CHAPTER 3

PRECIPITATION MEASUREMENT

3.1 GENERAL REQUIREMENTS: ACCURACY AND PRECISION

The total amount of precipitation reaching the ground during a stated period is expressed as the depth to which it would cover a horizontal projection of the Earth's surface, if any part of the precipitation falling as snow or ice were melted. Snowfall is also measured by the depth of fresh snow covering an even, horizontal surface.

The primary aim of any method of precipitation measurement is to obtain representative samples of the fall over the area to which the measurement refers. There is a critical need in hydrology for accurate measurement of precipitation. Therefore, for raingauges the choice of site, the form and exposure of the measuring gauge, the prevention of loss by evaporation, and the effects of wind and splashing are important considerations. More complex methods such as the use of weather radar and satellites require detailed understanding of error characteristics. This chapter discusses the facets of precipitation measurement that are most relevant to hydrological practice. A more general discussion of the topic can be found in the *Guide to Meteorological Instruments and Methods of Observation* (WMO-No. 8).

3.2 RAINGAUGE LOCATION

In a perfect exposure, the catch of the raingauge would represent the precipitation falling on the surrounding area. However, this is difficult to attain in practice because of the effect of the wind. Much care has to be taken in the choice of the site.

Wind effects are of two types: the effects on the gauge itself, which generally reduce the amount of water collected, and the effects of the site on the wind trajectories, which are frequently more important and can give rise to either an excess or a deficiency in measured precipitation.

The disturbance created by an obstacle depends on the ratio of the obstacle's linear dimensions to the falling speed of precipitation. This effect is reduced, if not entirely overcome, by choosing the site so that the wind speed at the level of the gauge orifice is as small as possible, but so that there is not any actual blocking of precipitation by surrounding objects, and/or by modifying the surroundings of the gauge so that the airflow across the orifice is horizontal. All gauges in any area or country should have comparable exposures, and the same siting criteria should be applied to all.

The gauge should be exposed with its orifice remaining horizontal over ground level.

Where possible, the gauge site should be protected from wind movement in all directions by objects, such as trees and shrubs, of as nearly uniform height as possible. The height of these objects above the orifice of the gauge should be at least half the distance from the gauge to the objects, but should not exceed the distance from the gauge to the objects (to avoid interception of precipitation that should reach the gauge). The ideal situation is to have the angle from the top of the gauge to the top of the encircling objects between 30° and 45° to the horizontal (Figure I.3.1).

Objects such as windbreaks, consisting of a single row of trees, should be avoided as protection for gauges, as they tend to increase turbulence at the gauge site. Isolated or uneven protection near the gauge should also be avoided because of variable and unpredictable effects on the gauge catch. When adequate protection from the wind is not possible, individual objects should not be closer to the gauge than a distance equal to four times their height. Subject to these limitations, a site that is sheltered from the full force of the wind should be chosen to avoid wind-induced measurement errors. Caution

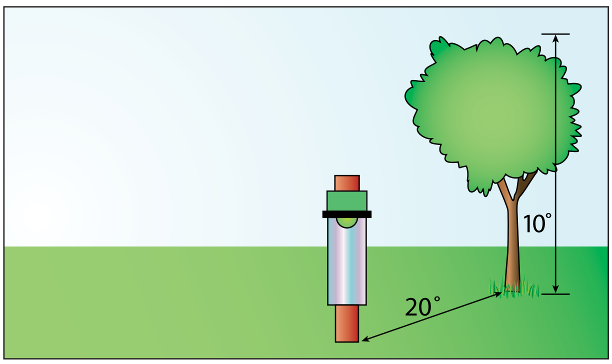


Figure I.3.1. Positioning of raingauge

should always be exercised, so that the site chosen does not produce significant disturbances in the wind. Sites on a slope, or with the ground sloping sharply away in one direction, especially if this direction is the same as the prevailing wind, should be avoided.

The ground surrounding the gauge can be covered with short grass or be of gravel or shingle, but a hard flat surface, such as concrete, gives rise to excessive splashing.

The height of the gauge orifice above the ground should be as low as possible because the wind velocity increases with height, but it should be high enough to prevent splashing from the ground. A height of 30 cm is used in many countries, in those areas that have little snow and where the surroundings are such that there is no risk of the ground being covered by puddles, even in heavy rain. Where these conditions are not satisfied, a standard height of 1 m is recommended.

In very exposed places, where there is no natural shelter, it has been found that better results can be obtained for liquid precipitation if the gauge is installed in a pit, so that the gauge rim is at ground level (Figure I.3.2). A strong plastic or metal anti-splash grid should span the pit with a central opening for the gauge funnel. The anti-splash grid should be composed of thin slats about 5 to 15 cm deep, set vertically at about 5 to 15 cm spacing in a square symmetric pattern. The area surrounding the gauge should be level and without unusual obstructions for at least 100 m in all directions.

An alternative installation, which is not quite so effective, is to install the gauge in the middle of a circular turf wall. The inner wall surface should be vertical with a radius of about 1.5 m. The outer surface should slope at an angle of about 15° to the horizontal. The top of the wall should be level with the gauge orifice.

Provision should be made for drainage. The pit gauge has been developed to measure liquid precipitation and should not be used for snowfall measurements.

An alternative way of modifying the surrounding of the gauge is to fit suitably shaped windshields around the instrument. When properly designed, these enable much more representative results to be obtained than with unshielded gauges fully exposed to the wind. An ideal shield should:
(a) Ensure a parallel flow of air over the aperture of the gauge;
(b) Avoid any local acceleration of the wind above the aperture;
(c) Reduce to the degree possible the speed of the wind striking the sides of the receiver. The height of the gauge orifice above the ground is then much less important;
(d) Prevent splashing towards the aperture of the receiver. The height of the gauge orifice above the ground is then much less important;
(e) Not be subject to capping by snow.

Precipitation in the form of snow is much more subject to adverse wind effects than is rainfall. In exceptionally windy locations, the catch in a gauge, with or without a windshield, may be less than half the actual snowfall. Sites selected for measurement of snowfall and/or snow cover should, as far as possible, be in areas sheltered from the wind. Windshields attached to the gauges have been shown to be quite effective in reducing precipitation catch errors due to wind, especially for solid precipitation. No shield yet developed, however, will entirely eliminate wind-induced measurement errors.

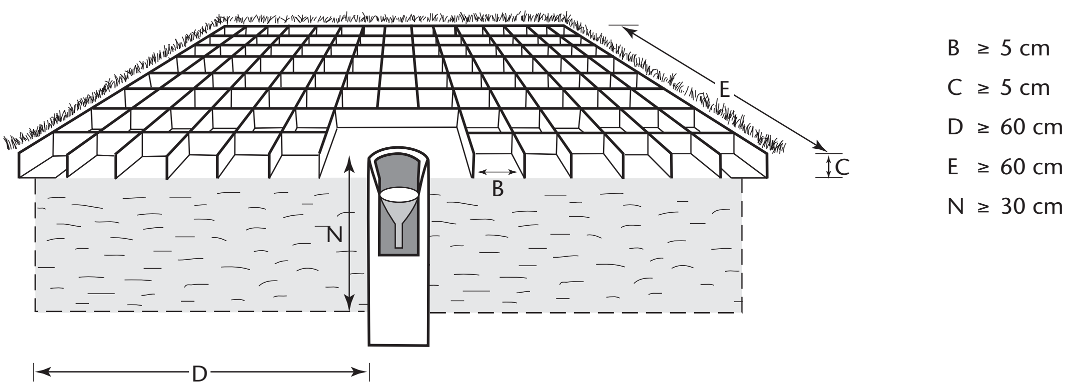


Figure I.3.2. Pit gauge for the measurement of liquid precipitation

3.3 NON-RECORDING RAINGAUGES [HOMS C27]

3.3.1 General

The non-recording raingauges used by most Hydrological and Meteorological Services for official measurements generally consist of open receptacles with vertical sides, usually in the form of right cylinders. Various sizes of orifice and height are used in different countries and, therefore, measurements are not strictly comparable. The depth of precipitation caught in a gauge is measured by means of a graduated flask or dipstick. In gauges having other than vertical sides, the measurement is made either by weighing or measuring the volume of the contents, or by measuring the depth with a specially graduated measuring stick or scale.

3.3.2 Standard raingauges

The ordinary raingauge used for daily readings usually takes the form of a collector above a funnel leading into a receiver (Figure I.3.3). The size of the opening of the collector is not important. A receiving area of 1 000 cm^2 is used in some countries, but an area of 200 to 500 cm^2 will probably be found most convenient. The area of the receiver may be made to equal 0.1 of the area of the collector. Whatever size is chosen, the graduation of the measuring apparatus must be consistent with it. The most important requirements of a gauge are as follows:

Figure I.3.3. Ordinary raingauge

(a) The rim of the collector should have a sharp edge and should fall away vertically inside and be steeply bevelled outside. The design of gauges used for measuring snow should be such that errors due to constriction of the aperture, by accumulation of wet snow about the rim, are small;
(b) The area of the aperture should be known to the nearest 0.5 per cent and the construction should be such that this area remains constant;
(c) The collector should be designed to prevent rain from splashing in or out. This can be done by having the vertical wall sufficiently deep and the slope of the funnel sufficiently steep (at least 45°);
(d) The receiver should have a narrow neck and should be sufficiently protected from radiation to minimize loss of water by evaporation;
(e) When a part of precipitation comes in the form of snow, the collector should be deep enough to store the snowfall that can be expected during at least one day. This is also important to avoid the snow drifting out of the collector.

Raingauges for use at places where only weekly or monthly readings are taken should be similar in design to the type used for daily measurement, but with a receiver of larger capacity and stronger construction.

3.3.3 Storage raingauges

Storage raingauges are used to measure total seasonal precipitation in remote, sparsely inhabited areas. These gauges consist of a collector above a funnel, leading into a receiver large enough to store the seasonal catch. The criteria for exposure and shielding given in previous sections should also be considered in the installation of these gauges.

In areas where heavy snowfall occurs, the collector must be placed above the maximum expected depth of snow cover. This may be accomplished by mounting the entire gauge on a tower or by mounting the collector on a 30-cm diameter steel pipe of sufficient length to place its catch ring above the maximum accumulated snow.

An antifreeze solution is placed in the receiver to convert the snow that falls into the gauge into a liquid state. A mixture of 37.5 per cent of commercial-grade calcium chloride (78 per cent purity) and 62.5 per cent water by weight makes a satisfactory antifreeze solution. Alternately, an ethylene glycol solution can be used. While more expensive, the ethylene glycol solution is less corrosive than calcium chloride and gives protection over a much wider range of dilution caused by ensuing precipitation. The volume of the solution placed in the receiver should not exceed one third the total volume of the gauge.

An oil film should be used in the gauge to prevent loss of water by evaporation. An oil film about 8 mm thick is sufficient. Low viscosity, non-detergent motor oils are recommended. Transformer and silicone oils have been found unsuitable.

The seasonal precipitation catch is determined by weighing or measuring the volume of the contents of the receiver. The amount of antifreeze solution placed in the receiver at the beginning of the season must be taken into account with either method.

3.3.4 Methods of measurement

Two methods are commonly used for measuring the precipitation caught in the gauge: a graduated measuring cylinder and a graduated dip-rod.

A measuring cylinder should be made of clear glass with a low coefficient of expansion and should be clearly marked with the size of gauge for which it is to be used. Its diameter should not be more than about one third of that of the rim of the gauge.

The graduations should be finely engraved. In general, markings should be at 0.2 mm intervals with whole millimetre lines clearly indicated. It is also desirable that the line corresponding to 0.1 mm should be marked. Where it is not necessary to measure rainfall to this accuracy, every 0.2 mm up to 1.0 mm and every millimetre above that should be marked, with every 10-mm line clearly indicated. For accurate measurements, the maximum error of the graduations should not exceed ±0.05 mm at or above the 2-mm graduation mark and ±0.02 mm below this mark.

To achieve this accuracy with small amounts of rainfall, the inside of the measuring cylinder should be tapered at its base. In all measurements, the bottom of the water meniscus should be taken as the defining line. It is important to keep the measure vertical and to avoid parallax errors. It is helpful, in this respect, if the main graduation lines are repeated on the back of the measure.

Dip-rods should be made of cedar or other suitable material that does not absorb water to any appreciable extent, and should have a capillarity effect that is small.

Wooden dip-rods are unsuitable if oil has been added to the collector, and rods of metal or other material from which oil can be readily cleaned must then be used.

They should be fitted with a brass foot to avoid wear and be graduated according to the relative areas of cross-section of the gauge orifice and the receiving can, making allowance for the displacement due to the rod itself. Marks at every 10 mm should be shown. The maximum error in the dip-rod graduation should not exceed ±0.5 mm at any point. Although the measurement may be made with a dip-rod, it should be checked by using a rain measure as well, whenever possible.

It is also possible to measure the catch by weighing. There are several advantages to this procedure. The total weight of the can and contents should be weighed, and the weight of the can should then be subtracted. There is no danger of spilling the contents, and none is left adhering to the can. However, the common methods are simpler and cheaper.

3.3.5 Errors and accuracy of readings

The errors involved in measuring the catch collected in a gauge are small compared with the uncertainty due to the effect of the exposure of the instrument if reasonable care is taken in the readings. Daily gauges should be read to the nearest 0.2 mm and preferably to the nearest 0.1 mm. Weekly or monthly gauges should be read to the nearest 1 mm. The main sources of error likely to arise are the use of inaccurate measures or dip-rods, spilling of some water when transferring it to the measure, and the inability to transfer all the water from the receiver to the measure.

In addition to these errors, losses by evaporation can occur. These are likely to be serious only in hot dry climates, and with gauges visited only at infrequent intervals.

