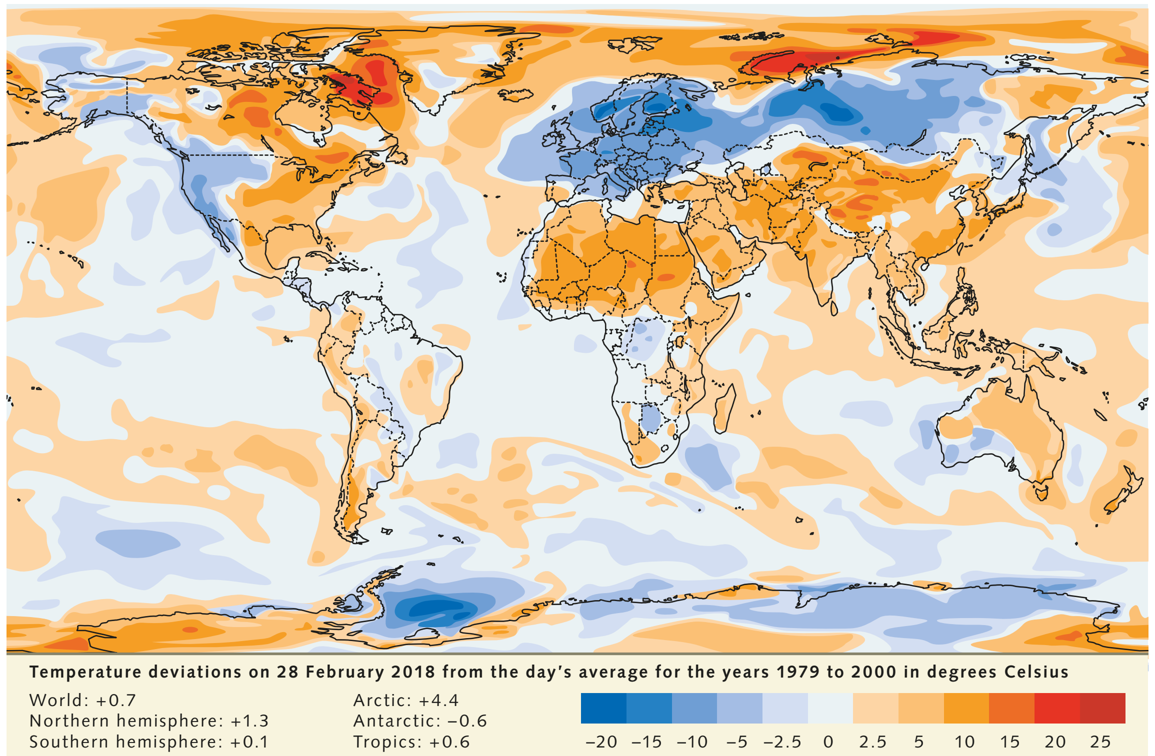
So far, however, researchers have not been able to predict which waves can be destructive to the stratospheric polar vortex or when a surge can be expected. In the southern hemisphere, in contrast, there are no extremely high mountains except for the Andes. Moreover, large portions of the southern hemisphere are covered by seas, which impede the formation of planetary waves. Since observations began there has only been one case of an abrupt warming of the Antarctic winter stratosphere. That was in September 2002.

A wall of wind

Besides the stratospheric polar vortex and its associated polar night jet, the polar regions possess another atmospheric protective shield called the tropospheric polar vortex. This also has an accompanying strong wind that is variously referred to as the polar front jet stream, polar jet, or simply the jet stream. Depending on the prevailing weather conditions, this belt of wind lies about eight kilometres above sea level between the 40th and 60th degrees of latitude, i.e. above the zone of westerly winds. In the northern hemisphere the wind speeds here can range from 200 to 500 kilometres per hour and they are sustained throughout the year.

As long as the northern jet stream blows around the Arctic at full strength parallel to the equator, it prevents warm, moist air masses from the south from moving into the northern polar region, just as the polar night jet does in the stratosphere. At the same time, it blocks cold Arctic air from advancing into the mid-latitudes. The jet stream is constantly affected by the Rossby waves. These ultimately contribute to our weekly changing weather conditions, as they are sometimes stronger, sometimes weaker and change their position. The high- and low-pressure areas that define our weather form in the peaks or valleys of the waves.

In winter, the strength and durability of the jet stream in the troposphere depend in part on the stability of the



2.10 > Exceptional atmospheric situation: At the end of February 2018 there was an unusual heat wave in the Arctic. The polar vortex split, causing the jet stream to lose strength, and allowing the penetration of warm air far into the Arctic. In the Labrador Sea and in the Siberian area of the Arctic Ocean the air temperatures rose to as high as 15 degrees Celsius above normal. At the same time, Central Europe was experiencing spells of extreme cold.

polar vortex in the overlying stratosphere. Rossby waves, which can destroy the stratospheric polar vortex and cause an abrupt warming of the polar stratosphere, also change the jet stream in the troposphere. The wind in the troposphere weakens and assumes a meandering course over the northern hemisphere. The result is that over North America and northern Europe the tropospheric vortex expands to the south, and cold polar air can penetrate deeper into North America and Central Europe. Over East Greenland it retreats back to the far north, allowing damp, warm air to migrate into the Arctic region.

In February 2018, for example, the northern hemisphere experienced this kind of exceptional atmospheric situation. At that time, Rossby waves were able to split the polar vortex, which caused a rapid warming of the stratosphere above the Arctic region to temperatures as high as 50 degrees Celsius. This, in turn, caused a weakening of the polar front jet stream in the underlying troposphere, which had far-reaching impacts. While Central Europe suffered from extreme cold in February, with snowfall even in Rome, mild spring temperatures were predominant in the Arctic in spite of the polar night. In Siberia the temperatures at times reached values up to 35 degrees Celsius above the normal average for February. The weather station at Cape Morris Jesup, the northernmost point of Greenland, recorded ten winter days in a row in which the temperature did not drop below the freezing point. And off the west coast of Alaska, one-third of the sea ice that is normally present at this time of year melted within a period of eight days.

Home of the blizzards

Antarctica is not only the coldest continent in the world, it also tops the list of windiest regions. At the French research station Dumont d'Urville, for example, scientists recorded a peak wind speed of 327 kilometres per hour in July 1972. This is more than double the strength of hurricane winds. A hurricane is defined as winds with average speeds of 120 kilometres per hour or more.

2.11 > The French research station Dumont d'Urville on Pétrel Island in Adélie Land is one of the windiest places in Antarctica. Particularly in winter, air masses that have cooled over the East Antarctic Ice Sheet gush down onto the sea as descending (katabatic) winds.

The documentation of such high wind speeds in the coastal region of Antarctica is not a coincidence. Apart from the global wind systems, the icy continent produces its own local wind system, which forces researchers to remain confined inside their stations, especially in winter, and which is largely responsible for the formation of sea ice in the Southern Ocean.

Winds normally develop when air masses flow from a high-pressure area into a low-pressure area to compensate for the difference in pressure. However, an air mass can also begin to move due to its own weight – for example, when it becomes colder and heavier than the surrounding air masses and sinks as a result. The near-surface air layer above the Antarctic Ice Sheet is particularly dense and heavy due to its altitude, low solar radiation input and high radiational cooling from the ice. The cooled air masses form a heavy 300-metre-thick layer above the central ice sheet. Because the ice sheet does not have a flat surface, but falls away at the edges, this extremely cold and heavy air from Central Antarctica at some point begins to slide down the slopes toward the coasts. It gains velocity exclusively through its own weight and the steepness of the slope.

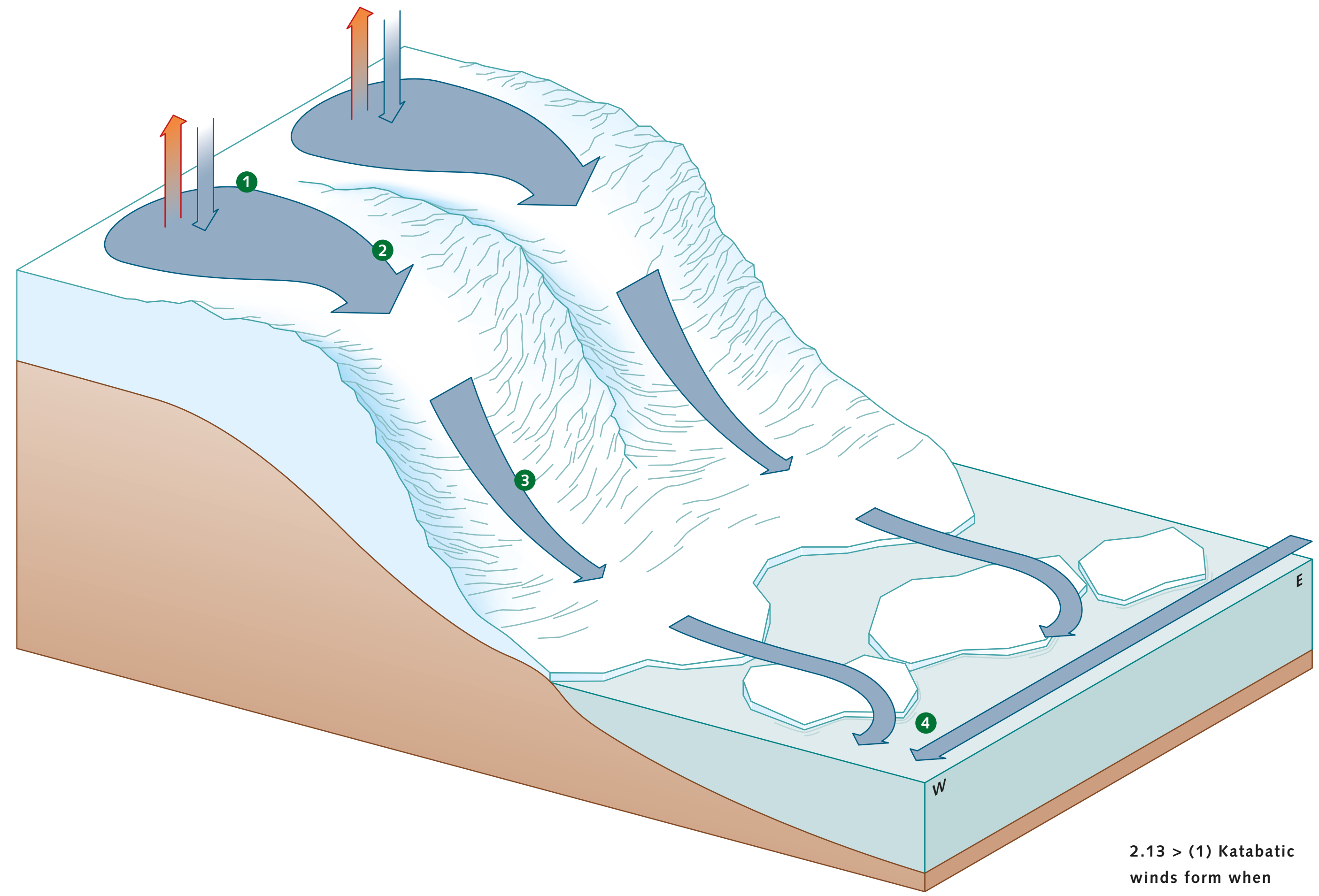
However, the cold rushing air really begins to gain speed when its path toward the coast is partially blocked by mountains. When this happens the total air mass has to squeeze through narrow valleys, which accelerates the air current enormously. In extreme cases, when these winds, called katabatic winds, reach the coast they can attain storm or even hurricane strengths. This technical term derives from the Greek prefix *kata*, which means "descending" or "downwards".

Polar researchers report that katabatic winds can arise unexpectedly from nowhere. At one moment the working conditions on a glacier or ice shelf can be totally windless and five minutes later, with no warning, a hurricane can sweep across the ice and lead to a condition known as whiteout.

Just as abruptly, however, the wind can again subside – depending on the area of the ice surface above

Whiteout
Polar scientists refer to "whiteout" as the condition where fog, clouds or a snowstorm restrict visibility to such an extent that neither contours nor the horizon can be recognized in any direction.

2.12 > A Piteraq blows through the Greenland settlement of Tiniteqilaaq. Gusts of this katabatic wind can reach speeds of up to 300 kilometres per hour.



2.13 > (1) Katabatic winds form when near-surface air above an ice sheet cools and thus becomes denser and heavier. (2) The air mass then slides down the slope due to its own weight, (3) gains speed along its way through narrow valleys, and (4) pushes loose ice floes off the coast into the sea, where the air mass is deflected and slowed by coastal winds.

which the cold air formed. In extreme cases, however, this kind of wind can persist for several days with sustained high velocities.

These exceptional winds occur most notably in the coastal region of Adélie Land, the windiest part of Antarctica and location of the French research station Dumont d'Urville. This is due to the topography of the region. Here the cold air from a large area of Eastern Antarctica flows down the ice sheet, and newly formed cold air masses can flow down at any time, especially during the winter. In addition, mountains channel the air currents through narrow valleys, which amplifies their strength each time.

Katabatic winds, incidentally, also occur in regions outside Antarctica – for example, at the margins of the vast Greenland Ice Sheet, where the near-surface air layer above the high plateaus is cooled to a temperature between minus 20 and minus 40 degrees Celsius in winter. Greenland's strongest winds occur on the southeast coast, in the region near the town of Tasiilaq. Storm winds can gust at speeds up to 300 kilometres per hour, and because of the great danger they are called "piteraq" by the natives, which in their local language means „that which attacks you".

On 27 April 2013 a wind like this not only blew snow from large areas of the ice sheet. Blasting through the 85-kilometre-long Sermilik Fjord, the piteraq also pushed all of the sea ice and glacier ice floating in the fjord out into the sea, so that the fjord was virtually ice-free after the storm.

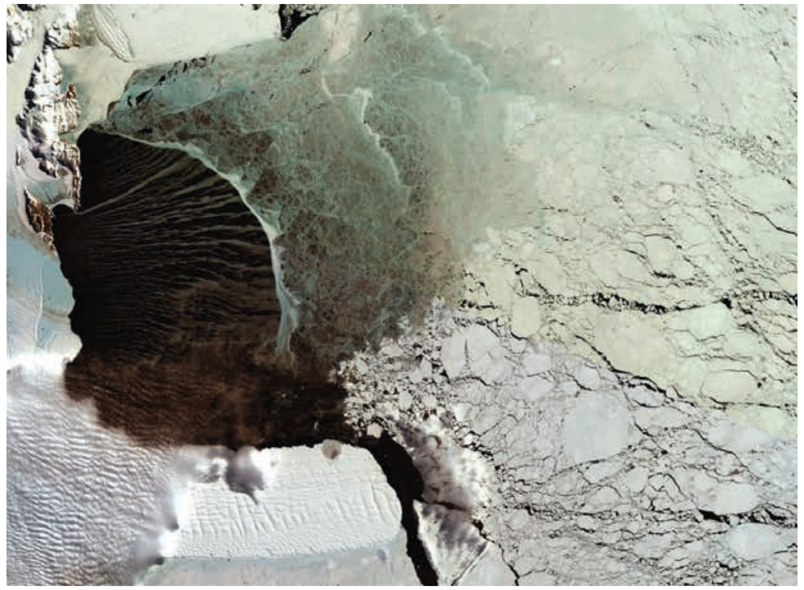
Ice floes, ice sheets and the sea

> There are large areas in the polar regions where water occurs predominantly in its frozen state. It either falls as snow to contribute to the growth of ice sheets and glaciers, or it drifts on the sea as ice floes. In both cases the fate of the ice depends largely on the ocean and its currents. The water masses can provide protection or accelerate melting, depending on the path that heat follows.

