that can be swept by landfalling tropical cyclones is unlikely to remain undamaged for more than a few decades, at most. Thus, some damage can be avoided by not building in vulnerable areas. As another example, there are several ways in which homes can be built to resist tornado damage (**Figure 6**), unless the homeowner is unlucky enough to be hit by the most intense winds in a violent tornado. Even within the whole violent tornado damage area, only a few places will actually experience the most violent winds; most of the rest of the structures will encounter winds that can be resisted through appropriate construction.

Mitigation of casualties can also be a complex undertaking. In some instances, as with tropical cyclones, evacuation is possible and may be the best way to protect lives when it is feasible. For tornadoes, access to a suitable shelter is preferred; in situations where proper shelter is not available, the alternatives during tornadoes are not very good. In flooding situations, evacuation to higher ground is the appropriate way to prevent casualties, when time permits. Clearly, our ability to detect and predict severe storms is also important for casualty mitigation. In the United States, there has been a gradual reduction in weather-related fatalities with time, in part because there are fewer 'surprise' storms today and in part because education about severe storm hazards has led to improved public preparations. Nevertheless, we continue to be vulnerable to disasters caused by severe storms, and complacency can be a fatal error.

See also

Air–Sea Interaction: Storm Surges. **Bow Echos and Derecho**. **Convective Storms:** Overview. **Cyclogenesis**. **Cyclones, Extra Tropical**. **Downslope Winds**. **Flooding**. **Lake Effect Storms**. **Mesoscale Meteorology:** Mesoscale Convective Systems. **Microbursts**. **Orographic Effects:** Lee Cyclogenesis. **Polar Lows**. **Tornados**. **Waterspouts**.

Further Reading

Agnone JC (ed.) (1995) *Raging Forces: Earth in Upheaval*. Washington DC: National Geographic Society.

Anthes R (1982) *Tropical Cyclones. Their Evolution, Structure and Effects*. Boston: American Meteorological Society.

Church C, Burgess D, Doswell C and Davies-Jones R (eds) (1993) *The Tornado: Its Structure, Dynamics, Prediction, and Hazards*. Washington, DC: American Geophysical Union.

Doswell CA III (ed.) (2001) *Severe Convective Storms*. Boston: Amererican Meteorological Society.

Foote GB and Knight CA (eds) (1977) *Hail: A Review of Hail Science and Hail Suppression*. Boston: American Meteorological Society.

Hill CE (ed.) (1986) *Nature on the Rampage: Our Violent Earth*. Washington, DC: National Geographic Society.

Junger S (1997) *The Perfect Storm*. New York: W.W. Norton.

Lamb H (1991) *Historic Storms of the North Sea, British Isles and Northwest Europe*. Cambridge: Cambridge University Press.

Lorenz EN (1993) *The Essence of Chaos*. Seattle: University of Washington Press.

Ludlam FH (1980) *Clouds and Storms*. Philadelphia: Pennsylvania State University Press.

Ray PS (ed.) (1986) *Mesoscale Meteorology and Forecasting*. Boston: Amererican Meteorological Society.

Shapiro M and Grønås S (eds) (1999) *The Life Cycles of Extratropical Cyclones*. Boston: American Meteorological Society.

SNOW (SURFACE)

M Sturm, US Army Cold Regions Research & Engineering Laboratory-Alaska, Fort Wainwright, AL, USA

Introduction

Snow blankets more than half of the Northern Hemisphere each winter, remaining in place for periods ranging from less than a month (typically south of 40° N) to more than 8 months of the year (typically north of 60° N). In the Southern Hemisphere, the coverage is less extensive, but still substantial. If the perennial snow covers of the Greenland and Antarctic ice sheets are included, along with the seasonal snow cover that forms on lake and sea ice, then the total percentage of the Earth's surface covered by snow during some period of each year is considerable. This blanket of snow is a complex, layered material that can exhibit a high degree of spatial heterogeneity. Year-to-year variations in coverage and properties can be large and they have a direct and immediate impact on the Earth's climate. In this article, the major types of snow cover are introduced and the layered nature of the snow is discussed. The role of the snow in moderating the exchange of energy and mass with the atmosphere is also described.

Snow Cover and Its Importance

The term 'snow cover' is directly analogous to the term 'formation' when discussing layered sedimentary or metamorphic rocks. Both the sequence and character of the layers, and the lateral variation of each layer (facies changes), contribute to the overall properties of the formation. Similarly, the bulk physical and thermal properties of a snow cover, the properties that are of importance in moderating the exchange of energy and mass between the Earth and the atmosphere, are an aggregate of the properties of the individual layers. For each layer, these properties are the result of the conditions (snowfall, wind, temperature) that prevailed when the layer was deposited, and the post-depositional conditions (temperature, temperature gradients, snow overburden, liquid water percolation, solar radiation) to which the layer was subjected after deposition. Because both deposition and post-deposition conditions vary across the landscape, the layers themselves vary. In order to understand the role of snow cover in atmospheric processes, the layered nature and spatial variability of the material need to be considered.

Much of the impact of snow on climate and atmospheric processes arises because of its high albedo and low thermal conductivity. Snow cover reflects up to 85% of incoming short-wave solar radiation, significantly reducing winter temperatures and retarding melting in the spring. At the same time, snow is an excellent insulator, so it can effectively lower the rate of heat loss from the ground or an underlying ice surface, thereby maintaining higher winter soil temperatures or retarding the rate of sea and lake ice growth. The total winter energy exchange across a snow cover is a complex balance between these two competing processes. Snow cover is also important because it traps aerosols and other atmospheric particulates like a filter, storing these until the snow melts, then releasing them abruptly. Snow can control, through thermal and physical means, the release of trace gases like CO_2 from subnivian plants and soils during the winter, and it functions as a temporary storage reservoir of water, stockpiling winter precipitation then allowing it to run off in a much shorter period of time than it otherwise would have had it not fallen as snow. In some cases in higher latitudes where the snow lasts many months, as much as 80% of the annual river discharge can be from snow melt, and this discharge may occur in a period of less than two weeks.

Perennial and Seasonal Snow Covers

Because of their fundamentally different layered structures, it is customary to distinguish between perennial and seasonal snow covers. Seasonal snow covers are deposited in the fall and melt away completely each spring; therefore they never become very deep. Perennial snow covers form at higher levels on glaciers and ice sheets, where the combined decrease in temperature and increase in snowfall precipitation with altitude is sufficient to allow winter snow accumulation to survive the summer melt. Snowfall of the following winter is deposited on the residual snow of the previous year, forming a sequence of annual layers of snow that can be tens to hundreds of meters thick before compaction at depth converts the snow into glacier ice.

