4

Snow and Ice

4.1 Introduction

The processes involved in the generation of precipitation, which includes snow, were discussed in Chapter 2. However, snowfall has a very large impact on the activities of humans, particularly on transport and in snowmelt flooding, and therefore it merits more detailed discussion. Indeed in many areas of the world snowfall is a primary factor in hydrological forecasting. For example in the United States it has been estimated (Castruccio et al., 1980) that a forecast improvement of 6% attained by the use of operational satellite measurement of snow covered areas resulted in an annual benefit to agriculture of \$28 million, and to hydroelectricity production of \$10 million.

4.2 Basic processes

4.2.1 *Formation of snow*

As warm moist air ascends, water vapour begins to condense to form cloud. When the cloud temperatures drop below freezing, conditions are suitable for the formation of snow. However, several processes are involved which govern the different types of snow that may be produced. Figure 4.1 summarizes these processes. At about –5 °C the aerosol nuclei (size 0.01 to 1 μm), which are always present in the atmosphere, form small crystals (diameters less than 75 μm) by ice nucleation. These crystals have simple shapes, but may continue to grow by sublimation (Chapter 2) to form snow crystals which often have very complex shapes. A number of snow crystals together form a snowflake. Sometimes snow crystals pass through parts of the cloud which have many cloud droplets (size 10–40 μm), and therefore *riming* (droplets freeze when they come into contact with the crystals) occurs if the crystals are larger than about 300 μm. This occurs at temperatures from –5 to –20 °C. If riming continues for a significant time then snow pellets (*graupel*) may be formed.

Hydrometeorology, First Edition. Christopher G. Collier.

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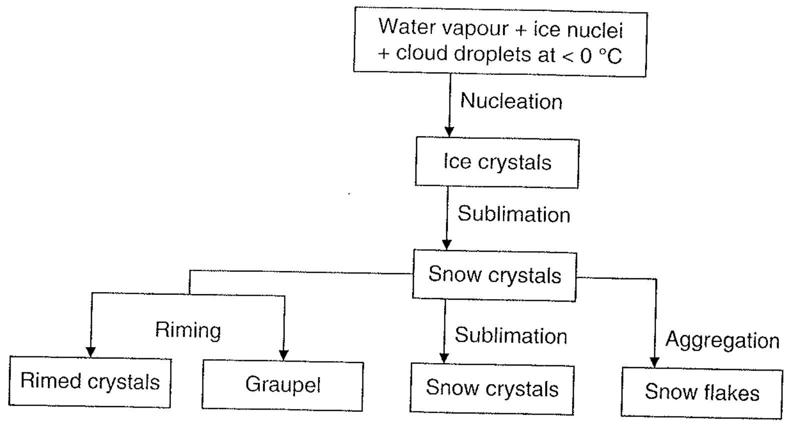

Figure 4.1 Flow diagram of the formation of different types of snow (after Gray and Male, 1981)

At temperatures appropriate for the formation of snow, a cloud may only be slightly supersaturated with respect to water, but 10–20% supersaturated with respect to ice. Hence there is a transfer of water vapour from the cloud droplets to the ice crystals, which consequently grow. This is the Bergeron process, discussed in Chapter 2. The shape of an ice crystal depends upon the temperature at which it grows, but its rate of growth and secondary crystal features depend upon the degree of supersaturation. The range of shapes is: 0 to –4 °C, plates; –4 to –10 °C, prism-like crystals, scrolls, sheaths and needles; –10 to –20 °C, thick plates, dendrites and sector plates; and –20 to –35 °C, sheaths and hollow columns. Figure 4.2 shows examples of the structure of snowflakes. The rate of increase with time t of a mass m of a crystal through diffusion of water vapour onto its surface is

$$\frac{\mathrm{d}m}{\mathrm{d}t} = 4\pi CDFA(\rho_\infty - \rho_0) \tag{4.1}$$

where C is a shape factor; D is the diffusivity of water vapour in air; F is a ventilation factor depending on the relative motion of the crystal with respect to the air; A is a function of crystal size; ρ_∞ is the vapour density (mass of water vapour per unit volume of moist air) at a large distance from the crystal; and ρ_0 is the vapour density at the surface of the crystal.

The mass growth rate due to riming depends upon the fall speed of the ice crystals relative to the cloud droplets, and the efficiency with which the droplets freeze and remain attached to the crystals. Hence

$$\frac{\mathrm{d}m}{\mathrm{d}t} = \pi r^2 abwW \tag{4.2}$$

where r is the radius of the ice crystals; a is the adhesion efficiency; b is the collision efficiency (ab is between 0.1 and 1); w is the fall speed of the ice crystals relative to the droplets (less than 5 m s^{-1}); and W is the liquid water content of the cloud.