Evaporation losses can be reduced by placing oil in the receiver or by designing the gauge so that only a small water surface is exposed, ventilation is poor, and the internal temperature of the gauge does not become excessive. Also, the receiving surface of the gauge must be smooth, so that the raindrops do not adhere to it. It should never be painted.

In winter where rains are often followed immediately by freezing weather, damage to the receiver, and consequent loss by leakage, can be prevented by the addition of an antifreeze solution. This again mainly applies to gauges visited infrequently.

Allowance for the solution added must, of course, be made when measuring the gauge catch. All gauges should be tested regularly for possible leaks.

3.3.6 Correction of systematic errors

Owing to the effects of wind, wetting, evaporation, blowing snow and splashing, the amount of precipitation measured is usually lower (by 3 to 30 per cent or more) than that which actually fell. This systematic error may be corrected if the readings are to be used for hydrological calculations (WMO, 1982). Before carrying out any corrections, the original measured data should be securely archived. Published data should be clearly labelled "measured" or "corrected", as applicable.

The corrections for these effects are generally based on relationships between the components of the error and the meteorological factors. Thus, the loss from wind field deformation near the gauge rim is related to wind speed and precipitation structure.

The latter can be characterized, depending on the time period used, by the proportion of rainfall at low intensity ($i_p \leq 0.03$ mm min^{-1}), by a logarithm of rainfall intensity, by air temperature and/or humidity, and the type of precipitation. Loss from wetting is related to the number of precipitation events and/or days, while loss from evaporation is a function of the saturation deficit and wind speed. Excess measured precipitation as a result of blowing or drifting snow is related to wind speed.

The above-mentioned meteorological factors may be derived from standard meteorological observations performed at the gauge site or in its vicinity, if daily corrections are to be applied. At sites without such meteorological observations, only estimates for time periods longer than one day, for example, one month, should be used.

The value of the correction varies from 10 to 40 per cent for individual months, depending on the type

Table I.3.1. Main components of the systematic error in precipitation measurement and their meteorological and instrumental factors listed in order of general importance

$$P_k = kP_c = k(P_g + \Delta P_1 + \Delta P_2 + \Delta P_3 \pm \Delta P_4 - \Delta P_5)$$

where P_k is the adjusted precipitation amount, k is the correction factor, P_c is the precipitation caught by the gauge collector, P_g is the measured precipitation in the gauge, and P_1 to P_5 are corrections for components of systematic error as defined below:

Symbol	*Component of error*	*Magnitude*	*Meteorological factors*	*Instrumental factors*
k	Loss due to wind field deformation above the gauge orifice	2–10% 10–50%[a]	Wind speed at the gauge rim during precipitation and the structure of precipitation	The shape, orifice area and depth of both the gauge rim and collector
$\Delta P_1 + \Delta P_2$	Losses from wetting on internal walls of the collector and in the container when it is emptied	2–10%	Frequency, type and amount of precipitation, the drying time of the gauge and the frequency of emptying the container	The same as above and, in addition, the material, colour and age of both the gauge collector and container
ΔP_3	Loss due to evaporation from the container	0–4%	Type of precipitation, saturation deficit and wind speed at the level of the gauge rim during the interval between the end of precipitation and its measurement	The orifice area and the isolation of the container, the colour and, in some cases, the age of the collector, or the type of funnel (rigid or removable)
ΔP_4	Splash-out and splash-in	1–2%	Rainfall intensity and wind speed	The shape and depth of the gauge collector and the kind of gauge installation
ΔP_5	Blowing and drifting snow		Intensity and duration of snow storm, wind speed and the state of snow cover	The shape, orifice area and depth of both the gauge rim and the collector

[a] Snow

of estimate of the meteorological factors employed.

The main components of the systematic error in precipitation measurement are given in Table I.3.1.

The correction factor k for the effect of wind field deformation above the gauge orifice, estimated experimentally for various gauges, is given in Figure I.3.4. It is a function of two variables, the wind speed during precipitation at the level of the gauge rim and the velocity of the falling precipitation particles. The latter depends on the structure of the precipitation.

The absolute value of wetting loss depends on the geometry and material of the gauge collector and container, on the number of measurements of precipitation, and on the amount, frequency and form of precipitation. It is different for liquid, mixed and solid precipitation, and can be estimated by

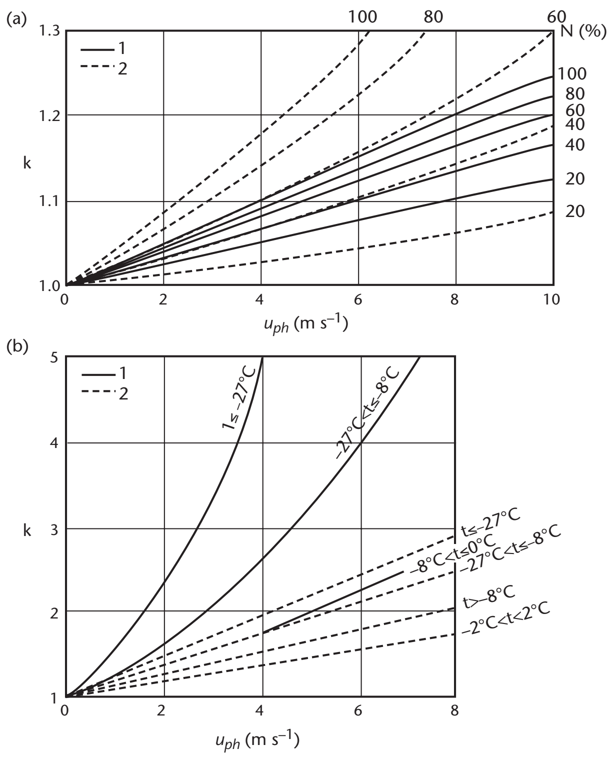


Figure I.3.4. Correction factor k as a function of the wind speed during precipitation at the level of the gauge rim (u_{ph}) and the parameter of the precipitation structure N and t for: (a) liquid precipitation; and (b) mixed and solid precipitation. 1 = Hellman gauge without windshield; 2 = Tretyakov gauge with windshield; t = air temperature during snowstorms; N = fraction in percentage of monthly totals of rain falling with an intensity smaller than 0.03 mm min^{-1} (UNESCO, 1978)

weighing or volumetric measurements in a laboratory. The wetting loss for solid precipitation is generally smaller than for liquid precipitation because the collector is usually wetted only once during snow melt.

The total monthly wetting loss, ΔP_1, can be estimated by using the equation:

$$\Delta P_1 = \bar{a}\, M \qquad (3.1)$$

where $\bar{a}$ is the average wetting loss per day for a particular collector and M is the number of days with precipitation.

In cases where the amount of precipitation is measured more than once a day, the total monthly wetting loss is:

$$\Delta P_{1,2} = a_x\, M_p \qquad (3.2)$$

where a_x is the average wetting loss per measurement of precipitation for a particular gauge and form of precipitation and M_p is the number of measurements of precipitation during the period concerned.

Evaporation loss can be estimated as follows:

$$\Delta P_3 = i_e\, \tau_e \qquad (3.3)$$

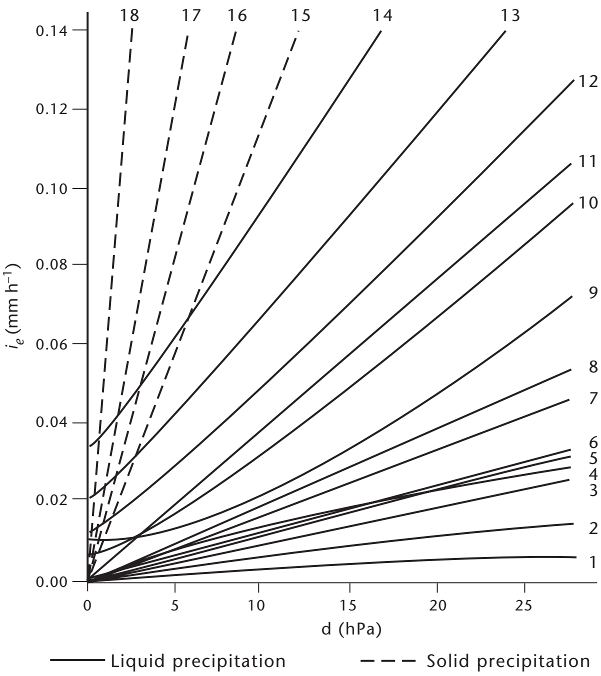
where i_e is the intensity of evaporation and τ_e is the time that elapsed between the end of precipitation and its measurement. The value of i_e depends on the construction, material and colour of the gauge, on the form and amount of precipitation, the saturation deficit of the air d (hPa), and on wind speed at the level of the gauge rim during evaporation. It is difficult to estimate i_e theoretically because of the complex configuration of a precipitation gauge. However, i_e can be computed by using empirical equations or graphical functions as shown in Figure I.3.5. The value of τ_e can be estimated by using precipitation recording gauges, but it also depends on the number of precipitation observations per day. It is three to six hours for liquid precipitation if measured twice per day and six hours for snow because the evaporation takes place during the snowfall.

The net error due to splash-in and splash-out of water can be either negative or positive, and therefore assumed as zero for most types of properly designed precipitation gauges (3.3.2). The error resulting from snow blowing into the gauge should be considered during snowstorms with wind speed larger than 5 m s^{-1}. The half-day values can be estimated at the gauge sites with visual observations of the duration of blowing snow, and in those with known data for wind speed and number of days with both blowing and drifting snow. The long-term monthly averages can be estimated from the graph in Figure I.3.6 if the duration of snowstorms and wind speed are known.

Besides these systematic errors there are random observational and instrumental errors. Their effect can often be neglected because of the high values of the systematic errors.

3.4 RECORDING RAINGAUGES [HOMS C53]

Five types of precipitation recorders are in general use: the weighing type, the float type, the tipping-bucket type, distrometers and the acoustic type. The only satisfactory instrument for measuring all kinds of precipitation utilizes the weight or the momentum/optical detection principle. The use of the other types is primarily limited to the measurement of rainfall.



Note: Intensity of evaporation (i_e) for various gauges: (a) Liquid precipitation: (i) Australian standard gauge 1, 2, 7, 11 for $P \leq 1$ mm; 1.1 to 20 mm; > 20 mm (all for wind speeds, $u_e > 4$ m s^{-1}), and for $u_e \geq 4$ m s^{-1}, respectively; (ii) Snowdon gauge in a pit 3, 6, 8 for $P \leq 1$ mm, 1.1 to 10 mm and ≥ 10 mm, respectively; (iii) Hellmann gauge 4; (iv) Polish standard gauge 5; (v) Hungarian standard gauge 9; (vi) Tretyakov gauge 10, 12, 13, 14 for wind speeds at the level of the gauge rim of 0 to 2, 2 to 4, 4 to 6 and 6 to 8 m s^{-1}, respectively; (b) Solid precipitation: Tretyakov gauge 15, 16, 17, 18 for wind speeds 0 to 2, 2 to 4, 4 to 6 and 6 to 8 m s^{-1}, respectively, where i_e is the intensity of evaporation in mm h^{-1} and τ_e is the time elapsed between the end of the precipitation and the measurement of precipitation.

Figure I.3.5. Evaporation losses from precipitation gauges

3.4.1 Weighing–recording gauge

In these instruments, the weight of a receiving can plus the precipitation accumulating in it is recorded continuously, either by means of a spring mechanism or with a system of balance weights. Thus, all precipitation is recorded as it falls. This type of gauge normally has no provision for emptying itself, but by a system of levers, it is possible to make the pen traverse the chart any number of times. These gauges have to be designed to prevent excessive evaporation losses, which may be reduced further by the addition of sufficient oil or other evaporation suppressing material to form a film over the water surface. Difficulties caused by oscillation of the balance in strong winds can be reduced by fitting the instrument with an oil damping mechanism. The main utility of this type of instrument is in recording snow, hail and mixtures of snow and rain. It does not require that the solid precipitation be melted before it can be recorded.

3.4.2 Float gauge

In this type of instrument, the rainfall is fed into a float chamber containing a light float. As the level of the water rises, the vertical movement of the float is transmitted by a suitable mechanism into the movement of the pen on the chart. By suitably

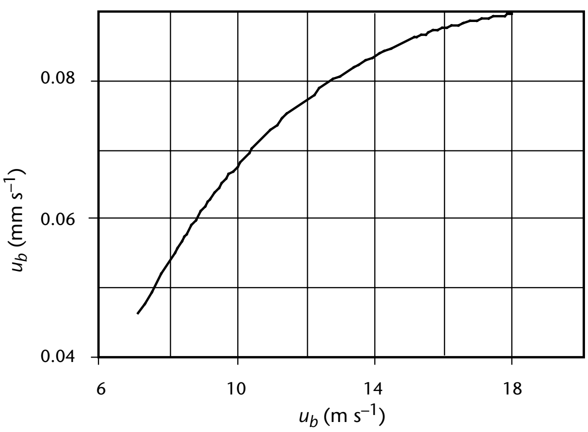


Figure I.3.6. Long-term seasonal intensity of blowing snow (i_b) as a function of long-term wind speed (u_b) at the level of the anemometer (10 to 20 m) during blowing snow

adjusting the dimensions of the receiving funnel and the float chamber, any desired scale on the chart can be obtained.

To provide a record over a useful period (at least 24 hours is normally required), the float chamber has either to be very large (in which case a compressed scale on the chart is obtained), or some automatic means has to be provided for emptying the float chamber quickly whenever it becomes full. The pen then returns to the bottom of the chart. This is usually done with some sort of siphoning arrangement. The siphoning process should start fully at a definite time with no tendency for the water to dribble over, either at the beginning or at the end of the siphoning, which should not take longer than 15 seconds. In some instruments, the float chamber assembly is mounted on knife edges so that the full chamber overbalances. The surge of the water assists in the siphoning process, and when the chamber is empty it returns to its original position. Other rainfall recorders have a forced siphon that operates in less than five seconds. One type has a small chamber separate from the main chamber which accommodates the rain that falls during siphoning. This chamber empties into the main one when siphoning ceases, thus ensuring a correct record of total rainfall.