Separate but related climate classification systems for perennial and seasonal snow covers have been suggested and are useful when thinking about both local and global variations in snow cover. For the perennial snow on glaciers and ice sheets, increasing elevation results in a decrease in melting. As a consequence, snow characteristics vary with elevation (**Figure 1**). At the lowest elevation, the melt removes all of the winter snow, and a seasonal rather than perennial snow pack forms each year. Higher, the snow pack survives the summer melt, but percolation of melt water into the snow pack and subsequent refreezing produce extensive icy features like the ice lenses and percolation columns. At the highest elevations, no melting takes place and the dry snow facies is observed. On a steep alpine glacier, the entire sequence is compressed into a distance of tens of kilometers. On ice sheets, sequence may spread over distances of hundreds of kilometers.

For seasonal snow covers, local climate rather than elevation determines the prevailing snow cover characteristics, and this local climate can be represented by three simple binary variables: winter temperature, winter precipitation, and wind. High and low values for each of these variables (**Figure 2**) define eight possible types of seasonal snow covers, most of which have a counterpart in the glacier facies system shown in **Figure 1**. For example, under warmer, wetter winter conditions, a maritime snow cover will develop. This snow cover tends to be deep (> 1 m) and warm (near or at freezing temperatures), and exhibits similar icy features to those observed in the percolation facies on glaciers (compare **Figure 3** to **Figure 1**). Similarly, alpine, tundra, and taiga snow cover classes exhibit features found in the dry snow facies on glaciers. The

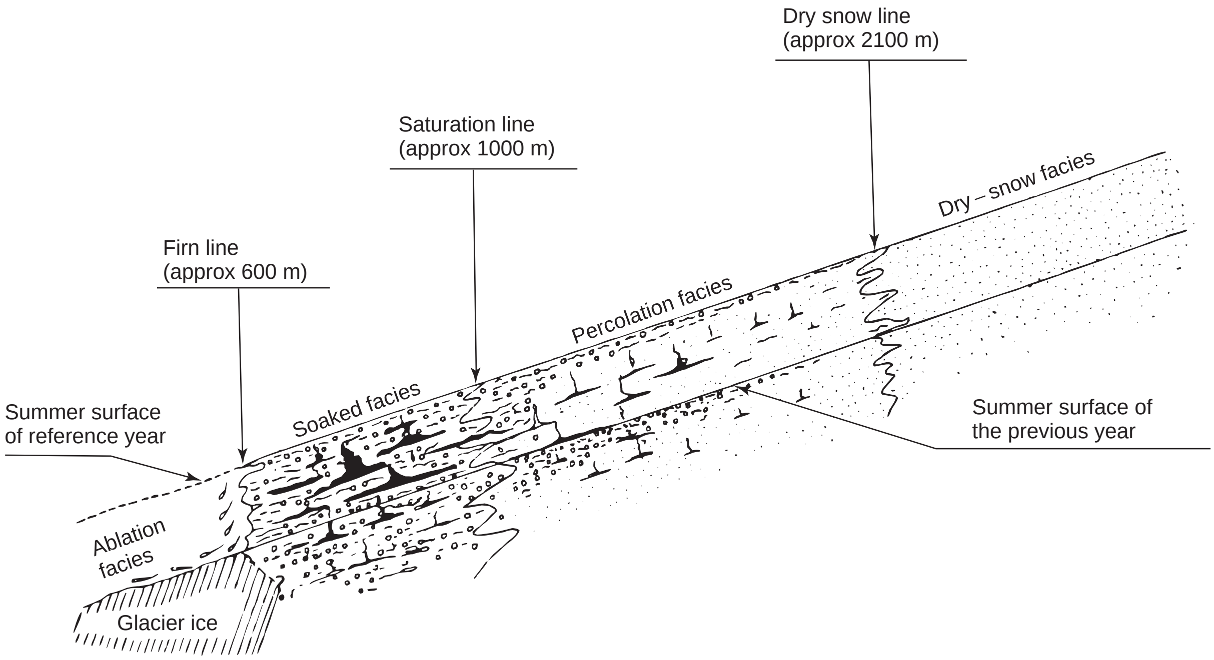

Figure 1 The glacier facies classification of Benson (1962), describing variations in the characteristics of the perennial snow cover found on glaciers and ice sheets. With increasing elevation, there is a decrease in the amount of melting and, as a consequence, a decrease in the amount of icy features in the winter snow pack. At the lowest level, all of the winter snow melts in the summer and the snow cover is essentially seasonal; at the highest level, no melting takes place and the snow has no features in it related to melting. (From Benson CS (1962) Stratigraphic studies in the snow and firn of the Greenland Ice Sheet. *SIPRE Research Report 70*, CRREL.)

stratigraphic diagram and key in **Figure 3** suggest the main snow cover characteristics associated with each climate class for seasonal snow.

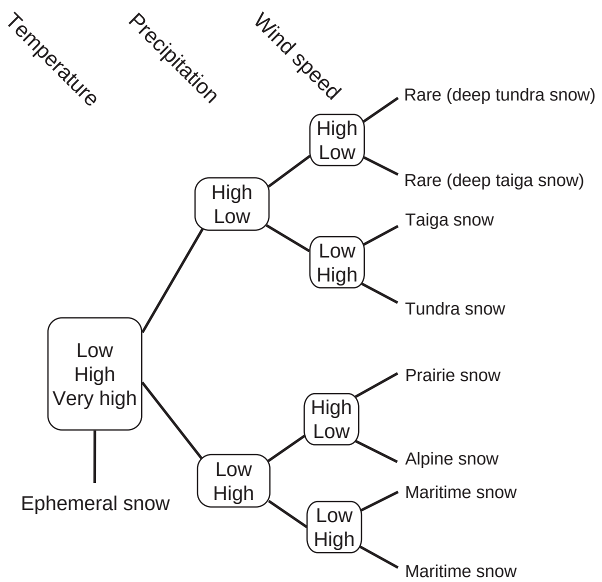

Figure 2 A dichotomous classification of seasonal snow covers based on winter temperature, precipitation, and wind. In **Figure 3**, a typical snow stratigraphy for each class is shown. Broad similarities in snow characteristics exist between the seasonal snow classes and the glacier facies shown in **Figure 1**. (From Sturm M, Holmgren J, Liston G (1995) A seasonal snow cover classification system for local to global applications. *Journal of Climate* 8: 1261–1283.)

Layer by Layer Development of a Snow Cover

Snow cover builds up layer by layer. The initial characteristics of each layer are determined by how much solid precipitation falls, whether the precipitation is accompanied by wind, and the prevailing temperature at the time of deposition. After deposition, each layer is subjected to mechanical and thermal metamorphic processes that alter the layer characteristics. These vary in intensity and duration depending on when the layer was deposited, its height in the snow pack and the number of overlying layers, the prevailing conditions at the snow surface, and the temperature and temperature gradients in the snow pack as a whole. At any given time, the characteristics of each layer in the snow are a product of its initial deposition and post-depositional metamorphism.

