

INTERNATIONAL STANDARD

NORME INTERNATIONALE



**Classification of environmental conditions –
Part 2-2: Environmental conditions appearing in nature – Precipitation and wind**

**Classification des conditions d'environnement –
Partie 2-2: Conditions d'environnement présentes dans la nature – Précipitations
et vent**



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INTERNATIONAL
ELECTROTECHNICAL
COMMISSION

COMMISSION
ELECTROTECHNIQUE
INTERNATIONALE

ICS 19.040

ISBN 978-2-8322-9873-2

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Precipitation and wind****FOREWORD**

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IEC 60721-2-2 has been prepared by IEC technical committee 104: Environmental conditions, classification and methods of test. It is an International Standard.

This third edition cancels and replaces the second edition published in 2012. This edition constitutes a technical revision.

This edition includes the following significant technical changes with respect to the previous edition:

- a) the layout of the information provided has been re-organized;
- b) the information provided has been extensively enhanced and revised;
- c) new information on wind severities has been included.

The text of this International Standard is based on the following documents:

Draft	Report on voting
104/1066/FDIS	104/1074/RVD

Full information on the voting for its approval can be found in the report on voting indicated in the above table.

The language used for the development of this International Standard is English.

This document was drafted in accordance with ISO/IEC Directives, Part 2, and developed in accordance with ISO/IEC Directives, Part 1 and ISO/IEC Directives, IEC Supplement, available at www.iec.ch/members_experts/refdocs. The main document types developed by IEC are described in greater detail at www.iec.ch/publications.

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INTRODUCTION

This part of IEC 60721 presents fundamental properties, quantities for characterization, and a classification of environmental conditions dependent on precipitation and wind relevant to electrotechnical products. The information presented is intended to be used as background material when selecting appropriate severities of parameters related to precipitation and wind for product applications.

Precipitation encompasses all forms of hydrometeors, both liquid and solid, which are free in the atmosphere, and which reach the Earth's surface. At altitudes below the freezing level, precipitation can occur as liquid or solid particles but above this level snow or hail will predominate. For this document, the different forms of hydrometeors are addressed separately and under the more commonly referred to meteorological conditions of rain, snow and hail. Also encompassed are icing conditions but only that occurring at ground level.

This document additionally and separately addresses wind.

The majority of the information presented in this document has been assembled by the UK Met Office from published sources as well as historical and forecasting weather records. The information has been assembled and maintained for the UK Ministry of Defence for equipment design and testing purposes [1]¹. The historical meteorological data employed for this work meets World Meteorological Organization criteria for validity. However, such data are only available from a limited number of worldwide locations (typically a few hundred). Forecasting weather records, which were extensively utilized for this work, are available from a significant number of locations (typically tens of thousands) but are not necessarily verified. Whenever the latter information has been used, an appropriate strategy was adopted to remove spurious data.

¹ Numbers in square brackets refer to the Bibliography.

CLASSIFICATION OF ENVIRONMENTAL CONDITIONS –

Part 2-2: Environmental conditions appearing in nature – Precipitation and wind

1 Scope

This part of IEC 60721 presents fundamental properties, quantities for characterization, and a classification of environmental conditions dependent on precipitation and wind relevant to electrotechnical products.

The information presented within this document is intended to be used as background material when selecting appropriate severities of parameters related to precipitation and wind for product applications.

For the purpose of this document, precipitation is considered to encompass all forms of hydrometeors, both liquid and solid, which are free in the atmosphere, and which reach the Earth's surface. The different forms of hydrometeors are addressed separately and under the more commonly referred to meteorological conditions of rain, snow and hail. Whilst icing conditions are additionally considered, only that occurring at ground level, is addressed.

This document separately addresses the climatic condition of wind and provides methodologies and quantitative information to enable wind severities and frequencies to be estimated worldwide.

2 Normative references

There are no normative references in this document.

3 Terms and definitions

No terms and definitions are listed in this document.

ISO and IEC maintain terminology databases for use in standardization at the following addresses:

- IEC Electropedia: available at <https://www.electropedia.org/>
- ISO Online browsing platform: available at <https://www.iso.org/obp>

4 Rain

4.1 General

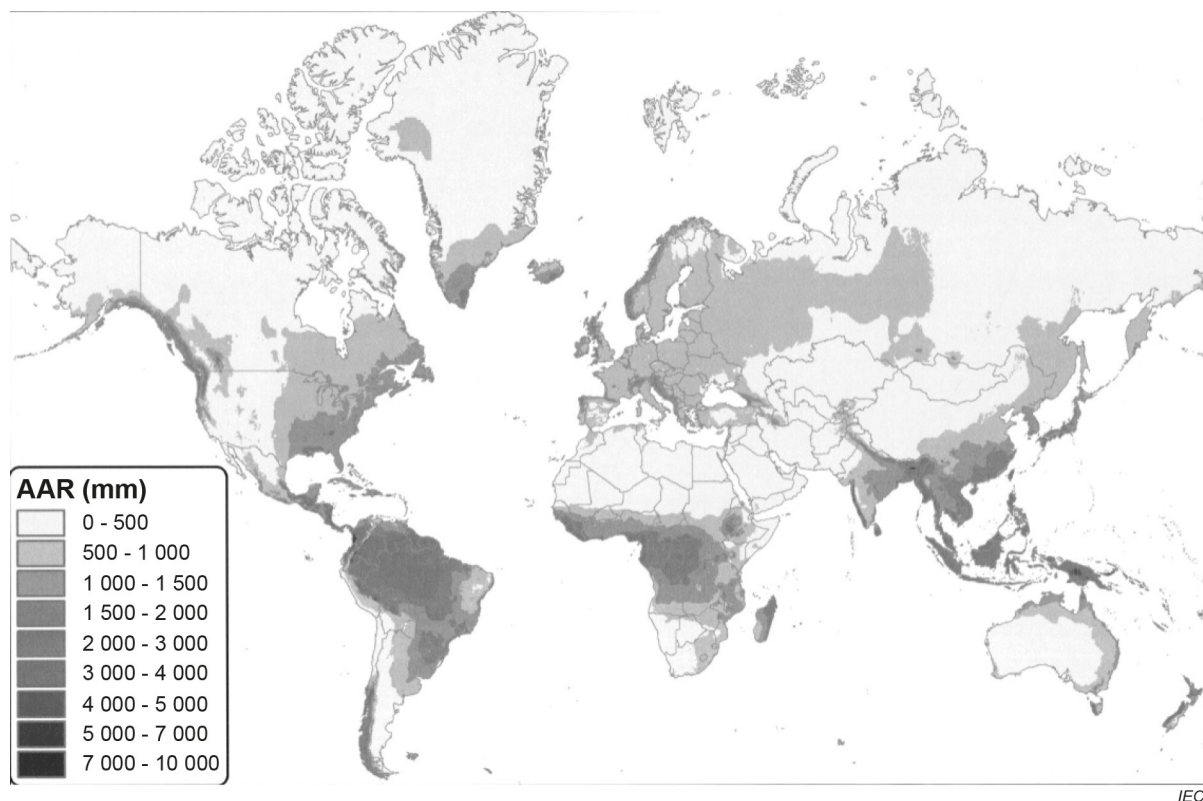
Rain is the primary focus of Clause 4 as it is the dominant meteorological condition associated with the wetting of electrotechnical products.

4.2 Global distribution of rainfall

Compared with meteorological air temperature which, at any particular time is often substantially the same ($\pm 5\text{ }^{\circ}\text{C}$) over relatively large regions, rainfall is a much more spatially variable climatic condition. In particular, the precipitation intensity that constitutes a near extreme value is peculiar to the highly localized area. Even a relatively short distance away, the intensity can differ by a factor of two or more. Thus, it is impracticable to relate precipitation intensity to specific geographical areas of the world, as is the case with temperature.

Precipitation intensity is defined as the rate at which precipitation falls. Although values of precipitation intensity may be considered as instantaneous rates, in practice they are averages taken over periods of one minute or longer. When using any rainfall data, it should be remembered that as the pattern of rainfall is infinitely variable both in time and space, only general information can be given by means of maps or diagrams.

For most places, readily available rainfall data are limited to observations of the rainfall catch made once daily. Where precipitation is of snow, the observations record the rainfall equivalent of that snow. The daily observations may be summarized to provide average monthly, seasonal or annual amounts. Figure 1 represents such a summary for annual amounts, based on observations from 27 075 locations worldwide over the period from 1961 to 1990.



NOTE At higher latitudes, an increasing proportion of this "rainfall" will fall as snow.

Figure 1 – Average annual rainfall (AAR) for global land areas, based on 1961 to 1990 data [1]

The fundamental requirements for precipitation to fall in significant amounts are high atmospheric moisture content and a mechanism for the uplift of air. Ascending air cools by expansion due to the decreasing atmospheric pressure with height. Given that the lower the temperature the less moisture the air can retain in vapour state, then if cooling and moisture content are sufficient, precipitation is the result. As a generalization, the wetter land regions of the world belong to one of the following three broad geographical categories:

- a) Along the equator $\pm 15^\circ$ of latitude, e.g. Indonesia, equatorial Africa, and the Amazon rain forest. The high rainfall of these regions is primarily due to convection, triggered by solar heating and accentuated by the convergence of the northern hemisphere tropic's north-easterly winds and the southern hemisphere tropic's south-easterly winds along the "inter-tropical convergence zone". Here, copious moisture is provided by either rain forest or warm, tropical ocean.
- b) The western side of continents in mid latitudes, e.g. UK, Western Europe, the north-western coastal fringe of North America and the south-western coastal fringe of South America (southern Chile). At these latitudes, winds blow predominantly from the west and therefore reach the western side of continents having picked up moisture over a long ocean track. The lifting mechanisms are varied and include convection and orographic uplift, but the dominant lifting mechanism is cyclonic or frontal uplift in weather disturbances that develop on, and move eastwards along, the boundary between polar and tropical air masses. In North and South America, the inland penetration of high rainfall is severely limited by the high, north-south aligned mountain chains of the American Cordillera. By contrast, the western fringe of Europe has no north-south aligned barrier of such proportion, so that moderate rainfall is able to penetrate well inland across the European plain. Iceland, the Falkland Islands, Tasmania, and the exposed west of South Island New Zealand also belong to this regime.
- c) Extending poleward from the equatorial regions along the eastern seaboard of continents, e.g. Eastern Asia from India to Kamchatka, North America from the Gulf of Mexico to Quebec (including the southern extremity of Greenland), South America from southern Brazil to north-eastern Argentina, the east of South Africa and the eastern fringe of Australia. The reasons for high rainfall in these areas are complex but include the predominantly easterly moist onshore winds of tropical latitudes and, at sub-tropical latitudes, the drawing of moist summer monsoon winds of tropical ocean origin inland towards a heat-generated continental low-pressure area, as in the Indian and south-east Asian summer monsoons. Further poleward, at mid latitudes, the prevailing wind is from the west, blowing offshore; however, the eastern seabords of both North America and Asia are favoured regions for the development of precipitation-bearing cyclonic weather systems which then move north-eastwards close to the mid-latitude coastline.