Sea-ice nurseries

When strong winds in the Arctic and Antarctic regions force icebergs and sea ice away from the coasts and out to sea, areas of open water remain where air and water are in direct contact with each other. These areas are called coastal polynyas, and they are the places where sea ice is created. Scientists sometimes refer to them as ice factories. Especially in winter, when the air temperature sinks far below zero degrees Celsius and offshore winds blow constantly, sea ice is produced in the Arctic and Antarctic polynyas in assembly-line fashion.

Sea-ice production follows the same routine everywhere. First, frigid winds cool down the areas of open water so intensely within a short time that the surface freezes over. Because seawater contains salt, its freezing point lies below zero degrees Celsius. In the Arctic and Antarctic, seawater has to be cooled to minus 1.9 degrees Celsius before the first ice crystals begin to form. For comparison, in the Baltic Sea, where the salinity is lower, the water begins to freeze at minus 0.5 degrees Celsius.



2.14 > Terra Nova Bay in the Antarctic Ross Sea is an example of a coastal polynya. Sea ice is created there. The light streaks are a result of the cold offshore winds blowing over the water from the Ross Ice Shelf, forming an icy sludge on the surface. This freezes to form thin ice, which is then pushed offshore by the wind where it ultimately develops into pack ice.

The first sea-ice crystals look like small, delicate needles or discs. They increase in number as more heat is removed from the water. At this point, the new ice resembles a fine slurry of needles and discs. Under calm wind conditions, a contiguous cover of thin ice forms from this still relatively transparent ice sludge. In the presence of strong winds, however, a typical structure called pancake ice forms in the wake of the rolling waves. This is composed of round, plate-sized ice slabs whose edges are curved slightly upward as a result of the wave impact. The ice therefore actually looks like an agglomeration of freshly baked pancakes before it freezes into a thin ice cover.

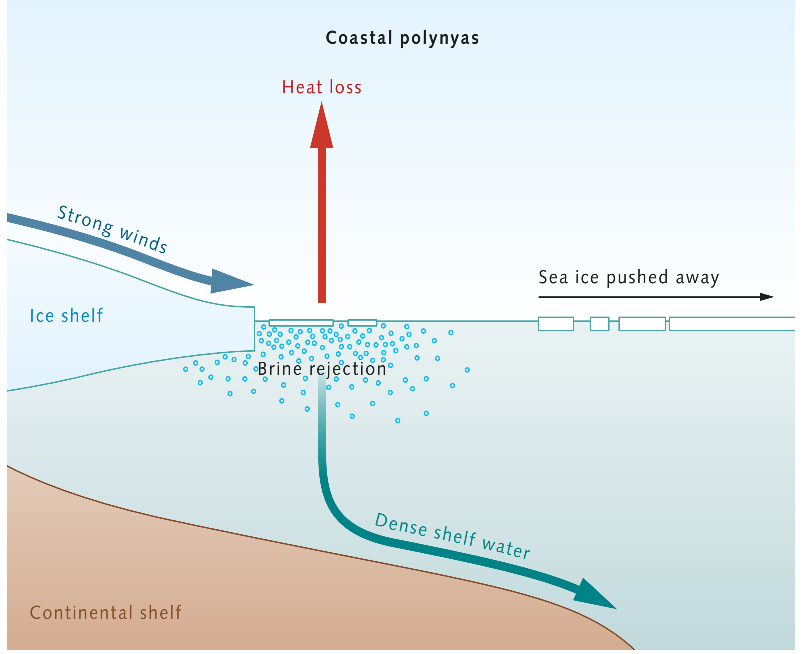
This young sea ice has a distinctive quality that distinguishes it from ice cubes, or from the ice on frozen freshwater lakes. Instead of forming a compact solid ice block, it is interspersed with small channels and cavities. The salt contained in seawater collects in these spaces because it cannot be incorporated into the lattice structure of the ice crystals during the freezing process. Instead, the salt flows as a highly concentrated brine through the small cavities, eventually seeping into the sea on the underside of the ice.

Because the density of frozen water is lower than that of liquid water, ice always floats on the surface. In the polynyas it thus presents a solid face above the sea surface for the wind to strike. The wind propels the young, thin ice out to sea where it collides with the older and thicker pack ice that is already floating off the coast. This compresses the young ice. It becomes thicker, breaks into individual floes depending on the conditions, and is pushed out ever further by the wind. Meanwhile, ice production in the coastal area of the polynyas starts over again from the beginning.

The coastal polynyas in the Antarctic can be from ten to a hundred kilometres wide, whereby scientists are not in complete agreement as to whether the term "polynya" should only refer to the completely ice-free water surface, or if the zone with new thin ice should be included. Satellite surveys show that Antarctic polynyas are almost completely frozen over in winter. The only exception, depending on the location of the polynya, is a strip of water about one kilometre wide directly off the coast or ice-shelf margin, which is kept free of ice by the offshore winds. Similar observations have been made in the Arctic. When air temperatures there drop to minus 40 degrees Celsius in the winter, the shallow-water polynyas (water depths less than 50 metres) off the coast of Siberia freeze up so quickly that a strip of water only a few hundred metres wide remains free of ice due to the wind. But as spring approaches the air becomes warmer. The surface waters are not cooled as intensely, and they freeze more slowly. Because the wind can now push the ice further out to sea, the polynya expands again to a width of several kilometres.

The most productive sea-ice factories in the Southern Ocean are the polynyas off the Ross Ice Shelf (producing 253 cubic kilometres of sea ice per year), the Cape Darnley polynya in the East Antarctic (127 cubic kilometres), and the polynya offshore of the Mertz Glacier (125 cubic kilometres). The Arctic Ocean sea ice is mostly formed in polynyas off the Siberian coast. The main suppliers of new ice are the Russian shelf seas, especially the Kara and Laptev Seas. This new ice is transported towards the Fram Strait by the wind and by the transpolar drift. But some sea ice is also produced off the coasts of Greenland and North America. However, because the wind on many segments of these coasts blows onshore instead of offshore, it pushes the sea ice toward the coasts, where it can become particularly thick.

Sea ice can also become thicker when seawater freezes on its underside. However, this can only happen when a sufficient amount of heat is somehow dissipated from the water on the underside of the ice into the atmosphere.



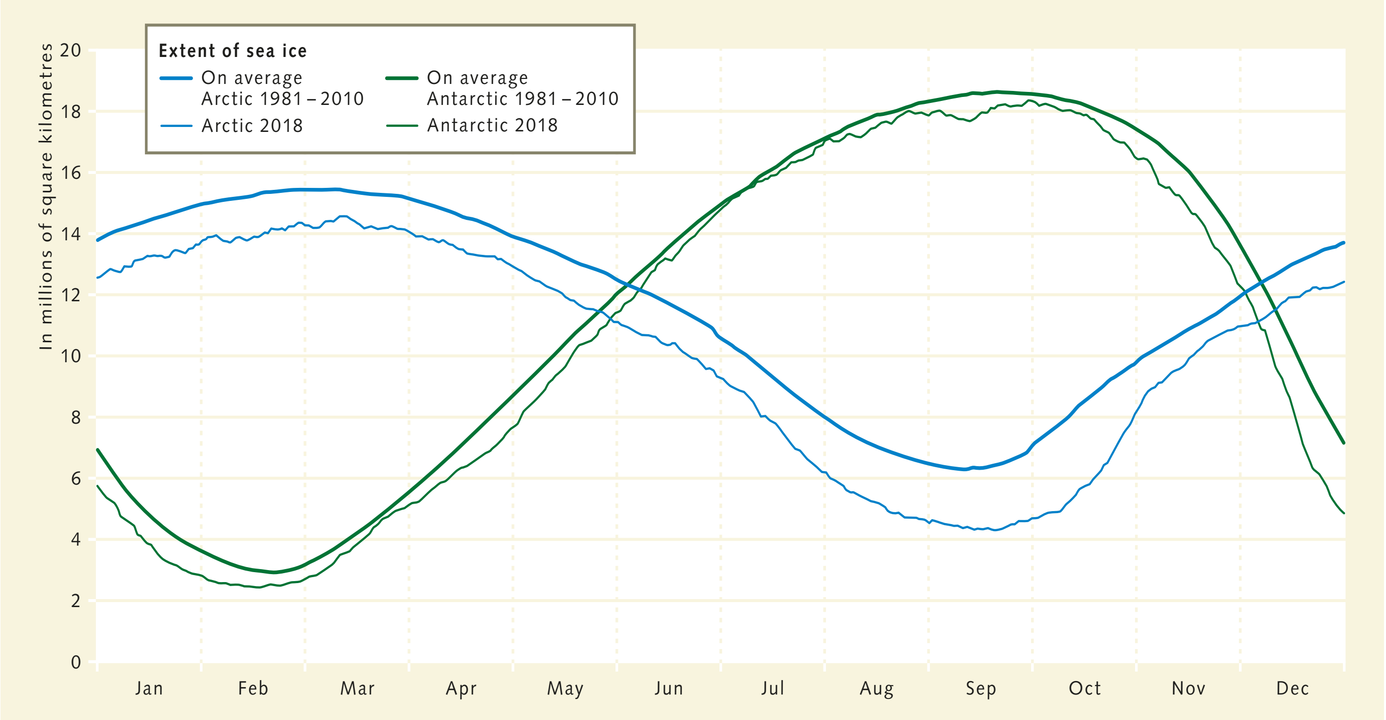
2.15 > When sea ice forms in Antarctic coastal polynyas, heat dissipates into the atmosphere and brine is released, creating cold, heavy water masses. These sink to greater depths where they feed the deepest layers of the world ocean as Antarctic Bottom Water or Circumpolar Deep Water.

This process is called thermodynamic growth. The heat lost by the water has to actually be transported through the ice from below and up into the atmosphere. This heat transmission works very well initially, when the ice is still relatively thin. Arctic sea ice, for example, can grow to a thickness of one to two metres within a single winter. As the ice becomes thicker, however, heat conduction is less effective and the ice floes grow more slowly. Thick, multi-year pack ice therefore acts in a way similar to a lid on a cooking pot: it inhibits the heat in the sea from escaping upward into the atmosphere.

Location matters

Adding all of the sea-ice areas in the world will give an annual average total of around 25 million square kilometres, which is about two-and-a-half times the area of Canada. The global distribution of this total sea ice, however, is not limited to the Arctic and Antarctic regions. During particularly cold winters the sea also freezes off the coast of China. From the Bohai Gulf, for example, thin ice floes drift on the sea at least for a short time to as far south as 38 degrees latitude. The distance to the equator

2.16 > Calving on the front of Lamplugh Glacier in the US state of Alaska. When ice masses weighing many tonnes break off and fall into the sea, fountains of water soar upward. When the blocks melt, the resulting water mixes with the surface water of the sea, reducing its salinity and density.



2.17 > The extent of sea ice expands and shrinks with the changing seasons both in the Arctic and Antarctic, whereby a greater proportion of sea-ice in the Southern Ocean consistently melts than does the ice cover of the Arctic Ocean.

from here is shorter than to the North Pole. In the southern hemisphere, seawater only freezes in the regions south of 55 degrees latitude. Due to the geographical situation (a large ocean surrounding a continent), Antarctic sea ice differs in a fundamental way from sea ice in the Arctic Ocean (continents surrounding a small ocean). It drifts much faster, for example, in part because the vast Southern Ocean offers more open area for the ice, allowing it to move more freely than it can in the Arctic. For the same reason, however, the sea ice in the Southern Ocean is less likely to form metres-high pressure ridges. These walls of ice, often kilometres-long, are common in the Arctic because the wind piles up the densely packed ice floes to heights of 25 metres or more, particularly near the coasts. These thick barricades of ice represent impregnable barriers, even for modern icebreakers.

Because of the great breadth of the Southern Ocean, the air masses coming from the northwest are also able to collect a great deal of moisture on their way to the Antarctic continent. This ultimately falls along the coast of the Antarctic continent, which is why the Antarctic sea ice is often covered by a thick blanket of snow. In the Arctic, by contrast, the incoming air masses pass over large areas of land before reaching the ocean in the centre. This air is therefore relatively dry and rarely produces snow.

Because ice floes in the Southern Ocean have room to drift into areas of warmer water, almost all of the freely moving ice in the Antarctic melts during the summer. The only exceptions are the sea-ice areas that are frozen along the coastline, or areas where icebergs have run aground and block the path of the sea ice to the open sea. This grounded sea ice, which is found both in the Arctic and Antarctic, is also called land-fast ice or bay ice. It provides, among other things, resting places and nursery areas for seals and penguins.

The migration of Antarctic sea ice to warmer northern latitudes means that it rarely survives for more than a year, and it is therefore thinner on average than Arctic sea ice. The ice cover in the Southern Ocean is usually one to two metres thick. In the Arctic, on the other hand, scientists often measure thicknesses of four to five metres, especially in regions with multi-year ice.

New sea ice is mostly produced during the winter months. Sea ice in the Arctic Ocean attains its greatest areal extent in March. When satellite measurements began in 1979, it amounted to slightly more than 16 mil-

2.18 > Researchers have set up camp on an ice floe to investigate the meltwater ponds. These often form on ice floes in the Arctic where meltwater collects. Because the dark water surfaces absorb more solar energy than the sea ice around them, the ice below the ponds melts especially rapidly.

lion square kilometres – an area about one-and-a-half times the size of the USA. This figure has since dropped to around 14.5 million square kilometres. The Antarctic sea ice freezes between March and September. At the end of the Antarctic winter it covers an area that averages more than 18 million square kilometres.

When spring comes

By the beginning of spring at the latest, however, the growth of sea ice comes to an end in both polar regions. Because of the increasing air temperatures, the formation of new ice begins to slow down and then at some point it stops completely. Now the sea ice begins to melt. There are a number of basic processes involved in this. When the air temperature rises above freezing, the sea ice begins to melt first on the upper surface, so it becomes thinner. In the summer of 2018, on an expedition to the central Arctic Ocean, German sea-ice researchers investigated the rates of surface melt there. After weeks of observation, they were able to confirm that, by the end of summer, the original two-metre-thick ice floes had lost up to 60 centimetres in thickness due to melting processes at the surface alone.

The water produced by surface melting can either seep through the porous sea ice or run off the edge of the floe into the sea. In the Arctic, the meltwater often collects on top of the ice to create meltwater ponds. Because the dark water surface absorbs more solar energy than the sea ice around the pond, the ice at the bottom of the pond melts more quickly.

In the Antarctic, on the other hand, researchers have rarely observed meltwater collecting on the ice. There are two reasons for this. For one, the snow cover on the Antarctic sea ice is much thicker than that on Arctic floes. The meltwater therefore seeps deeper into the snow and often refreezes to form an intermediate layer of ice. For another, the cold offshore winds in the coastal regions of the Antarctic generally prevent the sea ice and its snow cover from melting as quickly at the surface as on the ice floes in the high northern latitudes. Instead, a certain amount of snow in the Antarctic evaporates in the cold, dry air without ever melting. Scientists refer to this direct transformation of a substance from a solid to a gaseous state as sublimation.

Ice floes, however, do not only melt on the upper surface. The solar radiation absorbed there is also transferred through the ice. As a result, the floe becomes warmer overall and also begins to melt in the centre. The small brine channels become larger and the ice becomes more porous and brittle. Thus, at a certain point in this process, sea-ice researchers refer to it as "rotten ice", because these ice floes can disintegrate or crumble like a very rotten log.

Finally, sea ice can also melt from below. This is primarily caused by warm water masses that flow directly under the ice. In the Southern Ocean, these may well up from greater depths, or wind and ocean currents can transport the mobile pack ice northward into areas with comparatively warm water. Conversely, in the Arctic Ocean, the sun warms the surface water, which can then release its heat to the ice and accelerate the melting process.