Layer Deposition and Densification

Almost 80 different types of falling snow crystals have been identified. The particular crystals that accumulate at the Earth's surface in a snow storm are determined by the temperature and humidity in the layers of air through which the crystals fall and grow. However, crystal form is far less important than the rate of snowfall, the wind speed, and the temperature in determining the initial characteristics of a snow

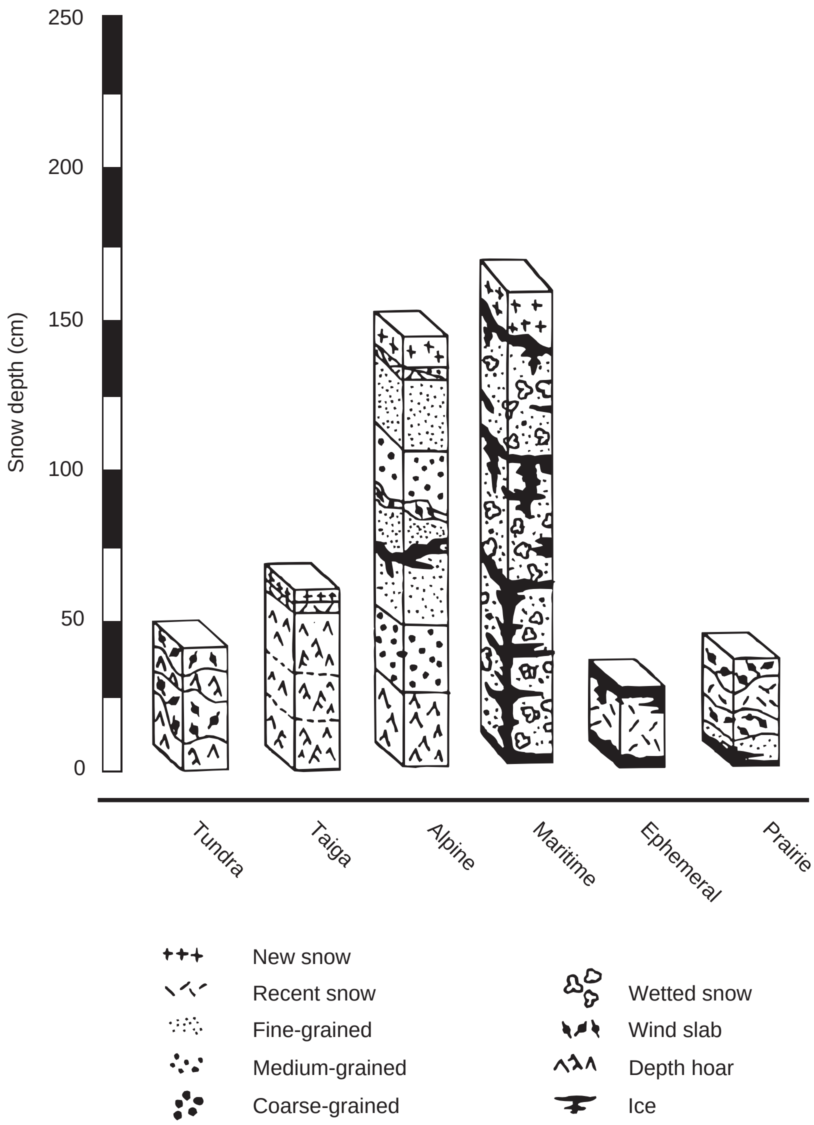

Figure 3 Typical snow stratigraphy for the six seasonal classes listed in **Figure 2**. (From Sturm M, Holmgren J, Liston G (1995) A seasonal snow cover classification system for local to global applications. *Journal of Climate* 8: 1261–1283.)

layer. In general, low temperatures, low wind, and low rates of snow fall produce the lowest-density layers of new snow (**Table 1**).

Once deposited, new snow layers densify rapidly. Initially, much of this densification is a result of thermodynamic instability. The sharp points and intricate branches of newly fallen snow crystals have high radii of curvature; the water vapor pressure over these highly curved surfaces is greater than elsewhere, so there is a net loss of water molecules from pointed areas to the air spaces in the snow, or to other areas on crystals that have lower degrees of curvature. The crystals rapidly break down and the resulting fragments become more rounded (**Figure 4**). The breakdown reduces the size of the crystals, increases the number of individual snow grains, and decreases the degree to which the crystals interlock. As a result, the entire snow layer settles.

Table 1 The density of newly deposited snow

Deposition conditions	*Density (g cm^{-3})*
No wind, low rate of snowfall, cold	0.02–0.05
Low wind, low rate of snowfall	0.05–0.10
Moderate wind, high rate of snowfall	0.20–0.35
Moderate wind, low rate of snowfall	0.35–0.40
High wind	0.40–0.55

As additional new layers of snow are added to the snow pack, the overburden load (σ) on buried layers