The important influence of topography on rainfall is demonstrated by the heavier rainfall in mountainous regions, particularly where a mountain range runs parallel to the coast and intercepts moisture laden winds as they blow onshore. Mountains also usually reduce rainfall downwind – the "rain-shadow" effect.

Many of the great deserts of the world lie within or close to latitudes 20° to 30° of latitude, where relatively high atmospheric pressure dominates e.g. the Sahara Desert and Saudi Arabia, the deserts of California and Arizona, the Atacama Desert in Chile, the Namibian and Kalahari Deserts of southern Africa and much of interior and western Australia. The dryness of some deserts is accentuated by the rain-shadow effect of adjacent mountain barriers (e.g. the inland deserts of California and Arizona). An additional factor in some coastal deserts is a cold ocean current offshore that suppresses convection, for example the narrow Atacama Desert of Chile is trapped between high mountains to the east and a cool ocean current offshore.

In Asia the circum-global belt of high pressure at sub-tropical latitudes is displaced by the Asian monsoon, which blows outwards from intense high pressure over Siberia in winter and blows into low pressure over southern interior Asia in summer. This effectively transfers the latitudinal desert belt by approximately 15° of latitude poleward to lie north and inland of the areas reached by the Indo-Asian monsoon e.g. the Gobi Desert of Mongolia and China.

Precipitation in polar regions is generally not particularly high on account of the reduced amount of water vapour in the air at low temperatures.

4.3 Characteristics of rain

4.3.1 Formation

Clouds are formed when air is cooled below its dewpoint, usually as a result of lifting and consequent expansion. At first the cloud droplets grow by the condensation of water on to them, but it can be shown that this process alone cannot produce drops of the size found in rain. Two mechanisms are thought to be important in the formation of raindrops.

Firstly, droplets which are slightly larger than the average will fall, relative to the air, and towards neighbouring smaller droplets, and so can collide and coalesce with some of them to become larger still. This process can continue until a droplet eventually falls out of the base of the cloud. This mechanism is confined mainly, but not exclusively, to the tropics, where clouds can remain devoid of solid precipitation throughout their depth. Theoretical studies have shown that a significant amount of rain can be produced in this way, provided the cloud is several kilometres deep.

Secondly, when a cloud top becomes appreciably colder than 0 °C, it contains a mixture of ice crystals and supercooled water drops. At first the crystals grow by direct sublimation of water vapour on to them, but as they become larger, they can collide with the supercooled droplets and other ice crystals to form snowflakes, and when these snowflakes have fallen below the level at which the temperature is 0 °C they will melt to form raindrops. This is the dominant mechanism in middle and high latitudes, but it also occurs within the tropics and applies to clouds with a top colder than about –10 °C. In convective (cumuliform) cloud, graupel or small hail, rather than snowflakes, can be produced.

4.3.2 Types of rain

Rainfall is often classified according to the process causing the uplift of air initiating the rain formation; there are three main types of rain which are not mutually exclusive, and these are known as orographic, cyclonic, and convective.

Orographic rain is caused by one, or sometimes both, of two primary mechanisms. The most commonly known mechanism is the forced ascent of a moist airstream over the physical barrier of the high ground. The ascending airstream cools by expansion, often to the temperature at which saturation occurs, above which altitude cloud forms. This can result in drizzle or rain over the high ground when there is none on the adjacent low ground, but more often it enhances cyclonic cloud and rain that are also affecting adjacent low ground. This enhancement is often due primarily to raindrops scavenging additional water as they fall through the perpetually reforming layer of low cloud formed by the ascent. This "seeder/feeder" mechanism can massively augment precipitation on, particularly, windward slopes of mountains exposed to moist airstreams, and this type of rainfall often continues for many hours or even days. The effect often extends for some distance upwind, and for a lesser distance downwind (spill over), of the foot of the mountain barrier. The other type of orographic rainfall does not require a flowing airstream and is caused by the convective (i.e. non-forced) ascent of moist air from sun-heated mountain slopes and summits. This type of orographic rainfall is a daytime, and particularly summer, phenomenon and is typically more intense but of shorter duration, an example being the afternoon thunderstorms to which the Alps are prone in summer. Such convection occurs more readily over mountains than over adjacent lowlands because convective air bubbles, formed over sun-heated mountain surfaces, more readily attain a temperature higher than that of the surrounding atmosphere at that altitude. Sometimes orographic convection is triggered, not by sun-heated surfaces, but by the initial forced orographic ascent of an unstable airstream, such triggering not necessarily confined to daytime.

In the case of cyclonic rainfall, large scale forced uplift is associated with features of the general weather situation, such as fronts and depressions. As with rainfall due to orographic forced ascent, cyclonic rainfall is often relatively steady but of long duration. However, as in the case of orographic forced ascent, cyclonic forced ascent can also trigger more intense pulses of convective rainfall given an unstable atmosphere.

Convective rain falls from cumulonimbus clouds forming within an unstable air mass, this being one in which temperature falls relatively rapidly with altitude. The formation of a cumulonimbus cloud requires a trigger to initiate the convective ascent of a large bubble of moist air which then, despite cooling by expansion as it rises, cools less quickly than the temperature of the surrounding atmosphere, so remaining buoyant and continuing to rise through a considerable depth of atmosphere. This forms clouds of relatively small lateral extent but great vertical depth. The trigger is often the daytime heating of the ground surface by the sun or the passage of a cold air mass over a relatively warm sea surface. However, convection can also be initiated by cyclonic uplift, particularly near some cold weather fronts, where the approaching cold air mass sometimes overrides the warmer air in advance of the front, causing extreme instability. Another trigger can be the initially forced ascent of an unstable air mass over high ground. Convective rainfall is typically more intense, but of shorter duration, than cyclonic or orographic rain, individual areas of rain being relatively small (20 km diameter or less), with dry areas close by.

4.3.3 Distribution of raindrop sizes

It is known that drops of a diameter greater than about 6 mm break up before reaching their terminal velocity, and this fixes the upper limit for raindrop diameter. Table 1 presents estimates of the number and size of raindrops for various rates of rainfall. It should be noted that the values quoted represent averages based on a large number of observations of size distributions, and that individual measurements of drop sizes can show wide variations about these average values. Additionally, the table was derived using rain from predominantly stratiform clouds at mid-latitudes and therefore it is possible that it does not represent the predominantly convective rainfall of tropical regions as accurately.

Table 1 – Estimated number of raindrops per cubic metre for various rates of rainfall [1]

Estimated number of raindrops per cubic metre for various rates of rainfall					
Rate of rainfall mm/h or l/m ² h	Drop diameter				
	1 mm to 2 mm	2 mm to 3 mm	3 mm to 4 mm	4 mm to 5 mm	> 5 mm
1	32	0,53	0,01	0,00	0,00
5	140	7,5	0,41	0,20	0,00
10	230	19	1,5	0,12	0,01
25	415	51	6,4	0,75	0,11
50	615	101	17	2,8	0,55
100	850	179	38	7,9	2,1
200	1 140	259	113	20	6,9
500	1 590	525	173	57	28
1 000	1 970	754	289	111	69

4.3.4 Terminal velocity

The rate at which a raindrop falls through still air depends both on the size of the drop and on the resistance offered by the air. The speed of fall will increase until the air resistance is equal to the weight of the drop, after which it will fall at a steady rate known as the terminal velocity.

Table 2 shows the variation of terminal velocity with drop size, for the following still air conditions:

air pressure 101,3 kPa;
 temperature 20 °C;
 relative humidity 50 %.

Table 2 – Terminal velocity of raindrops in still air [1]

Terminal velocity of raindrops in still air	
Drop diameter	Terminal velocity
mm	m/s
0,1	0,27
0,5	2,06
1,0	4,03
1,6	5,65
2,0	6,49
2,6	7,57
3,0	8,06
3,6	8,60
4,0	8,83
4,6	9,03
5,0	9,09
5,8	9,17

4.4 Rates of rainfall

4.4.1 Instantaneous rates and clock-hour totals

Data on the frequency and duration of rainfall rates are relatively scarce, although the use of recording rain gauges with suitable time scales provides some information on the occurrence of specific rates of rainfall, and radar data contributes in some areas. Indirect estimates have been made from daily rainfall summaries, using regression equations based on a limited number of stations; these methods can be misleading when marked orographic or seasonal effects occur, especially at high rainfall intensities.

For many stations, especially in North America and Europe, routine tabulations of clock-hour rainfall amounts have been made but are only available from respective national meteorological organizations. Empirical formulae have been developed to make use of these clock-hour totals to derive frequencies of occurrence of specified rainfall rates.

The clock-hour totals do not give probabilities for the near instantaneous rates of fall which are important in many design problems since the rate of fall can vary widely inside any one hour. Studies have been made of the distribution of near instantaneous rates about the clock-hour total, for different rain intensities. Figure 2 may be used to convert clock-hour values to probabilities of instantaneous rainfall rates, though this curve should be interpreted with caution.