In the past, melting on the upper surface was the primary cause for the summer shrinking of sea-ice cover in the Arctic. But in recent years the amount of melting on the underside of the ice has increased significantly because, due to its long-term decrease in sea-ice cover, the Arctic Ocean is absorbing more solar energy and the surface waters are getting warmer. The heat supply is not yet sufficient for the sea ice in the Arctic Ocean to disappear completely. But even now, well over half of the winter ice cover is already melting in summer.

Standing on the Antarctic sea ice in winter, one might easily imagine that one is on a gigantic white land mass. Ice covers the Southern Ocean as far as the eye can see. There is usually a blanket of freshly fallen snow on the ice that increases the reflectivity of the surface to as much as 90 per cent. However, the reflection of incident solar energy is not the only critical function of sea ice within the Earth's climate system. It is also, in a sense, a driving force behind the conveyor belt of the world's ocean currents, because the brine that enters the ocean when the ice freezes plays an important role in a gigantic chain reaction.

What drives the ocean currents

The temperature differences between polar regions and the tropics effectively drive not only the air currents in the atmosphere in the global wind system but also, to a large degree, the worldwide system of ocean currents. These, in turn, influence the Earth's weather and climate in two important ways:

- The ocean currents transport an immense amount of heat energy and distribute it around the world.
- Variously warm air currents and water currents regulate the Earth's water cycle through the evaporation of seawater and the absorption or release of heat at the sea surface, depending on whether the overlying atmosphere is colder or warmer than the water.

Vertical transport of water in the oceans is involved when water from great depths reaches the surface at upwelling areas, while elsewhere surface waters sink to greater depths. The descending currents carry heat, oxygen and dissolved trace gases down from the sea surface with them. As a result of this process, the world's oceans have become our planet's most important heat-storage reservoir. In the past 50 years, they have absorbed 90 per cent of the excess heat that has been retained in the Earth system due to rising greenhouse-gas concentrations.

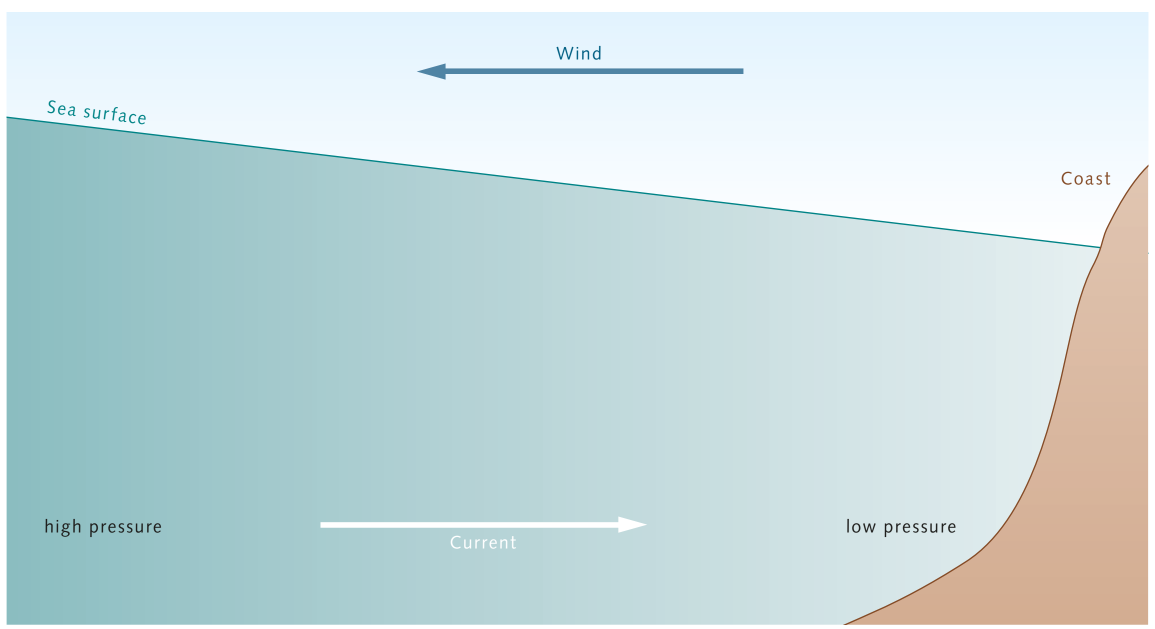
At right angles to the wind

Ocean currents generated by the motions of high and low tides are a familiar phenomenon around the world. But the large marine currents around the world are primarily driven by the density differences between water masses or by the power of the wind. When the wind blows over the water surface, friction is produced. The wind energy is transferred to water particles near the surface and sets them in motion. This produces waves and turbulence. The energy is distributed within the upper several metres of the water column, and a wind-driven surface current is created.

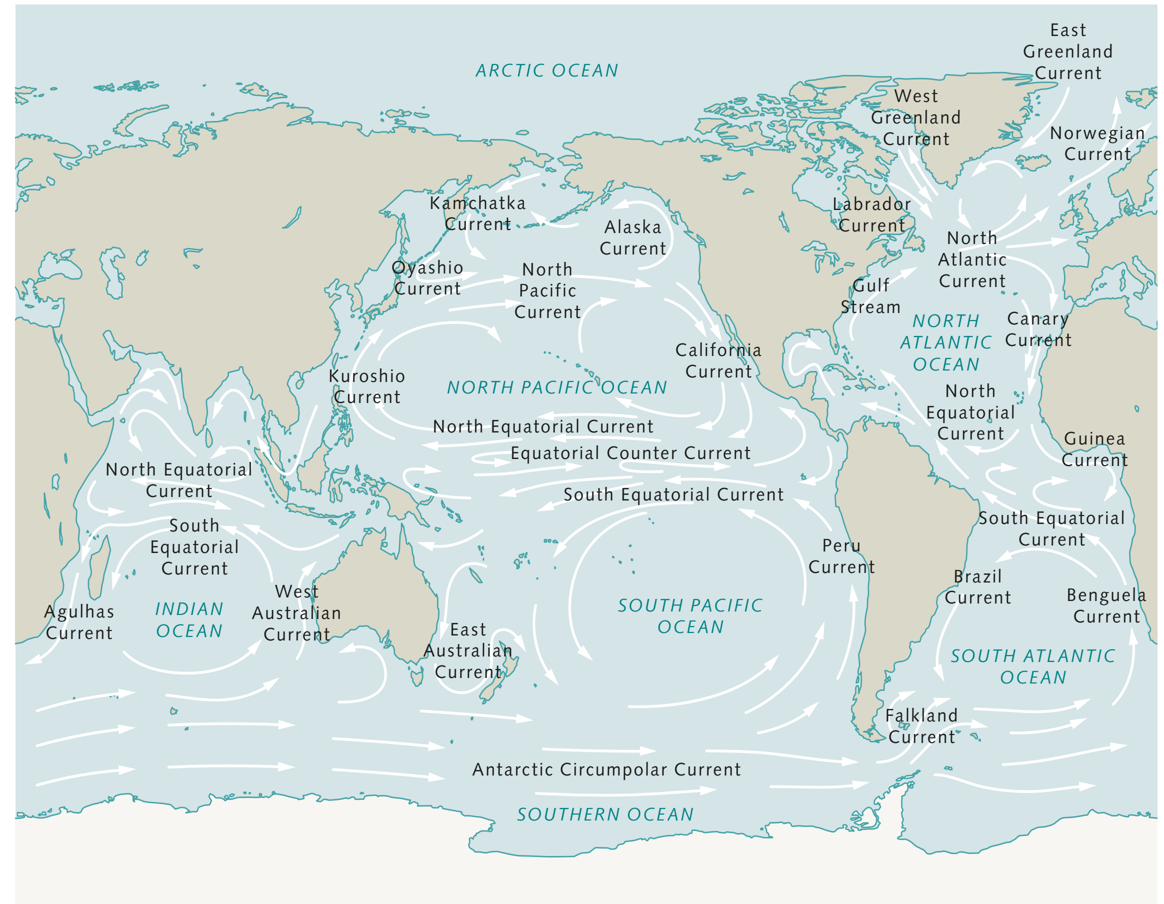
Contrary to reasonable expectation, perhaps, this current does not flow in a straight line parallel to the wind. Because the Earth is turning, the Coriolis force operates here to deflect the current. The total deflection, however, is only 45 degrees because the surface water driven by the wind pulls the immediately underlying, more static water layer with it to some extent. This means that the deeper water masses likewise sheer off and are diverted. With increasing depth, therefore, the flow angle with respect to the surficial wind direction increases and the flow velocity decreases. A schematic drawing, with the flow direction and speed of each of these successively deepening water layers represented as arrows, reveals a spiral-shaped, vertical velocity profile that resembles a corkscrew and is called the Ekman spiral. It was named for the Swedish oceanographer Vagn Walfrid Ekman (1874–1954). He was the first to recognize that the wind-driven near-surface water layers flowed more slowly with increasing depth and that their flow direction deviated increasingly from the wind direction. When all of these progressively changing flow directions in the water column are combined and the mean value is calculated, the result is that, for purely wind-driven ocean currents, the overall water transport is at right angles to the wind direction.

This phenomenon is known as Ekman transport, and it helps to explain, among other things, how water rises from great depths in upwelling areas such as the Benguela Current off the west coast of South Africa. This kind of upwelling occurs in coastal areas where the wind blows parallel to the coast and the Ekman transport it generates forces the near-surface waters out to the open sea at a right angle. Deep waters then flow in from below, replacing these surface waters.

Such upwelling currents are of crucial importance for life in the sea and for the climate in the coastal regions where they occur. For one, nutrients brought up with the deep water promote the growth of algae and microorganisms, which in turn become food for many larger marine organisms. That is why the most important worldwide fishing grounds are always in upwelling areas. For



2.19 > When winds at the sea surface push water masses in one direction and pile them up, a counter current develops at depth due to the changing pressure conditions.



2.20 > All of the ocean currents and gyres illustrated here are part of the surface circulation of the world ocean, and are driven by winds.

another, the cold water masses at the surface flow toward the equator as a part of the eastern boundary currents of subtropical gyres, and have an effect on the air temperatures and amounts of precipitation in the coastal regions. Worldwide, there are five of these currents. They are the California Current, the Peru Current, the Canary Current, the Benguela Current and the West Australian Current.

The five subtropical ocean gyres are among the most prominent surface currents in the world ocean. They are driven by the trade winds and the west winds, and they differ only by the fact that, due to the Coriolis force, the water masses in the gyres in the northern hemisphere rotate clockwise and those south of the equator flow in a counter-clockwise direction. A piling up of water masses on the western side of these ocean gyres results in the formation of western boundary currents. These include, among others, the Gulf Stream off the east coast of the USA and the Agulhas Current in the southern Indian Ocean. The western boundary currents, as a rule, are significantly narrower than the boundary currents on the eastern side of the gyres, and they also flow faster.

Density changes – ascending or descending?

In addition to the wind as a driving force, there is another mechanism that sets enormous currents into motion: a global-scale overturning circulation that transports the water masses on a kind of conveyor belt through all the world's oceans. The motion along this conveyor belt is maintained by differences in the temperature and salinity of the water masses, which is why scientists also refer to it as thermohaline circulation (*thermo*: driven by temperature differences; *haline*: driven by differences in salinity). To understand the mystery of its function, one has to know two things about the world oceans in general and about water specifically, because water behaves differently from most other chemical substances. In almost all other substances, the atoms and molecules move closer together the colder it gets, but this is not strictly the case with water. Normal freshwater exhibits its maximum density at a temperature of four degrees Celsius, because at that temperature the water molecules are closest together.

When it contains dissolved salt, however, the chemical and physical properties of water are different. The density of saltwater continues to increase steadily with falling temperature, and reaches its maximum at the freezing point. For this reason, saltwater at two degrees Celsius is significantly more dense and heavier than freshwater at the same temperature.

There is another important factor: the saltier the water is, the heavier it is. This means that the actual density of seawater is determined by a rather complex relationship between temperature and salinity. In principle, the water masses of the ocean are layered one above the other according to their density. The heaviest and usually the coldest water is found in the deep sea, while the lightest water is found at the surface.

As a rule, winds and waves are only capable of mixing the upper 200 metres of the water column. The deeper water masses, on the other hand, remain virtually unmixed. This is why scientists can speak in terms of the stable stratification of the oceans. Similar to the way that horizontal density differences between high- and low-pressure areas in the atmosphere cause winds, the horizontal pressure differences in the ocean, with a small intermediate step, are responsible for the creation of currents.

A question of salinity

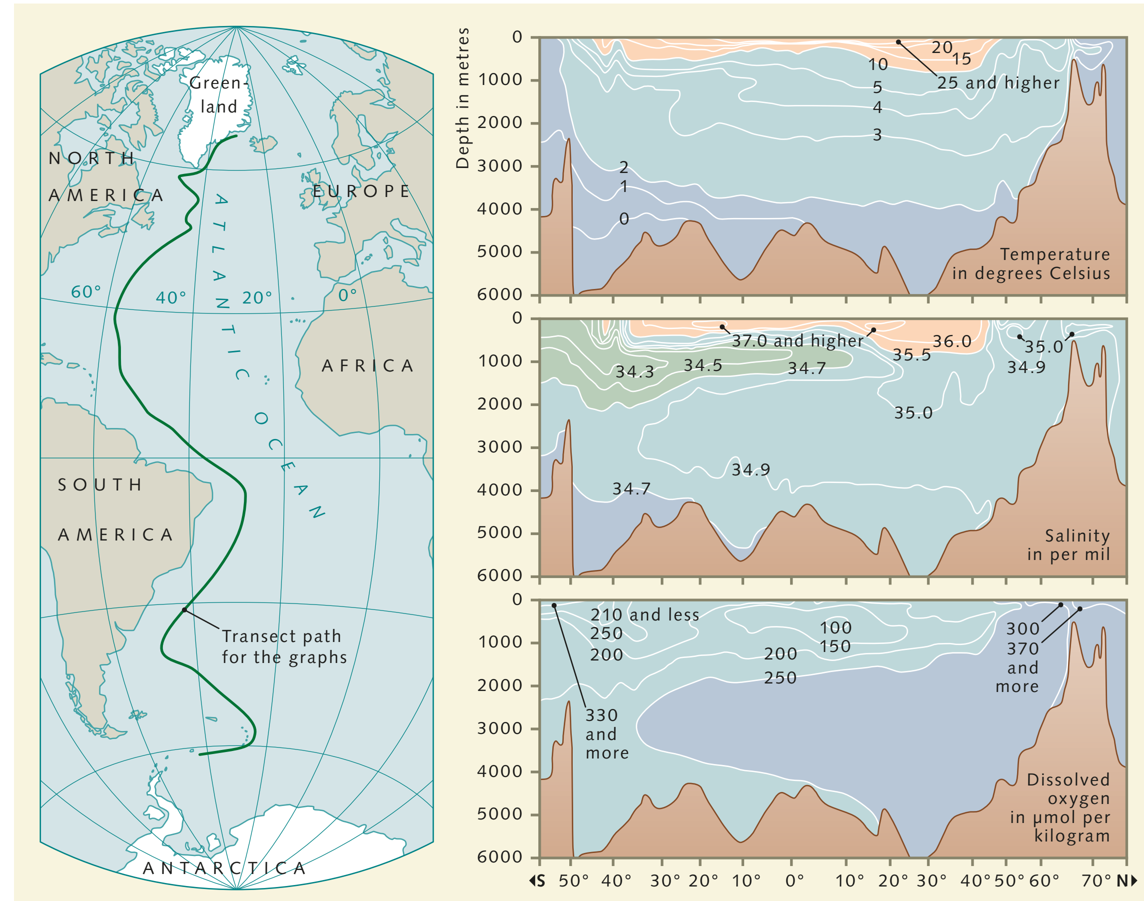
The temperature and salinity of the water, and therefore its density, are determined by processes at the sea surface. When water cools, its density increases. It becomes heavier and sinks to a greater depth. This process is called thermal convection. But when the surface water warms up, it becomes less dense. It becomes lighter, and the difference between its density and that of the underlying water increases. As a result, the warm, light water remains at the sea surface unless a mixing of the two water layers is induced by the wind.