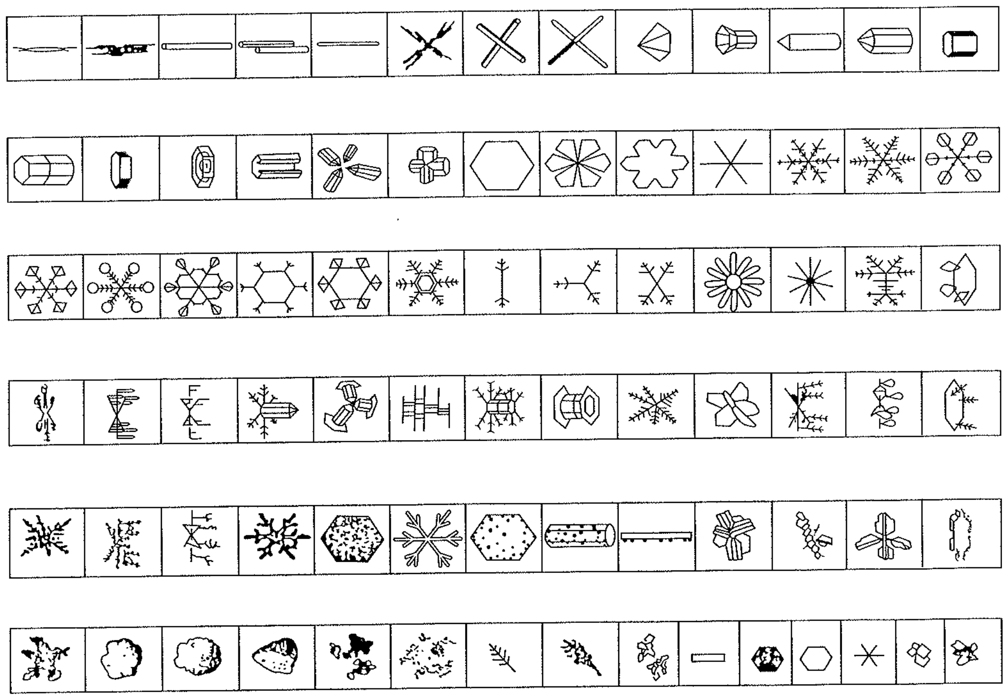

Figure 4.2 Snowflake structures (from Gray and Male, 1981)

4.2.2 *Formation of snow cover and its effects on the atmosphere*

Several conditions have been found to be necessary though not sufficient for the occurrence of significant snowfall:

1. sufficient moisture and aerosol nuclei for the formation and growth of ice crystals
2. sufficient depth of cloud to permit snow crystal growth
3. temperatures below 0 °C in most of the layer through which the snow falls
4. sufficient moisture and aerosol nuclei to replace losses caused by precipitation.

Operational weather forecasts tend to use the temperature of the lower part of the atmosphere from 500 hPa to 1000 hPa, known as the thickness temperature, as an indicator of likely snowfall. However, the other factors are very important, and currently numerical prediction models cannot represent the processes involved accurately enough to provide completely reliable guidance. Fulkes (1935) indicated the rate of precipitation obtained from adiabatically ascending air for a 100 m layer with a vertical velocity of 1 m s^{-1} (Figure 4.3). Such information is useful, but the vertical velocity of the air is determined by the nature of the atmospheric system (see for example Browning and Mason, 1981), orographic effects (Browning, 1983) and topographic effects (Lavoie, 1972). Hence the estimation of the likely snowfall on particular occasions is a very difficult problem, although the recent developments of high resolution numerical models hold out much promise (see Chapter 8).