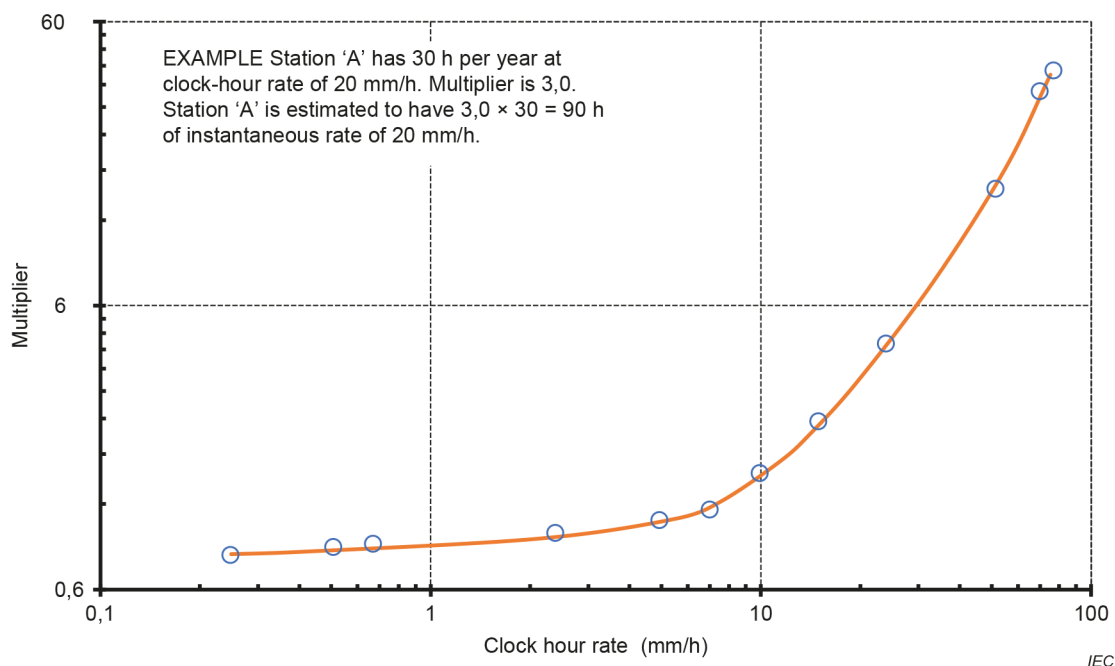


Figure 2 – Estimated conversion factor for converting to hours at instantaneous rate [1]

4.4.2 Frequency of instantaneous rates

Table 3 presents some estimates of the frequency of specified instantaneous rates of rainfall in an average year, for a number of localities. These estimates have been based, in the main, on clock-hour data obtained over a number of years. To convert to percentage probability, the number of hours should be divided by 87,66.

Table 3 – Duration in the average year of instantaneous point rainfall equalling or exceeding specified rates [1]

Duration (h) in the average year of instantaneous point rainfall equalling or exceeding specified rainfall rates						
Place	Rainfall rate					
	1 mm/h	5 mm/h	10 mm/h	25 mm/h	50 mm/h	100 mm/h
Akrotiri	95,0	27,0	10,0	2,7	0,7	0,15
Birmingham Airport	187,0	35,0	9,5	0,7	0,09	0,01
Changi	208,0	79,0	44,0	16,0	6,0	1,3
Freetown	380,0	137,0	81,0	29,0	11,0	2,8
Gibraltar	91,0	29,0	13,0	3,2	0,9	0,16
Hong Kong	517,0	138,0	49,0	14,0	2,9	0,6
London Heathrow	234,0	41,0	11,0	1,1	0,17	0,01
Malta	111,0	33,0	13,0	3,8	1,0	0,20
Manchester Airport	308,0	52,0	13,0	1,1	0,16	0,02
Miami	250,0	70,0	32,0	15,0	5,0	1,5
New Orleans	260	80,0	49,0	26,0	10,0	2,5
Tiree	205,0	34,0	9,0	0,5	0,04	0,0
Washington	226,0	40,0	15,0	4,0	1,4	0,5
NOTE Values of 0,0 indicate less than measurable quantity.						

4.5 Heavy rates of rainfall

4.5.1 Thunderstorm rain

The heaviest falls of rain in short periods of time are usually associated with thunderstorms; since these are experienced at some time in most parts of the world, peak intensities of rainfall can be almost as high in temperate climates as in the wetter regions of low latitudes. The relative incidence of thunderstorms over the world is, in general, most frequent in tropical and sub-tropical areas, they are common during the summer half of the year in temperate latitudes and are infrequent only in the Arctic and Antarctic.

Thunderstorms occur most frequently inland and are much less common over the open sea. Exceptions to this are some of the equatorial ocean areas and the Mediterranean, which, having relatively high temperatures in autumn and winter can set up active convection when cold air masses move south from Europe.

4.5.2 Prolonged heavy rain

Prolonged periods of heavy rainfall occur mostly in monsoon areas or in association with tropical cyclones, although heavy orographic rainfall can occur in most mountainous regions.

4.5.3 Worldwide extremes

Recorded rainfalls over periods ranging from one minute to one year, taken at many places, have been used to determine an envelope which provides an approximation to the worldwide extremes of rainfall at any one point. This envelope may be expressed in the form:

$$R = 363D^{-0,5} \quad (1)$$

where

R is the rate of rainfall expressed in millimetres;

D is the duration expressed in hours.

Table 4 shows some worldwide extreme hourly rainfall rates, covering different durations ranging from one minute to several hours. These are predicted values based on the formula for R .

Table 4 – Predicted worldwide extremes of rainfall [1]

Predicted worldwide extremes of rainfall	
Duration min	Rate of rainfall mm/h or litre/m ² /h
1	2 900
5	1 200
10	900
30	550
50	400
60	380
100	280
200	200
400	140

For comparison, Table 5 lists the world's highest recorded short period rainfall amounts (WMO 1994), with corresponding average rainfall rates (mm/h). The observed values in Table 5 are within the envelope provided by Equation (1), with the exception of the observed record 60 min rainfall of 401 mm, which is slightly more extreme than predicted by the formula.

Table 5 – Observed world maximum rates of rainfall [1]

Observed world maximum rates of rainfall		
Period min	Maximum observed rainfall mm	Corresponding rainfall rate mm/h
1	38	2 280
8	126	945
15	198	792
20	206	618
42	305	436
60	401	401

4.6 Spatial variations of rainfall rate

4.6.1 General

Large gradients exist in the rainfall rates observed over short periods between points not very far apart, but the longer the period over which the rainfall rate is determined, the less is the variation with distance. For durations of 1 min and a whole storm, Figure 3 shows the rate of decay of correlation between the rainfalls at two points as the distance between them increases. The figure is intended to illustrate only the general principle as, in reality, the shape and slope of the "total storm rainfall rate" curve will depend on the duration, area extent and intensity of the storm.

There is clearly a substantial improvement in correlation when accumulated rainfall is considered over a period. The rapid decay of correlation, when short period rates of rainfall are being considered, shows the need for high density observational networks if small scale details of the rainfall pattern are of importance. A spacing of about half a kilometre is needed in order to account, on the average, for 75 % of the variance which occurs in the short period rates of rainfall during heavy storms.

The decay of correlation coefficient with distance is much more rapid for convective rainfall than for cyclonic or frontal rainfall, an implication being that the relationship between correlation coefficient and distance also varies with region and season, according to the prevalent rainfall type. Similar plots to Figure 3, but for more precisely defined circumstances, are contained in: [2], [3], [4], [5], [6] and [7].

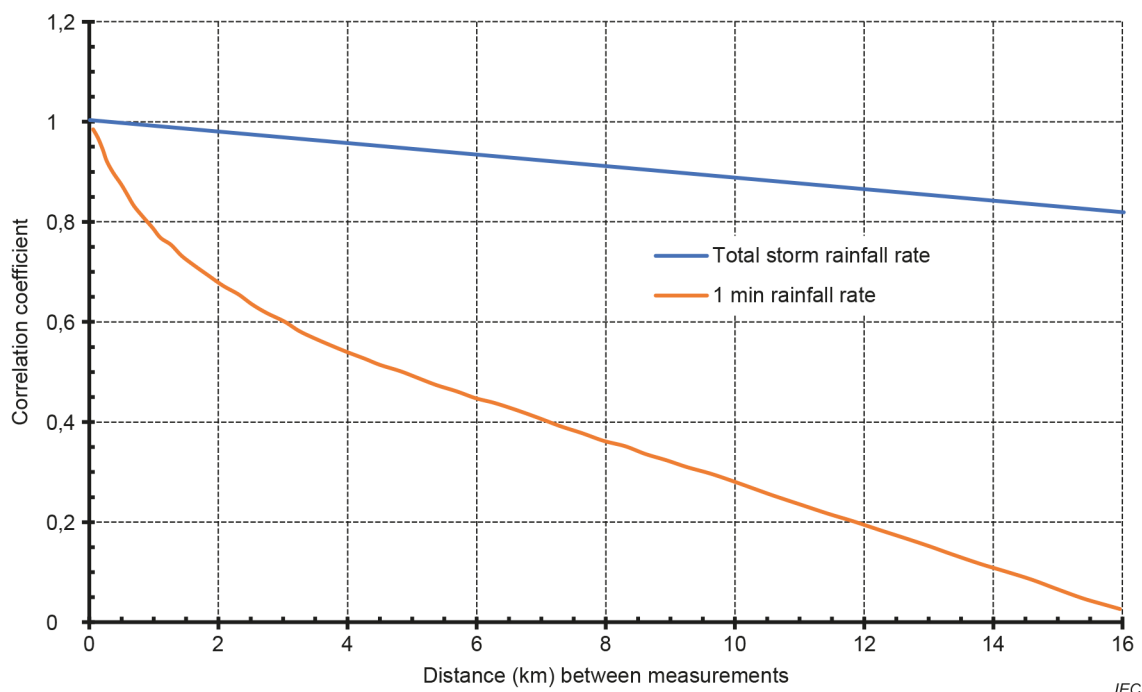


Figure 3 – Average decay of correlation with distance, of 1 min rainfall rate and total storm rainfall rate [1]

4.6.2 Rainfall rates below 2 mm/h

Rain areas associated with these low rates of rainfall are often extensive, areas of 1 000 km to 1 500 km by up to 3 000 km being quite common, in association with large scale features of the pressure pattern. In extreme cases the length of a rain belt can extend to as much as 6 000 km. In suitable conditions, orographic rainfall can extend along the whole windward side of mountain chains, such as the Atlantic coasts of Great Britain and Norway, the Pacific coast of North America, and especially the mountain chains of Asia during the summer monsoons. These rain belts seldom exceed 250 km in width and are usually much narrower.

4.6.3 Rainfall rates between 2 mm/h and 10 mm/h

In temperate climates, rainfalls rates between 2 mm/h and 10 mm/h are unlikely to exceed 250 km in length. Lengths of 150 km to 250 km can occur on fairly numerous occasions. In tropical latitudes, rain belts with this rainfall rate can extend up to 300 km or more in hurricanes. Monsoon rains in West India and West Africa frequently cover lengths of 1 000 km, and 2 000 km has been exceeded at times during the summer monsoon on the southern and western slopes of the Himalayas.

4.6.4 Rainfall rates between 10 mm/h and 25 mm/h

Rainfalls between 10 mm/h and 25 mm/h are nearly always associated with convection. Individual convective clouds are rarely more than 15 km across but on occasions, for example along a cold front, the clouds can become closely linked and form a rain belt of some 150 km to 250 km in length although the width is unlikely to exceed about 15 km. Similar linkages can also occur along a mountain barrier and during the most intense phase of the tropical summer monsoon, rainfalls of these intensities can occur virtually anywhere along the southern and western slopes of the Himalayas and associated ranges for 2 000 km or more. Monsoon rains of these intensities can also fall in West Africa and West India, over areas some 1 000 km across. Similar heavy rains over an area of perhaps 300 km across can also be experienced with tropical hurricanes.