A similar case is observed for salinity. It increases when water evaporates at the sea surface. But when it rains, or where rivers or glaciers deliver fresh water into the sea, the salinity of the surface water decreases along with its density. In this case again, the light water remains at the sea surface. If a water mass becomes more saline, however, and thus heavier, then haline convection commences. The heavier water sinks. In this way, immense amounts of water are overturned to depths of several kilometres.

The salinity of surface water also changes when sea ice forms. For example, when the coastal regions of the Southern Ocean freeze in early winter, salt is effectively spread over large areas in the sea, as the brine that collects in the small channels and chambers of the porous sea ice gradually seeps out into the water.

Scientists have found that 70 to 90 per cent of the salt contained in the surface water is released into the underlying water layer during the freezing process. With decreasing temperature or increasing salinity of this layer beneath the sea ice, the water becomes heavier. It sinks to the sea floor, collecting there as dense shelf water. It then spreads out and, at some point, flows down the continental slope into the deep sea. There, at a depth of several kilometres, it feeds the Antarctic Bottom Water, which is the lowest level of the world ocean. Above this flows the somewhat warmer, and thus lighter North Atlantic Deep Water coming from the north.

There are presently four known regions in which Antarctic Bottom Water is created: in the Weddell Sea, the Ross Sea, off the coast of Adélie Land, and in the Cape Darnley polynya west of the Amery Ice Shelf. The heavy, cold water is an important component in the worldwide conveyor belt of ocean currents. In somewhat simple terms, the cycle functions as follows: Warm water from the tropics flows into the polar regions. There it releases its heat into



2.21 > The individual water masses of the Atlantic Ocean can be distinguished by their temperature as well as by their salt and oxygen content. The vertical cross sections show how the water masses are layered one above the other, following the course of the line drawn on the map on the left, from the Antarctic to East Greenland.

the cold polar atmosphere. As a result, the water cools, becomes heavier, and descends to greater depths, whereupon it flows back toward the equator. At the sea surface, new warm water continues to flow in and cool down so that the overturning motion continues uninterrupted.

Even from this simplified explanation it is clear that the polar seas play a key role in the global water-mass circulation. They are the driving force behind the global conveyor belt, although the processes controlling the turnover of water masses differ greatly between the Arctic and Antarctic.

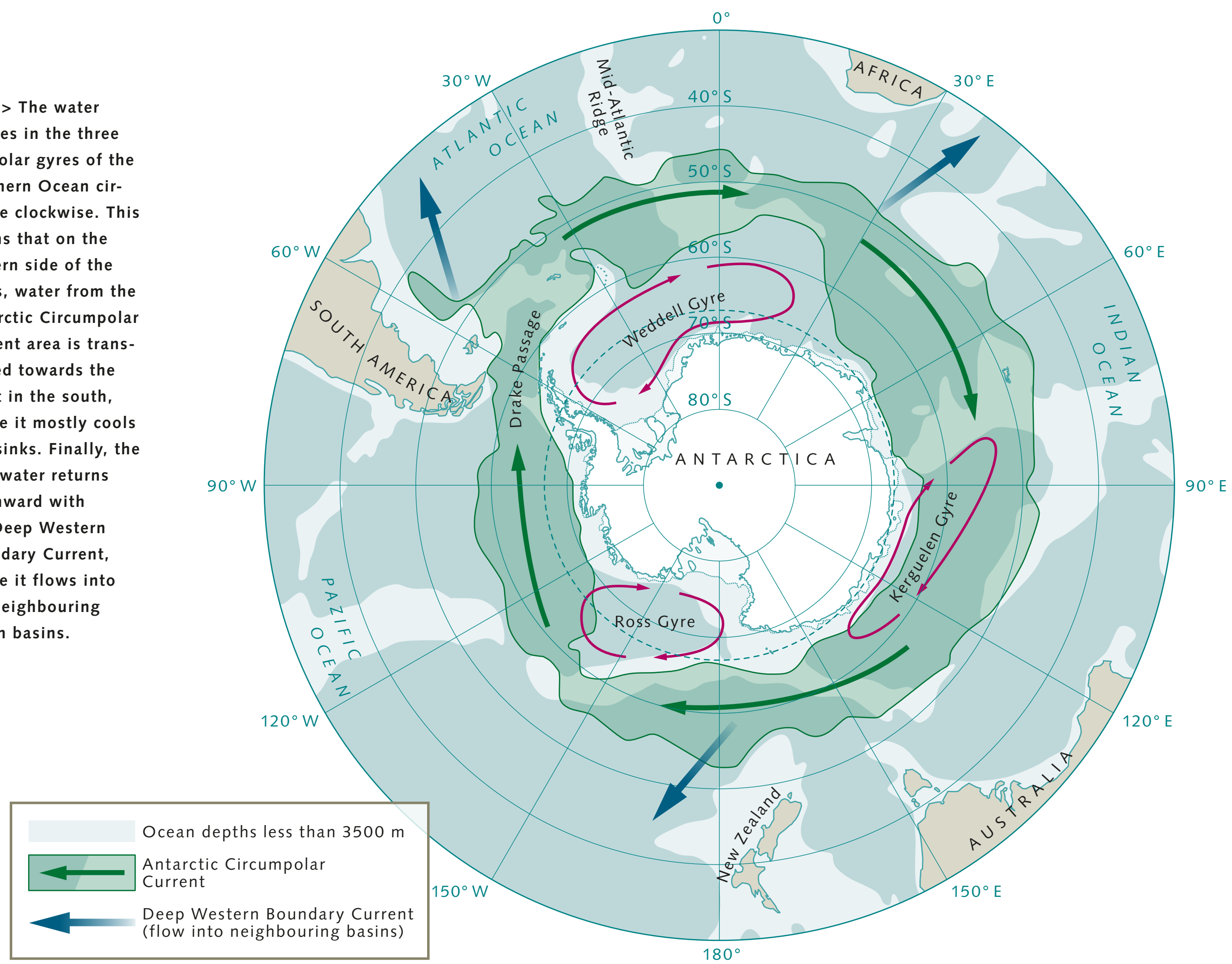
Overturning in the wild Southern Ocean

As a sea that circles the globe, the Southern Ocean connects the world's three large ocean basins and thus facilitates the global circulation of water masses. Hydrographically, it can be broken down into the Antarctic Circumpolar Current in the north, the coastal current on the continental margin in the south, and the three large subpolar gyres situated in between. These gyres, rotating clockwise, are located in the areas of the Weddell Sea (Weddell Gyre), the Ross Sea (Ross Gyre) and the Australian-Antarctic Basin (Kerguelen Gyre).

The sea-surface characteristics of the individual water masses are primarily controlled by conditions in the atmosphere. The air temperature over the oceans in the southern hemisphere drops strongly toward the south, which has an impact on the air pressure and thus on the wind conditions. Over the near-coastal parts of the Southern Ocean, easterly winds blow as well as offshore fall winds in some areas, which are known as katabatic winds. The zone of circumpolar west winds is located further to the north. These loosely defined bands are known as the "roaring forties", the "wild fifties", and the "howling sixties", and they provide the driving force behind the marine currents in the Southern Ocean. Like the air temperature, the temperature of the water also falls to the south. In the subtropics the water temperature at the surface is a warm 25 degrees Celsius. In the Antarctic coastal waters it is near the freezing point of salty seawater, which is minus 1.9 degrees Celsius.

The Antarctic Circumpolar Current is driven by a large-scale band of west winds. It transports water masses more than one hundred times greater than all the world's rivers combined, and is the most powerful current system on the Earth. Immense amounts of water are involved here because the Circumpolar Current is up to 2000 kilometres wide and extends far below the surface. While other wind-driven currents move the water to maximum depths of only 1000 metres, the Circumpolar Current can extend down to depths of 2000 or even 4000 metres. The current velocity in many places, however, is only 20 centimetres per second or less. This makes it a comparatively slow ocean current.

The Circumpolar Current is not a unified cohesive belt, but is subdivided into a number of smaller segments connected by what are known as fronts. To a large extent,



2.22 > The water masses in the three subpolar gyres of the Southern Ocean circulate clockwise. This means that on the eastern side of the gyres, water from the Antarctic Circumpolar Current area is transported towards the coast in the south, where it mostly cools and sinks. Finally, the cold water returns northward with the Deep Western Boundary Current, where it flows into the neighbouring ocean basins.

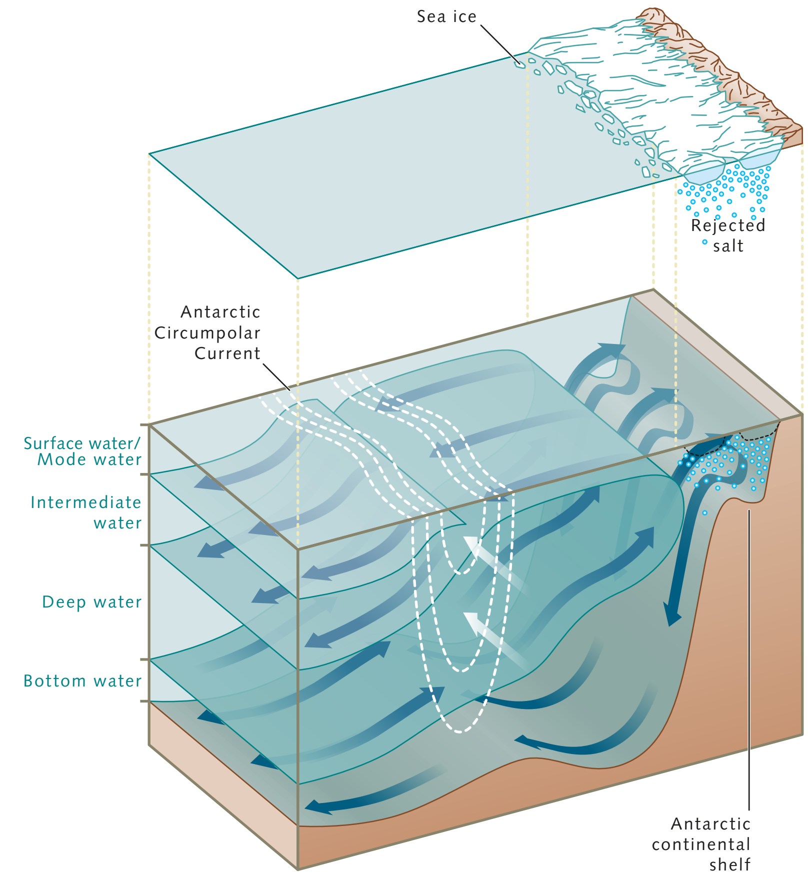
it prevents warm surface water from the Tropics from penetrating directly into the Antarctic region. But this barrier is not completely impregnable. Eddies with diameters typically around 100 kilometres repeatedly break away from the fronts, migrate a bit to the north or south depending on the direction of rotation, and then dissipate again after a few weeks. The eddies thus allow for a certain amount of horizontal mixing of the water-mass properties by permitting deep water coming from the north to penetrate southward beyond the Circumpolar Current at a depth of 2000 to 3000 metres.

Here, scientists distinguish between the Upper Circumpolar Deep Water, which has average temperature and salinity values and contains little oxygen because it has been circulating for centuries through the deep Pacific Ocean with no surface contact, and the high-salinity Lower Circumpolar Deep Water, which originates from the North Atlantic Deep Water and is not as old.

Both of these deep water masses are initially carried along with the Circumpolar Current. They make a couple of revolutions around the continent of Antarctica, slowly rise upward, and are eventually able to break away to the south with the help of the subpolar gyres. Upon reaching the sea surface, the water masses release their heat to the atmosphere. At the same time, snow, rain and melting icebergs all contribute to reducing their salinity. A portion of this ascending water subsequently flows to the north and sinks to intermediate depths again as Antarctic Intermediate Water on the northern flank of the Circumpolar Current. The remaining portion is transported southward to the coast by the subpolar gyres. There the surface water freezes and, through the process of ice formation, it is again enriched with salt.

Its subsequent path back into the depths is thereby predestined. The cold, heavy water sinks and thus triggers a convective mixing. The more intensive the cooling and salt enrichment process is at the sea surface, the deeper the heavy water sinks. In some situations, it can even flow beneath the relatively warm Circumpolar Deep Water lying on the continental slope.

While the Circumpolar Current is driven by westerly winds in the northern part of the Southern Ocean, the near-coastal easterly winds further to the south propel a counter current, the Antarctic Coastal Current. This flows westward above the Antarctic continental slope as a boundary current and includes the southern segments of the subpolar gyres. The Antarctic icebergs drift with the Coastal Current. One reason why researchers are interested in this current is that warm, relatively salt-rich Circumpolar Deep Water lurks on its underside and, in the course of climate change, this is becoming increasingly threatening for the Antarctic ice masses.



2.23 > With the overturning in the Southern Ocean, the deep water coming out of the north wells up just off the coast, cools down almost to the freezing point at the surface, and then, saturated with brine, it sinks to the deepest level of the ocean as heavy bottom water. Further to the north, on the other hand, upwelling water flows back toward the equator, changing to lighter mode water or intermediate water underlying the surface waters.

Overturning in the Arctic Ocean

The formation of deep water in the Antarctic is not the only process that keeps the global conveyer belt of ocean circulation in motion. A second driving force, the Atlantic Meridional Overturning Circulation (AMOC), acts in the northern part of the Atlantic Ocean. Simply put, this is a mechanism that transports warm surface water from the tropics to the North Atlantic, where it cools, sinks, and then flows southward again as cold North Atlantic Deep Water at a depth of two to three kilometres. At the surface it is continuously replaced by water flowing in from the subtropics.

In order to understand the decisive role that the Arctic plays in this overturning process, it is helpful to take a closer look at the individual steps involved. The warm surface water is transported to the west coast of Ireland by the northern branch of the Gulf Stream, which scientists call the North Atlantic Current. There the current divides and about one-third of the water is entrained by the sub-

2.24 > The current system of the Arctic Ocean is controlled to some extent by the influx of warm, salt-rich water from the North Atlantic Current. In addition, some cold water masses form in the Siberian marginal seas and in the Norwegian Sea, especially in winter, and subsequently flow out of the Arctic region toward the Atlantic Ocean.