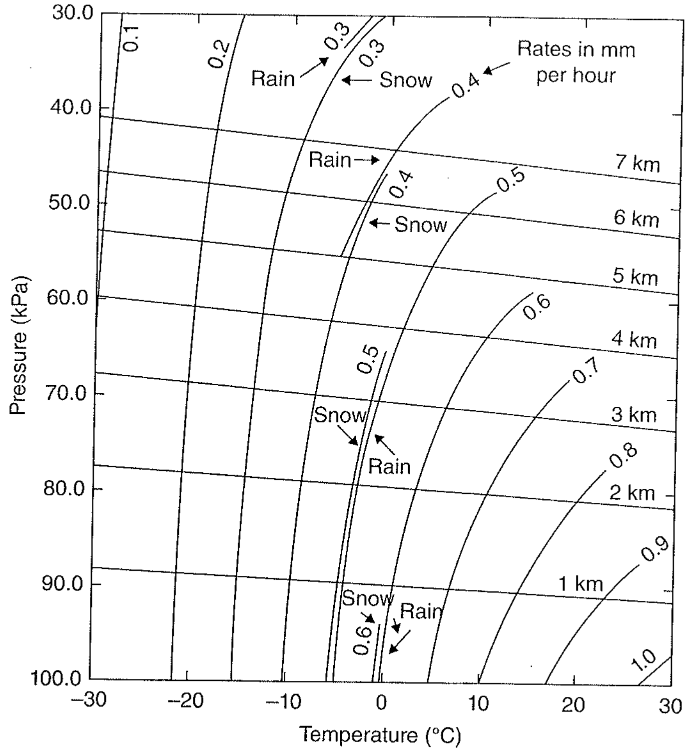

Figure 4.3 Rates of precipitation from adiabatically ascending air for a 100 m layer with a vertical velocity of 1 m s^{-1} (Fulkes, 1935)

Heavy falls of snow are not always associated with high latitudes, although clearly the lower temperatures experienced in these regions are conducive to snowfall for much of the year. Other areas where very cold air occasionally crosses relatively warm stretches of water are also likely to be subjected to heavy snow. Hence, in winter, snow cover can persist for some time in mid-latitudes. This persistence is prolonged if the snow cover is over land at high latitudes. The Rocky Mountains, the Alps and the Himalayas all retain snow cover at the highest elevations all year round. Even the Scottish Highlands, only about 1 km high, lose their snow cover for just one or two months each year.

Both land and sea (the Arctic) areas are covered by snow. The areas involved vary with the seasons, with night and day, and with the day-to-day weather over particular areas. It is not surprising to discover that snow cover is an important factor in determining climate and climatic change. Snow cover influences the atmospheric circulation by interacting with, and affecting, overlying air masses, which results in either the amplification or the stabilization of circulation anomalies that often cause the weather (Rango, 1985).

Snow strongly reflects visible and near infrared light, that is it has a high *albedo*. Consider Table 4.1, which compares the albedo of snow with that of other natural surfaces. The albedo of snow varies with the age of the snow, but is considerably higher than the albedos of most natural surfaces. Therefore because of the seasonal changes in the extent of snow cover, the albedo of the surface of the Earth varies from

Table 4.1

Surface
Fresh snow
Old snow
Sea ice
Open water
Clouds
Coniferous fc
Deciduous fo
Desert
Desert sand

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Table 4.1 Percentage of incident short wave radiation reflected by some surfaces

Surface	Albedo (%)	Surface	Albedo (%)
Fresh snow	80–95	Bare fields	15–25
Old snow	50–60	Field crops	3–25
Sea ice	50–70	Barley	23
Open water	5–30	Sugar beet	26
Clouds	1–80	Grass	20–25
Coniferous forest	15	Tundra	15–20
Deciduous forest	18	New concrete	55
Desert	24–30		
Desert sand	40	Average over all Earth	30–35

season to season. Since snow also has a high thermal emissivity, a low thermal conductivity, and a low water vapour pressure, the energy balance within the atmosphere will change as the seasons change. As melting occurs, the water vapour pressure and the thermal conductivity of the snow increase, and the latent heat of the snow must be taken into account in the energy balance.

Since the snow cover at higher latitudes persists for much of the year, less solar energy is absorbed at these latitudes, causing a profound effect upon the meridional flux of energy. Higher latitudes act as source regions for cold air masses which move to mid-latitudes, resulting in the atmospheric systems discussed in Chapter 2 (Walsh, 1984). Deviations from the regular seasonal changes in snow cover may be one factor in triggering climatic change (Chapter 13).

4.2.3 *Formation of ice*

As the surface of an area of water cools, the cooler, denser water near the surface sinks, and is replaced by less dense water from below. For temperatures less than about 4 °C, the less dense, cooler water at the surface begins to freeze even though the water below it is relatively warmer.

The initial orientation of the ice crystals is random, but as the ice thickens some crystals grow more quickly downwards than others. In sea ice more horizontally oriented crystals are favoured, whereas for ice on lakes this is not the case (see for example Hobbs, 1974). The water below the ice continues to lose heat by conduction through the ice sheet, albeit much more slowly, and as a consequence the ice thickens downwards. If snowfall is low then the layer of ice may be over 3 m thick and is known as *black ice*.

As snow accumulates on the ice its density increases with depth, and the existing black ice is depressed below the original water level. The temperature within the lower layers of the snow cover may rise slightly, so that sublimation (Chapter 2) occurs, and a layer of *white ice* is formed on the black ice, separated by a layer in which some melting occurs due to the flooding caused by the snow cover depression. This area of *slush* may also be caused by warmer water moving upwards by capillary action through cracks in the black ice, but it quickly refreezes.

Like a snow cover, the ice on lakes delays runoff from a catchment area, and therefore may have a significant impact upon hydrological forecasting. Melting will