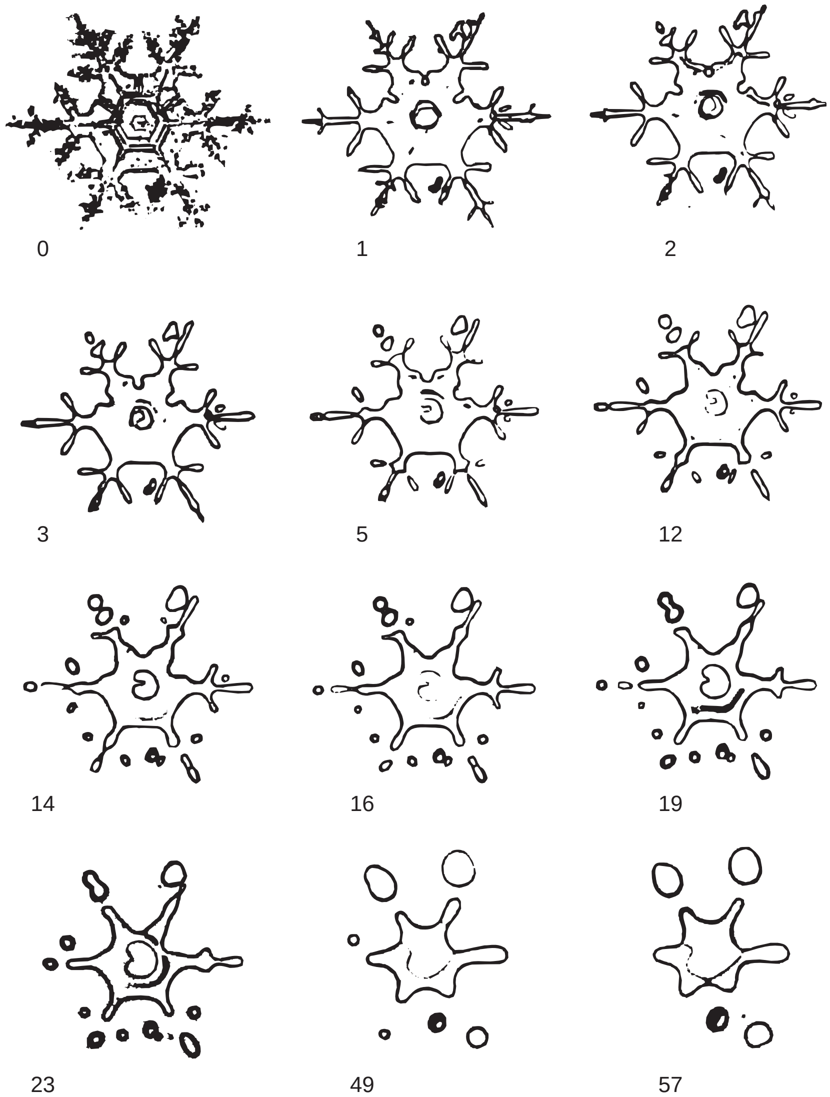

Figure 4 Changes in a snow flake held at a constant temperature of $-11.5°C$ for a total period of 57 days (indicated by small numbers). The snow flake grew in the atmosphere under conditions of supersaturation with respect to water vapor. Once deposited, the sharp points and thin branches were thermodynamically unstable and the snow flake metamorphosed, even in the absence of a temperature gradient or overburden stress. (From Bader H, Haefeli R, Bucher E, Neher J, Eckel O, Thams C (1939) *Der Schnee und seine Metamorphose* (Snow and its Metamorphism), US Army SIPRE Translation 14, 1954.)

increases. For these layers, compaction due to vertical stresses begins to dominate the snow densification process. The response of the snow to these stresses has been modeled by assuming the snow layer behaves like a viscous fluid (eqn [1]).

$$-\frac{1}{h}\frac{\mathrm{d}h}{\mathrm{d}t} = \frac{1}{\rho}\frac{\mathrm{d}\rho}{\mathrm{d}t} = \frac{\sigma}{\eta_c} \qquad [1]$$

In eqn [1] h is the thickness of the layer (m), t is time (s), ρ is the layer density ($\mathrm{kg\,m^{-3}}$), and η_c is the compactive viscosity. Values of η_c (Pa s) have been determined from observations of the settlement of natural snow layers, from uniaxial strain compressive tests, and from depth–density profiles on glaciers and ice sheets. The combined results show wide scatter, but individual sets of data are usually fitted to the relation in eqn [2], where k is a factor that depends on the type of snow cover (**Figures 1** through **3**).

$$\eta_c = \eta_0 e^{k\rho} \qquad [2]$$

The effective viscosity term incorporates a number of physical mechanisms including gravity-driven movement of snow grain centers of mass toward each other, vapor and volume diffusion, and sintering.

Table 2 Compactive viscosity factors for three classes of snow cover

Snow cover type	*k-value* ($m^3\,kg^{-1}$)
Maritime	18–22
Alpine/taiga	35–60
Tundra	>70

Not surprisingly, viscosity factor values vary widely depending on the temperature, liquid water content, and grain characteristics of the snow – i.e., the snow cover class (**Table 2**). Colder, drier, finer-grained layers of snow tend to be more viscous than warmer, wetter, layers with larger grains, and therefore compact more slowly.

In the absence of melting or the introduction of liquid water, snow layers will continue to densify until they reach a limiting density of about $0.6\,g\,cm^{-3}$. By this time, the snow grains will have metamorphosed until they have become highly rounded, a shape that minimizes their surface free energy. The rounded grains will be in close contact with each other, and the grain arrangement will approximate that of hexagonal close-packing of ice spheres. Further densification will require actual deformation of the individual grains of snow, or the influx and refreezing of melt water in pore spaces. The overburden stresses required to achieve this further deformation are only realized in the deep perennial snow packs found on glaciers and ice sheets.

Snow layers deposited during windy conditions (wind slabs) have much higher initial densities than other new snow layers. The wind tumbles snow crystals as it transports them, breaking the more fragile crystal junctions and pulverizing the crystals in general. The resulting grains are actually crystal fragments, often less than 0.1 mm in length, and these shardlike grains (**Figure 5**), when they come to rest, pack well and sinter together into a cohesive slablike layer. Initial densities for wind-transported layers of new snow range from 0.35 to 0.6. The upper limit occurs for the same physical reasons as discussed before. Due to their high initial densities and cohesiveness, wind slabs are highly resistant to compaction and often remaining at a fixed density after deposition.

There has been much discussion and experimentation to determine the wind speed necessary to transport snow. The transport takes place through three mechanisms: creep, saltation, and suspension. Creep consists of the rolling movement of grains along the snow surface under the action of the wind. Saltation is the movement of grains along the surface by jumping and ricocheting after impact by other grains. Suspension is the movement of grains in the wind stream at some level above the snow surface. The threshold shear velocity, u^*, at which transport occurs is usually

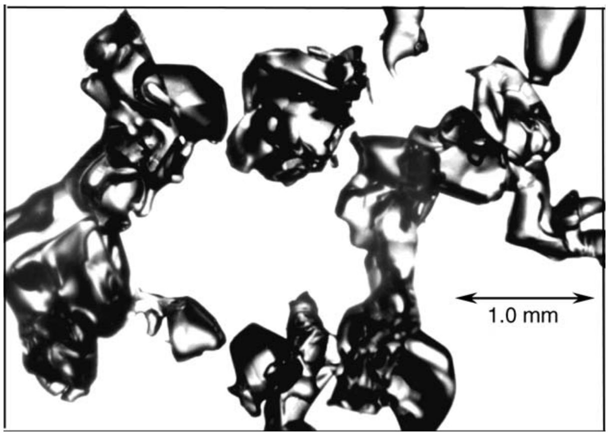

Figure 5 Wind-pulverized snow grains from Arctic Alaska, showing irregular shapes and thick bonds due to rapid sintering after deposition.

estimated by assuming a logarithmic-shaped wind profile and projecting the 10-m high wind speed (u_{10}) down to the snow surface (u^*). In general the value of u_{10} is between 18 and 30 times the value of u^*. Experimental studies indicate that when u_{10} is greater than $6\,m\,s^{-1}$ transport will occur if the snow has fallen recently. If the snow is new and falling while there is wind, transport will occur with wind of 5 or even $4\,m\,s^{-1}$. If the snow is aged, was previously transported by the wind, or has undergone some melt–freeze processes, speeds in excess of $30\,m\,s^{-1}$ may be needed before the snow will start to be tranported (**Figure 6**).