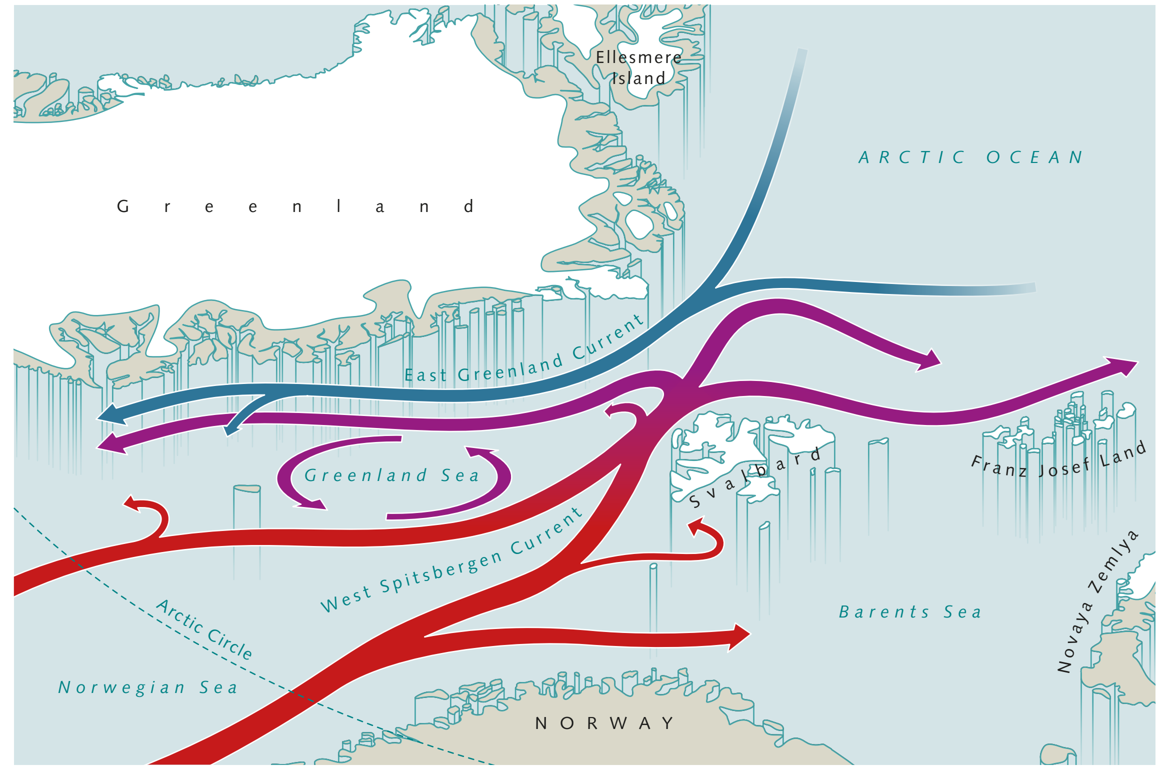
polar gyre. As the Norwegian Current, this flows along the west coast of Scandinavia, then from the Norwegian Sea into the Barents Sea, a marginal sea of the Arctic Ocean. The remaining water branches off toward Greenland, then divides again into the West Spitsbergen Current, which flows into the Fram Strait, and another arm that transports the warm water into the Labrador Sea between Greenland and Canada.

On their northward pathways, all of these currents cool down and are diluted by rainwater. With the release of heat energy into the atmosphere, they significantly influence the climate of northern Europe. Without the heat transport of the Gulf Stream and its extensions the climate would be much colder in northern Europe, especially in the winter.

The water loses particularly large amounts of heat in the Barents Sea. As an Arctic marginal or shelf sea it is only 50 to 400 metres deep, and therefore cools down fairly rapidly. Furthermore, there is a great extent of mixing of the inflowing water masses here. A number of rivers, the Russian Kola, for example, transport large amounts of freshwater into the Barents Sea. The water masses flow back and forth with the tides, which causes the entire water column to lose a great deal of heat energy, especially in winter. If the water also freezes to form sea ice, and the brine created increases the density of the shelf water, three different kinds of water are formed:

- cold, low-salinity surface water that is driven by the wind and distributed into the central Arctic;
- cold, high-salinity water that sinks to intermediate depths and spreads out there; and
- very salty, heavy water masses that either flow directly through the Norwegian Sea back to the Atlantic or take the longer route as Arctic Bottom Water through the Arctic Basin and the Fram Strait.

The tides play a less important role In the Fram Strait, but even here the three- to six-degrees-Celsius warm Atlantic Water cools down by large increments. The current system of the Fram Strait can be envisioned as a major road with a turning lane. In the right lane to the



2.25 > The Fram Strait is one of three marine regions in the Arctic where water is overturned. Here the West Spitsbergen Current transports warm, salt-rich water northward where it first cools and then eventually sinks to a depth of 200 to 800 metres. One part of the current then flows into the Arctic Ocean. The other part reverses direction and flows southward as North Atlantic Deep Water beneath the cold East Greenland Current.

east, the warm, saline Atlantic Water first flows northward on the surface in the West Spitsbergen Current. The colder it becomes, the heavier it becomes. At a certain point the current sinks to a depth of 200 to 800 metres, where it splits. One branch of the current continues on its path into the Arctic Ocean. The remaining water turns to the west, making a 180-degree-turn, and moves into the opposing lane, where it flows back to the south as North Atlantic Deep Water on the eastern edge of the Greenland shelf. However, along this opposing lane, called the East Greenland Current, there is also a second current that flows one level higher at the sea surface. It comes from the Arctic and transports cold, minus 1.8-degree-Celsius water with relatively low salinity and abundant ice floes into the North Atlantic.

Together, these water masses cross the shallow thresholds, only 800 metres deep, between Greenland, Iceland and Scotland, and then flow downward like giant waterfalls into the deep basin of the North Atlantic. A third current, with deep water from the Labrador Sea, now flows above them. During its winter cooling it has sunk to a depth of about 2000 metres and now completes the Arctic cold-water stream, which flows as deep water along the east coast of America toward the South Atlantic.

A comparison of the overturning circulation in the North Atlantic with deep-water formation in the Southern Ocean reveals an important difference. The water masses in the north sink because they lose heat to the atmosphere in ice-free marine regions like the Labrador Sea, the Norwegian Sea and the Siberian shelf seas, and thereby become colder and heavier. At the same time, in the central Arctic Ocean hardly any convection takes place. Here the sea-ice cover insulates the ocean too well for it to be able to release much heat into the atmosphere.

In the Antarctic, on the other hand, deep-water formation is mainly driven by the freezing of sea ice and the associated release of brine. Although the prior heat loss of the water also plays a role, the formation of sea ice is more significant here.

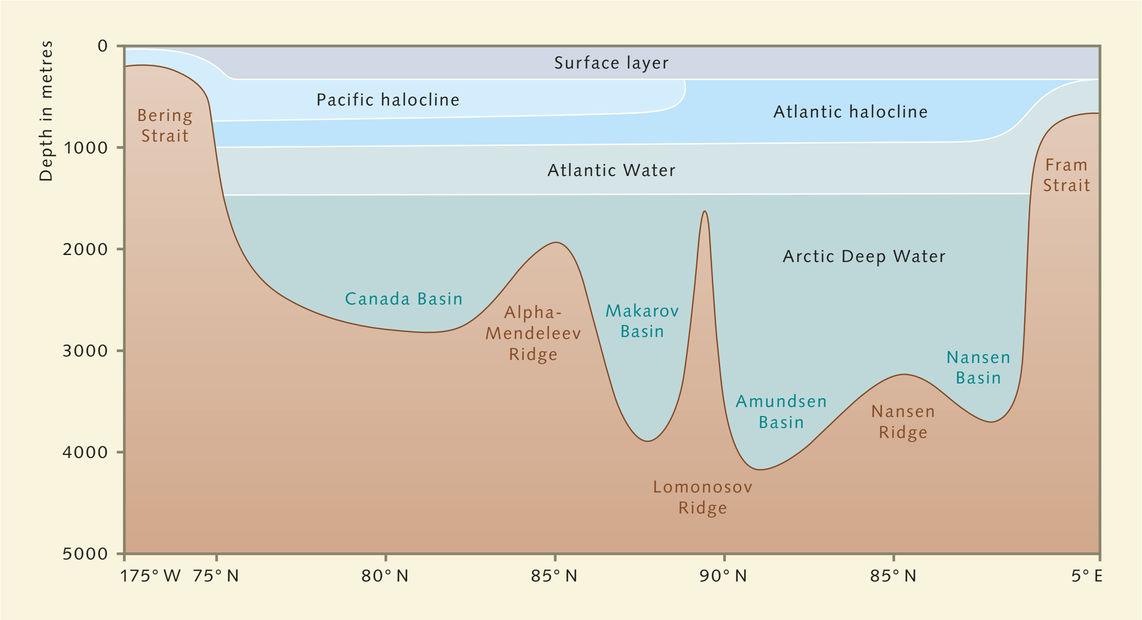
A protective layer for the sea ice

Overturning of the Atlantic Water, however, is not the only role played by the Arctic Ocean in the global conveyor belt of ocean currents. It also represents an important link between the Pacific and Atlantic Oceans. Through the Bering Strait, only 85 kilometres wide and 50 metres deep, relatively warm, low-salinity Pacific water flows into the Arctic Ocean. The inflow is only one-tenth of the amount that enters through the Fram Strait and the Barents Sea from the North Atlantic, but it definitely has an influence on the course of events here. The water masses from the Pacific transport heat into the high north, which has an impact on the formation of sea ice in the Chukchi Sea north of the Bering Strait. Because of its low salinity, the Pacific water reinforces the stratification of the Arctic Ocean. Looking at a profile of its water column, the following characteristic features can be recognized from top to bottom:

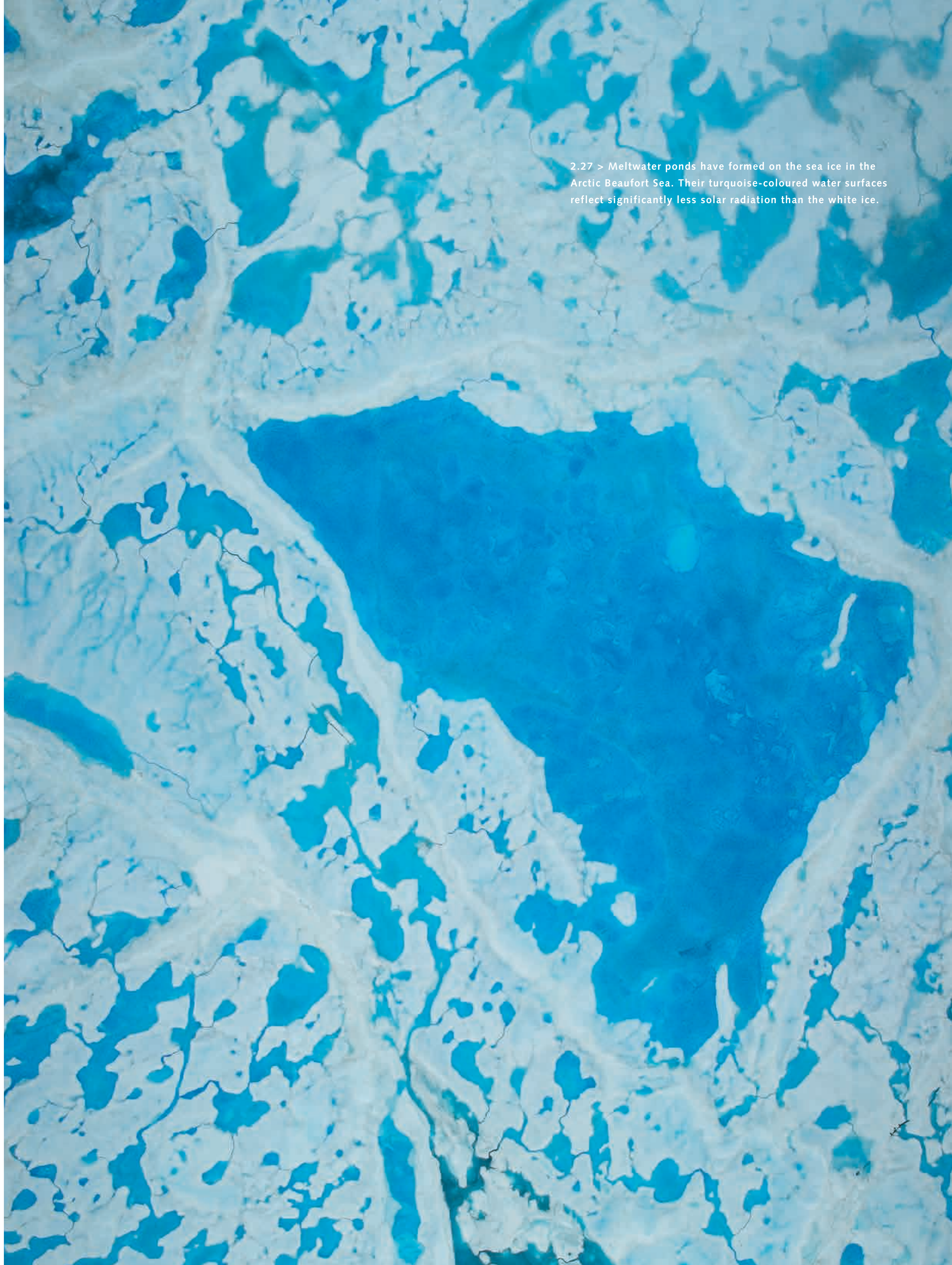
The surface layer

Wherever sea ice floats on the Arctic Ocean, it is underlain by a 5- to 50-metre-thick layer of low-salinity water. This uppermost water layer is fed by freshwater that primarily comes from the many rivers that empty into the Arctic Ocean. The northern European, Siberian and North American rivers alone transport around 3300 cubic kilometres of water into the Arctic Ocean annually. This is equal to about eleven per cent of the world's continental runoff, and explains why the water of the Arctic Ocean contains significantly less salt than, for example, the water masses of the Atlantic Ocean.

The freshwater carried in by rivers mixes with seawater in the shallow shelf seas and then, driven by the wind, spreads into the central Arctic. In the shelf seas, as well as in the central Arctic Ocean, this surface layer can be relatively warm in summer, especially where the ice cover has broken up into individual floes or even completely melted. Where no sea ice is present the solar radiation can warm the surface water, which in many places leads to more enhanced melting of the remaining floes from below. As a consequence of melting and the associated freshwater input, the surface layer becomes less saline and thus more stable as the summer progresses. As a result, this water tends to mix less readily with the underlying, higher salinity water masses. The incoming heat radiation thus remains trapped within the uppermost water layer. In the autumn and winter,



2.26 > Schematic representation of the water masses in the Arctic Ocean. Warm water masses from the Atlantic circulate above the Arctic Deep Water. Above these, in turn, are the cold Atlantic and Pacific Haloclines, which, together with the surface layer, protect the floating sea ice from the heat of the Atlantic Water.



2.27 > Meltwater ponds have formed on the sea ice in the Arctic Beaufort Sea. Their turquoise-coloured water surfaces reflect significantly less solar radiation than the white ice.

however, the surface layer again cools down and, with the initiation of ice formation, becomes more saline.

The halocline

Beneath the surface layer, especially in the deep basin of the Arctic Ocean, lies a second well-defined layer called the cold halocline. The term “halocline” comes from the Greek and indicates a transitional zone between water layers that have different salt contents, which is why the halocline is also sometimes called the salinity discontinuity layer. The salinity of the water increases from the base of the surface layer to a depth of around 200 metres, until it has the same value as the underlying Atlantic Water. This kind of salinity layering is not at all unusual in the world’s oceans. A special feature of the Arctic Ocean, however, is that, although the salinity of the water in the Arctic halocline does increase with depth, the water temperature remains fairly close to freezing throughout, despite the fact that the Atlantic Water below the halocline is significantly warmer, with a temperature of approximately one degree Celsius.

The temperature in the Arctic halocline is relatively low because its water originates in the shelf seas, where the surface waters cool down considerably in winter, and large amounts of ice are formed in the coastal polynyas. Furthermore, numerous rivers dilute the shelf water with freshwater, which is why its salinity is very low. During the winter, however, the salt content increases as a result of the constant formation of sea ice.

This cold water, which is still fairly low in salinity at the beginning of winter, flows from the shelf seas into the central Arctic. There it spreads in all directions, flows beneath the even lower-salinity surface layer because of its density, and provides an additional layer of insulation against the deeper warm Atlantic Water.

Together, the surface layer and halocline provide a degree of stability in the stratification of the Arctic Ocean, such that neither the wind nor convection are able to produce the turbulence necessary to transport significant amounts of warmer Atlantic Water up to the sea surface from below.