A heating device should be installed inside the gauge if there is a possibility of freezing. This will prevent damage to the float and float chamber owing to water freezing and will enable rain to be recorded during that period. A small heating element or electric lamp is suitable where a supply of electricity is available, but, if not, other sources of power have to be employed. One convenient method is to use a short length of heating strip wound around the collecting chamber and connected to a large capacity battery. The amount of heat supplied should be kept to the minimum necessary in order to prevent freezing, because the heat will affect the accuracy of the observations by stimulating vertical air movements above the gauge and by increasing evaporation losses.

3.4.3 **Tipping-bucket gauge**

The principle of this type of recording gauge is very simple. A light metal container is divided into two compartments and is balanced in unstable equilibrium about a horizontal axis. In its normal position the container rests against one of two stops, which prevents it from tipping completely. The rain is led from a conventional collecting funnel into the uppermost compartment. After a predetermined amount of rain has fallen, the bucket becomes unstable in its normal position and tips over to its other position of rest. The compartments of the container are so shaped that the water can then flow out of the lower one leaving it empty. Meanwhile, the rain falls into the newly positioned upper compartment. The movement of the bucket, as it tips over, is used to operate a relay contact and produce a record that consists of discontinuous steps. The distance between each step represents the time taken for a pre-specified amount of rain to fall. This amount of rain should not be greater than 0.2 mm if detailed records are required. For many hydrological purposes, in particular for heavy rainfall areas and flood-warning systems, 0.5 to 1.0 mm buckets are satisfactory.

The main advantage of this type of instrument is that it has an electronic pulse output and can be recorded at a distance, or for simultaneous recording of rainfall and river stage on a water stage recorder. Its disadvantages are:

(a) The bucket takes a small but finite time to tip, and during the first half of its motion, the rain is being fed into the compartment already containing the calculated amount of rainfall. This error is appreciable only in heavy rainfall (Parsons, 1941);
(b) With the usual design of the bucket, the exposed water surface is relatively large. Thus, significant evaporation losses can occur in hot regions. This will be most appreciable in light rains;
(c) Because of the discontinuous nature of the record, the instrument is not satisfactory for use in light drizzle or very light rain. The time of beginning and ending of rainfall cannot be determined accurately.

3.4.4 **Rainfall-intensity recorder**

A number of rainfall-intensity recorders have been designed and used for special purposes. However, they are not recommended for general purposes because of their complexity. A satisfactory record of rainfall intensity can usually be determined from a float- or weighing-type recorder by providing the proper timescale.

3.4.5 **Distrometers**

Distrometers measure the spectrum of precipitation particles either through the momentum transferred to a transducer as the hydrometeors hit a detector, or through the image/reflectivity of the hydrometeors illuminated by light or microwaves (Bringi and Chandrasekar, 2001). They have the advantage of providing comprehensive information on hydrometeor size distributions (Figure I.3.7). These devices are available commercially albeit at a high cost compared with tipping-bucket raingauges.

3.4.6 Acoustic type

The measurement of rainfall over lakes and the sea is particularly problematic. However, the noise raindrops make as they hit a water surface may be detected using a sensitive microphone. The noise spectrum reveals the raindrop size distribution, and hence rainfall amount. Such systems are now available commercially. Over land, acoustic profilers, actually designed to measure wind profiles, may also make measurements of rainfall.

3.4.7 Methods of recording the data

Whether the rainfall recorder operates by the rise of a float, the tipping of a bucket or some other method, these movements must be converted into a form that can be stored and analysed later. The simplest method of producing a record is to move a time chart by a spring or an electrically-driven clock, past a pen that moves as the float or weighing device moves. There are two main types of charts: the drum chart, which is secured around a drum that revolves once a day, once a week, or for such other period as desired, and the strip chart, which is driven on rollers past the pen arm. By altering the speed of the strip chart, the recorder can operate for periods of one week to a month or even longer. The timescale on the strip chart can be large enough for intensity to be calculated with ease.

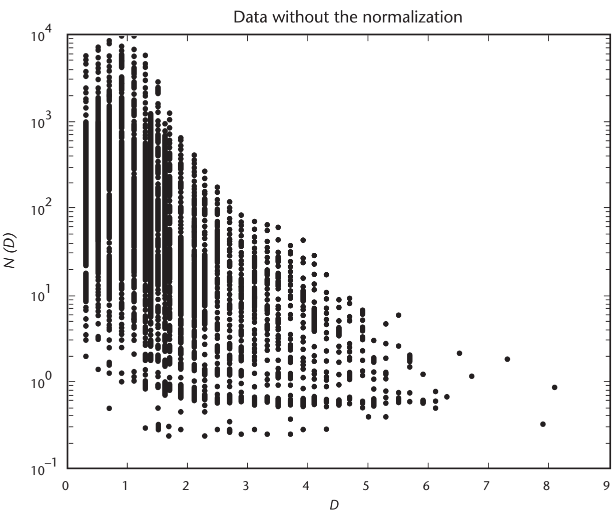


Figure I.3.7. *N (D)* [$mm^{-1}m^{-3}$] versus *D* [mm] for 70 two-minute average drop size distributions measured by a 2D-video distrometer in convective rain cells in Florida

The value to be recorded may also be mechanically or electronically converted to digital form and recorded on magnetic media at uniform time intervals for later automatic reading and processing. A wide range of magnetic media and solid state recorders are available.

The movement of a float, bucket, or weighing mechanism can also be converted into an electric signal and transmitted by radio or wire to a distant receiver, where records may be collected from a number of recorders on data-logging equipment (2.5.5).

3.5 SNOWFALL AND HAIL [HOMS C53]

Snow that accumulates in a drainage basin is a natural storage reservoir from which a major part of some basin's water supply is derived. This section describes procedures for the measurement of snow cover. Discussion on snow surveys and on snow cover network design is provided in 2.4.2. Guidance for the use of satellite remote-sensing of snow cover is covered in 3.13. Additional discussion of snow cover measurement is given in *Snow Cover Measurements and Areal Assessment of Precipitation and Soil Moisture* (WMO-No. 749).

3.5.1 Depth of snowfall

Snowfall is the amount of fresh snow deposited over a limited period. Measurements are made of depth and water equivalent. Direct measurements of fresh snow on open ground are made with a graduated ruler or scale. A mean of several vertical measurements should be made in places where there is considered to be an absence of drifting snow. Special precautions should be taken so as not to measure old snow. This can be done by sweeping a suitable patch clear beforehand or covering the top of the snow surface with a piece of suitable material (such as wood, with a slightly rough surface, painted white) and measuring the depth down to this. On a sloping surface (to be avoided if possible), measurements should also be made with the measuring rod in a vertical position. If there is a layer of old snow it would be incorrect to calculate the depth of a snowfall from the difference between two consecutive measurements of the total depth of the fresh and old snow, because of the continuous settling of the old snow. Where strong winds have occurred, a large number of measurements should be made to obtain a representative depth.

The depth of snow may also be measured in a fixed container of uniform cross-section after the snow has been levelled without being compressed. The container should be well above the average snow level, for example, at least 50 cm above the maximum observed level, and not exposed to drifting snow. The receiver should be at least 20 cm in diameter and should either be sufficiently deep to protect the catch from being blown out or else be fitted with a snow cross, that is, two vertical partitions at right angles, subdividing it into quadrants.

Ordinary unshielded receivers are unreliable when the wind is strong because of the wind eddies around the mouth of the receiver. Their catch is usually much less than that of a shielded gauge. However, large errors may be caused, in spite of the use of a shield, by the collection of drifting snow. Such errors can be reduced by mounting the gauges 3 to 6 m above the surface.

3.5.2 Water equivalent of snowfall

The water equivalent of a snowfall is the amount of liquid precipitation contained in that snowfall. It should be determined by one of the methods given below. It is important to take several representative samples:

(a) Weighing or melting – Cylindrical samples of fresh snow are taken with a suitable snow sampler and either weighed (the column of snow is known as a snow pillar) or melted;
(b) Using raingauges – Snow collected in a non-recording raingauge should be melted immediately and measured by means of an ordinary measuring cylinder graduated for rainfall.

The weighing-type recording gauge may also be used to determine the water content of snowfall. During snowfall periods, the funnels of the gauges should be removed so that any precipitation can fall directly into the receiver. Snow pillars are widely used in the western United States, where the SNOw TELemetry (SNOTEL) network uses over 500 gauges. High melt rates can take place under high wind speeds with the passage of a warm front.

3.5.3 Snow cover

3.5.3.1 Snow courses

A snow course is defined as a permanently marked line where snow surveys are taken each year. Snow courses must be carefully selected so that measurements of water equivalents will provide a reliable index of the water in snow storage over the entire basin.

In mountainous areas, the selection of appropriate locations for snow courses may be a challenging exercise because of the difficult terrain and serious wind effects. Criteria for the ideal location of a snow course in mountainous areas are:

(a) At elevations and exposures where there is little or no melting prior to the peak accumulation if the total seasonal accumulation is to be measured;
(b) At sites sufficiently accessible to ensure continuity of surveys;
(c) In forested areas where the sites can be located in open spaces sufficiently large so that snow can fall to the ground without being intercepted by the trees;
(d) At a site having protection from strong wind movement.

Criteria for suitable snow course locations are the same as those for siting precipitation gauges for measurement of snowfall.

In plain areas, the snow course locations should be selected so that their average water equivalents will represent, as nearly as possible, the actual average water equivalent of the area. Thus, it is desirable to have snow courses in typical landscapes, such as in open fields and forests, with different snow accumulation conditions.

If the snow cover in an area under consideration is homogeneous and isotopic and if there exists a spatial correlation function for the depth or water equivalent of the snow, the length of the snow course or the number of measuring points along it needed to determine a mean value to a given accuracy can be determined.

3.5.3.2 Points of measurement

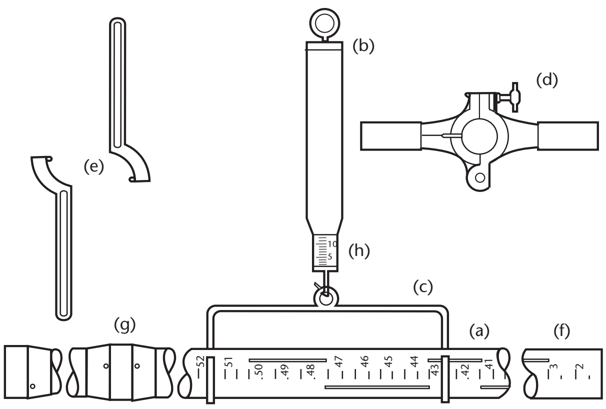
Measurements at a snow course in a mountainous terrain usually consist of samples taken at points spaced 20 to 40 m apart. More samples will be required in large open areas where snow will tend to drift because of wind action. Because sufficient knowledge of the tendency of the snow to drift may be initially lacking, it may be necessary to provide for an extensive survey having long traverses and a large number of measurements. Once the prevailing length and direction of the snowdrifts have been ascertained, it should be possible to reduce the number of measurement points. In plain regions, the distance between points of snow density sampling should be 100 to 500 m. Depth of snow along the snow course should also be measured at about five equally spaced points between the density samples.

Each sampling point should be located by measuring its distance from a reference point marked on a map of the snow course. Stakes should be set high enough to extend above the deepest snow and offset from the course far enough not to affect the snow cover. They may be placed as markers opposite each point where snow samples are to be taken, or at as many points as necessary, to minimize possible error in locating the sampling point. The ground surface should be cleared of rocks, stumps and brush for 2 m in all directions from each sampling point.

Watercourses and irregular ground surfaces should also be avoided by at least 2 m. If a course meanders through timber and if small openings are used as places of sampling, each point should be located with respect to two or three marked trees.

3.5.3.3 Snow-sampling equipment [HOMS C53]

Snow-sampling equipment commonly consists of a metal or plastic tube (sometimes in sections for portability) with a snow cutter fixed at its lower end and with a length scale stamped on its exterior surface throughout its length, a spring or level balance for determining the weight of the snow cores, a wire cradle for supporting the tube while it is being weighed and tools for operating the snow sampler. A typical set of equipment for deep snow, shown in Figure I.3.8, is described in the following way:



(a) snow-sampling tube
(b) tubular spring balance
(c) cradle
(d) driving wrench
(e) spanner wrenches
(f) cutter
(g) screw couplings
(h) scale

Figure I.3.8. Snow-sampling equipment

(a) Cutter – The cutter must be designed to penetrate various types of snow, through crusted and icy layers and, in some cases, through solid ice layers of appreciable thickness that may form near the surface. The cutter must not compact the snow so that an excessive amount of snow is accepted by the interior of the cutter. The cutter must seize the core base with sufficient adhesion to prevent the core from falling out when the sampler is withdrawn from the snow.
Small diameter cutters retain the sample much better than large cutters, but larger samples increase the accuracy in weighing. The shape of the cutter teeth should be designed to allow sufficient back feed on the cutter to remove the ice chips. The cutter should be as thin as practicable but somewhat larger than the outside diameter of the driving tube. This construction allows the chips to find a dumping area when carried backward by the feed on the cutter. The horizontal cutting surface on the cutter blade should be sloped slightly backward to carry the chips away from the interior of the cutter and should be kept sharp so that there is a definite separation of the snow at the inner wall. A large number of teeth provide a smooth cut and keep the cutter free of large chunks of ice;
(b) Sampler tube – In most cases, the inside diameter of the driving tube is larger than the inside diameter of the cutter. The core, therefore, is able to proceed up the tube with a minimum of interference from friction on the wall. However, in normal snow, the core will tend to move over and rub on the walls of the driving tube. Therefore, the walls should be as smooth as possible so that the core may proceed upward without undue friction. In most cases, samplers are constructed of anodized aluminium alloy. While the surface may appear smooth, it cannot be assumed that this will assure non-adhesion of the snow, especially when sampling is made of wet spring snow with a coarse-grained structure. The application of wax may minimize sticking. Some samplers are provided with slots so that the core length may be determined. In general, especially with wet snow, the core length inside may be considerably different from the true depth of the snow measured on the outer markings on the sampler. The slots also provide an entrance for a cleaning tool. The advantage of the slotting arrangement is that errors due to plugging may be immediately detected and erroneous samples may, obviously, be discarded at once. However, the slots may allow extra snow to enter the sampler and increase the measured water equivalent;

(c) Weighing apparatus – The standard way to measure the water equivalent of snow samples is to weigh the snow core collected in the sampler. The core is retained in the sampler, and the sampler and core are weighed. The weight of the sampler is known.