4.6.5 Rainfall rates exceeding 25 mm/h

Rainfall of these high rates is thought to be limited everywhere to one or two neighbouring convection cells. The horizontal extent of these cells is generally in the range of 1 km to 5 km and is unlikely to exceed 30 km even in tropical storms. In the cores of these cells, over diameters of 2 km or less, rainfall rates will occasionally exceed 100 mm/h even in temperate climates. Extreme rates recorded are of the order of 1 000 mm/h occurring over a period of about 1 min and thought to be associated with cells only about 0,5 km in diameter.

4.7 Rainfall probabilities along a line

The information given in this subclause relates mostly to rainfall probabilities at a specific point, but for many purposes it is necessary to assess the chances of specified rainfall rates occurring simultaneously along given distances of horizontal tracks. Direct information of this kind is rarely obtainable, but assessments can be based on point probabilities and on assumed shower models. Table 6 gives some values obtained in this way for tracks in Southern England and in Singapore.

Table 6 – Estimated duration (h) in the average year when stated distances along given tracks simultaneously have rainfall at or exceeding specific rates [1]

Estimated duration (h) in the average year when distances along given tracks simultaneously have rainfall at or exceeding specific rates							
Location	Track length km	Distance along track km	Rainfall rate				
			5 mm/h	10 mm/h	25 mm/h	50 mm/h	100 mm/h
Southern England	25	0	172	81	12,5	2,9	0,2
		5	107	20	0	0	0
		10	40	0	0	0	0
		15	0	0	0	0	0
	50	0	303	151	25	5,6	0,4
		5	221	60	6	0	0
		10	120	5	0	0	0
		15	50	0	0	0	0
		20	5	0	0	0	0
	100	0	562	292	48	11	0,9
		5	440	140	25	2	0
		10	290	50	0	0	0
		15	170	0	0	0	0
		20	60	0	0	0	0
Singapore	25	0	330	280	162	82	29
		5	220	130	0	0	0
		10	100	0	0	0	0
		15	0	0	0	0	0
	50	0	570	510	310	159	56
		5	430	250	100	20	0
		10	280	100	0	0	0
		15	180	0	0	0	0
		20	100	0	0	0	0
	100	0	1 050	980	600	311	111
		5	880	550	200	80	0
		10	650	280	10	0	0
		15	480	100	0	0	0
		20	300	0	0	0	0

5 Snow

5.1 General

Snow is precipitation in the form of ice crystals and these can exist singly but more often they coagulate to form snowflakes. The crystals can be formed initially by the spontaneous freezing of supercooled liquid cloud droplets at a temperature below $-40\text{ }^{\circ}\text{C}$ but are more often formed initially by the deposition of water vapour onto suitable freezing nuclei at higher threshold temperatures. Once ice crystals are present in a cloud, they grow by further deposition of water vapour at the expense of liquid cloud droplets, which evaporate to replenish the supply of water vapour. This is known as the Bergeron-Findeisen process, which occurs because air becomes saturated with respect to ice at a higher temperature than it becomes saturated with respect to liquid water. When the ice particles have grown to become heavy enough, they start to fall and often join together to form snowflakes. In certain circumstances, usually in shower clouds, the ice crystals can grow predominantly by collision with supercooled liquid cloud droplets, coating them in opaque rime to form denser "snow pellets" or "graupel" of diameter 2 mm to 5 mm.

After falling below the altitude at which the temperature reaches $0\text{ }^{\circ}\text{C}$, snowflakes begin to melt. Therefore, the occurrence of snow at ground level is usually associated with temperatures of $2\text{ }^{\circ}\text{C}$ or lower, with persistent snow cover occurring at temperatures of $0\text{ }^{\circ}\text{C}$ or below.

Freshly fallen snow has a density ranging from 70 kg/m^3 to 150 kg/m^3 whereas the density of old snow can be in a range of 400 kg/m^3 to 500 kg/m^3 , even up to 910 kg/m^3 . If the density exceeds 910 kg/m^3 , snow is usually considered as ice. Firm snow will normally have a density of 600 kg/m^3 . Wind exposure will often increase the density by breaking down the snowflakes, temperature will also increase the density as will settling over time.

5.2 Global distribution of the incidence of snowy weather

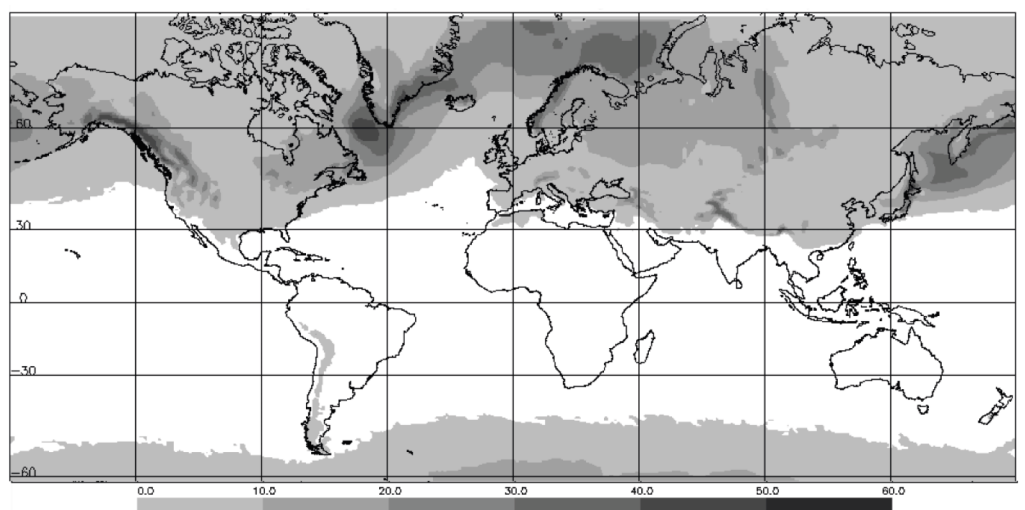
A global climatology of snow is difficult to produce because large variations in snow climatology occur on distance scales that are much shorter than the spatial separation of surface observing stations. This is the case even though, on a broad scale, the incidence of snow is strongly weighted towards higher latitudes where temperatures are close to or below $0\text{ }^{\circ}\text{C}$ for longer periods. Not only does temperature fall by an average of $0,6\text{ }^{\circ}\text{C}$ for each 100 m of ascent, but also precipitation amount increases with altitude, except at very high altitudes. As a consequence, the incidence of snow is particularly sensitive to variations in altitude. Other factors, such as urban heat islands and distance from sea, also cause local variation in snow incidence. The sensitivity to all these factors is enhanced in temperate climates where a small change in one of them can make the difference between precipitation falling as snow or rain. Regression relationships between snow parameters and geographical parameters have been used to interpolate between observation stations, but these may only be valid for a specific region, making it difficult to combine climatologies from different countries.

Recently, numerical modelling and satellite remote sensing techniques have begun to provide greater insight into the global climatology of snow, enabling additional information to be presented. In describing the climatology of snow, it is important to distinguish between falling snow and lying snow. This is because falling snow does not necessarily lie and because lying snow, depending on the severity of the winter, can be intermittent, melting within a few days, or can persist and accumulate over successive snowfall events, in some regions until spring. The regions where it snows the most are in many respects different from where accumulated snow cover is deepest or most persistent, snowy mountainous areas being an exception.

An indication of where it snows the most is given by Figure 4. These figures show estimates, derived from numerical atmospheric model output, of the percentage of days on which an amount of snow equivalent to 1 mm of rainfall falls. Very approximately 1 mm of rainfall is the equivalent of approximately 10 mm of lying snow, but it is possible that this snow will not necessarily lie. The spatial distributions for spring and autumn lie between those shown for winter and summer. When interpreting these maps, it should be noted that they do not highlight regions where it snows frequently but in negligible amounts, as is the case in some of the coldest continental winter climates.

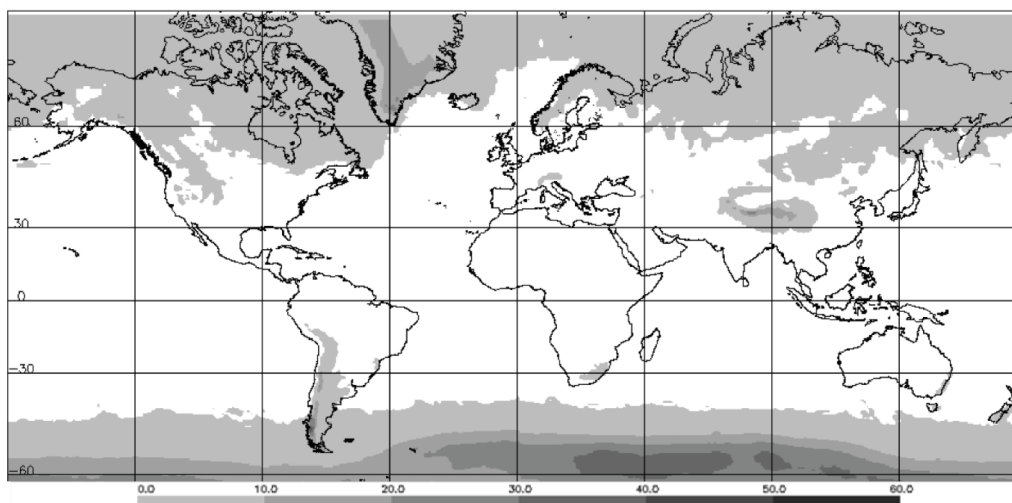
Some of the snowiest regions of the world are at high latitudes over, or adjacent to, open sea which provides a ready source of moisture but where, of course, the snow does not lie. Temperatures here, though often low enough for snow, are much higher in winter than over high latitude continental interiors. At these high latitude continental interiors, temperatures are often so low that the moisture content of the air is greatly reduced. In the southern hemisphere the most prone areas lie almost concentrically around the globe at 60 °South to 70 °South, just poleward of the main winter storm track along the boundary between polar and tropical air masses. In the northern hemisphere the distribution is contorted by the irregular distribution of land masses and warm ocean currents, with the most prone areas lying in two arcs. A "North Atlantic" arc extends from Labrador north eastwards between Greenland and Iceland towards Spitzbergen, curving eastwards to the Barents Sea and the northern shoreline of Europe. A "Pacific" arc extends north eastwards along the eastern seaboard of Asia from northern Japan across the Khamchatka Peninsula and then eastwards across the Bering Sea. Although winter storm systems will be responsible for much of this snow, there is also a convective component due to very cold air from Siberia, Canada, the Greenland Ice Cap, or the Arctic ice sheet flowing out over a relatively warm sea.