The water masses from the Pacific Ocean flowing through the Bering Strait into the Arctic Ocean have a fate similar to that of the shelf water. They are also relatively low in salt content, and experience a similar development in the shallow Chukchi Sea as the water masses from the other shelf seas. Ultimately, the ocean water from the Pacific, because of its density, is integrated into the layering scheme of the central Arctic as the Pacific Halocline.

The Atlantic Water

The Atlantic Water, which has already been mentioned numerous times, flows into the Arctic through the Fram Strait and the Norwegian Sea. It originates in the Gulf Stream far to the south, but cools down markedly on its northward journey. By the time it has reached the central Arctic, its temperature is only about one degree Celsius, but it is still by far the warmest water there. It circulates counter-clockwise through the Arctic as a narrow boundary current. One part of this boundary current flows along the continental slope throughout the entire Arctic. Additional portions branch off at the three submarine ridges that divide the central Arctic into the Canada, Makarov, Amundsen and Nansen Basins. If this Atlantic Water were to rise to the surface, the days of the sea ice would be numbered, because its heat energy would be sufficient to melt great volumes of ice.

The Arctic Deep Water

Beneath the Atlantic Water, at the greatest depths of the Arctic Ocean, flow the coldest and most saline water masses of the Arctic Ocean: the Arctic Deep Water. This is heavy water that has travelled down the continental slope in narrow, shallow channels from the shelf seas, and along its way mixed with the salty Atlantic Water. These descending streams are also affected by the Coriolis force. It deflects the water to the right so that it travels through the entire Arctic Ocean along the continental shelf on its way to the deep sea. The upper part of these water masses, in turn, ultimately leaves the Arctic Ocean through the Fram Strait.

The relatively stable stratification of the water masses in the Arctic Ocean has so far prevented the heat coming in from the Atlantic from rising to the sea surface, where it would present a serious threat to the Arctic sea ice. In the course of climate change, however, researchers expect to see far-reaching changes in the interactions between the ocean and sea ice.

Continental-scale ice sheets

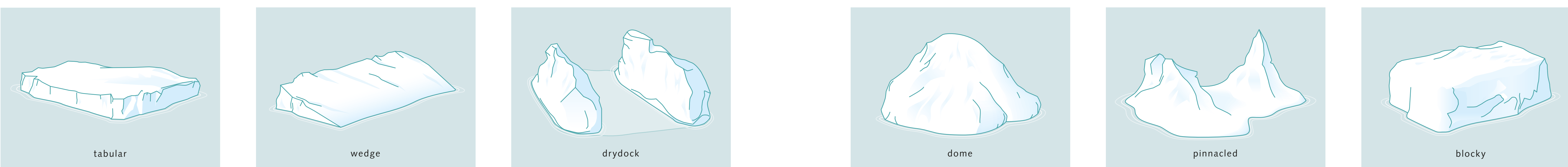
The amount of ice incorporated in the ice sheets of Greenland and Antarctica is difficult for the human mind to conceive. The polar ice sheets are the largest contiguous ice masses on the Earth. In order to illustrate their magnitude, impressive statistics of their mass are often cited. They incorporate around 99 per cent of the Earth’s total ice mass and, with a total area of 15.6 million square kilometres, they cover around 9.5 per cent of the land area of our planet. For illustration, the entire area of Germany could be covered almost five times by the Greenland Ice Sheet and almost 39 times by the inland ice of Antarctica.

The ice sheet of Greenland is up to 3300 metres thick and that of Antarctica as thick as 4900 metres. Together they store a volume of ice that, if completely melted, would cause global sea level to rise by around 65 metres. The ice sheets of West and East Antarctica have a combined ice volume of 26.37 million cubic kilometres, and the inland ice of Greenland around three million cubic kilometres.

Each of the ice sheets is surrounded by glaciers, through which the ice formed in the continental interior flows towards the sea. Researchers have counted 13,880 glaciers in Greenland alone. Many of them culminate in fiords, where icebergs can break off at the glacier’s leading edge, called the calving front. By contrast, in Antarctica the ice masses of multiple glaciers often converge on a coastal segment to form a large ice tongue that protrudes out into the sea. These floating extensions of the glaciers are called ice shelves. Icebergs also break off at these calving fronts, but as a rule they are considerably larger than those in Greenland. Because of their shape, Antarctic icebergs are commonly referred to as tabular icebergs.

The significance of the ice sheets for the climate of the polar regions is primarily due to the high albedo effect of the seemingly endless white ice surfaces. In regions

Tabular iceberg
A freshly calved tabular iceberg has a flat, level surface and nearly vertical flanks. In the Antarctic these icebergs can be up to 160 kilometres long and tens of kilometres wide. As a rule, they are 200 to 400 metres thick and rise 30 to 50 metres above the water surface. Similarly shaped icebergs in the Arctic are generally much smaller.



2.28 > No two icebergs are the same, largely because waves and winds always affect the ice masses in different ways. Nevertheless, six fundamentally different types of iceberg can be distinguished.

2.29 > At the centre of every snow crystal there is a small dust particle, around which water vapour from the air condenses and freezes to ice. Snow, however, does not usually fall in the form of single crystals, but as flakes that are composed of several interconnected snow crystals.

where freshly fallen snow lies on the ice sheet, up to 90 per cent of the incident solar radiation may be reflected. Even without a snow cover this value is still around 55 to 60 per cent. Through the glaciers and ice shelves, continental ice sheets also have an impact on the oceans. Where glaciers are calving, where meltwater is flowing into the sea, or where ice shelves and floating glacier tongues melt on the underside, fresh water is released directly into the ocean. Conversely, the growth of ice sheets and glaciers also removes large amounts of moisture from the water cycle. In the Antarctic, for example, the amount of snow that falls on the inland ice annually would be enough to raise global sea level by six millimetres. In the Southern Ocean, the floating glacier tongues and ice shelves also play a decisive role in the formation of deep water, and thus also in driving the global ocean currents. And finally, the growth and shrinking of the ice masses on land can serve as an indicator of developments in the global climate. Shrinking of the ice sheets and glaciers is a fairly certain sign of global warming, while an increase in their masses would suggest cooling of the world climate.

From snow to ice in three steps

Ice sheets and glaciers form in polar or high-altitude regions where more snow falls in winter than melts, evaporates, or is otherwise lost, such as through the breaking-off of icebergs, in the summer. However, in order for compacted glacial ice to form from a loose powder of snow, pressure and a fairly large amount of time are necessary, as is illustrated by the formation of ice in Greenland.

When new snow falls on the inland ice of Greenland, it has a density of 50 to 70 kilograms per cubic metre. This is because new snow is a relatively light material that contains a great deal of air compared to water in its liquid form. Freshwater, for example, has a density of 1000 kilograms per cubic metre. As soon as the snow falls its metamorphosis begins, which proceeds in a similar way through three phases everywhere in the world.

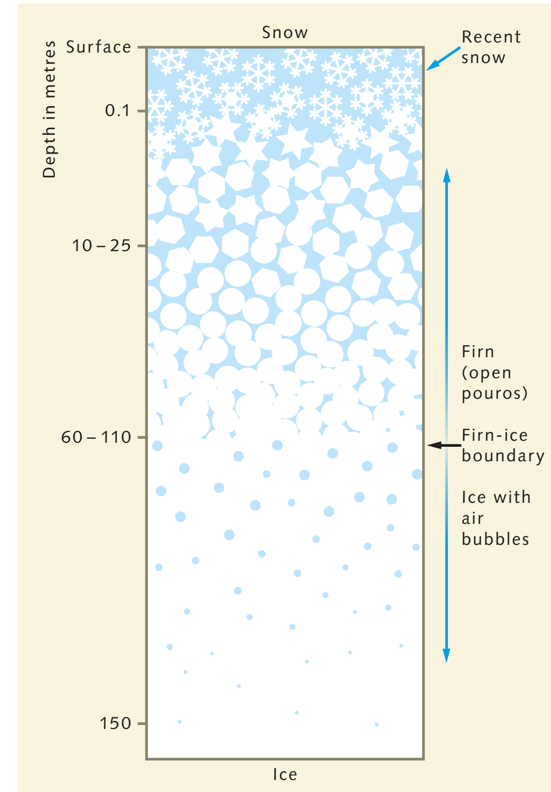
1. Snow compaction

First, the snow crystals are transported or blown about by the wind, which tends to break off their fine crystalline branches. In this, and other ways, every snowflake transforms to a granule of snow that resembles a tiny ball. This is driven by the physical principle of the minimization of surface energy. Spherical bodies have the minimum surface energy, and snow crystals, too, take on a spherical shape with time. Because of this shape, the snow can now also settle and be compacted. Many more spherical snow granules can fit into a given volume than fine-structured branching snow crystals. However, at this point the snow grains are not yet sticking together. If a shovel were used to dig into this top layer of snow, the individual snow grains would roll loosely off the blade.

2. Firn formation

Because the inland air temperature of Greenland rarely rises above zero degrees Celsius, even in the summer, the snow of a single winter generally remains unmelted. The following winter, when new snow falls onto the old snow, the weight of the new snow slowly compresses the underlying layers. The loose snow granules lying adjacent to one another now begin to bond with and adhere to their neighbouring grains. It almost appears as though the larger snow grains are consuming the smaller ones, because they continue to grow over the years. If one were to dig a pit in the snow at this point and repeat the shovel test at a depth of about one metre, the snow would remain on the shovel as a fairly solid block. Specialists call these coherent snow layers firn.

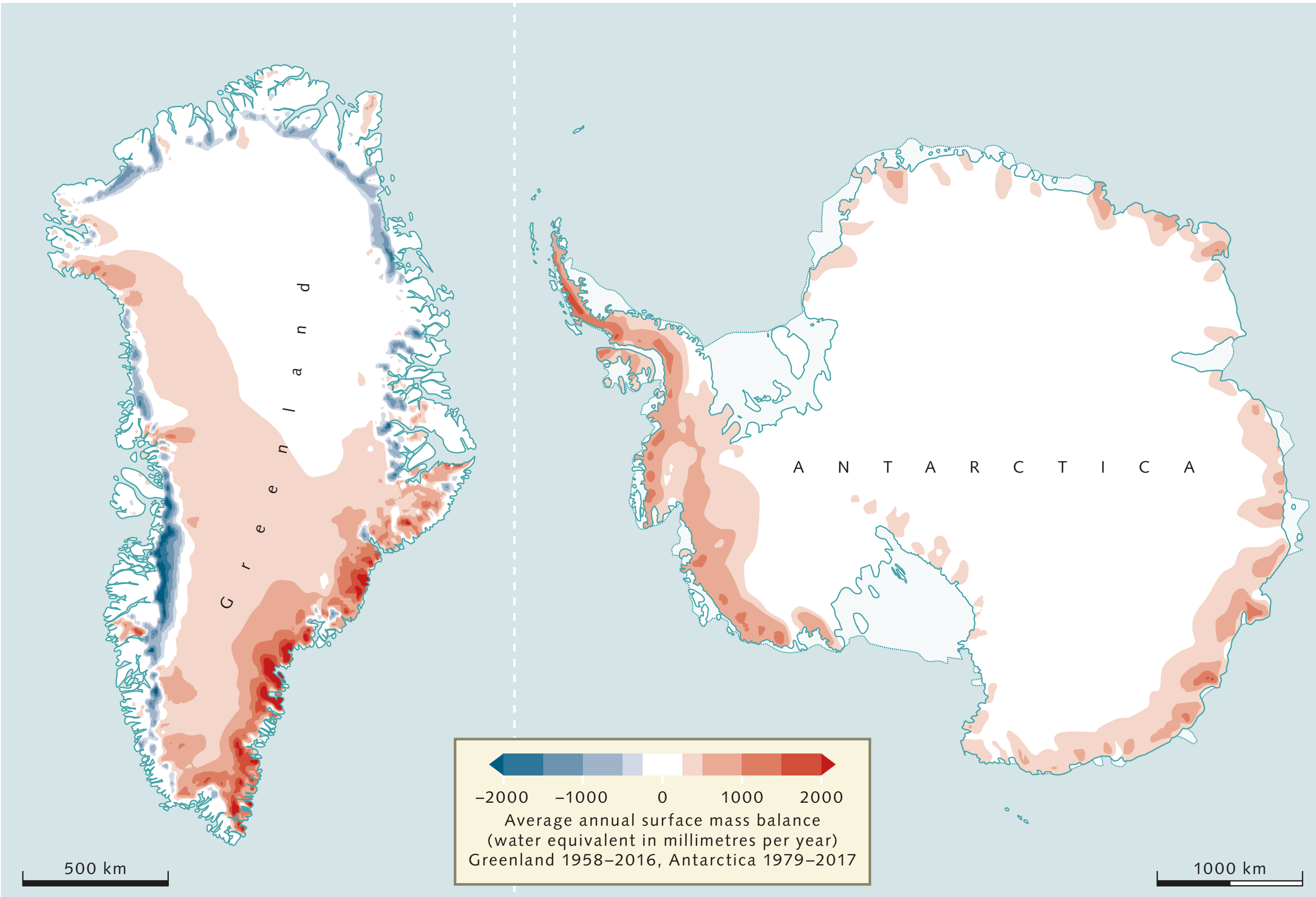
In the upper part of the firn layer the compressed material has a density of around 350 kilograms per cubic metre. In this phase it is still porous as a sponge, and air can circulate freely through it. But the more snow that falls on the surface of the ice sheet above, the greater the pressure on the deeper layers becomes. The ice crystals in the firn grow and press closer together, and the pore spaces become narrower.



2.30 > Glacial ice forms from snow, which initially compacts to firn and with continued compaction becomes ice. The speed and depth at which this process occurs depends, among other things, on how much new snow falls on the glacier's surface to increase the pressure on the underlying layers.

3. Ice formation

The compaction through settling processes and the growth of ice crystals ultimately produces a maximum density of 550 kilograms per cubic metre. However, as the snow load and resulting pressure from above continues to increase, pressure sintering commences. This means that the ice crystals fuse with each other. The pores close off and are sealed so that all of the air that was not able to escape is trapped in small bubbles. The point in time that this blockage of air flow occurs, and at what depth it occurs, depends on both the amount of annual snow accumulation and the temperature of the firn. In regions with higher snowfall, pore closure generally happens sooner than in areas with less snowfall. The same applies to firn that is warmer. The ice grains are cemented together more readily than they are in a very cold firn. In Greenland, as



2.31 > Whether there is a positive or negative surface mass balance in an ice sheet depends on how much snow has fallen and what portion of it is lost through melting, wind transport or sublimation.