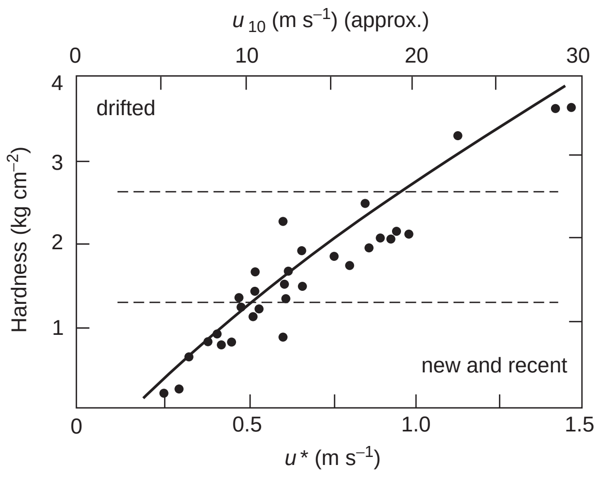

Figure 6 The critical wind shear velocity (u^*) as a function of snow hardness, which is a good measure of the type of snow. Increasing hardness, common for wind slabs and layers of snow that have undergone melt–freeze, requires considerably higher winds to mobilize these types of snow. u_{10} is the wind speed measured at a standard height of 10 m. (From Kind RJ (1981) Snow drifting. In: Gray DM, Male DH (eds). *Handbook of Snow*, pp. 338–359. Toronto: Pergamon.)

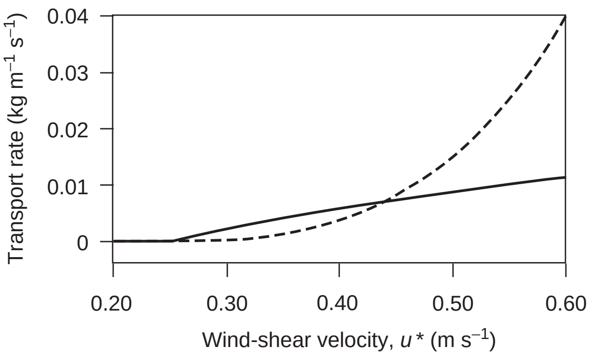

Figure 7 Snow transport rates for saltation (solid curve) and suspension (broken curve) as a function of wind shear velocity ($u*$). The wind speed at 10 m height is approximately 18–26 times $u*$. At $u* = 0.44$ (10 m height wind speeds of 8–11 m s^{-1}), suspension begins to transport the majority of the wind-borne flux of snow. (From Liston GE, Sturm M (1998) A snow-transport model for complex terrain. *Journal of Glaciology* 44: 498–516.)

In similar fashion, the flux of snow transported by the wind is a strong function of the wind speed, with increasing speeds producing a marked increase in the total amount transported (**Figure** 7). For values of $u*$ between 0.2 and 0.44 m s^{-1}, saltation dominates the transport, but for $u*$ values in excess of 0.44 m s^{-1}, suspension exceeds saltation in transporting snow.

One other consequence of wind transport of snow is the development of a wide range of drift deposit and erosion features at the snow surface. These features include ripple marks, dunes, barchans, and sastrugi. Surprisingly, little is known about the relationship between these features and the wind speed, the temperature, and the snow conditions necessary for their formation.

The final, and most efficient, method for densifying a layer of snow is through the infiltration of melt or rain water into the snow cover, followed by subsequent refreezing. Water can infiltrate, surround grains as thin films or lie in veins along grain junctions, and refreeze to produce large multiparticle grains. Water can also percolate downward in pipelike structures called percolation columns, or spread out along stratigraphic boundaries (owing to variations in the hydraulic conductivity of the snow). When this water refreezes, ice lenses and layers are created. Frequently, a single infiltration event will produce ice layers at multiple levels in the snow pack. Densities in excess of 0.6 g cm^{-3}, sometimes even as high as 0.9 g cm^{-3}, can result. This mechanism is commonly observed in ephemeral and maritime seasonal snow covers (**Figures 2** and **3**), and in the percolation facies for perennial snow (**Figure 1**).

Snow Metamorphism

In addition to compaction and densification, several other metamorphic processes can affect layers of snow. These processes result chiefly in changes in snow grain characteristics and bonding, which in turn affect the thermal conductivity, air permeability, and albedo of the snow. The processes are typically divided into 'wet' and 'dry' categories because different snow grain characteristics are produced depending on whether liquid water is present. Further metamorphic subdivisions are shown in **Table 3**.

For wet snow metamorphism, the degree to which grains and a snow layer are changed is mainly a function of how much water is present. For low liquid contents ($<5\%$ by weight), the water in the snow exists as thin films and isolated pockets or veins around grains; continuous ice grain and air space pathways still exist through the snow layer. This is called the pendular regime. Under this regime, snow grains will rapidly round, and clusters of grains, looking much like bunches of grapes, will form as a result of the minimization of surface free energy. The clusters themselves are quite strong because the bonds between the spherical grains are still intact and substantial. The wet snow pack will have considerable bearing strength. Spring skiing, which can be excellent, takes advantage of these ball-bearing like grain clusters and the general strength and cohesiveness of this type of wet snow metamorphism. If the temperature of the snow drops and the grain clusters freeze, they will take on the slightly more amorphous shapes of melt–grain clusters (**Figure 8**), while at the same time the strength of the layer will increase dramatically as all the interstitial water freezes. For higher liquid water contents, snow grains and air spaces become surrounded and isolated by the liquid water present in the layer. This water begins to drain downward under the influence of gravity and is called the funicular regime. Once again, when surrounded by water, the snow grains will round, but now boundaries between grains will not be thermodynamically stable and will melt rapidly, creating a slush. The slush has little or no bearing strength, and can even flow like a fluid under certain conditions. The grains themselves, if surrounded by water at 0°C for long enough (24–36 hours), will metamorphose into oblate spheroids (**Figure 9**).

In the absence of liquid water, snow will metamorphose in one of two ways depending on the temperature gradient imposed on the snow. Water vapor density over ice is a strong positive function of temperature, so temperature gradients in the snow give rise to water vapor density gradients in the air spaces in the snow and a diffusive flow of vapor from warmer to colder grain surfaces. For convenience the

Table 3 Metamorphic processes that affect the snow cover

Wet snow metamorphism	*Dry snow metamorphism*	*Dry snow metamorphism – older terms*
Melt-grain clusters and melt–freeze particles	Equilibrium or rounded growth	Equi-temperature metamorphism (ET)
Slush	Kinetic or faceted growth	Temperature-gradient metamorphism (TG)

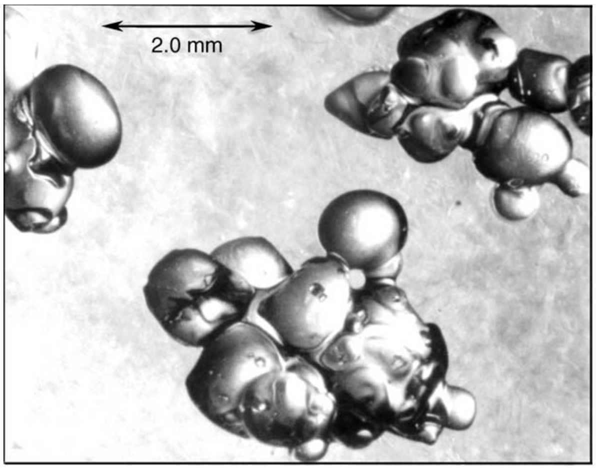

Figure 8 Melt-grain clusters showing the well-rounded grains and the high degree of contact between grains.