Generally, weighing is accomplished by means of a spring scale or by a special balance. The spring scale is the most practical approach as it may be easily set up and read even under windy conditions. However, the spring scale is accurate only to about 10 g, and the error in weighing by this method may be appreciable for small diameter samplers and shallow depths of snow. Scale balances, potentially more accurate, are very difficult to use in wind. It is doubtful if the intrinsically greater accuracy of this system can be realized except in calm conditions.

Another approach is to store the samples in plastic containers or bags and return them to a base station where they may be accurately weighed or melted and measured with a graduated cylinder. In practice, this procedure is difficult to carry out as the samples must be bagged without loss, carefully labelled, and carried back to the base. The advantage of measurement in the field is that any gross errors due to plugging the sampler, or losses due to part of the sample falling out, may be readily recognized, and repeat readings can be taken at once.

The results may be recorded on site with other pertinent observations and, if a good notebook is used, there can be little chance of confusion as to the location or the sampling conditions.

In all measurements of this type, the extremely difficult physical conditions under which observations must frequently be made should always be kept in mind, and practical consideration should prevail in sampler designs.

3.5.3.4 Snow-sampling procedures

Sampling points should be located by measuring from a reference mark, as indicated on the map of the snow course. Missing a point by more than a few metres may result in significant error.

In order to cut the core, the sampler is forced vertically downward through the snow cover until it reaches the ground. If snow conditions permit, a steady downward thrust, causing an uninterrupted flow of the core into the tube, is best. A minimum amount of turning in a clockwise direction is possible without interrupting the downward thrust. This brings the cutter into play, which is desirable for quick penetration of thin ice layers.

With the cutter at or slightly below ground level and the sampler standing vertically, the reading on the scale that corresponds to the top of the snow is observed.

When the depth that the sampler has penetrated beyond the bottom of the snow cover is ascertained and deducted from this reading, the result is recorded. This is an important reading because it is used in computing the snow density.

In order to prevent loss of core through the cutter while the sampler is withdrawn from the snow, sufficient soil is gathered in the cutter to serve as a plug. The extent to which this will have to be done depends on the condition of the snow.

About 25 mm of solid soil may be required to hold slush. A trace of ground litter on the lower end of the sampler indicates that no loss has occurred.

The length of snow core obtained is observed through the tube slots and read on the scale on the outside of the sampler. After this reading is corrected for any foreign matter picked up in the cutter, it is recorded. The purpose of this reading is to provide a means for judging quickly if a complete sample of the snow cover has been obtained.

The measurement is completed by carefully weighing the snow core in the tube.

The weight of the snow core in equivalent centimetres of water can be read directly on the scale of the balance. The density of the snow is computed by dividing the water equivalent of the snow by the depth of the snow. The density should be reasonably constant over the entire course. A large deviation from the average usually indicates an error in measurement at an individual point.

3.5.3.5 Accuracy of measurements

The accuracy of measurements of snow depth or the water content of snow cover at individual points of the snow course depends on the graduations of the scale being used and on instrumental and subjective errors.

3.5.3.6 Depth and extent of snow cover

Measurements of snow cover over extended areas together with an established local correlation with density make it possible to approximate the water content of the snowpack.

The most common method for determining the depth of snow cover, primarily in regions of deep snow, is by means of calibrated stakes fixed at representative sites that can be inspected easily from a distance. This procedure may be acceptable if the representativeness of the site is proven and if the immediate surroundings of the site (about 10 m in radius) are protected against trespassing. The readings are taken by sighting over the undisturbed snow surface.

The stakes should be painted white to minimize undue melting of snow immediately surrounding them. The entire length of the stake should be graduated in metres and centimetres.

In inaccessible areas, stakes are provided with crossbars so that they can be read from a distance with the aid of field glasses, telescopes, or from aircraft. In the case of measurements of snow depth from aircraft, visual readings of snow stakes may be supplemented by large-scale photographs of the snow stakes, which make the readings less subjective.

The vertical depth of snow cover is also measured by direct observation with a graduated snow tube, usually during the course of obtaining the water equivalent.

3.5.3.7 Radioisotope snow gauges

Radioactive gamma sources are used in various ways to measure water equivalent of snow. Attenuation of gamma radiation may be used to estimate the water equivalent of a snow cover between a source and a detector. One type of installation (vertical) is used to measure total water equivalent above or below a point source. A second installation (horizontal) measures water equivalent between two vertical tubes at selected distances above the ground.

Installation of isotope snow gauges requires relatively expensive and complex instrumentation. In addition, adequate safety measures should be part of any installation, especially where a comparatively high energy source is required. In all cases, consultation with the appropriate licensing or controlling agencies during the development stage is essential and will eliminate many difficulties later on. Although these constraints may limit the use of these gauges, they are a valuable tool and provide the possibility of continuous recording that is particularly useful in inaccessible regions.

Vertical radioisotope snow gauges

Measurement of snow density with the use of radioactive isotopes depends on attenuation of gamma rays traversing a medium. This attenuation is a function of the initial energy of the rays and the density and thickness of the substance traversed. A high energy source of gamma radiation is required, and cobalt-60 is frequently used for this purpose because of its high gamma energy and long half-life (5.25 years).

The source, contained in a lead shield, is placed so that the upper surface of the shield is on the same level as the ground surface, and the beams of gamma rays are directed on the radiation detector above the snow. The detector is a Geiger–Müller or scintillation counter. The impulses from the counter are transmitted to a scalar or, in the case of continuous recording, to an integrator and recorder.

The source of radiation may also be placed in the soil at a certain depth (50–60 cm) so that the gamma rays pass not only through the snow cover but also through a layer of soil. By this means, it is possible to obtain data during the melting of snow pertaining to the quantity of water permeating into the soil or flowing off the surface. There is also a third way of placing the system in the field. The radiation detector counter is placed under the ground surface and the source with shielding is placed above the expected maximum snow layer. This arrangement reduces temperature variations of the detector and provides a constant background count.

Horizontal radioisotope snow gauges

In France and the United States, various modifications of telemetering radioisotope snow gauges have been developed, giving a horizontal and vertical profile of the layer of snow and transmitting the results of measurements to base stations via land, radio or satellite. In both types, the measuring element consists of two vertical tubes of the same length, at a distance of 0.5 to 0.7 m from each other. One tube contains a source of gamma-radiation (Caesium-137 with a half-life of 30 years and activity of 10 or 30 mCi), while the other contains a detector (Geiger–Müller counter or scintillation crystal with photo-multiplier). In the process of obtaining a profile, a special motor, synchronous with the detector, moves the radioactive source upwards and downwards in the tube.

By recording the intensity of the horizontal flux of gamma pulses outside and at various levels inside the layer of snow and by suitably processing the data at a base station, it is possible to determine the depth of snow cover and the density and water content of the snow at a given depth. Furthermore, freshly fallen snow, liquid precipitation and the rate of melting of the snow can be determined.

3.5.3.8 Snow pillows

Snow pillows of various dimensions and materials are used to measure the weight of snow that accumulates. The most common pillows are flat circular 3.7-m diameter containers of rubberized material filled with a non-freezing liquid. The pillow is installed on the surface of the ground, flush with the ground, or buried under a thin layer of soil or sand. In order to prevent damage to the equipment and to preserve the snow cover in its natural condition, it is recommended that the site be fenced. Under normal conditions, snow pillows can be used for 10 years or more.

Hydrostatic pressure inside the pillow is a measure of the weight of the snow on the pillow. Measurement of the hydrostatic pressure is by means of a float-operated water-level recorder or a pressure transducer. Snow pillow measurements differ from those made with standard snow tubes, especially during the snow melt period. They are most reliable when the snow cover does not contain ice layers, which can cause bridging above the pillows. A comparison of the water equivalent of snow, determined by snow pillow, with measurements by the standard method of weighing, showed differences of 5 to 10 per cent.

3.5.3.9 Natural gamma radiation surveys

The method of gamma radiation snow surveying is based on attenuation by snow of gamma radiation emanating from natural radioactive elements in the ground. The greater the water equivalent of the snow, the more the radiation is attenuated. Measurement of gamma radiation can be made by terrestrial or aerial survey. The ratio of gamma radiation intensity measured above the snow cover to that measured over the same course before snow accumulation provides an estimate of the water equivalent.

Aerial gamma surveys of snow cover

While the snow course is a series of point measurements, the aerial survey is an integrated areal estimate of snow cover equivalent. The method is intended for mapping the water equivalent of snow in flat country or in hilly country with a range in elevation up to 400 m. In regions with more than 10 per cent of their areas in marshland, the measurements of water equivalent of snow cover are made only for those areas without marshes, and the integrated characteristics are applied to the area of the entire basin. The usual flying height for an aerial gamma survey is 25 to 100 m above the land surface.

Measurements consist of the total count for a large energy range and spectral counts for specific energy levels. The spectral information permits correction for spurious radiation induced by cosmic rays and radioactivity of the atmosphere. The accuracy of an aerial gamma survey of snow cover depends primarily on the limitations of the radiation measuring equipment (for example, the uniformity of operation of the measuring instruments), fluctuations in the intensity of cosmic radiation and radioactivity in the layer of the atmosphere near the ground, soil moisture variations in the top 15 cm, uniformity of snow distribution and absence of extensive thawing (for example steady flying conditions, or errors in setting course for successive flights). The expected error ranges between ±10 per cent, with a lower limit of approximately 10 mm water equivalent.

Detailed experiments have shown that the standard deviation of measurements of the water equivalent of snow made from an aircraft over a course of 10 to 20 km is about 8 mm and is of a random nature.

To obtain the water equivalent of snow over an area up to 3 000 km^2, with an error not exceeding 10 per cent, recommended lengths of course and distances between courses are given in Table I.3.2.

A great advantage of the gamma survey method is that it yields an aerial estimate of water equivalent over a path along the line of flight. The effective

Table I.3.2. Recommended lengths of flight courses (L) and distance between courses (S)

Natural regions	*S km*	*L km*
Forest-steppe	40–50	25–30
Steppe	40–50	15–20
Forest	60–80	30–35
Tundra	80–100	35–40

width of the path is approximately two to three times the altitude. A second advantage is that the attenuation rate of the gamma rays in snow is determined solely by the water mass independent of its state.

Ground surveys

A hand-carried detector provides a means of measurement of the averaged water equivalent for a band width of approximately 8 m for the length of the course. Water equivalents from 10 to 300 mm may be measured. The accuracy of the measurement ranges from ±2 mm to ±6 mm depending on changes in soil moisture, distribution of the snow, as well as the stability of the instrument system.

A stationary ground-based detector, such as a Geiger–Müller counter or scintillation crystal with photo-multiplier, may also be installed over a snow course area and be used to monitor the water equivalent of an area. However, the occurrence of precipitation carries considerable gamma radiating material to the snow cover, and measurements during and following precipitation are affected by this additional radiation.

Decay of the radiating material permits accurate readings of the water equivalent, approximately four hours after precipitation ceases. Comparison of readings before and after the occurrence of precipitation will provide information on the change in the water equivalent of the snow cover.

3.5.4 Hail pads

Direct measurements of the size distribution of hail are made using, for example, a 1 m x 1 m square of material such as polystyrene on to which hail stones fall leaving an indentation the size of which may be measured.

3.6 RAINFALL ESTIMATION FROM CATCHMENT WATER BALANCE

This chapter is concerned primarily with instrumentation, however, it is important to appreciate that integrated measurements of catchment rainfall can be derived from consideration of the water balance equation in ungauged basins where no instrumentation is available. The amount of water percolating into the soil is related to the effective rainfall, that is, the difference between the rainfall reaching the ground and that evaporated from the surface and vegetation. A simple input-storage-output hydrological model of the catchment may be used to relate the measured river hydrograph to the effective rainfall (Chapter 4).

3.7 OBSERVATION OF PRECIPITATION BY RADAR [HOMS C33]

3.7.1 Uses of radar in hydrology

Radar permits the observation of the location and movement of areas of precipitation, and certain types of radar equipment can yield estimates of rainfall rates over areas within range of the radar (Bringi and Chandrasekar, 2001). For hydrological purposes the effective radar range (European Commission, 2001) is usually 40 to 200 km depending on the radar characteristics, such as antenna beam, power output and receiver sensitivity. The hydrological range of the radar is defined as the maximum range over which the relationship between the radar echo intensity and rainfall intensity remains reasonably valid. The rate of rainfall in any area of precipitation within hydrological range can be determined provided the radar is equipped with a properly calibrated receiver gain control.

Precipitation attenuates the radar beam and this effect is greatest for short wavelength radar. On the other hand, long wavelength radar does not detect light rain and snow as readily as shorter wavelength equipment. The selection of a suitable wavelength depends on climatic conditions and the purposes to be served. All three of the radar bands given in Table I.3.3 are in use for observation of precipitation.