The snowiest land areas of the world, in terms of snow falling, are where the afore-mentioned ocean storm tracks meet coastal mountains on the western side of continents at high latitudes. These regions include the coastal mountains of Norway, Alaska and British Columbia and, to a lesser extent, the Andes of South America. The lack of a continuous high mountain barrier along the western shore of Europe allows weather systems and a relatively snowy zone to extend eastwards across the plains of northern Europe as far as the snow prone Ural Mountains of the Russian Federation.



IEC

**Percentage of days with snow falling >1 mm (snow water equivalent) –
Northern hemisphere (December, January, and February)**



IEC

**Percentage of days with snow falling >1 mm (snow water equivalent) –
Southern hemisphere (June, July, and August)**

SOURCE: Reference [8]. The maps are derived from ERA-interim model data (January 1979 to October 2013), courtesy of the European Centre for Medium Range Weather Forecasts (2011).

Figure 4 – Estimated percentage of days on which an amount of snow equivalent to at least 1 mm of rainfall falls in northern and southern hemispheres [1]

Equatorward of latitude 60° (approximately), winter conditions are often borderline between rain and snow. Winter conditions are more likely to err on the side of rain over the relatively warm winter sea and adjacent coastal lowlands than over inland regions of similar latitude. However, when snow does fall in coastal regions, it is often of greater intensity than inland. This is due to moisture laden maritime air moving over cold land, or Arctic air reaching land via a relatively warm sea trajectory, making it convectively very unstable. This effect can be increased if a range of hills lies parallel to the coast, causing additional uplift. Owing to the low atmospheric moisture content at very low temperatures, the interiors of large continents rarely experience high rates of snowfall. However, snow which does fall in these areas can remain on the ground for long periods or all winter.

Figure 4 also illustrates that there are large areas of the world where snow never falls. For example, snow is rare at mean sea level between latitudes 30° North and 40° South, although occasional snowfalls are experienced in South China in most winters. At these tropical and sub-tropical latitudes, snow does occur on ground above about 2 500 m.

5.3 Global distribution of the incidence and depth of lying snow

Historically the global distribution of snow has sometimes been estimated using indirect information such as height of ground above sea level and the probability of temperatures occurring below 0 °C. However, satellite remote sensing techniques have enabled a more specific indication of the global extent and amount of lying snow to be mapped in terms of the "snow water equivalent" (SWE) [9]. This is shown in Figure 5 for December to February, March to May, June to August and September to November. Nonetheless, as the techniques for estimating snow water equivalent from satellite data are still under refinement, and as the resolution of the charts is insufficient to show the large variation that can occur on a local scale. The values indicated in the figure should be taken as no more than a general indication. It should be stressed that the maps do not show literally the actual depth of snow but rather the rainfall equivalent of lying snow and ice or compressed snow.

As a convenient but very approximate guide, 1 mm of rain or "snow water equivalent" is equal to approximately 10 mm of fresh lying snow but only 1 mm of ice. The rainfall equivalent of fresh lying snow is actually very variable depending on the nature of the snow. For instance, for wet snow the snowfall to rainfall ratio is typically 8:1. For dry snow the ratio is more typically 15:1.

One of the most obvious features of the Figure 5 maps is the almost complete absence of significant snow cover in South America, Africa and Australasia in any season. A second notable feature is the strong seasonal cycle in the Northern Hemisphere, where large areas of the Eurasian and North American land masses are snow covered in winter and early spring. This difference is due to the higher latitudes reached by North America and Eurasia, and their vastness, allowing polar air masses to accumulate and spread in winter without the warming influence of the sea.

In areas subject to a period of permanent snow cover through the winter, the season of snow lying lags behind the season for falling snow and has a "saw-tooth" pattern. The lying snow typically accumulates in a series of increments throughout the winter before melting relatively rapidly at some stage during spring. Hence, average snow cover is greater in spring than in autumn.

Generally, the greatest amounts of lying snow and ice (in terms of snow water equivalent) are where temperatures are sub-zero for longest, i.e. continental interiors at polar latitudes. Superimposed on this pattern are some perturbations due to altitude. This is primarily a factor at temperate latitudes where altitude can make a critical difference between whether or not precipitation falls as snow and whether or not that snow lies.

The maps in Figure 5 do not show peripheral areas where small amounts of snow lay intermittently. Therefore, it should be borne in mind that snow can lay intermittently in winter, usually to depths of no more than a few centimetres, for a margin of approximately 5° of latitude south of the limiting boundaries in Figure 5.

5.4 Mass, size, and fall-speed of snow

Ice crystals have a large variety of shapes, sizes and densities. The ice crystal lattice is hexagonal under most atmospheric conditions, with a principal axis perpendicular to the axis of hexagonal symmetry. According to the temperature and humidity conditions within the cloud, hexagonal symmetry can present in a wide variety of forms, from flat hexagonal plates and flat dendrites (branched crystals) to "vertically extended" hexagonal prisms, columns and needles ("vertically extended" meaning extended along the axis of hexagonal symmetry). As a generalization, the flat shapes form in cloud temperatures between 0 °C and –3 °C and again between –10 °C and –21 °C, whereas the vertically extended shapes form mainly between –3 °C and –10 °C and again below –21 °C.

The term "snowflake" often refers to a cluster of up to many thousand ice crystals, joined by the process of aggregation, but can alternately refer to a single crystal. Aggregate snowflakes are often no more than 10 mm in diameter but, exceptionally, can reach 80 mm in diameter or more (in the longest direction). An insight into the physics of the various types of ice crystal grown in a laboratory is provided by Libbrecht [10]. However, real snowfalls seldom contain only perfect ice crystals and, more typically, comprise mainly broken crystals or aggregates of the same.

Table 7 gives the relationship between the mass and the maximum linear dimensions of the various crystal forms. These relationships represent mean values and any measurements made on individual crystals would show some variation from this mean.

Table 7 – Snow crystals: relation between mass (mg) and diameter (mm) of the sphere which just contains the crystal [1]

Crystal type	Relationship between mass (m) and diameter (d)
Snow pellets or graupel	$m = 0,065 d^3$
Rimmed plates and stellar dendrites	$m = 0,027 d^2$
Powder snow and spatial dendrites	$m = 0,010 d^2$
Plane dendrites	$m = 0,0038 d^2$
Needles	$m = 0,029 d$ (d = length)

The terminal velocity of a snowflake is determined by the balance between gravitational and drag forces. The gravitational force is constant for a given mass, but the drag force increases as the velocity of the snowflake increases until an equilibrium (terminal) velocity is attained. The terminal velocity is dependent on the shape, size and density of the flake. These parameters, in turn, are a function of the temperature and humidity environment in which the snowflake evolves, together with the cloud droplet concentration and the time duration over which the snowflake remains in the cloud. These factors regulate deposition, crystal type, aggregation and riming.

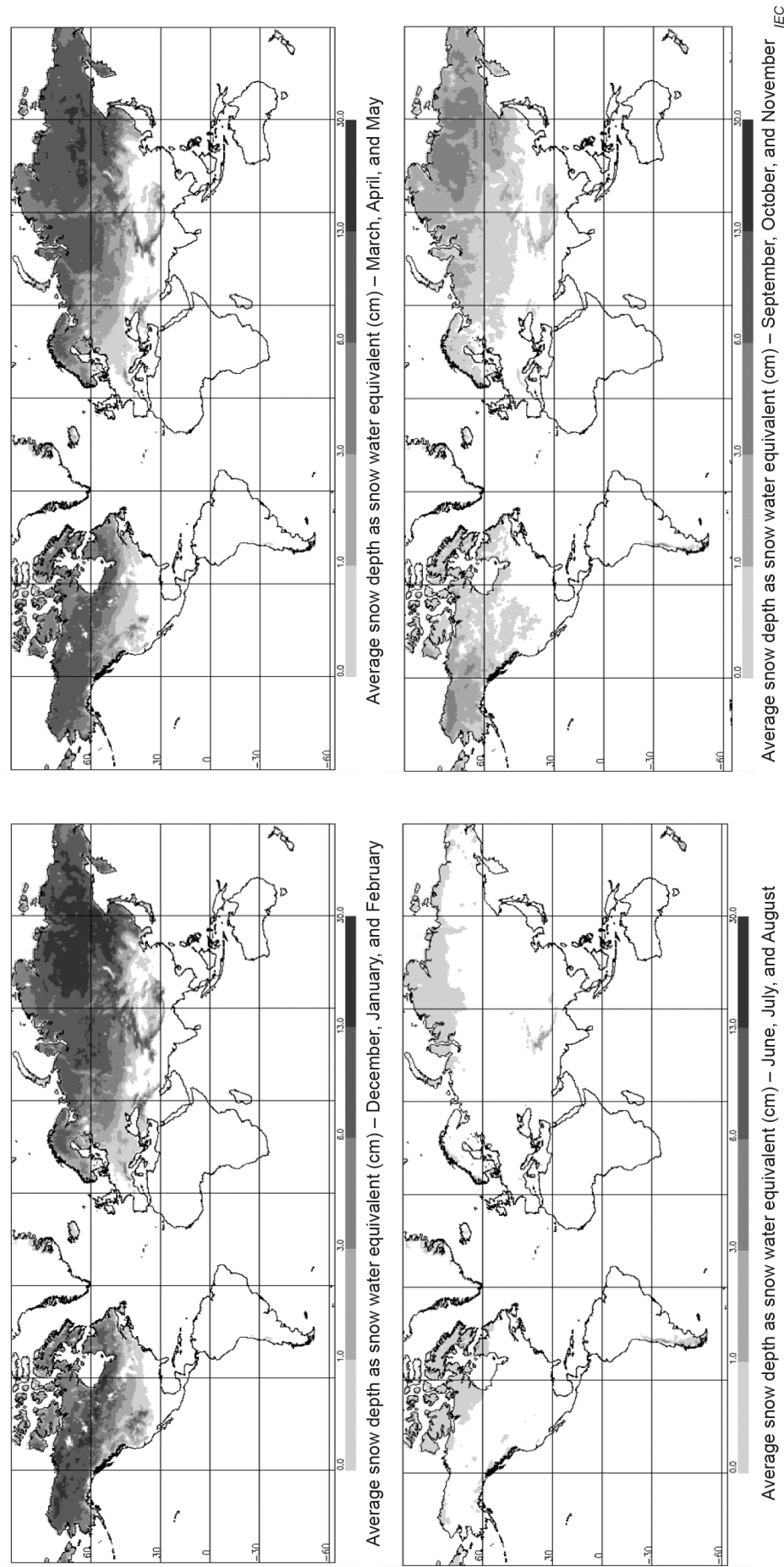
The relationship between terminal velocity (V) and the diameter (D) of the snowflake is usually represented by the relationship shown in the following equation.