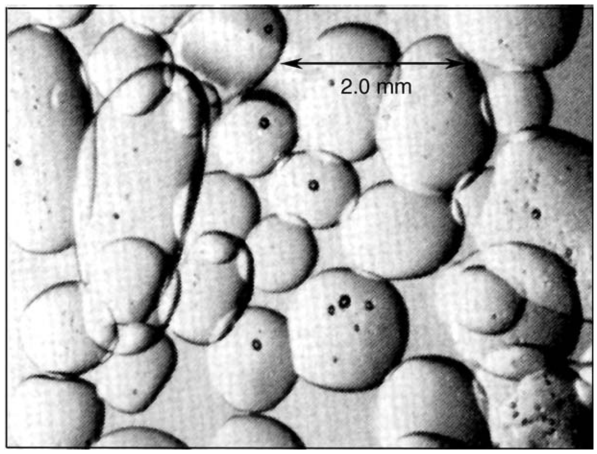

Figure 9 Snow slush, showing the oblate spheroid shape of the grains and the complete lake of bonding. (From Colbeck SC (1986) Statistics of coarsening in water-saturated snow. *Acta Metallargica* 34, 347–352.)

temperature gradient is often defined as the difference between the basal and surface temperatures of the snow cover, divided by the thickness of the snow (**Figure 10**), but in reality the actual temperature gradient varies continuously with both time and height in the snow. For example, rapid fluctuations in air temperature can produce very large temperature gradients near the snow surface, at least for short periods of time. Experimental work has shown that when the temperature gradient exceeds a magnitude of approximately $0.25^{\circ}\mathrm{C\,cm}^{-1}$, kinetic crystal growth will occur. If the gradient is lower, equilibrium growth takes place. Not surprisingly, temperature gradients in thick perennial snow covers tend to be lower than those in the thinner seasonal snow covers, particularly thin taiga, tundra, and alpine seasonal classes that can be subjected to very low air temperatures in the winter. As a result, kinetic growth is common in seasonal snow covers but occurs infrequently (often only in autumn) in perennial snow covers.

Equilibrium crystal growth, also widely known as 'equi-temperature metamorphism' (ET-metamorphism) occurs when temperature gradients in the snow pack are less than $0.25^{\circ}\mathrm{C\,cm}^{-1}$. These low temperature gradients produce weak water vapor density gradients in the snow and low rates of vapor diffusion. The rates are so low that the supply of vapor to a growing crystal, rather than crystal growth dynamics, controls the growth. Rounded, well-bonded grains result.

Kinetic growth, also widely known as 'temperature-gradient metamorphism' (TG-metamorphism) produces ornate, faceted crystals commonly referred to as 'depth hoar'. In this case, temperature gradients imposed on the snow are of a large enough magnitude to produce a flux of water vapor that exceeds the rate at which the crystal can grow. Crystal growth dynamics, rather than vapor supply, control both the growth rate and the crystal form, producing crystals with distinct sharp-edged facets, well-defined interfacial angles, and surface striae (**Figures 11** and **12**). Unlike the case for equilibrium growth, intergrain bonds are weakened and reduced in number during kinetic growth, producing layers that tend to be brittle and weak. This has two important ramifications: the brittle layers can shear easily and often create failure planes that are responsible for the release of avalanches. Second, the poor bonding creates layers that have low thermal conductivity. In absence of air movement in the snow, these layers provide excellent insulation that contributes to the retention of heat in the ground or ice underlying the snow cover.

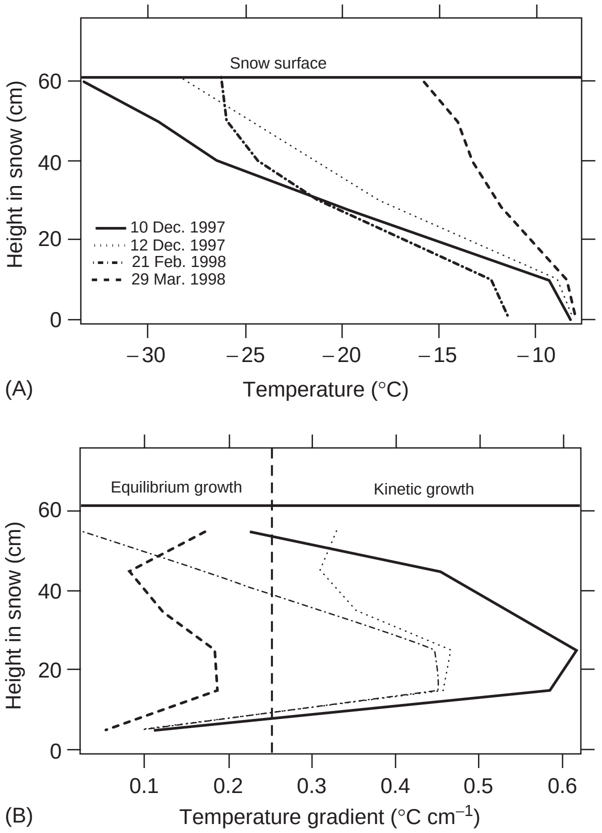

Figure 10 (A) Temperature profiles and (B) computed vertical temperature gradients from the snow cover on the ice of the Beaufort Sea north of Alaska. The temperature profiles are not linear, and as a consequence the temperature gradients vary in a complex way with height in the snow. Note that at some heights and times the gradient is below the critical magnitude of 0.25°C cm^{-1} and kinetic growth will not occur.

Energy and Mass Exchange across a Snow Cover

It is beyond the scope of this article to address in full the mass and energy exchange over a snow cover, but a few points particular to snow are discussed. The reader should also see articles on surface energy balance, albedo, turbulence, boundary layer meteorology, surface roughness, and solar radiation for more details.

Heat transfer across a snow cover occurs mainly by conduction through the ice network of grains, by conduction across the air-filled pore spaces in the snow, and by diffusion of vapor across the pore spaces. The thermal conductivity of ice is more than 100 times higher than that of air, so the conduction of heat across air spaces is thought to contribute relatively little to the total. The heat transported by vapor diffusion, in contrast, is thought to contribute as much as 40%, particularly at temperatures near freezing when the vapor flux is high. This diffusive vapor transport is envisioned as occurring in a 'hand-to-hand' manner across pore spaces, with vapor diffusion from the warm side of snow grains balanced by vapor condensation on the colder side.

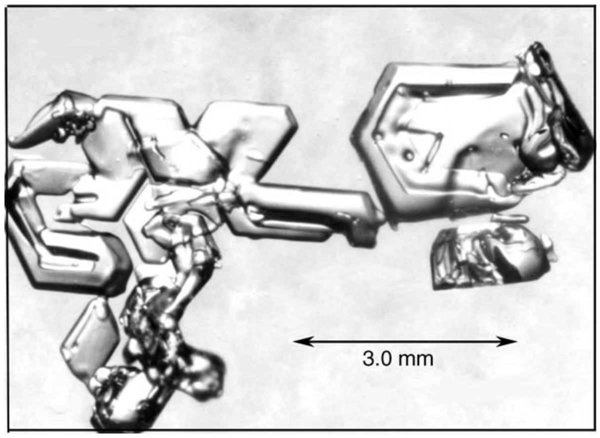

Figure 11 The initial stages of kinetic growth metamorphism. The grains are starting to exhibit distinct faceting.