3.7.2 The radar–rainfall equation

The radar equation is sometimes referred to as the free space maximum range equation (FSMR). This equation defines the maximum range that can be anticipated from a particular radar system. For precipitation targets, where rainfall is considered

Table I.3.3. Weather radar frequency bands

Band	*Frequency (MHz)*	*Wavelength (m)*
S	1 500–5 200	0.1930–0.0577
C	3 900–6 200	0.0769–0.0484
X	5 200–10 900	0.0577–0.0275

to have filled the radar beam, the equation has the form:

$$P_r = P_t \pi^3 G^2 \theta.\phi.h K^2 Z / 512(2 \ln 2) R^2 \lambda^2 \quad (3.4)$$

where P_r is the average power in watts received from a series of reflected pulses, P_t is the peak power transmitted in watts, G is the antenna gain, θ and ϕ are the horizontal and vertical beam widths, h is the pulse length in metres, R is the range in metres, λ is the wavelength in metres, K^2 is the refractive index term of rain (0.9313 for 10-cm radar equipment assuming a temperature of 10°C), and Z is the reflectivity.

It should be understood that equation (3.4) is only applicable under certain assumptions (European Commission, 2001; Meischner, 2003), and therefore is likely to be in error when these conditions are not met. Nevertheless, it is the basis of all radar estimates of precipitation from a single frequency radar.

The rainfall rate in mm h^{-1} is related to the median drop diameter, as follows:

$$\sum d^6 = aP_i^b \quad (3.5)$$

where P_i is the rainfall intensity in mm h^{-1} and a and b are constants. Many determinations have been made of the drop size distribution measured at the ground and the conversion by means of the fall speeds of different sized drops to a particular rainfall rate. The most common equation in use is:

$$Z = 200\, P_i^{1.6} \quad (3.6)$$

3.7.3 Factors affecting measurements

A summary of the factors affecting measurements is discussed in turn below.

3.7.3.1 Wavelength

The use of S-band frequency, as in the United States, removes problems associated with attenuation of the radar beam as it passes through precipitation. The use of C-band frequency in much of the rest of the world improves sensitivity, but does result in attenuation problems. C-band systems are presently a factor of about two cheaper than S-band systems for the same aerial dimensions, although this may change with the introduction of tuneable travelling wave tube (TWT) technology in the future. Correction procedures have been developed for attenuation at C-band (3.7.3.4).

3.7.3.2 Ground clutter

Both the main part of the radar beam and the side lobes may encounter ground targets. This will cause strong persistent echoes, known as ground clutter, to occur, which may be misinterpreted as rainfall. Although radars may be sited to minimize these echoes, it is not possible to remove them altogether, and other techniques such as the use of Doppler processing with clutter map removal (Germann and Joss, 2003) must be used.

In addition to producing permanent echoes, interception of the beam by the ground also causes occultation or screening of the main part of the beam. In this case only a fraction of the power illuminates the rain at longer ranges. This may be corrected provided at least 40 per cent of the beam is unobstructed. It is possible to simulate the visibility from a radar site using a digital terrain model, although the result is not perfect owing to small errors in the pointing angle, uncertainties in the simulation of the refraction of the radar beam and insufficient resolution of the digital terrain model particularly at close range.

3.7.3.3 Beam width and range

At 160 km, the radar beam may be several kilometres wide, depending on the beam width employed. Normally, there will be marked variations in the radar reflectivity within this large sampling volume. Thus, an average value over a large volume is obtained, rather than a point value. The radar equation is based on the beam being filled with meteorological targets. Therefore, one would not expect the values of rainfall rate obtained with a radar to be highly correlated with point raingauge measurements. However, the areal pattern displayed by radar should generally be much more representative of the true storm isohyetal configuration than that measured by most raingauge networks.

In showery conditions it has been found that the frequency of echoes recorded at 160 km was only about 4 per cent of that of echoes recorded at 64 km. Therefore, a shower which fills the beam at 64 km would only fill about one eighth of the beam at 160 km. This result is due to a combination of beam width and beam elevation factors.

3.7.3.4 Atmospheric and radome attenuation

Microwaves are attenuated by atmospheric gases, clouds and precipitation. The attenuation experienced by radio waves is a result of two effects: absorption and scattering. In general, gases act only as absorbers, but cloud and raindrops both scatter and absorb. For radar sets operating at the longer wavelengths, attenuation is not a problem and can usually be neglected. The generally accepted form of expressing attenuation is in decibels. The decibel (dB) is used as a measure of relative power and is expressed as:

$$dB = 10 \log_{10} P_t / P_r \qquad (3.7)$$

where P_t and P_r would be the power transmitted and power received. Signal attenuation as related to rate of rainfall and wavelength is given in Table I.3.4.

Corrections may be made to account for the distance from the radar site ($1/R^2$, R = range), for the attenuation due to beam dispersion by atmospheric gases (0.08 dB km^{-1} one way) and for signal attenuation through heavy rain (Table I.3.4). However, such procedures (Meischner, 2003; Collier, 1996) may be unstable in cases of severe attenuation and operational corrections are "capped" (limited) at an upper limit based on what is reasonable. In the future it may be that procedures based on the use of multiparameter radar (3.7.8) will be employed.

Table I.3.4. Radar signal attenuation due to precipitation (dB km^{-1})

Rate of rainfall (mm h^{-1})	*Wavelength (m)*			
	0.1	0.057	0.032	0.009
1.0	0.0003	0.002	0.007	0.22
5.0	0.0015	0.015	0.061	1.1
10.0	0.003	0.033	0.151	2.2
50.0	0.015	0.215	1.25	11.0
100.0	0.015	0.481	3.08	22.0

Distance (km) over which precipitation at a given rate of rainfall must extend to give an attenuation of 10 dB at various wavelengths

Rate of rainfall (mm h^{-1})	*Wavelength (m)*			
	0.1	0.057	0.032	0.009
1.0	33 000	4 500	1 350	45
5.0	6 600	690	164	9.1
10.0	3 300	310	66	4.5
50.0	600	47	8	0.9
100.0	300	21	3.2	0.4

3.7.3.5 Refraction of beam and multiple scattering

Radar waves are propagated through space with a refractive effect which gives the waves a curved path. The approximate mean radius is four thirds the mean radius of the Earth. As a result of vertical moisture discontinuities, additional refractive bending of the radar beam can occur. This produces what is often called ducting or trapping of the radar beam and either causes the radar beam to re-curve earthward or to be curved upwards overshooting precipitation 80 to 120 km away. The meteorological conditions favouring ducting (trapping) can be determined mathematically.

If the radar pulses are scattered by water-coated ice spheres, then a process known as three-body scattering may produce unusual precipitation signatures. The "hail spike" is such a signature. This process involves the combined scattering from the ground as well as from hydrometeors, but is not a common phenomenon.

3.7.3.6 Vertical velocity

Vertical velocity of rainfall in very intense convective systems may cause radar echoes, which in turn may cause the relationship between rainfall, R, and radar reflectivity, Z, to differ quite significantly from that in still air. For example, in a downdraft of 8 m s^{-1} the reflectivity value for a given rainfall rate would be about 3 dB less than in still air, producing an underestimate of the rainfall rate by 40 per cent.

3.7.3.7 Vertical profile of reflectivity

The main factor introducing bias into radar estimates of surface precipitation is the vertical measurement geometry of weather radars. At increasing range a radar measurement volume is located at increasing altitude above the Earth's surface. Hence a radar measurement of reflectivity aloft can be accurate, but not representative of conditions at the surface. This is not a measurement error, but a sampling problem.

When the radar beam intersects the level at which snow starts to melt the reflectivity is enhanced, and the result is known as the bright band. This occurs a few hundred metres below the freezing level (see Figure I.3.9). In this figure, when snow is present throughout the depth of the precipitation, the bright band is not present and the radar reflectivity decreases with increasing height.

The vertical profile of reflectivity (VPR) above each point on the Earth's surface can be denoted as $Z_e(h)$, where h is the height above the surface at range r from the radar site. The shape of VPR determines the magnitude of the sampling difference (Koistinen and others, 2003). Denoting the shape of the radar beam pattern f^2 then,

$$Z_e\,(h, r) = \int f^2\,(y)\,Z_e\,(h)\,dy \qquad (3.8)$$

The integration is performed vertically (y) from the lower to the upper edge of the beam. The vertical sampling difference (in decibels, or dB) is then

$$c = 10 \log\,(Z_e\,(0) / Z_e\,(h, r)) \qquad (3.9)$$

where Z_e (0) is the reflectivity at the surface in VPR. Hence by adding the sampling difference c to the measured reflectivity aloft (dBZ), the reflectivity at the surface $dBZ(0, r)$ is,

$$dBZ\,(0, r) = dBZ + c \qquad (3.10)$$

In snowfall the sampling difference increases as a function of range, indicating a significant under-estimation of the surface precipitation even at close ranges. However, in rainfall the radar measurement is relatively accurate up to a range of 130–140 km. When the height of the bright band is more than about 1 km above the radar antenna, the overestimation due to it will compensate for the underestimation effect of snow in the beam. Hence, a radar measurement is more accurate at longer ranges than it would be without the bright band.

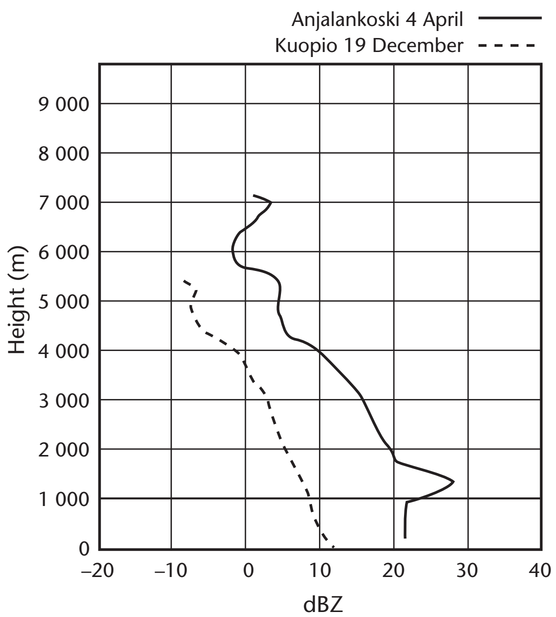


Figure I.3.9. Two vertical profiles of reflectivity averaged from single polar volumes at ranges of 2–40 km from the radar. The solid line represents rain and the dashed line snowfall
(from Koistinen and others, 2003)

In some parts of the world there exists orographic growth of precipitation at low levels over hills exposed to strong moist maritime airflows. This growth may be reflected in VPR, but sometimes may occur below the height of the radar beam. In this case, to some extent the growth may be estimated by applying climatological correction factors. In some synoptic situations, for example ahead of warm fronts, the opposite effect may be observed, mainly low level evaporation. In this situation it is more difficult to apply a correction, and it may be necessary to use the output from a mesoscale numerical forecast model. However, care should be taken in using model output to do this and it may be more reliable to use as low a radar beam elevation as possible to observe the evaporation in VPR.

3.7.4 Snow and hail

A radar is capable of measuring snowfall as accurately as rainfall. However, the accuracy depends very much upon VPR and, in particular the height of the bright band. As for rainfall the a and b (equation 3.5) in the $R{:}Z$ relationship may vary greatly depending upon, for example, whether the snow is wet or dry. Typical values often used are a = 2000 and b = 2.0.

When a radar observes hail, the backscattered power is no longer proportional to the sixth power of the particle size, and Mie theory is applicable. When only hail is observed in the pulse volume the number of hail in the pulse volume is directly related to the hail diameter D_H (mm) by (Auer, 1972):

$$N\,(D_H) = 561\,D_H^{-3.4} \qquad (3.11)$$

Hence assuming Rayleigh scattering in the hail at C-band:

$$Z = 10 \log_{10}\,(561\,D_H^{2.6})\ dBZ \qquad (3.12)$$

Problems occur in heavy rainfall which often contains hail. This increases the reflectivity. Whilst polarization radars are capable of detecting the presence of hail directly, single polarization radars are not, and other techniques must be used (Collier, 1996).

3.7.5 Scanning strategy

Automatic electronic radar digitizers, capable of sampling radar echoes at the rate of 80 range-increments for each 1–2° of azimuth, have been developed and are now standard on all commercial radars. Their data are recorded on magnetic tape or other magnetic media for immediate on-site computer analysis, transmission via data link to a remote computer, or for retention and later analysis. The results of this type of sampling are similar to those for manual methods, except that the number of discrete samples is larger by at least one order of magnitude than the finest grid overlay. The time required to sample and record the entire radar sweep is about 1 to 4 minutes and 8 to 14 elevations may be used.

The actual scan strategy deployed is dependent upon the use to which the radar data are to be put. Two types of scanning strategies, known as contiguous and interlaced, may be employed. For contiguous scanning the radar beam is scanned rapidly through all elevations after which the procedure is repeated. For interlaced scanning every other beam elevation is missed out in the first scan sequence, and the elevations omitted are implemented during the second scan sequence immediately following the first. The data from individual radar beams may be combined to use the best data at each particular bin based on beam height, terrain effects and beam blockage.

3.7.6 Summary of accuracy considerations

As discussed previously, a number of difficulties must be faced in retrieving estimates of surface rainfall from radar. Of particular importance is the vertical variability of radar reflectivity (VPR; see also 3.7.3.7).

Vignal and others (2000) discuss three approaches to the determination of VPR. They found that a correction scheme based on a climatological profile improves the accuracy of daily radar rainfall estimates significantly within 130 km of the radar site. The fractional standard deviation (FSE) is reduced from the uncorrected value of 44 per cent to a corrected value of 31 per cent. Further improvement is achieved using a single, mean hourly average VPR (FSE = 25 per cent) and a locally identified profile (FSE = 23 per cent). This analysis was carried out for both stratiform and convective rainfall, although better improvement is obtained for stratiform events.