$$V = \alpha \times D \times \beta \quad (2)$$

Numerous observational and theoretical studies have derived values of α and β for different types of crystal and degrees of riming, though often only for a particular range of temperature or crystal type. For example, the observational findings of Barthazy and Schefold (2006) [11], derived at approximately 1 600 m altitude in the Swiss Alps, are presented in Table 8, but were confined to surface (not necessarily cloud) and temperatures between $-1\text{ }^{\circ}\text{C}$ and $-5\text{ }^{\circ}\text{C}$. That study also provides a useful rule of thumb that beyond a certain size, the terminal velocity of snowflakes remains approximately constant at between 1 m/s and 2 m/s. In this instance, the intention is merely to give a pointer to the body of work that now exists in the literature. The first such study was by Langleben (1954) [12]. In that study, as in some subsequent ones, D refers to the "melted diameter" of the snowflake. However, in the Barthazy and Schefold study, D is the width of the enclosing box. A useful set of references is included in Szyrmer and Zawadzki (2010) [13] and, for any in-depth literature review, this paper and its sister paper [14] are currently recommended as a starting point.

Table 8 – Values of α and β for different crystal types [11]

Values of α and β for different crystal types in the snowflake terminal velocity formula of Equation (2)						
Crystal type	α			β		
	Degree of riming			Degree of riming		
	None	Moderate	Dense	None	Moderate	Dense
Needles	0,97	1,26	1,61	0,20	0,24	0,35
Plates	1,02	1,21	1,36	0,23	0,26	0,40
Dendrites	-	0,98	1,06	-	0,27	0,24
Irregular crystals	1,23	1,20	1,27	0,22	0,21	0,22
Snow pellets/graupel	-	-	1,54	-	-	0,61
NOTE Snow pellets/graupel are densely rimed by definition, but the reason for the absence of values for unrimed dendrites is simply that none were observed in this study.						



NOTE The analyses exclude the Greenland Ice Cap, sea ice sheets and land areas south of 60°S.

SOURCE: Reference [9]. The maps are based on the period August 1987 to May 2007 and the data are courtesy of the National Snow and Ice Data Centre, USA (Armstrong et al. (2005)).

Figure 5 – Satellite derived average rainfall equivalent of lying snow and ice [1]

6 Hail

6.1 General

Hail occurs in several forms but usually consists of spherical balls of ice of diameter of 5 mm to, occasionally, 50 mm or more and made up of alternate layers of clear and opaque ice. Hailstones can also be conical or formed into irregularly shaped lumps by agglomeration and can be formed entirely of clear ice or contain trapped liquid water. The density of hail can vary from 10 kg/m³ to 90 kg/m³ according to the structure, although ordinary hail is often assumed to have a density of 70 kg/m³ to 90 kg/m³, the latter value corresponding to clear ice. Snow pellets, formerly called soft hail, consist of a central crystal covered with frozen cloud droplets and are easily compressible, with a density of about 30 kg/m³ and a diameter of 2 mm to 5 mm.

Hailstones fall from a specific type of cloud, cumulonimbus, which has great vertical depth and a particularly high liquid water concentration, typically 1 g to 3 g of liquid water per cubic metre of air. This is approximately 10 times greater than in most clouds and approximately 100 times greater than in cirrus clouds. It is this particularly high-water content that enables ice particles within the cloud to grow, primarily by collision with supercooled water droplets. This distinguishes the growth of hail from the growth of snow crystals, in which the dominant process of ice crystal growth is the deposition of water vapour.

However, not all cumulonimbus clouds produce hailstones. The convection has to be sufficiently strong and sustained in order for ice particles to be prevented from falling from the cloud before they have grown into hailstones. The cloud has to be sufficiently deep to contain the range of temperatures required for both the initial deposition of ice onto hygroscopic nuclei (< –20 °C) and then the rapid growth of hailstones (0 °C to –5 °C). It is in the latter temperature range that the cloud contains the highest concentration of supercooled water droplets on which the hailstones feed and grow. The cloud, as a whole, should not be too warm for hailstones to form, as can be the case in the tropics. Likewise, the atmosphere outside of the cloud should not be so warm or dry that any hailstones melt or evaporate before reaching the ground.

The conditions for hail formation are also broadly those required for the development of thunderstorms, and the presence of thunderstorms can indicate the possibility of hail. However, there is no simple relationship between thunderstorms and hail, with hail sometimes forming in convective clouds which do not develop into thunderstorms.

6.2 Global distribution of hailstones

Hail is mostly a localized and short duration phenomenon with any individual hailstorm often affecting only a few square kilometres for time periods of only a few minutes. Occasionally, multiple hail cells can form along a weather front or squall line up to several hundred kilometres in length. However, it is likely that many hailstorms are missed by the surface observing network, particularly where there is a large separation between stations. Even where there is a good observing network, the number of occurrences of significantly large hail in the data series is often too small to determine a statistically reliable estimate of average frequency. Hail climatology derived from surface observations often relies on additional observations from outside the official meteorological network. The quality of these observations is very variable, and the availability of such observations is strongly dependent on population density. As a result of this, across many parts of the world, hailstorm frequencies derived from surface observations are, to some extent, misleadingly a function of population density. Recent attempts at global mapping have used either numerically modelled data or satellite techniques to overcome the inadequacy of surface observations.

The map of Figure 6 shows, very approximately, the worldwide land distribution of large hailstorm frequency. The map is based on empirical post processing of archived UK meteorological data from a global numerical model. This approach errs on the side of overestimation and excludes oceans and seas but has the advantage over surface observations of being consistent and objective globally. Hail is most likely in the sub-tropics and middle latitudes as well as on high ground nearer the equator. The almost total absence of even large hail on low ground in the tropics may be accounted for, primarily, by the greater height of the 0 °C isotherm and in tropical deserts such as the Sahara, the infrequency of any form of precipitation. Care should be taken in interpreting Figure 6 in quantitative terms, as it includes only hail of diameter ≥ 15 mm. If all sizes of hail were to be included, the annual number of hail days would be considerably greater. Continental areas in the sub-tropics and middle latitudes are the most prone to large hail but small hail appears to be relatively frequent over the oceans.

Although the methodology for Figure 6 is objective, it is based on an empirical algorithm that will possibly not apply with consistent accuracy across all types of hail generation, such as the small hail that is characteristic of oceanic regions. Attention is also drawn to one other clear inaccuracy in Figure 6 and that is the underestimation of the frequency of large hail in the tornado belt of central USA (around Oklahoma). In this region, very large hailstones are observed on at least six days per year on average. The most likely explanation for this discrepancy is that the algorithm used is not suited to the unique atmospheric temperature and humidity profiles responsible for these exceptionally severe storms.

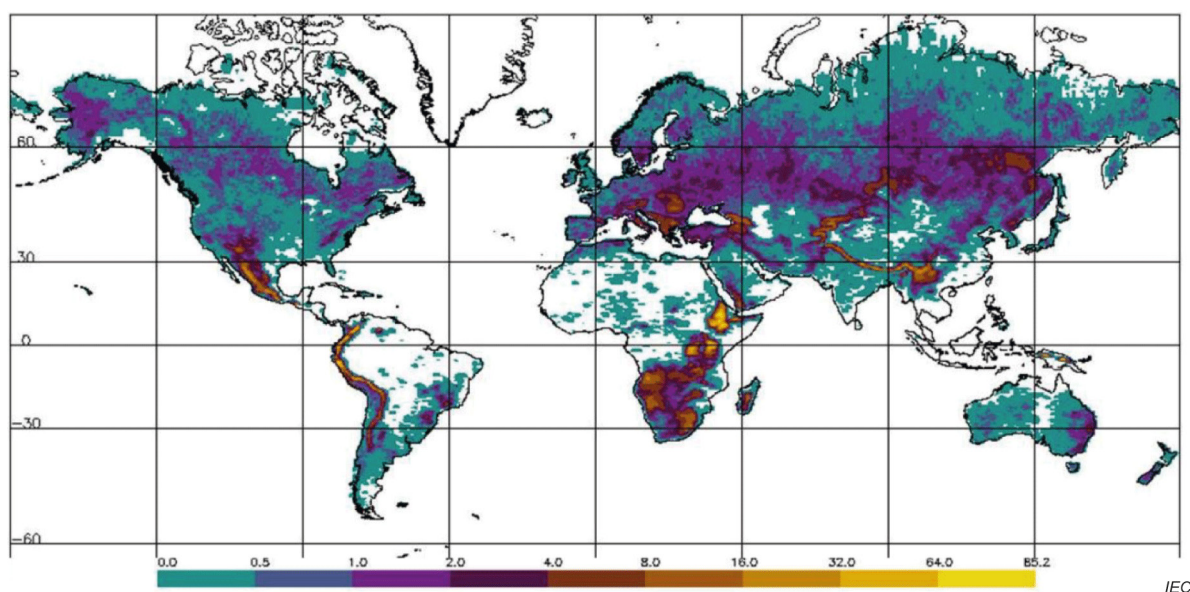


Figure 6 – Estimated average annual number of days per year with hail of diameter ≥ 15 mm [1]

6.3 Seasonal variations in global distribution

The global distribution of hail has a strong seasonal cycle, being heavily weighted towards spring and, particularly, summer in each hemisphere, as illustrated by Figure 7. These are the seasons when, over land, convection is at its most pronounced. In this convection, note that the atmosphere over land is generally more unstable in spring than it is in autumn. This is because in spring, the atmosphere is generally being heated from below as it warms up after the winter, whereas in autumn the atmosphere is generally being cooled from below as it cools down after the summer. During December, January and February, most of the events occur over the Andes, New South Wales and the southern hemisphere portions of Africa. However, during this time, some events can also occur in the Himalayas and adjacent mountains, some mountainous areas in Mexico and the far southwest of the USA. In June, July and August there is very little activity in the southern hemisphere, but hail is widespread in the northern hemisphere from 30° North to 70° North.