Because the contributions of these three individual mechanisms are difficult if not impossible to separate,

Figure 12 At-depth hoar cup, shown in typical growth position. The hexagonal pyrimidal cup opens downward because the flow of water vapor is upward. Heavy striae can be seen on all crystal facets. This is the late stage of kinetic growth metamorphism.

in practice they are always lumped together by reporting an 'effective' thermal conductivity for the snow. Both solid body conduction through the ice network and vapor diffusion are driven by the temperature gradients in the snow, suggesting that a simple heat flow equation can be used to model the flux of heat across the snow (eqn [3]).

$$q = k_{\mathrm{eff}} \frac{\mathrm{d}T}{\mathrm{d}z} \qquad [3]$$

Here q is the vertical heat flow through the snow cover, $\mathrm{d}T/\mathrm{d}z$ is the temperature gradient across the snow, and k_{eff} is the effective thermal conductivity of the snow. However, the driving temperature gradient in the ice network may be quite different from the gradient across pore spaces that drives vapor diffusion, in which case eqn [3] may be an oversimplification. Be that as it may, it is customary to describe the heat transfer using eqn [3] and assigning an appropriate value for k_{eff}.

Figure 13 shows compilation of most measured values of k_{eff} as a function of density. As the density of the snow increases, so in general does the value of k_{eff}. In many climate models, regression equations relating k_{eff} to density (often using the viscous snow compaction (eqns [1] and [2]) to determine the snow density) are used to set the thermal conductivity of the snow. However, as the figure shows, the scatter in k_{eff} at any given density is large and real. It is the result of differences in the bonding of the snow, and perhaps also due to variations in snow temperature. For a given density, higher temperatures and better bonding between grains lead to higher values of thermal conductivity. Given the scatter, care should be exercised when choosing a value of k_{eff} for modeling. The values should be consistent with the type of snow cover (**Figures 1** through **3**) as well as a k_{eff}–density relationship. For improved accuracy, a value of k_{eff} for each layer of snow should be determined; then the bulk value for the entire snow cover should be computed using a series-type solution.

Convective heat transfer is also known to operate in snow and complicates the energy exchange across a snow cover. Two types of convection have been reported: buoyancy-driven convection, and convection

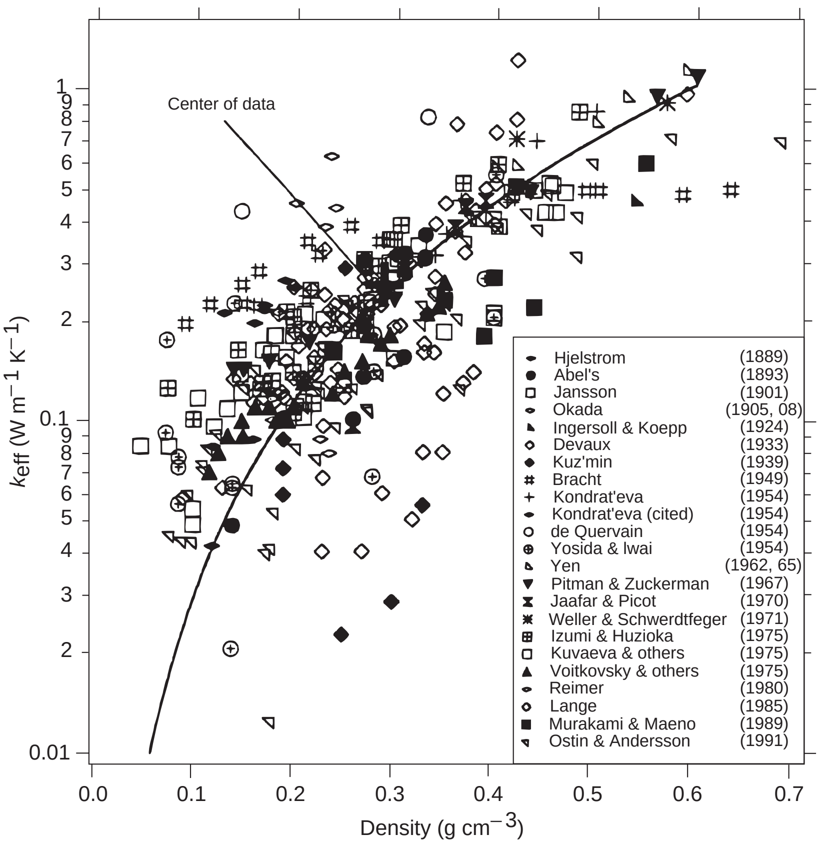

Figure 13 A compilation of most published values of the thermal conductivity of snow. There is nearly an order of magnitude scatter at any given density, and this scatter is real. It arises from differences in snow cover characteristics. (From Sturm M, Holmgren J, König M, Morris K (1997) The thermal conductivity of seasonal snow. *Journal of Glaciology* 43: 26–41.)

forced by the wind (wind-pumping). The former has been documented only in a highly permeable snow covers like taiga snow. This snow cover often wholly comprises layers of large, poorly bonded kinetic growth crystals called depth hoar. The layers have extremely high values of air permeability and, owing to low winter air temperatures, are subjected to temperature gradients of high magnitude, both conditions favorable for buoyancy-driven convection. Convective air flow velocities of several millimeters per second have been computed based on observations of temperature fields in the snow, and these air flow speeds are sufficient to increase the heat transfer rate by a factor of 3. The prevalence of buoyancy convection in other types of snow covers may be low, but this has not been shown experimentally.

Forced convection also probably occurs in some snow covers. Theory indicates that pressure differences arising when wind blows across surface irregularities like dunes and sastrugi are most likely to produce a flow of air that can move both heat and mass (in contrast to turbulence or other aspects of the wind over snow). Flow rates are probably on the order of a few millimeters per second and are likely to be confined to near-surface layers of snow. Observations of the mixing depth of aerosols and particulates in snow layers indicate that wind pumping is definitely effective in moving mass, but the magnitude of the effect of wind-pumping on heat transfer has yet to be demonstrated. In addition, it appears that near-surface and surface wind and melt crusts in the snow can effectively eliminate any wind-pumping by reducing the air permeability of the snow creating barriers in the form of impermeable wind on melt crusts that can effectively shut off all air movement.

As neither wind-pumping nor buoyancy-driven convection are state properties of the snow, they pose difficulties when one is trying to model heat transfer in snow. Both processes depend on external conditions for their onset and strength, and they can transport anything from zero to several times the conductive heat flux, depending on the snow characteristics, the temperature structure in the snow, and the wind speed and direction.