Although it has now been recognized that the application of a VPR correction is an essential first step after the removal of ground clutter echoes in estimating surface precipitation, bias errors may remain. The appropriateness of subsequent raingauge adjustment (Meischner, 2003) to mitigate residual errors remains uncertain. However, the application of time-integrated raingauge data does produce improvements, particularly in mountainous terrain (Collier, 1996).

It seems clear that a single polarization radar can be used to measure daily rainfall to an accuracy approaching 10 per cent, provided VPR is adjusted carefully. Such a level of accuracy is similar to that provided by raingauges. However, sub-daily rainfall is much more problematic, particularly at C-band and shorter wavelengths for which attenuation is a serious problem in convective rainfall. Hourly rainfall over catchments of about 100 km^2 may be measured to a mean accuracy of 20 per cent in stratiform rain, but only 40 per cent or so in heavy convective rain. The current aim for point measurements of instantaneous rainfall is within a factor of two although this has not yet been achieved reliably.

3.7.7 Doppler radar

3.7.7.1 Basics

To measure the absolute speed of movement (or velocity) of a raindrop and its direction of movement instantaneously, it is necessary to use a radar with a very precise transmitter frequency and a receiver system sensitive to the changes of frequency induced by a moving target, even though in the case of meteorological targets these changes may be small. This type of radar is sometimes referred to as a coherent radar but more frequently as a Doppler radar because it uses the well-known Doppler effect. A more detailed discussion of this topic along with added references is found in publications by the European Commission (2001) and Meischner (2003).

Doppler radars have been used for research purposes for many years, both singly and, more recently, in multiple networks consisting usually of two or three radars. They have played a considerable part in the investigation of the atmosphere and are considered by some radar meteorologists to be indispensable in the study of the dynamics of air masses, particularly of convective clouds. However, problems of interpretation of data still exist, and it is only in recent years that serious consideration has been given to their use in operational systems. In certain parts of the world, particularly those subject to violent weather, they constitute operational systems and are now regarded as a highly desirable form of radar. They are inherently more complex though not more expensive than conventional radars, and they require greater processing power and more maintenance effort. Despite this, radars with Doppler capability exist in large national network in the United States and elsewhere. Doppler radars can be used for general forecasting purposes to provide data that may reveal signatures useful for the advanced warning of such phenomena as tornadoes and severe storms. Moreover, they can provide more information on the intensity and structure of these phenomena than any other practical means.

3.7.7.2 Clutter cancellation

Most systems measure precipitation intensities in a conventional way as well as providing Doppler data. One important advantage is that it is possible to determine with some degree of accuracy the position and extent of permanent, and to some degree anomalous propagation, echoes, which are, by definition, stationary from the Doppler channel. This information can then be used in an attempt to ensure that only precipitation data are measured by the non-Doppler channel. As with any other system of clutter removal, the method is unlikely to be totally successful alone since, under some transmission and weather conditions, permanent echoes can appear to move and, conversely, precipitation is sometimes effectively stationary. Doppler clutter cancellation is usually accompanied by other procedures for removing clutter such as clutter maps and use of VPR.

To obtain echoes from refractive inhomogeneities and for the purpose of measuring precipitation intensity to the greatest possible ranges compared with conventional non-Doppler radar or for studying the structure of severe storms, longer wavelengths are necessary, preferably 10 cm.

3.7.7.3 Measuring winds

A number of different techniques for estimating winds using a single Doppler radar have been developed (Bringi and Chandrasekar, 2001; European Commission, 2001; Meischner, 2003). Commercial radar manufacturers now offer some of these techniques, and they are used to produce both mean profiles of horizontal wind velocity and radial winds under certain assumptions. These data are not yet in operational use to help in the estimation of precipitation, although this may change in the near future as such data begin to be assimilated routinely into numerical weather prediction (NWP) models (3.17).

3.7.8 Multi-parameter radar

Development of multi-parameter radar hardware with which to measure the properties of hydrometeors has been slow since the initial production of high-speed switches enabling the rapid alternate transmission of horizontally and vertically polarized microwave radiation. However, in recent years following work on other polarization states such as circular, and the recognition of the potential of multi-parameter radar for measuring precipitation, attention to the design of the hardware has increased.

Versatile research radars such as the CSU-CHILL installation in the United States and Chilbolton in the United Kingdom have provided the test beds from which to consider what polarization base is most effective in measuring rainfall and hydrometeor type. It is now possible to make simultaneous transmissions of horizontal (H) and vertical (V) radiation without the need for a high-power polarization switch. This form of simultaneous transmission is now being implemented on the National Severe Storms Laboratory's (NSSL) research WSR-88D S-band radar, and is being considered as the basis of the polarimetric upgrades to the operational WSR-88D radars in the United States.

3.8 GROUND-BASED RADAR AND RAINFALL MONITORING TECHNIQUES

Ground-based weather radars have been used operationally for more than 20 years in some countries, mostly in combination with raingauge networks, which are often used to calibrate them. Rainfall estimates made by weather radars are sometimes more useful than those made by raingauges because they are continuous in time and space and provide areal coverage (D'Souza and others, 1990). However, associated with them are problems dealing with backscatter, attenuation, absorption and reflection, particularly in areas of varying relief and signal calibration. Although the weather radar networks operated in European and North American countries by the national meteorological agencies adequately service the primary meteorological requirement of daily weather reporting and forecasting, the need for quantitative estimates of rainfall to support applications in hydrology and water resources, especially flood forecasting, has not been so well serviced. Radar is in widespread use as an informal means of initial alert of impending flooding utilizing the moving images of storm systems that it provides. However, quantitative use of radar data is much less common and is constrained by accuracy limitations, especially in mountainous areas and at times when bright band effects are apparent (WMO, 1998). Although the informal use of radar for flood warning is widespread, only two or three countries make quantitative use of the data as part of a flow forecasting system. Even then, the radar is used in a complementary way with additional information being provided by networks of raingauges. The development of distributed flood forecasting models (specially tailored to utilize the radar grid data through a model grid formulation) are still at a pre-operational stage. Furthermore, real-time implementation should only be contemplated following detailed offline assessment and proven improved performance relative to simpler and more conventional lumped models (WMO, 1998, 1999). Weather radar data are less immediately suited to applications for design rainfall estimation, compared with their use for flood forecasting, because of the shortness of radar records relative to the storm periods to be inferred. However, this is in part compensated for by the complete spatial coverage that radars provide. Furthermore, the advantages of the high temporal and spatial resolution of radar data ought to be of particular importance for short duration design estimates.

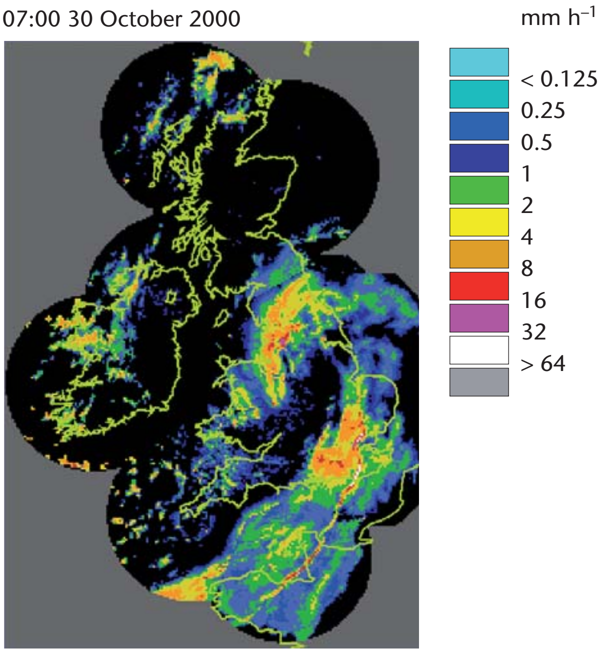


Figure I.3.10. United Kingdom radar network image comprising radar data from the United Kingdom and Ireland at 0700 UTC on 30 October 2000. The different colours represent different rainfall rates in mm h^{-1} as shown. The coastline is shown.
(Courtesy Met Office, United Kingdom)

3.9 OPERATIONAL RADAR NETWORKS

Operational radar networks now exist in many countries. In the United States S-band Doppler radars are used, whereas in Europe C-band systems form most of the networks. An example of a radar network image from the United Kingdom is shown in Figure I.3.10. The boundaries between individual radars are designated by consideration of the heights of the radar beams, presence of ground clutter, areas where the best accuracy is required, etc. The importance of regular maintenance and calibration of radars cannot be overestimated, even though radar technology has now become very reliable, with downtime in some countries being only a few per cent per month, usually planned to occur in no-rain situations.

3.10 DUAL FREQUENCY MICROWAVE LINK ATTENUATION MEASUREMENTS OF RAINFALL

It has been shown (Holt and others, 2000) that good estimates of path-averaged rainfall may be obtained from the difference in signal attenuation caused by rainfall at two frequencies along a microwave communications link. The specific attenuation K (dB km^{-1}) along the link is estimated according to:

$$K = c\, R^{d} \tag{3.13}$$

This relationship depends critically upon the signal frequency, and the parameters c and d which are unknown, but sensitive to temperature, rain drop shape and drop-size distribution. However, if a dual frequency link is used two frequencies and polarization states may be selected for which the specific attenuation differences are relatively insensitive to these unknown parameters. After the raw attenuation measurements have been adjusted for gaseous absorption, there is a linear relationship between this parameter and rainfall rate. An example of the success of this technique in measuring line integrated rainfall is shown in Figure I.3.11.

The installation of a dual frequency microwave link along a radial from a weather radar offers the opportunity to measure the integrated rainfall along the

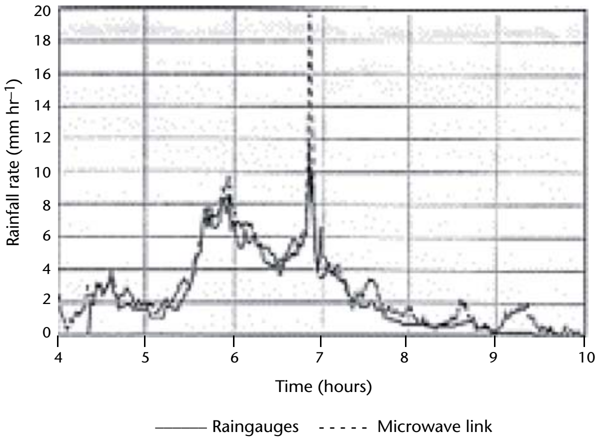


Figure I.3.11. Time series of path-integrated rainfall estimates from raingauges (solid line) and dual frequency microwave link (dashed line) attenuation measurements over north-west England on 10 February 2000
(Holt and others, 2000)

link which may be compared with the same quantity measured by the radar. If the radar operates at an attenuating frequency, then this comparison provides a method of measuring attenuation through rain, or from rain on the radar radome albeit in only one azimuthal direction. This technique is not operational at present.

3.11 OBSERVATIONS OF RAINFALL BY SATELLITE

3.11.1 Basics

Rainfall estimation from space is based on measuring the amount of radiation that is reflected and emitted through cloud tops. Most of the radiation does not penetrate deep into cloud regions containing particles with similar or greater size than the radiation wavelength. Therefore, except for the longest wavelengths, most of the radiation comes from the upper regions of precipitating clouds and can therefore only indirectly be related to surface rainfall. Consequently there are very many techniques using a range of procedures.

3.11.2 Visible and infra-red

Rain intensities vary with the rate of expansion of cold ($T < 235°K$) cloud top areas. It is assumed that the expansion of the cloud top is an indicator for the divergence aloft and, hence, to the rate of rising air and precipitation. However, when used over a large area, this method does not show significant improvement with respect to the simplest possible method which assumes that all clouds with tops colder than a given threshold temperature T precipitate at a fixed rate G mm h^{-1}, where T = 235°K and G = 3 mm h^{-1} is typical of the eastern equatorial Atlantic. This method was developed into the Global Precipitation Index (GPI), which has been used extensively.

Such area methods work well only for a time-space domain that is large enough to include a large number of storms that provide a good representation of the full evolution of convective rain-cloud systems (for example, 2.5° x 2.5° x 12 hours). Classification of clouds into convective and stratiform by the texture of the cloud top temperature has shown some improvement for tropical rainfall over land. However, it fails (along with the rest of the infra-red methods) in mid-latitude winter systems, because there the "convective" relation between cold cloud top area and surface rainfall does not apply to largely non-convective cloud systems.

The use of visible wavelengths to characterize the strength of convection works well when used with infra-red wavelengths to indicate the height of the cloud. However, such procedures can be misleading by the presence of bright cirrus cloud, or the presence of low-level orographic rain.

The atmospheric window of about 10 microns is split into two closely spaced wavebands, centred at 10.8 and 12 microns. Clouds have large absorption and emissivity in the longer waveband. Therefore, the 10.8 micron radiation in thin clouds will be contributed from lower and warmer levels as compared with the 12 micron waveband, creating a brightness temperature difference between the two channels. It has been shown that cirrus clouds can be distinguished from thicker cloud by having larger brightness temperature difference. This helps in eliminating thin clouds from consideration as precipitating cloud.

Very cold cloud top temperature is not always a requirement for precipitation, in which case the infra-red threshold technique breaks down. The precipitation formation processes require the existence of large cloud droplets and/or ice particles in the cloud, which often spread to the cloud top. These large particles absorb the 1.6 and 3.7 micron radiation much more strongly than small cloud droplets. This effect makes it possible to calculate the effective radius (r_{eff} = integral volume divided by integral surface area) of the particles. It has been shown that r_{eff} = 14 microns can serve to delineate precipitating clouds, regardless of their top temperature (Figure I.3.12).