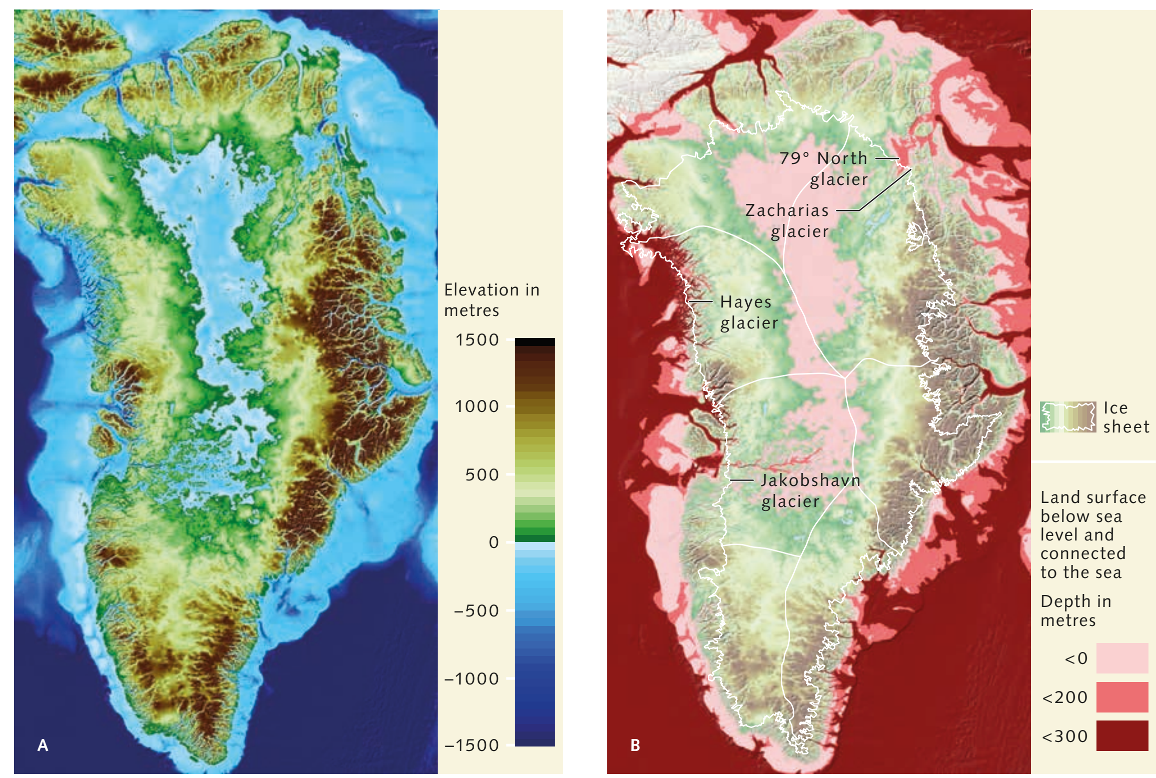
a rule, sealing occurs at a depth of 60 to 110 metres. At this point the material has a density of around 830 kilograms per cubic metre.

When the air can no longer escape, the state of ice has been reached. On the sub-Antarctic islands, researchers can recognize the firn-to-ice transition zone by a thick layer of refrozen meltwater in the ice body. During the summer there, snow on the glacier surface melts and the meltwater seeps down into the firn as deeply as the pores in the material allow. At the firn-ice transition it is blocked and then freezes again.

But the formation of glacial ice does not end with the sealing of the pore spaces. When the sheet of snow, firn and ice is several hundred metres thick, there is so much weight on the lower layers, and especially on the air bubbles, that the air within them crystallizes out. This means that all of the molecules contained in the bubbles are incorporated into the crystal structure of the ice. This applies to the gas molecules as well as to dust particles or other impurities in the air. Ultimately, a very dense, bubble-free ice forms that is characterized by its blue colour. When natural light shines on this ice, it absorbs a small portion of the red light, so that humans perceive the ice as having a slightly bluish hue. Glacial ice that appears to be more white, on the other hand, generally still contains many air bubbles.

How rapidly a glacier or ice sheet grows depends, among other things, on the amount of precipitation that falls on it. In West Antarctica up to four metres of new snow fall annually, with as much as six metres in the northern Antarctic Peninsula and on the coast of Wilkes Land, although these are only approximate values. Researchers always specify the amount of precipitation in terms of water equivalent (WE). This refers to the height of a water column that would result if the snow were to melt. In West Antarctica the precipitation would have a water equivalent of up to 1200 millimetres, or that same number of litres per square metre, while on the Antarctic Peninsula and in Wilkes Land it would come to 1800 millimetres (or litres). In order to derive the precise snow thicknesses from this, one would have to accurately know the density of the snow, which is seldom possible for large areas. Therefore, an estimate is commonly applied. One cubic metre of fresh snow yields a water column with a height of about 300 to 350 millimetres. The snow depths for the coastal areas of West Antarctica and Wilkes Land given above are derived by applying this approximation. In the centre of the continent, on the other hand, only a few centimetres of new snow fall each year. At the US American Amundsen-Scott South Pole Station, for example, between 1983 and 2010, meteorologists documented an annual snow accumulation of 27.4 centimetres, whereby this value also includes snow that was blown by the wind into the measurement field.

Most of the snow in Greenland falls on the southeast coast. Satellite data indicate that the new snow there drifts to heights of up to ten metres. The northern part of the island, by comparison, is very dry. Here, for the most part, less than 30 centimetres fall annually. The question then immediately arises as to how much of this new snow melts or evaporates during the summer and how much remains. Up until 30 years ago, this was just under half of the total snow in Greenland. Of the 750 gigatonnes of snow that fell during the winter about 350 gigatonnes survived to the end of summer. Today only around 200 gigatonnes remain through the summer. On the continent of Antarctica, excluding the ice shelf, around 2236 gigatonnes of snow fall each winter, of which about 50 gigatonnes are lost due to melting, particularly on the Antarctic Peninsula. An additional 84 gigatonnes of snow



2.32 > The topography of the land surface beneath an ice sheet has a great influence on how the ice masses flow and whether warm sea water can pose a threat to them. These two ice-free depictions of Greenland illustrate that the central region of the island actually lies below sea level (A) and it is directly connected by fiords to the sea (B). Also clearly recognizable are the deep channels through which the warm Atlantic Waters can encroach onto the shelf and into the fiords.

evaporate by sublimation. Most of the remaining snow compacts into ice.

Why does ice flow?

When a glacier or ice sheet has reached a certain size, the ice masses begin to move. Alpine glaciers, which are found in high mountainous areas such as the Alps or the Rocky Mountains, always travel down towards the valleys, a phenomenon that every skier and sledge rider can confirm from personal experience. But why do ice masses that lie on level terrain or in a valley also move? The Greenland Ice Sheet, for example, largely rests in a kind of basin, as the map of the island's underlying land surface shows. Still, its ice masses flow toward the outer margins.

The explanation for this is rather complex. Basically, large masses of ice move because the glacial ice either deforms under its own weight or because it glides on a slick subsurface. Usually it is a combination of the two processes. A key difference between them, however, is that gliding always requires a thin film of melt water on which the ice can slide, while the deformation can occur in a completely frozen state.

To help in understanding the deformation process, an ice sheet can be compared to a huge, viscous mass of cake batter piled onto a flat working surface, to which more dough is added one spoonful at a time. With the initial additions, the shape of the mound will not change substantially. Over a longer time, however, the mass in the centre will become so great that the dough begins to flow away towards the edges.

A large ice sheet responds in a similar way. With every new layer of snow the total amount of material increases. The pressure on the underlying ice masses increases, causing them to deform and flow toward the edges. As the deformation progresses, shear heat is generated within the ice sheet. This warms the ice and thus further accelerates the flow, because warmer ice deforms more easily.

The deformation processes alone, however, are not sufficient to cause ice streams, and especially glaciers, to move at the rapid speeds that scientists are observing today. The ice masses primarily gain speed by basal sliding. On a thin film of lubricating meltwater they glide down an incline like a sledge. In Greenland, under certain conditions, this meltwater can originate from the ice surface. In the summer it collects there in large meltwater lakes. In some of these lakes, the water then drains through cracks, crevices or tunnels in the ice down to the underside of the glacier, where it becomes the gliding film responsible for acceleration. As a rule, however, melting at the base of the ice sheet primarily occurs due to geothermal heat from the Earth below. This does not require a large amount of heat because, beneath thousands of metres of ice, the melting point at its base is reduced due to the high load pressure. It can therefore melt at a temperature of around minus two or minus 1.5 degrees Celsius. Still, however, the ice sheets only lose a few millimetres of ice on their underside each year due to melting.

Ice streams

To date, scientists are only beginning to understand the processes of gliding glacial ice. On the one hand, movement is influenced by the nature of the underlying landscape. On the other hand, sliding generates frictional heat, which melts a small amount of ice and warms the lower ice layers. As a result, these ice layers deform more easily, which can further accelerate the flow of ice.

In the Antarctic, about 30 outlet glaciers and ice streams transport ice into the sea. Researchers refer to large bands of flowing ice within an ice sheet as ice streams. These are generally distinguished from the surrounding ice by their flow velocity and direction, and they flow into glaciers at the outer margins of the ice sheet. Science still has no clear explanation as to why ice streams form or what mechanisms regulate their ice-mass transport, because hardly any two streams are alike. Some flow constantly, for example, and others only intermittently. Ice streams can also change their flow direction, abruptly increase their speed, or slow down significantly. There must therefore be a number of influencing factors. Researchers have identified the following seven parameters:

2.33 > Icebergs breaking off from the Jakobshavn Ice Stream in western Greenland are not uncommon. In July and August 2015, however, Greenland's fastest-moving glacier lost an unusually large amount of ice, which ultimately drifted out to sea.

1. Topographic constraint

The presence of a valley in the underlying bedrock restricts the ice masses at depth. In order to maintain speed with the upper layers, the ice masses at depth have to flow faster. Furthermore, the total friction surface at the base is larger. This generates more heat, which causes the ice on the underside to melt, and likewise increases the speed of flow. The best-known example of an ice stream whose origin can be related to topographic constriction is the Jakobshavn Ice Stream in western Greenland. At depth, its ice masses flow through a valley that is up to 2000 metres deep in some places, facilitating a velocity of the ice stream of as much as 17 kilometres per year.

2. Topographic steps

When an ice sheet flows across a steep cliff or similar abrupt topographic step, the deformation and acceleration of the ice is especially enhanced because of its great weight and the pull of gravity. At the same time, it is warmed, which further facilitates the amount and rate of deformation. The total flow speed of the ice thus increases. Well known glaciers that accelerate in this manner include the Byrd and the Thwaites Glaciers in West Antarctica.

3. Unevenness of the bedrock

There is still very little known about the topography below the large ice sheets. Researchers assume with some confidence, however, that various surface features such as rock outcrops, hills and small ditches can have a considerable influence on the flow velocity and direction of an ice stream. The more of these that are present, producing an uneven gliding surface, the slower the ice masses flow. In other words, the ice masses can flow more easily on a smooth bedrock surface than on a coarse one. This effect is observed in the Möller Ice Stream in West Antarctica.

4. Break-off of icebergs

When icebergs break off at the calving front of a floating glacier or ice shelf and ice is lost, a self-sustaining process is initiated. First, the ice masses in the stream behind the front accelerate. With this motion the ice of the entire stream warms up so it deforms more easily. Furthermore, on its underside, due to the increased friction, more lubricating meltwater is produced on which the ice masses can glide. These two processes result in an increase in the speed of the ice stream. The Jakobshavn Ice Stream in West Greenland again shows how effective this self-reinforcement can be. After an unusually high number of icebergs calved at its head between 1992 and 2004, such that the glacier tongue barely reached the fiord, its flow speed tripled to 17 kilometres per year.

5. Deformable sediments beneath the ice sheet

The ground below the ice sheet is not composed of hard bare rocks everywhere. In many places the upper ground layer is predominantly made up of gravel or other fine-grained sediment deposits. On this kind of soft ground the ice masses of an ice sheet glide much more easily than on a hard surface for three reasons:

- Firstly, the sediments, as a covering layer, smooth out existing irregularities in the subsurface and thus reduce its unevenness.
- Secondly, a sediment layer saturated with melt water creates an optimal sliding surface. Anyone who has slipped in the mud as a child knows that this sliding effect is not experienced on a dry, paved or asphalted surface.
- And thirdly, sediment deposits are easily deformed under the weight of the ice. Under some circumstances, they may even slip themselves and thus clear the way or accelerate the ice flow.

Researchers have found evidence for these three explanations in the Whillans Ice Stream in West Antarctica. Its ice masses lie atop a five-metre-thick unfrozen sediment layer that separates the ice stream from the rocky substrate below and accelerates its flow speed in the ways described above. Some scientific opinion now even suggests that a soft subsoil is one of the basic prerequisites for the generation of ice streams. According to one hypothesis, only where the ice sheets are underlain by sediment deposits, usually in basins, can ice streams originate and their ice masses glide at great speeds. This, however, is not consistent with the fact that there are also ice streams overlying bare rock that still carry large amounts of ice with them. The true role of the subsurface will therefore remain an object of research for some time to come.

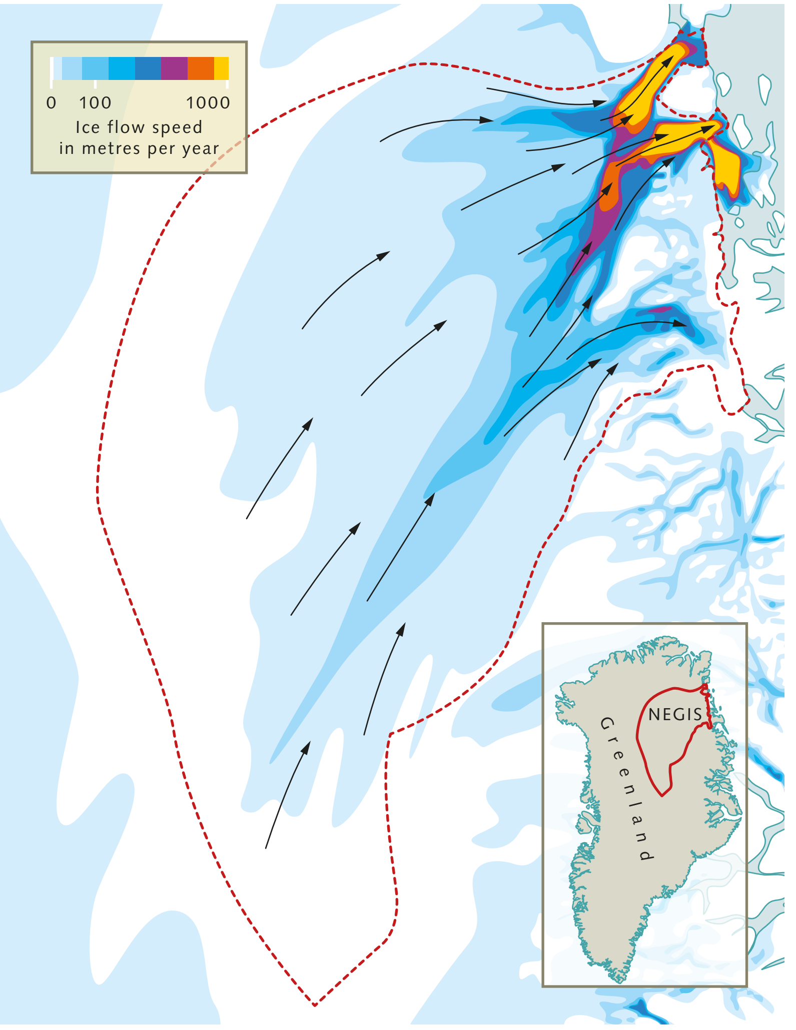
6. Geothermal heat

The larger the meltwater film is on its underside, the faster glacial ice moves. Meltwater, in turn, is produced by heat, which can also originate from within the Earth. This geothermal heat plays an important role, particularly in regions where active volcanoes are located beneath the ice sheet, or where the Earth's crust is especially thin. Scientists have found evidence for both of these phenomena in West Antarctica. Geothermal heat has also been suggested as a possible explanation for the origin of the Northeast Greenland Ice Stream (NEGIS). This is Greenland's only ice stream. Its catchment area covers twelve per cent of the total area of Greenland's inland ice and it is the connection between the ice and the ocean. In the area of its origin, the Earth's crust releases almost 20 times more heat than Greenland's overall average.

7. Meltwater lakes and rivers

As more details about the topography of the land surface beneath the ice sheets in Antarctica and Greenland are discovered, it is becoming clear that some ice streams originate in regions where the subsurface gradient alone is not sufficient to initiate the flow of ice. The area where the Recovery Ice Stream begins in East Antarctica is just one example of many. Theoretically the ice in this part of East Antarctica should hardly move at all. In fact, however, the stream transports its ice masses at a speed of ten to 400 metres per year from the high plateau of the East Antarctic Ice Sheet down towards the Weddell Sea. Its catchment area spreads inland for about 1000 kilometres from the Filchner-Ronne Ice Shelf on the coast, and is equal to an area almost three times as large as Germany. It is an enormous ice stream that researchers previously thought might be receiving the decisive impetus for its formation from overflowing meltwater lakes beneath the ice sheet. The basic idea was that these lakes occasionally overflow, creating a lubricating film on which the ice sheet slides like an aquaplaning car.