Over the oceans, and on coasts when the prevailing wind is onshore, seasonal variation is less marked than over land. This is because the annual temperature variation of the ocean's surface is much less than over the continents and is also time lagged, with the highest ocean temperatures typically in late summer or early autumn. In some enclosed seas, such as the Mediterranean, the flow of cold air from land at higher latitudes across the relatively warm water produces a maximum hail occurrence in autumn and winter.

6.4 Diurnal variation

The large hail of continental areas is most likely to fall during the late afternoon and evening, and it is probable that some 85 % of damaging hail occurs between 13:00 and 22:00 hours local time. In contrast the small hail of coastal regions and open seas exhibits little diurnal variation.

6.5 Size of hail

The size of hail reported as reaching the ground is usually that of the largest stones in any given fall. No accurate observations are available of the distribution of hailstone size in a particular fall, but estimates have been made. Table 9 indicates the number of stones of a given diameter likely to be found per unit volume, based on the assumptions that the stone density is 90 kg/m^3 , that the mass of airborne ice is equally divided amongst the size ranges (an assumption based on some observations of snow pellet distribution in convective clouds) and that the equivalent rainfall rates are reasonable.

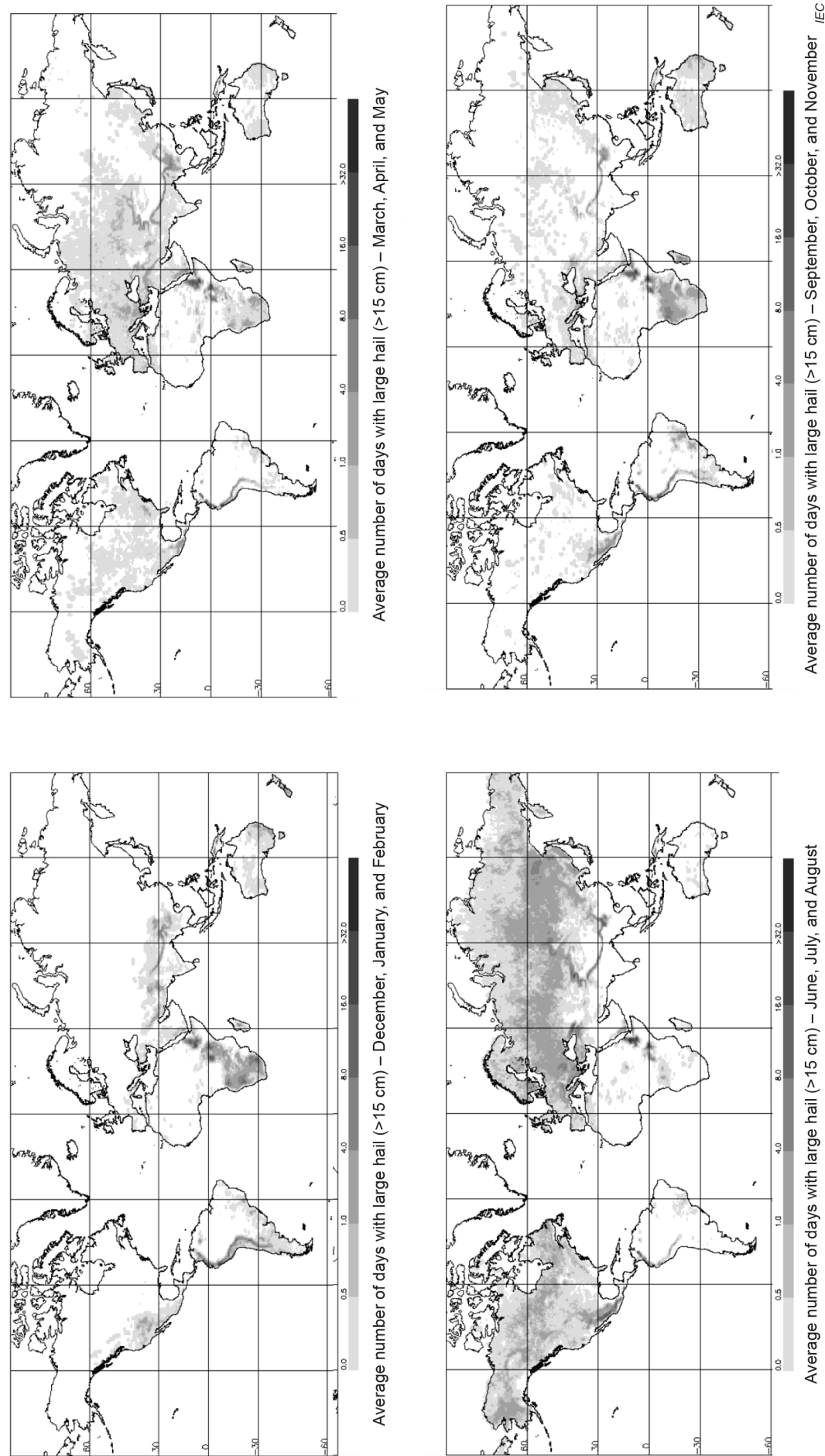


Figure 7 – Estimated average seasonal number of days per year with hail of diameter ≥ 15 mm, based on post-processed global model data (oceans excluded) [1]

Table 9 – Number of hailstones per cubic metre (at about 4 000 m) for specified maximum hailstone size and specific ranges of stone size [1]

Range of stone diameter mm	Size of largest stone in hailstorm		
	30 mm	60 mm	100 mm
	Numbers per cubic metre		
0 to 10	22	9	7
10 to 20	0,8	0,35	0,27
20 to 30	0,17	0,07	0,06
30 to 40	-	0,03	0,02
40 to 50	-	0,01	0,01
50 to 60	-	0,000 7	0,005
60 to 70	-	-	0,000 3
70 to 80	-	-	0,002
80 to 90	-	-	0,001 5
90 to 100	-	-	0,001
Equivalent rainfall (mm/h)	250	300	500

On theoretical grounds the maximum size of hailstones is about 130 mm and stones up to this diameter have been reported. The Coffeyville, Kansas hailstone of 3rd September 1970 was irregular in shape with a diameter of about 138 mm. Since warm, moist, and unstable air is required for the formation of hail, the size observed varies with locality. In Arctic regions, true hail at ground level is rare and graupel is more likely, with stones less than 10 mm in diameter. Large hail is also rare over the open seas, the maximum diameter being about 15 mm.

The largest and most damaging hail is usually confined to the interiors of continents and most likely in middle latitudes and the sub-tropics. The probability of large hail does not continue to increase towards the equator, and in equatorial regions hail occurs only infrequently near the coasts and at low levels. However, larger land areas and greater elevation both favour an increase in large hail and stones up to 25 mm in diameter have been known to fall at the higher tropical stations.

6.6 Terminal velocity

The terminal velocity of large hail may be determined from the equation below. However, it should be noted that the terminal velocity is relative to the air movement.

$$V = \frac{(2\sigma g d)^{0,5}}{(3\rho C_D)^{0,5}} \quad (3)$$

where

d is the stone diameter;

σ is the stone density;

ρ is the air density;

g is the acceleration due to gravity;

C_D is the drag coefficient;

V is the terminal velocity.

Empirical values of C_D , for normal sea level values for g and σ , a value of $\rho = 0,9 \text{ g/cm}^3$ leads to the terminal velocities presented in Figure 8, which is based on the following equation.

$$V = kd^{0,5} \quad (4)$$

where

V is the terminal velocity in m/s;

d is the hailstone diameter in cm;

k is a constant = 14,7.

A better estimate of the constant k has been given as 16,3 and using this value, the variation of kinetic energy (E_k) of the hailstone at terminal velocity with its diameter is expressed in joules and given by the equation:

$$E_k = 0,0626d^4 \quad (5)$$

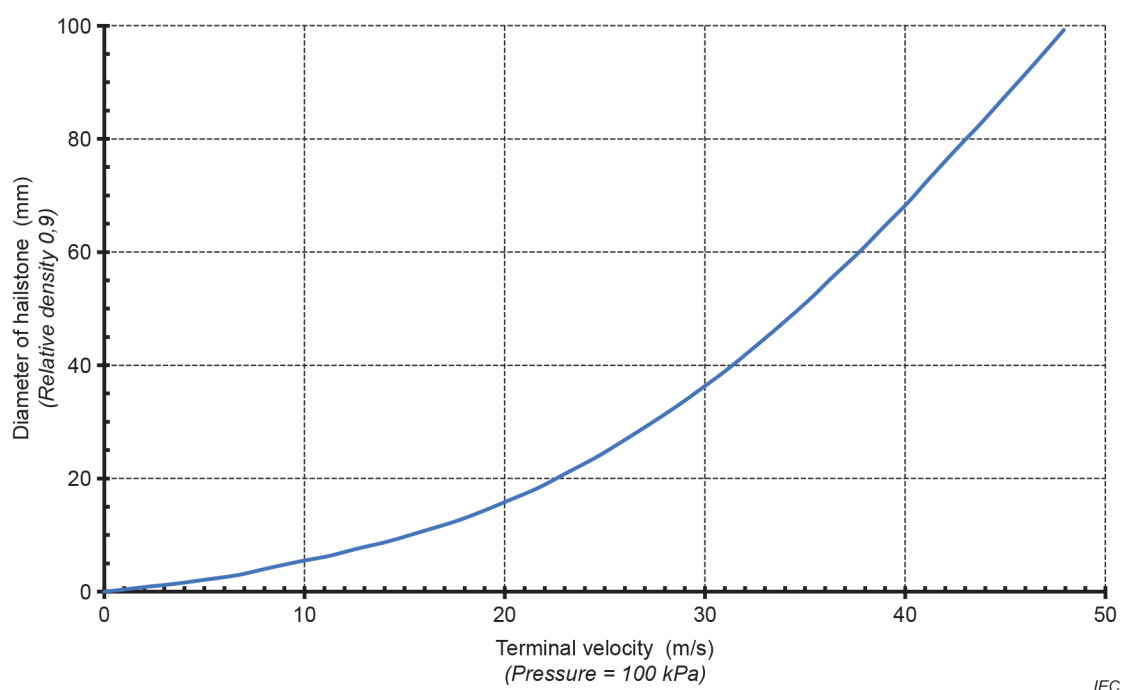


Figure 8 – Relationship between diameter and terminal velocity of spherical hailstones [1]

6.7 Duration and diameter of hail cell

The most frequent duration of a hailstorm at any point on the ground is about 2 min, with extremes in the range 10 s to 45 min. Long durations of hail are probably due to more than one storm cell but, excluding such occurrences, the typical duration is 5 min to 6 min.