Water, water vapor, CO_2, methane, and aerosols and particulates are all transferred across a snow cover and the transfer process for each is complicated. In general, mass transfer is controlled by the air permeability of the snow, the surface topography of the snow cover (for wind-pumping), and the supply rate of particles, gases, or chemicals. As discussed previously, both diffusive and convective transport of air are possible, and the chemicals and gases move with the air. The air permeability of naturally occuring snow (**Figure 14**) ranges over two orders of magnitude. It is a

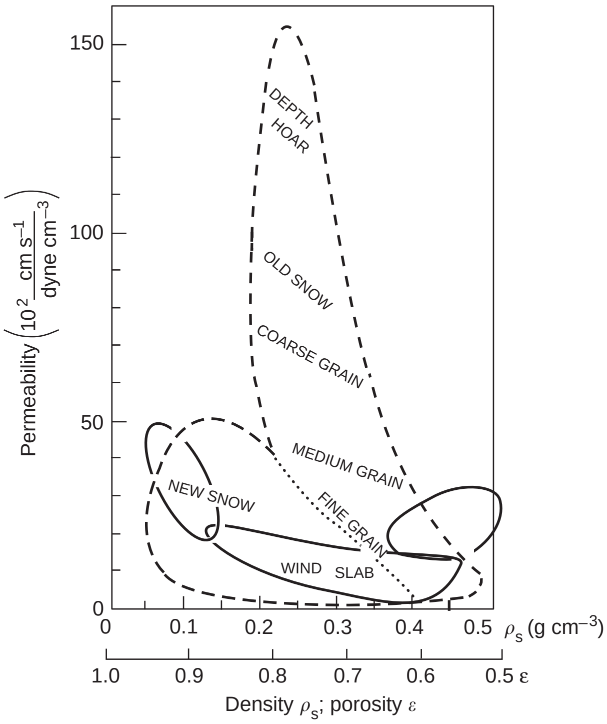

Figure 14 The air permeability of snow. Again, there is a greater variation by snow type than by density. (From Shimizu H (1970) Air permeability of deposited snow. *Low Temperature Science*, Series A, 1–32.)

major control on deposition and transfer rates, which vary widely with chemical species and environmental conditions. For aerosols, when the residence time of the air in the snow is greater than 15 seconds, the filter efficiency of the snow can be almost 100%.

See also

Boundary Layers: Overview. **Energy Balance Model, Surface**. **Land–Atmosphere Interactions:** Canopy Processes; Overview; Trace Gas Exchange. **Reflectance and Albedo, Surface**. **Solar Terrestrial Interactions**.

Further Reading

Colbeck SC (1986) Classification of seasonal snow cover crystals. *Water Resources Research* 22(9): 59S–70S.

Gray DM and Male DH (1981) *Handbook of Snow*. Toronto: Pergamon Press.

LaChapelle ER (1969) *Field Guide to Snow Crystals*. Seattle: University of Washington Press.

Magono C and Lee CW (1966) Meteorological classification of natural snow crystals. *Journal of the Faculty of Science, Hokkaido University* 2(4): 321–335.

Seligman G (1936) *Snow Structure and Ski Fields*. (Reprinted by the International Glaciological Society, Cambridge, 1980.)

Shimizu H (1970) Air permeability of deposited snow. *Low Temperature Science* Series A (22): 1–32.
Sommerfeld RA (1970) The classification of snow metamorphism. *Journal of Glaciology* 9(55): 3–17.
Sturm M, Holmgren J, *et al*. (1997) The thermal conductivity of seasonal snow. *Journal of Glaciology* 43(143): 26–41.
Waddington ED and Harder SL (1996) The effects of snow ventilation on chemical concentrations. In: Wolff EW and Bales RC (eds) *NATO ASI Series*, vol. I–43, pp. 403–451. Berlin: Springer-Verlag.
Warren SG (1982) Optical properties of snow. *Reviews of Geophysics and Space Physics* 20(1): 67–89.

SOLAR TERRESTRIAL INTERACTIONS

J D Haigh, Blackett Laboratory, Imperial College of Science, Technology and Medicine, London, UK

Introduction

Many studies have shown an apparent response in weather or climate indicators to solar variability. At various locations temperature, rainfall, surface pressure, cloud cover, storms, and droughts, among other meteorological parameters, have been found to correlate with measures of solar activity over the 11-year solar cycle and over periods extending from decades to centuries and longer. Some of these studies do not stand up to rigorous statistical analysis and some only appear to hold only over limited time periods, but there is mounting evidence of solar influence on climate on many time scales.

The radiant energy output of the Sun varies by about 0.1% over the 11-year solar cycle. If simple radiation balance estimates are used, this does not appear to be large enough to explain some of the apparent solar signals, in particular those in lower-atmosphere temperatures, although it is consistent with observed decadal variations in sea surface temperature of order 0.1 K. Comparison of solar activity, reconstructed back to the Maunder Minimum in sunspot numbers at the end of the seventeenth century, with estimates of Northern Hemisphere temperature suggests that the Sun has made a significant contribution to climate variability since that time but cannot alone account for the warming of the latter half of the twentieth century.

The amount of solar radiation reaching the Earth is also modified by variations in the Earth's orbit around the Sun. These variations take place over periods of tens to hundreds of thousands of years and may be responsible for the occurrence of ice ages.

Variations in factors other than the total amount of solar radiant energy affect the atmosphere and possibly influence weather and climate. Solar ultraviolet emission varies by several percent over the solar cycle and influences ozone production and temperatures in the middle atmosphere. The resulting change in thermal structure of the stratosphere may influence the climate of the lower atmosphere through dynamical and radiative processes. The chemical structure of the stratosphere is also affected by high-energy protons and electrons emitted during solar flares and coronal mass ejections.

Alterations in the solar magnetic field affect the flux of galactic cosmic rays reaching the Earth and thus the strength of the Earth's electric field and ionization rates in the lower stratosphere. It is plausible, but unproven, that these result in changes in thunderstorm activity or cloud cover.

Observations

The effects on the atmosphere of varying solar insolation can be observed clearly in diurnal and seasonal variations. However, direct observation of the effect of changing solar activity on weather and climate is more difficult, so the detection of solar signals in meteorological records has usually relied on statistical analysis. One approach is to perform a spectral analysis of a time series of data to see if periodicities associated with solar variability (e.g., around 11, 22, or 80 years) are present. Another simple statistical approach is to estimate the degree of correlation between a meteorological parameter and some measure of solar activity. More sophisticated methods take into account other possible factors influencing the state of the atmosphere and also information on the variability of all the parameters involved.

Solar 11-Year Cycle

Cycles of 10–12-year periodicity have been isolated in many data records, including global surface temperature, surface temperature at many land stations across the globe, rainfall in the United States and Africa, surface pressure in the North Atlantic and North Pacific Oceans, North American forest fires, Atlantic tropical cyclones, tropical corals, and the Southern Oscillation. An example of a 10–12 year