The existence of subglacial lakes in Antarctica is known from Russian and British research projects at Lake Vostok and Lake Ellsworth. Both of these water bodies formed in depressions beneath the ice sheet. Over the course of many millennia, so much melt water has accumulated in them that, as a rule, they are larger than Lake Constance. But the assumption that these kinds of huge lakes are present in abundance beneath the Antarctic Ice Sheet, and that they are responsible for initiating the ice streams could not be confirmed by German polar scientists in a field study of the Recovery Ice Stream. Every-

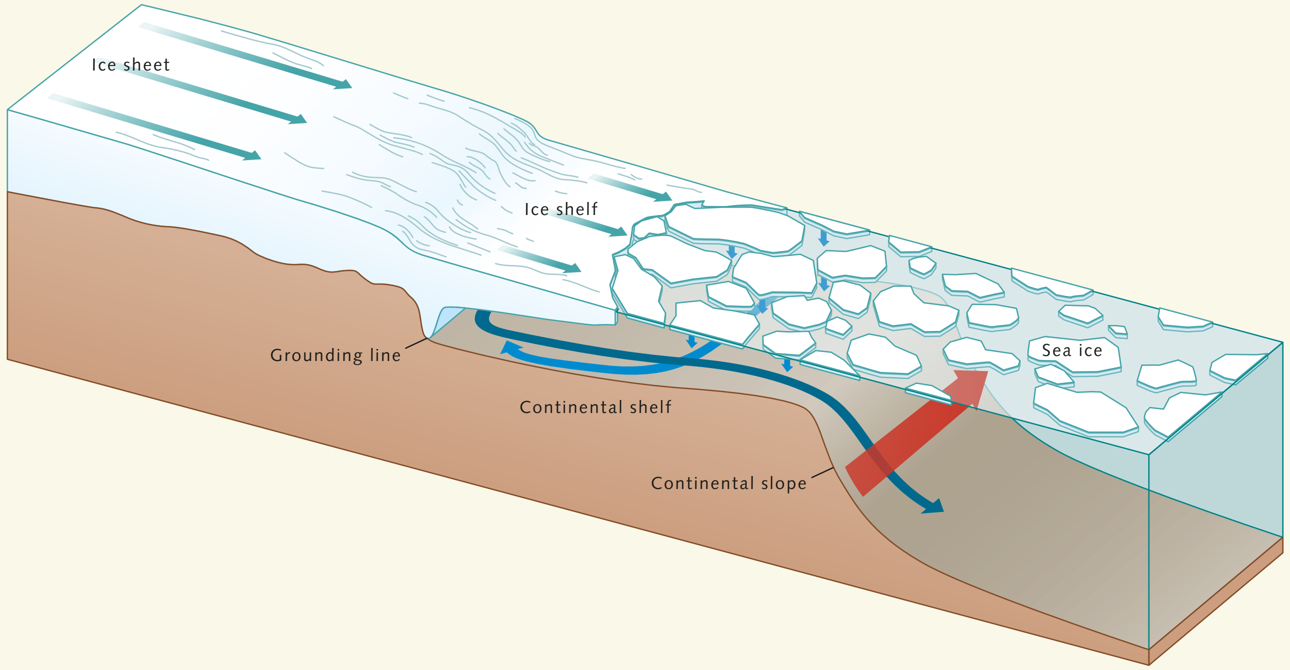


2.34 > The NEGIS catchment area extends far into the interior of Greenland. From there it transports the ice masses towards the northeast where they reach the sea through three glaciers.

Ice shelf and ocean – a give-and-take relationship

By definition, ice shelves come into large-scale contact with the sea. The Southern Ocean water masses have a decisive influence on the stability and mass balance of the ice sheets. As an example, on the continental shelf of the southern Weddell Sea, where the sea floor slopes towards the land, highly saline, cold shelf water with a temperature of about minus 1.9 degrees Celsius flows just above the seabed for a distance of up to a thousand kilometres, reaching far under the Filchner-Ronne Ice Shelf. The further the water masses penetrate under the ice sheet, the more destructive they are for the ice. This is because the water sinks deeper with every metre that it travels towards the coast. The water pressure under the ice shelf therefore increases and, correspondingly, the freezing point of the water drops from minus 1.9 degrees Celsius to minus 2.5 degrees Celsius. The result of this change is that the cold shelf water deep under the ice sheet does not freeze, but releases its residual heat to the ice and cools down even further. The loss of heat has two consequences: First, the coldest water masses in the world, known as ice-shelf water, form underneath the ice shelf. Its initial temperature is minus 2.5 degrees Celsius. Second, the ice shelf melts from the bottom (basal melting) because of the heat released by the inflowing water. When the ice melts, freshwater is released, thus diluting the super-cold ice-shelf water. Its density decreases and it rises up to meet the underside of the ice shelf. It then flows back to the shelf-ice margin.

On its way there, the freezing point of the ice-shelf water continues to rise due to decreasing pressure. As a result, salt-free ice crystals form in the super-cold ice-shelf water, which then rise to attach to the underside of the ice and freeze there. Scientists refer to this new ice as marine ice. The remaining water masses continue to drift further. Measurements at the shelf-ice edge have revealed that the ice-shelf water flows out from beneath the ice with a temperature below minus 2.2 degrees Celsius, and that it ultimately becomes a part of the Antarctic Deep Water. Ice shelves are therefore intimately interconnected with the deep ocean. While the ocean regulates the thickness of the ice shelf, the ice shelf cools the migrating water masses of the shelf sea and contributes to driving the thermohaline circulation.



2.35 > When sea ice forms in the Weddell Sea, cold, highly saline shelf water is created that flows beneath the Filchner-Ronne Ice Shelf where it cools still further. It ultimately flows down the continental slope as super-cold ice-shelf water.

where the researchers predicted the presence of water under the ice they were unable to detect it.

The exact role of subglacial lakes is thus still uncertain, as is the question of the general distribution of meltwater under the ice sheets. The paths of presumed streams and rivers beneath the ice have so far only been predicted by computer simulations. Initial efforts have also been made to derive this information from satellite data. However, measurement methods for detecting and mapping large-scale lake chains or river networks are not yet available.

How ice shelves retard glacial flow

More than half of the Antarctic coast is bounded by ice shelves. The more than 300 floating glacier tongues are all extensions of one or more glaciers that slowly push their coherent ice masses out into the Southern Ocean. The leading edge of the Larsen C Ice Shelf in the western Weddell Sea, for example, moves at a rate of around 700 metres annually. Expansion of the ice sheet is limited only by the loss of ice due to icebergs breaking away from the calving front at regular intervals. On large ice shelves, it can take more than a thousand years for an ice crystal to travel through the entire ice shelf and commence the final stage of its journey aboard an iceberg.

Ice shelves in the Antarctic are, as a rule, between 300 and 2500 metres thick, although they become thinner the further they extend out into the sea. They are thickest at the grounding line, the furthest seaward point where the ice is still in contact with the bottom and where it begins to float. In the Antarctic region ice shelves cover a total area of 1.3 million square kilometres. The largest ice shelf, the Ross Ice Shelf in the Ross Sea, is almost as large as Spain.

Ice shelves are fed primarily by the ice of the glaciers and ice streams behind them. However, their volume can also increase when snow falls on the ice shelf or the offshore sea ice and subsequently condenses in some areas to form firn and ice. In other places seawater can freeze onto the underside of the ice shelf and contribute to the growth of the ice tongue. Ice shelves lose ice through the calving of icebergs, but warm water masses can also melt the ice tongues from below. Researchers refer to this process as basal melting of the ice shelf.

The ice shelf is considered to be in a state of equilibrium if it loses the same amount or less ice than flows in through the glaciers. In this state, the floating ice tongues can survive for several millennia. But if the rate of ice loss increases abruptly there is reason for concern, because the ice shelves perform a critical and elementary function in the Earth's climate system. They inhibit the flow of further ice masses from the interior and thus also slow the rise in sea level.

To clearly understand this role, one has to look again at their formation. As floating extensions of one or more glaciers, the ice masses of the ice shelf have a long journey behind them, from the high plateau in the Antarctic interior, through ice streams and glaciers down to the sea. Then, extending out from the coast as large floating sheets and pushed out into the sea, the ice can get caught up on islands or rocks. Ice shelves can sometimes skim over flat obstacles, or they may collide with an island that abruptly applies the brakes to the ice flow. The thicker the ice shelf is, the more effective it is at holding back the inland ice masses.

The amount of pressure the ice shelves have to withstand is perhaps best illustrated by the fact that, through the glaciers and ice shelves, 74 per cent of Antarctica's inland ice reaches the sea. When the Larsen B Ice Shelf on the Antarctic Peninsula broke into thousands of icebergs in 2002, which led to a loss of its braking function, the flow rate of the glaciers behind it increased by a factor of three to eight times within the following 18 months.

Floating glacier tongues are also found in the Arctic, of course, especially in Greenland and on the coast of Canada's Ellesmere Island. These ice areas, however, which are firmly attached to the land, are not usually referred to as ice shelves because they primarily occur in fiords and the width of their expansion is thus limited by land. For this reason, specialists refer to this floating ice from the land as ice tongues. The Ward-Hunt Ice Shelf off the coast of Ellesmere Island is an exception. It is made up of consolidated sea ice, onto which snow fell and was compacted into ice. This ice mass is therefore not land ice and so it is distinctively different from the large ice sheets of Antarctica.

The drifting paths of icebergs

The calving of substantial icebergs at the leading edge of a glacier or ice shelf is a completely natural process. At regular intervals, ice shelves in the Antarctic release tabular icebergs with surface areas that can be as large as cities such as Hamburg or Berlin. The size of an iceberg also determines its subsequent fate, at least in the Antarctic. Icebergs that are less than two kilometres long or wide drift away from the edge of the ice shelf or glacier and out of the coastal region within a few months. Thereupon, offshore winds force them out onto the open sea, where they break into smaller pieces and melt within one to two years.

But the offshore winds play a less important role for icebergs that are larger than two kilometres in diameter. Their movement, in contrast to their larger siblings, is primarily driven by their own weight. To understand this phenomenon, one needs to realize that the Southern Ocean is not actually a flat surface. Because of the prevailing winds, its surface may be as much as 50 centimetres higher near the coast. Large, freshly calved icebergs can slide down this incline of the sea surface. Their path does not follow a straight line, however, but forms an arc due to the Coriolis force. So the icebergs are deflected towards the coast. This means that they remain within the cold coastal current for a long time and often do not reach the warmer, more northerly waters until years later, when they finally break apart and melt.

The speed at which the icebergs travel along their paths can be influenced by the topography of the sea floor. Large icebergs may often run aground and remain trapped for an indeterminate length of time. In addition, the giant icebergs often freeze onto sea ice, so that the waves can no longer strike their flanks and the effect of erosion is reduced. Scientists have tracked the pathways of drifting Antarctic icebergs and produced computer models to calculate them. Depending on the marine area in which the giant icebergs have calved, they take one of four major routes that all drifting ice follows, both sea ice and icebergs, into warmer climes. GPS data have shown that one large iceberg has even been able to completely circumnavigate Antarctica. It started in the Weddell Sea, was driven northward along the east coast of the Antarctic Peninsula, then turned back to the east and drifted once around the continent before finally melting north of the Antarctic Peninsula.

From the Arctic glacier tongues, it is more common for a fleet of numerous smaller icebergs to calve instead of a few large ones. The winds drive them out of the fiords onto the open sea where they normally drift southward with the coastal current. Many of the icebergs that reach the shipping lanes off the southern coast of Newfoundland originate from the Jakobshavn Ice Stream in western Greenland. In 2018 alone, more than 500 icebergs from the west coast of Greenland drifted into the coastal areas of Newfoundland and Labrador. In the record year of 1984 there were 2002 icebergs. For most of them this journey lasted from one to three years.

Researchers believe that there is a correlation between the prevailing atmospheric current conditions over the North Atlantic and the number of icebergs drifting so far to the south. If onshore winds blow along the coast of Labrador in winter, warmer sea air reaches this region. That air prevents the formation of sea ice. As a result, the drifting icebergs are exposed to greater levels of destructive wave power. In addition, the onshore winds push them the into shallower waters where the ice masses run aground.

If the large air-current patterns reverse, a strong cold westerly wind blows over Labrador. Icy air reaches the region in its wake. Sea ice forms from the seawater, protecting the icebergs from extensive destruction. In the following summer, they then begin their southward journey unhampered and in large numbers. But icebergs also calve on the east coast of Greenland. On 22 June 2018, for example, the Helheim Glacier lost a six-kilometre-long strip of ice in a single stroke. Greenland-wide, it was the largest iceberg to break off in the past ten years.

Conclusion

A chain reaction with an icy end

It is so extremely cold in the polar regions because of a self-reinforcing process that involves several factors. Fundamentally, it is due to the fact that much less solar energy reaches the Earth's surface in the polar regions than it does, for example, in Central Europe or at the equator. The reasons for this are the low angle of incoming sunlight, the tilt of the Earth's axis and the orbit of our planet around the sun. This combination of factors results in the polar regions being generally undersupplied with energy compared to the rest of the world, and being completely cut off from the sun's heat during the polar nights.

While the polar regions receive sparse solar radiation, the tropics receive a great deal, which results in a marked temperature contrast between the two regions. The large air and ocean currents that we see today are generated as responses to compensate for this difference. They distribute heat from the tropics around the globe and thereby determine the weather conditions throughout the world. Without the cold regions in the far north and south, these global circulation patterns of air and water masses would not exist. It is also important to note that strong bands of wind form in both hemispheres that act as protective walls to prevent heat from the tropics from reaching deep into the polar regions.

However, the freezing conditions in the Arctic and Antarctic also mean that precipitation in these regions is primarily in the form of snow, and that large areas of the polar surface waters freeze over in winter. Because the white snow and ice covers have a high reflective capability, called the albedo, a large proportion of the solar radiation is not absorbed, and thus cannot contribute to warming the surface of the Earth. In this way, the snow and ice surfaces amplify cooling in the polar regions. To scientists this kind of effect is known as positive feedback.

The fact that cold air cannot hold large amounts of water vapour is another factor that facilitates the low temperatures. Especially over central Antarctica, the air masses lack this important heat reservoir, and thus also the capability to form a thick cloud cover. This could otherwise help to limit the cooling. Instead, the dry air amplifies the cooling effect and, in concert with the other factors, helps to create the ideal conditions for the formation of immense ice sheets, glaciers and areas of sea-ice.

The various forms of polar ice and their strong albedo are fundamental components of the cooling and climate system of our Earth. They regulate chemical and biological cycles and interact very closely with the ocean, the atmosphere and the land. However, there are numerous geographical differences between the Arctic and Antarctic regions. In the Arctic Ocean, with its shallow shelf seas, the water masses circulate in a completely different way than they do in the Southern Ocean, a ring ocean that surrounds a large continent. These regional differences likewise affect the polar ice masses. In the Antarctic the conditions result, among other things, in large-scale melting of the Southern Ocean sea ice in summer. In the Arctic, on the other hand, slightly less than half of the sea ice survives the summer, and researchers there refer to a permanent sea-ice cover. The amounts of precipitation that fall in the Arctic and Antarctic are also different, resulting in different rates of growth on the ice sheets of Greenland and Antarctica. There is, however, one thing that the ice masses of the two regions have in common – they both react very sensitively to increasing temperatures.