At ground level the diameter of a hail cell is most frequently about 2 km to 3 km. Evidence from hailstorm damaging indicates a range from a few metres to 120 km, with 50 % within 2 km to 5 km. This suggests that the typical hail core of a mature storm is roughly circular in section, with a diameter of about 3 km, with occasional merging of neighbouring cells forming more extensive areas of hail.

7 Ice

7.1 General

Clause 7 deals mainly with ice formation on ground objects. Ice formation is also important to aircraft at altitude, however, this aspect is only discussed briefly herein.

Ice forms on objects when fog, cloud or rain is present at temperatures below 0 °C, or when the object itself is below this temperature and below the frost point of the air surrounding it. In addition, ice in the form of wet snow adheres to structures and to itself when snow falls at air temperatures just above freezing point. This phenomenon is termed "wet snow accretion" and can lead to catastrophic build-up of snow on structures such as overhead power lines. In addition to wet snow accretion there are four main types of icing, the other three types being hoar frost, rime ice and glazed frost, although intermediate and mixed types do occur.

7.2 Hoar frost

Hoar frost is a white, feathery deposit of ice crystals and has a density of about 10 kg/m³ or less. It forms on objects which are at temperatures below 0 °C by sublimation of water vapour, where the water changes directly from vapour to solid without passing through the intermediate liquid stage. For this process to occur, the temperature of the object should also be below the frost point. Any water vapour, in excess of that required to achieve saturation with respect to ice, will be deposited to produce the hoar frost, but the weight of such deposits is usually small.

Hoar frost requires cooling of the air and of the object on which it is to form, and so is usually associated with conditions favouring radiation cooling such as clear still nights.

7.3 Rime ice (including in-cloud icing and freezing fog)

Ranging in density from 30 kg/m³ to 90 kg/m³, rime is harder and denser than hoar frost though less so than glazed frost. Rime ice occurs when supercooled fog or cloud droplets freeze on contact with objects at temperatures below 0 °C forming opaque white ice, which grows out windward on exposed objects.

The amount of rime deposited increases with wind speed, the duration of time in cloud or fog, the amount of free water in the air, the droplet size distribution, and the aerodynamic flow past the object (which determines the efficiency of catch). Because of the positive dependency of rime accumulation on wind speed and the duration of exposure to foggy conditions at sub-zero temperatures, prolonged and heavy icing is most likely over the summits and windward slopes of hill and mountain ranges, particularly in cooler maritime climates with winds off the sea. On low ground, fog is more likely to be accompanied by very light winds, and in stronger winds temperatures on low ground are usually higher than on high ground and so less often sub-zero. Localities where very large deposits of rime ice can build up include the hills and mountains of western Scotland, western Norway, western Japan, and Greenland.

The catch efficiency is greatest for objects with a small radius of curvature and therefore, ice collects most readily on such items as the supporting guys of a mast. In addition to the weight of ice, increased wind speed increases the strain on guys and cables. However, as the ice accumulates, the effective shape of the object changes and so alters the catch efficiency, which then becomes more difficult to define.

There are two types of rime, "hard rime" and "soft rime". Soft rime consists of white, feathery needles or spikes and is similar in appearance to hoar frost, though formed by a different process (the freezing of water droplets rather than water vapour). It is more likely to form than hard rime at extreme low temperatures or if winds are light and the liquid water content of the fog or cloud is low.

Hard rime is harder, denser and often less milky in colour than soft rime. It has a comb-like appearance similar to that of the icing of ships and shoreline structures by supercooled sea spray. Hard rime is more likely to form than soft rime when the temperature, wind speed and liquid water content of the fog or cloud are higher, albeit if the temperature is too high (within only a few degrees Celsius below freezing) glaze frost can form in preference to rime, especially in strong winds. There is also a dependence on size of object, small objects being more prone than large objects to hard rime rather than soft rime. The dependence of rime type on these factors is illustrated by Figure 9 and Table 10.

Table 10 – Meteorological parameters controlling atmospheric ice accretion [15]

Type of ice	Air temperature °C	Wind speed m/s	Droplet size	Water content in air	Typical storm duration
In-cloud icing					
Glazed	See Figure 9	See Figure 9	Medium	High	Hours
Hard rime	See Figure 9	See Figure 9	Medium	Medium	Days
Soft rime	See Figure 9	See Figure 9	Small	Low	Days
Precipitation icing					
Glaze (freezing rain or drizzle)	$-10 < t_a < 0$	Any	Large	Medium	Hours
Wet snow	$0 < t_a < +3$	Any	Flakes	Very high	Hours
NOTE Value t_a refers to the air temperature (dry-bulb °C).					

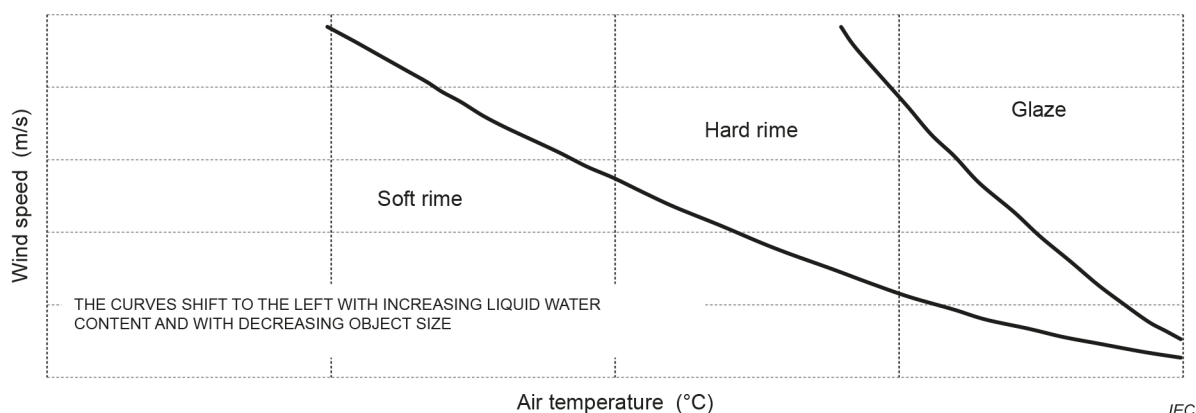


Figure 9 – Air temperature and wind speed criteria for the formation of different types of ice [1] [15]

7.4 Glazed frost (including freezing rain and freezing drizzle)

Both glazed frost and rime occur when supercooled water droplets strike an object at a temperature below freezing, however, glazed frost formation is confined to air temperatures within a few degrees below freezing and is favoured by stronger winds and larger droplet size (see Figure 9). Glazed frost, also known as black ice or glazed ice, is harder and denser than either rime or hoar frost and consists of a layer of clear ice. Its most well-known manifestation is when rain or drizzle falls on a freezing surface, typically but not necessarily, at the end of a prolonged period of severe frost, for example with the approach of a warm front. The smaller sized water droplets that make up cloud or fog more often produce rime icing but can produce glaze if the temperature and wind criteria indicated for glaze in Figure 9 are met. The ice deposit continues to grow until warmer air brings the temperature of the surface up to 0 °C. This type of ice is tough and hard to remove, and can reach a considerable thickness, although the conditions which favour its formation are normally transient and occur only infrequently.

Generally, snow is more likely than liquid precipitation at temperatures below freezing and liquid precipitation is confined to temperatures above $-10\text{ }^{\circ}\text{C}$ and unlikely at temperatures below $-6\text{ }^{\circ}\text{C}$.

A prerequisite for freezing rain and freezing drizzle is a sub-zero surface layer of air to supercool the falling water droplets, with warmer air aloft at temperatures higher than $0\text{ }^{\circ}\text{C}$, typically above a warm frontal surface. Freezing precipitation is rarer than rime and often restricted in area but affects low ground as much as high ground. In the northern hemisphere winter, it has a relatively high incidence (typically two days per year to four days per year, locally more) in eastern Canada and north-eastern United States where, occasionally, severe ice storms occur along the boundary between polar and tropical air masses. It also has similarly high incidence in parts of central and eastern Europe where cold but shallow continental air masses are sometimes overridden by mild, moist air masses from the Mediterranean or Atlantic. It should not be overlooked that in many Arctic regions, freezing rain is more likely to be a spring, summer or autumn phenomenon, as in winter the vertical profile of the atmosphere is usually too cold for precipitation to be of freezing rain rather than snow. Some locations towards the Arctic coastlines of Canada and the Russian Federation appear to be particularly prone to freezing rain in the northern hemisphere spring, summer and autumn.

7.5 Wet snow accretion

Wet snow merits specific mention under the heading of icing because in some parts of the world wet snow accretion is the dominant form of disruption due to icing.

Meteorological observations generally do not distinguish between dry snow and wet snow so that in addition to snow being observed, a further criterion is essential to estimate whether the snow is of the wet variety. Wet snow accretion is confined mostly to the narrow range of air temperatures $0\text{ }^{\circ}\text{C}$ to $+3\text{ }^{\circ}\text{C}$, within which precipitation can alternatively be of rain. Such snow has often begun to melt on its descent, resulting in high liquid water content and adhesive properties. At temperatures below freezing, dry snow is more likely and is less likely to adhere to structures. The rate of accretion of wet snow onto structures increases with precipitation rate and the density of accreted wet snow increases with wind speed.

The liquid water content of wet snow that has clung to structures will refreeze if temperatures subsequently fall below freezing point, as can well occur overnight after a snowfall.

The most prone regions for wet snow, occurring on over 45 days per year on average, include Iceland, the Falkland Islands, northern Scandinavia, and north-western Russia. More extreme cold climates are unlikely to experience this type of icing in their winter, when temperatures are too low, but can do so in their autumn, spring or summer. Hence parts of Alaska, northern Canada, Greenland, and the far north of Russia experience 10 days to 15 days with wet snow but confined to the summer half of the year.

7.6 Water content of fog and cloud at temperatures below $0\text{ }^{\circ}\text{C}$

Strictly, fog differs from cloud only in that fog is at the earth's surface whereas cloud is above the earth's surface; this leads to the ambiguity that a weather station below cloud level reports as cloud what a higher altitude weather station within this cloud reports as (hill) fog. The median diameter of supercooled cloud droplets is about $20\text{ }\mu\text{m}$, similar to that in freezing fog. However, not all types of fog and cloud are interchangeable in this way and there are important differences in the liquid water content of different types of fog and cloud.

As a generalization, the higher liquid water contents occur when the causative cooling and subsequent condensation of water vapour are due to the ascent of air. Except for hill fog, the various types of fog do not involve significant uplift of air, so that the liquid water content of most types of fog is typically lower than that of hill fog and many types of cloud, seldom exceeding $0,3\text{ g/m}^3$ and more typically approximately $0,05\text{ g/m}^3$.