3.11.3 Passive microwave

Microwaves provide the measurements that are physically best related to the actual precipitation, especially in the longest wavebands. The interactions of passive microwave with precipitation clouds and the surface are illustrated in Figure I.3.13, using two wavebands, shorter (85 GHz) and longer (19 GHz). Measurement techniques are based on the two physical principles of absorption and scattering.

Absorption-based measurements

Water drops have relatively large absorption/emission coefficient, increasing for the higher frequencies. The emission is proportional to the vertically integrated cloud and rainwater in the low frequencies, but due to the increased emissivity for the higher frequencies the emission saturates for light rain intensities.

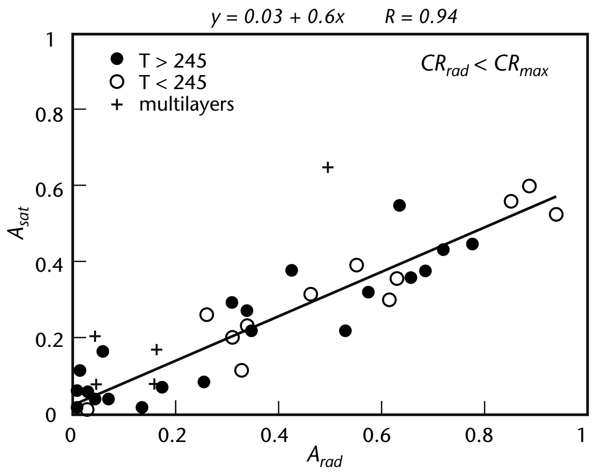


Figure I.3.12. Fraction of precipitating area, defined by the area with $r_{eff} \geq 14$ microns (A_{sat}), as a function of the fraction of precipitating area detected by radar (A_{rad}) for convective clouds. Windows with multi-layered clouds are marked with crosses, windows with cloud-top temperatures higher than 245 K are marked with solid circles, and windows with cloud-top temperatures lower than 245 K are marked with hollow circles. CR_{sat} is the cloud radius parameter and CR_{max} is the maximum cloud radius for a given depth.
(Rosenfeld and Gutman, 1994; Lensky and Rosenfeld, 1997)

Scattering-based measurements

Ice particles have relatively small absorption/emission, but they are good scatterers of the microwave radiation, especially at higher frequencies. Therefore, at high frequencies (85 GHz), the large scattering from the ice in the upper portions of the clouds makes the ice an effective insulator, because it reflects back down most of the radiation emitted from the surface and from the rain. The remaining radiation that reaches the microwave sensor is interpreted as a colder brightness temperature. A major source of uncertainty for the scattering-based retrievals is the lack of a consistent relationship between the frozen hydrometeors aloft and the rainfall reaching the surface.

The two physical principles of absorption and scattering described above have been used to formulate a large number of rain estimation methods. In general, passive microwave rainfall estimates over the ocean were of useful accuracy. However, over the equatorial Pacific, passive microwave does not show significantly improved skill when compared with the simplest infra-red method (GPI).

Over land the passive microwave algorithms can detect rain mainly by the ice scattering mechanism, and this indirect rain estimation method is less accurate. Moreover, rainfall over land from clouds which do not contain significant amounts of ice aloft, goes mostly undetected.

3.11.4 Active microwave (rain radar; Tropical Rainfall Measurement Mission)

A major limiting factor in the accuracy of passive microwave methods is the large footprint, which causes partial beam filling, especially at the higher frequencies. The resolution is greatly improved with the Tropical Rainfall Measurement Mission (TRMM) satellite, with a corresponding improvement in the expected accuracy of the microwave rain estimates. The TRMM satellite has a radar transmitting at a wavelength of 2.2 cm (active microwave) and microwave radiometers (19 to 90 GHz) (Figure I.3.14). The resolutions of these instruments range from about 1 km for the visible and infra-red radiometer, about 10 km for the microwave radiometers and 250 m for the radar. The radar has provided an improvement in the accuracy of instantaneous rain estimates over those previously achieved from space. Since TRMM samples each area between 35 degrees north and south, at best, twice daily, the sampling error is the dominant source of inaccuracy.

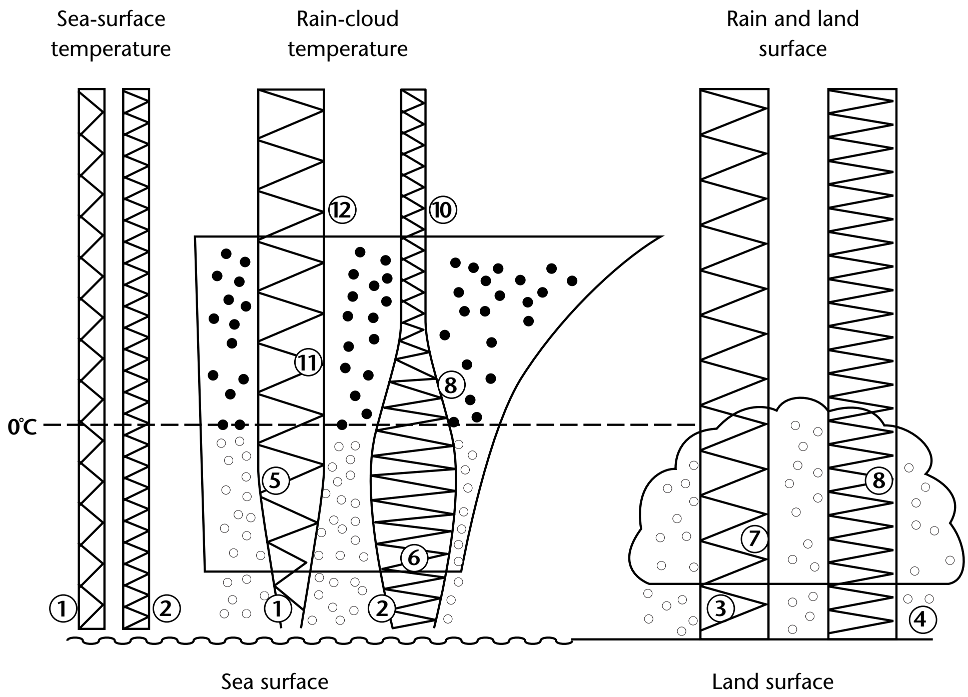


Figure I.3.13. The interaction of high (for example 85 GHz) and low (for example 19 GHz) frequency passive microwave with precipitation clouds and the surface. The width of the vertical columns represents the intensity of temperature of the upwelling radiation. The illustrated features and their demarcations are: (a) the small emissivity of sea surface for both low (1) and high (2) frequencies; (b) the large emissivity of land surface for both low (3) and high (4) frequencies: (c) the emission from cloud and rain drops, which increases with vertically integrated liquid water for the low frequency (5), but saturates quickly for the high frequency (6); (d) the signal of the water emissivity at the low frequency is masked by the land surface emissivity (7); (e) the saturated high frequency emission from the rain (8) is not distinctly different from the land surface background (4); (f) ice precipitation particles aloft backscatter down the high-frequency emission (9), causing cold brightness temperatures (10), regardless of surface emission properties; (g) the ice lets the low frequency emission upwell unimpeded (11), allowing its detection above cloud top as warm brightness temperature (12).
(Rosenfeld and Collier, 1999)

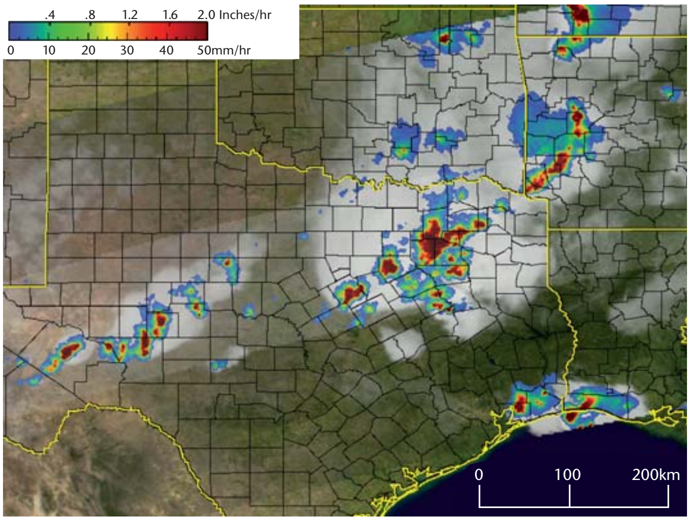


Figure I.3.14. Heavy rainfall over Texas derived from the TRMM Microwave Imager and Precipitation Radar on the TRMM satellite at 0439 UTC 1 May 2004 (Courtesy NASA)

A combination of the measurements from TRMM-like and geostationary satellites provides the best potential for accurate global precipitation estimates from space. Currently plans are being developed to implement such systems under the general title of the Global Precipitation Mission (GPM).

3.11.5 Summary of accuracy considerations

In tropical regions there can be a significant diurnal cycle in rainfall activity, and the phase and intensity of the cycle may vary from region to region. The low inclination orbit used for TRMM will process in such a way as to sample a full diurnal range of Equator crossing times over the course of a month. This is not the case for satellites in polar orbit for which the Equator crossing time is always the same. The diurnal cycle may therefore increase the errors due to sampling.

For monthly averages over a 280 km^2 and a sampling interval of 10 hours, appropriate for the TRMM satellite, the sampling error is about 10 per cent. However, for convective systems in other regions, which have shorter decorrelation times than observed for tropical rain, the sampling error is likely to be larger.

The validation of satellite algorithms for estimating rainfall accumulations is complex and must be undertaken in ways that ensure that different techniques provide data with similar characteristics, that is, integration times and coverage.

The best accuracy for areal rainfall measurements from space at present is obtained over the tropical oceans, where GPI performs as well as passive microwave techniques for long period (in the order of several months) integrated rainfall. However, errors for individual events may be large because "warm rain" from shallow clouds is common in some places in the tropics. The passive microwave techniques become increasingly advantageous towards higher latitudes where convective rainfall occurs less frequently. Here the best accuracy is achieved by combining passive microwave with infra-red from geostationary satellites. Somewhat lower accuracies of infra-red techniques are achievable in convective rain over land, due to large dynamic and microphysical diversity of rain-cloud systems. This causes a larger variability between the rainfall and the properties of the upper portions of the clouds. The skill of passive microwave techniques is also reduced over land, because its emissivity reduces greatly the usefulness of frequencies lower than 35 GHz. Nevertheless, results over land at 88.5 GHz are encouraging.

3.12 REMOTE-SENSING MEASUREMENTS OF SNOW

Remote-sensing of snow can be accomplished using gamma rays, visible and near IR, thermal IR and microwaves. An overview of the relative sensor band responses to various snowpack properties shows that the microwave band has the greatest overall potential followed by the visible and near-infra-red band. The gamma ray portion is extremely limited by the fact that the sensing must be carried out with low altitude aircraft, and to a lesser extent that it is really only sensitive to a finite snow-water equivalent. Thermal infra-red is also limited in potential, but it can be used from space in night-time situations (Rango, 1993; WMO, 1999). Different approaches for determining snow area, water equivalent and snow properties have been developed. These have been driven for the most part by the availability of data from existing satellites or from experimental aircraft and truck programmes. Remote-sensing data are currently being used operationally in snow cover and snow-water equivalent assessments, and seasonal snow melt runoff forecasts. The potential of satellites to provide usable information on snowpack dynamics is now widely recognized and today many schemes exist that employ satellite-derived snow measurements for snow melt runoff prediction (Lucas and Harrison, 1990).

Even more important than snow extent and location for various snowpack processes is the vertical dimension of the snowpack. This vertical dimension essentially provides the information needed for estimating snow volume which relates directly to the potential for snow melt runoff.

Although the airborne gamma-ray spectrometry approach is a very accurate remote-sensing method for measuring the snow-water equivalent, its previously mentioned drawbacks limit its use. However, airborne gamma-ray data and weather satellite data together provide good possibilities for operational snow cover mapping (Kuittinen, 1989; Carroll, 1990).

Airborne gamma-ray spectrometry can be used to determine the snow-water equivalent values, because snow attenuates the terrestrial gamma radiation (WMO, 1992*b*). Background gamma radiation of the soil is obtained before snowfalls, and subsequent flights are flown to measure the gamma radiation through the attenuating snow cover. The degree of attenuation is related to the snow-water equivalent through various calibration graphs.

As resolutions of passive microwave sensors improve (a Russian microwave radiometer at about 0.8 cm wavelength with approximately 8-km resolution was launched in 1996), all-weather capability will increasingly be exploited. The final advantage of the microwave spectrum is that night-time measurements are easily made because of the reliance on emitted microwave radiation as opposed to reflected visible microwave radiation. Emittance and backscattering of microwave radiation are affected by almost all snow parameters, which complicates the measurement of the most needed parameters: water equivalent, areal extent and amount of free water.

Good relationships have been established between snow depth and microwave emission and backscatter for snowpacks that are dry and uniform with little evidence of layering. Such relationships are not so clear once the snowpack has been subjected to thaw and refreeze cycles, whilst the presence of unfrozen water anywhere in the snowpack results in marked changes in microwave response. In general, the use of microwave radiometry appears more reliable than radar for this type of measurement (Blyth, 1993).

The new Special Sensor Microwave/Imager (SSM/I) data are being used operationally to produce snow-water equivalent maps of the Canadian prairies, which are now supplied operationally to Canadian users (Goodison and Walker, 1993).

The active microwave region has a potential similar to the passive microwave region. However, it should be noted that not only are active microwave observations of the snowpack very sparse and almost non-existent, but the analysis of active microwave data is more complex than that of passive data because of the confusion caused by the effect of surface characteristics (including soils) and geometry considerations on the reflected radar wave. The higher resolution (10 m from space) of the active microwave is a considerable advantage over passive microwave. The major problem is the lack of sensors at about 0.8-cm wavelength for experiments on any kind of platform. Although satellite Synthetic Aperture Radar (SAR) can provide high-resolution data, current single-frequency systems such as ERS-1 are likely to be limited to the recognition of the onset of melt and the delineation of wet snow extent. Some of these problems may be overcome by using multifrequency and multipolarization SAR measurements.

Some of the more promising research on remote sensing measurements of snowfall is included in the list of references at the end of this chapter.

3.13 SATELLITE REMOTE-SENSING OF SNOW COVER

Remote-sensing data are currently being used operationally in snow cover and snow-water equivalent assessments, and seasonal snow melt runoff forecasts. The potential of satellites to provide usable information on snowpack dynamics is now widely recognized, and today many schemes exist that employ satellite-derived snow measurements for snow-runoff prediction.

Only satellites enable seasonal snow cover to be monitored periodically, efficiently and on a sufficiently large scale. Significant remote-sensing data for operational snow mapping are available from satellites such as Satellites pour l'observation de la terre (SPOT), Landsat, National Oceanic and Atmospheric Administration (NOAA), Geostationary Operational Environmental Satellite (GOES), Earth Observation Satellites (EOS) and Defense Meteorological Satellite Program (DMSP). The choice of satellite for snow mapping depends upon the smallest partial area of the region to be monitored. While the accuracy of the snowpack delineation and snow area estimation depends upon the spatial resolution of the sensors involved, operational snow mapping schemes are rarely necessary for such small areas (Lucas and Harrison, 1990). As a result, the Landsat Thematic Mapper (TM) sensor is usually applied in the context of research projects and in some cases, aerial photographs may be preferred to TM imagery as these can be collected for selected cloud-free days and for similar sized areas.

The areal extent of the snow cover is mapped operationally in many countries using weather satellite data. Although snow cover can be detected and monitored with a variety of remote sensing devices, the greatest application has been found in the visible and the near-infra-red region of the electromagnetic spectrum (EMS). The reason is that "the reflectance of snow in the visible and near-infra-red parts of the EMS is much greater than that of any other natural material on the ground and thus snow can easily be detected and the extent of snow cover determined. The reflectivity (albedo) depends upon snow properties such as the grain size and shape, water content, surface roughness, depth and presence of impurities. In particular, the visible red band (0.6–0.7 m) of the multispectral scanner (MSS) on the Landsat has been used extensively for snow cover mapping because of its strong contrast with snow-free areas. It is to be noted that although Landsat and SPOT may provide adequate spatial resolution for snow mapping, their inadequate frequency of coverage hinders their snow mapping capabilities. As a result, many users have turned to the NOAA polar-orbiting satellites with the Advanced Very High Resolution Radiometer (AVHRR); although characterized with a much higher frequency of coverage (every 12 hours as opposed to every 16 to 18 days), the problem with the NOAA-AVHRR data is that the resolution of 1 km (in the visible red band (0.58–0.68 m)) may be insufficient for snow mapping on small basins.

The current EOS AM and PM satellites carry the Moderate Resolution Imaging Spectroradiometer (MODIS) instrument which provides daily data at fairly high spatial resolutions. The EOS programme is also backed up with a series of rather robust snow algorithms. Despite the spatial and temporal resolution problems associated with visible aircraft and satellite imagery, they have proven to be very useful for monitoring both the build-up of snow cover and the disappearance of snow-covered areas in the spring. Meteor (which has been used to delineate snow/no snow lines for river basins and other areas in the then Union of Soviet Socialist Republics) and NOAA data were combined to map snow cover area in basins ranging from 530 to more than 12 000 km^2 (Shcheglova and Chemov, 1982). Although snow can be detected in the near-infra-red band, the contrast between a snow and a no-snow area is considerably lower than with the visible region of EMS. However, the contrast between clouds and snow is greater in the Landsat TM Band 5 (1.57–1.78 m). Thus the near-infra-red band, when available, serves as a useful discriminator between clouds and snow. Visible/near-infra-red difference data from NOAA-9 imagery of the United Kingdom has been used to locate areas of complete or partial snow cover and identify melt and accumulation zones. Daily snow area maps were produced and were subsequently composited to generate weekly estimates of snow distribution. This technique is currently being considered for operational use in the United Kingdom and elsewhere.

Thermal infra-red data has limited importance for snow mapping and measuring properties because it is hindered by cloud cover, and the surface temperature of snow is not always that much different from the surface temperatures of other adjacent areas with different cover, such as rock or grass. However, thermal infra-red data can be useful to help identify snow/no snow boundaries, and for discriminating between clouds and snow with AVHRR data because the near-infrared band has not been available on this sensor. Furthermore, Kuittinen (WMO, 1992*b*) stated that the best result in snowline mapping can be achieved by combining the information of the thermal emission and the reflectance in the visible part of EMS.

Although there are currently many problems with using microwave sensing for mapping snow cover, one major advantage of the microwave approach is the ability to penetrate cloud cover and map snow extent. Owing to its cloud penetration or all weather capability, the microwave wavelength at about 1 cm has the greatest overall potential for snow mapping. However, the current major drawback is the poor passive microwave resolutions from space (about 25 km) so that only very large areas of snow cover can be detected. Large-scale snow cover extent maps are currently produced using geophysical algorithms on the data from satellite microwave radiometers such as Special Sensor Microwave Imager (DMSP) SSM/I. These maps are most reliable over large flat regions with little or low-lying vegetation when snow is dry. The resolution problem can potentially be solved with the use of high-resolution active microwave sensors. Unfortunately, few, if any, experiments with the short wavelength region of the microwave spectrum (about 1 cm wavelength) that is sensitive to snow have been reported.

3.14 OPERATIONAL SATELLITES

Remote-sensing techniques from space provide the capability of observing precipitation and snow cover in real- or near-real-time over large areas, and thus complement the conventional more accurate point measurements or weather radar. Useful data can be derived from satellites used primarily for meteorological purposes, including polar-orbiting NOAA and DMSP and the geostationary GOES, Geostationary Meteorological Satellite (GMS) and Meteosat (Engman and Gurney, 1991).

Operational polar-orbiting satellites also carry sounders such as the TIROS-N operational vertical sounder (TOVS) and the advanced microwave sounding unit (AMSU), which provide data for numerical weather prediction models used for forecasting rainfall. The NOAA series carrying these instruments has now been replaced by the European Organization for the Exploitation of Meteorological Satellites (EUMETSAT) METOP satellites. While passive microwave radiometers have been operational to date, following the success of the TRMM satellite, there are now well advanced plans to launch a series of satellites carrying visible, IR, passive microwave and active microwave instruments (EPS).

Whereas the ERS-1 and -2 satellites provided semi-operational satellite data, the EUMETSAT ENVISAT satellite has now replaced the ERS satellite and is now in operation providing a range of sensors including SAR. This satellite is complemented by the Japanese Advanced Earth Observation Satellite (ADEOS) system. Meteosat Second Generation (MSG) is also now operational providing high resolution in space and time visible and IR imagery. Multispectral data are also available from LANDSAT, SPOT and, most recently, MODIS.

3.15 DEW

Although the deposition of dew, essentially a nocturnal phenomenon, is not spectacular as a source of moisture, being relatively small in amount and varying locally, it could be of significant interest in arid zones, where it could even be of the same order of magnitude as rainfall. As the process by which moisture is deposited on objects largely depends on the source of moisture, it is necessary to distinguish between dew formed as a result of downfall transport of atmospheric moisture condensed on cooled surfaces, known as dewfall, and that formed by water vapour evaporated from the soil and plants and condensed on cooled surfaces, known as distillation dew. Both sources generally contribute simultaneously to observed dew, although at times they operate separately.

A further source of moisture results from fog or cloud droplets collected by leaves and twigs and reaching the ground by dripping or stem flow. There has been a great tendency to overestimate the average dew over an area, and this is due primarily to overlooking the physical limits on possible quantities of dew. Examination of the energy–budget equation reveals that the latent heat of dewfall and/or distillation dew is unlikely to exceed net radiation and should be less if sensible and soil-heat transfers are taken into consideration. Under favourable conditions there is a definite limit, at the rate of about 1.1 mm h^{-1} for the average rate of dew over an area. However, dew may be substantially increased in local areas where mean temperatures are not horizontally homogeneous and there is small-scale advection from relatively warmer and moister areas to cooler areas. Moreover, the one dimensional form of energy-flux computations should be modified when applied to isolated plants because the pattern of radiation and moisture flux is quite different from that of a homogeneous source. This does not mean that the average deposit over a large horizontal area is affected, but only that some parts gain at the expense of others.

Actual deposition rates will generally fall well below the upper limit.

Much effort has been devoted, but without much success, to devising a means of measuring leaf wetness from artificial surfaces in the hope of yielding results comparable to those for natural conditions. A review of the instrumentation designed for measuring duration of leaf wetness and an assessment of the extent to which various instruments give readings representative of plant surface wetness is given in the Appendix to the *The Influence of Weather Conditions on the Occurrence of Apple Scab* (WMO-No. 140). Any of these devices can only be used as a qualitative guide in any particular situation, or as a crude means of regional comparison. Careful interpretation is required in either role. Unless the collecting surface of these gauges is more or less flush with the surface and of very similar properties, it will not correctly indicate the amount of dew that the natural surface receives.

Theoretically, the flux technique should give reasonable average values over an area, but lack of knowledge of transfer coefficients under very stable conditions makes it extremely difficult to implement. The only certain method of measuring net dewfall by itself is by a sensitive lysimeter. However, this method does not record distillation dew, since no change in weight accompanies distillation dew.

The only generally accepted means of measuring total amount of dew is by the blotting technique, that is, by weighing a number of filter papers both before and after being thoroughly pressed against leaves. A brief outline of dew measurement methods is given in the *Guide to Meteorological Instruments and Methods of Observation* (WMO-No. 8).

3.16 SAMPLING FOR PRECIPITATION QUALITY

In recent years it has become increasingly apparent that deposition of atmospheric pollutants is of major ecological significance. Most notable have been the effects resulting from acidic precipitation in the United Kingdom, Scandinavia, eastern Canada and the north-eastern United States. For a complete picture of the atmospheric transport of toxic substances, both the wet and dry precipitation must be sampled and analysed as well as the air itself. This section discusses the criteria necessary for the collection of liquid and frozen precipitation samples and of surface deposition. For the analysis of atmospheric deposition over periods of tens to hundreds of years, several other substrates have been found useful in providing a record. These include naturally growing mosses, which quantitatively retain some metals, ice cores from glaciers and bottom sediments. Sampling for precipitation quality is further discussed in 7.2.3.

3.16.1 Rain and snow collectors [HOMS C53]

Many types of collectors have been used to sample precipitation, from a plastic, stainless steel or glass container placed on location at the beginning of a precipitation event, to a sophisticated sequential sampler designed to collect precipitation samples automatically at selected intervals during an event.

A common device for the collection of both wet and dry deposition separately is the double bucket collector. One bucket is used to collect precipitation, while the other bucket collects the dry deposition. The collector is equipped with an automatic sensing system that detects precipitation, liquid or frozen. At the onset of a precipitation event, a cover is moved from the wet bucket to the dry bucket. On cessation of the event, the cover automatically returns over the wet bucket. The sample container normally used is a black polyethylene vessel. It consists of two parts. The top part is a removable rim that has been specially fabricated to ensure a sharply defined uniform area of collection. The second part is the bucket itself. Both the rim and the bucket must be rinsed with distilled, de-ionized water each time a sample is removed. When sampling precipitation for organic contaminants, a stainless steel or glass bucket must be used.

When directional information is desired, associated meteorological instruments can be utilized. Equipment has been designed in which precipitation is directed to one of a number of bottles, depending on the direction of the wind, by means of a wind vane.

Modern snow collectors are similar to rain collectors, except that they are heated to thaw and store the entrapped snow as liquid in a compartment beneath the sampler (HOMS C53).

3.16.2 Dry deposition collection

Many of the problems associated with snow collection also apply to the collection of dry deposition. The double-bucket collector provides a measure of the amount, but considerable controversy exists about the relevance of such a measurements. The air turbulence around such a device is not the same as at the surface of a lake, for example, which leads to differences both in absolute collection efficiency and relative efficiency between different particle sizes. Other methods, such as glass plates coated with sticky materials and shallow pans with liquids, aqueous ethylene glycol or mineral oil, have been suggested.

3.17 ASSIMILATION OF RAINFALL INTO HYDROLOGICAL AND HYDRAULIC MODELS: COMBINING AND ASSESSING DATA FROM DIFFERENT SOURCES

Increasingly work is being undertaken to assimilate measurements of radar reflectivity and surface rainfall into numerial weather prediction models. However, there are a number of inherent difficulties, such as the fact that models deal with features in the weather at scales much larger than the observational data scale and that radar reflectivity is not a direct model diagnostic. Work to use the radar data as a proxy for humidity information has met with some success (Meischner, 2003), but the problem is far from solved. It is likely that advanced three- and four-dimensional variational assimilation techniques (3D-VAR, 4D-VAR) will be needed.

The input of rainfall data to hydrological models also presents difficulties. Quality control of the radar or satellite-based inputs is essential. Also, it is necessary to employ advanced statistical techniques to ensure that the error characteristics of the input data are represented in model output flows. The hydrograph produced using such input must never be used without some accompanying measure of uncertainty.

3.18 GLOBAL PRECIPITATION CLIMATOLOGY PROJECT

The Global Precipitation Climatology Project (GPCP) has provided, since 1979, monthly global rainfall estimates over areas of 2.5° latitude x 2.5° longitude grids (Adler and others, 2003). Independent rainfall estimates, usually based upon raingauge observations, provide essential assessment of the accuracy of the GPCP rainfall estimates, although sampling errors between these systems of measurement require statistical evaluation. This was achieved recently by the decomposition of the variance of the satellite and raingauge difference into the error of the satellite sensor and the raingauge sampling error (Gebremichael and others, 2003).

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