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PRECIPITATION AND MEASUREMENTS OF PRECIPITATION

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ORIGIN OF PRECIPITATION

F.H. SCHMIDT

Royal Netherlands Meteorological Institute, De Bilt

In Western Europe, precipitation is normal phenomenon; it is of importance to all aspects of society, particularly to agriculture, in cattle breeding and, of course, it is a subject of hydrological research. Precipitation is an essential part in the hydrological cycle. How disastrous local conditions may be when precipitation holds off was obvious recently in the Sahel territory, where for several years precipitation, which had previously been marginal, stayed well under normally expected quantities. This introductory paper aims at conclusions that may be drawn with respect to the distribution of rainfall, when judged from a scrutiny of the mechanism that is responsible for the very origin of precipitation. (See Fletcher (1962) and Mason (1971)).

Precipitation is exclusively caused by clouds, and the origin of clouds is in itself something special. The condensation of water vapour, which results into clouds, cannot take place in pure air. For, relative humidities of 300 to 400% would then be necessary and such high humidities never occur in the atmosphere. That is why condensation of water vapour in the free atmosphere always takes place on small particles, the so-called condensation nuclei. These condensation nuclei can have different dimensions, generally in the order of 10^{-1} to 10μ. It is interesting to note that larger condensation nuclei generally lead to larger cloud-drops.

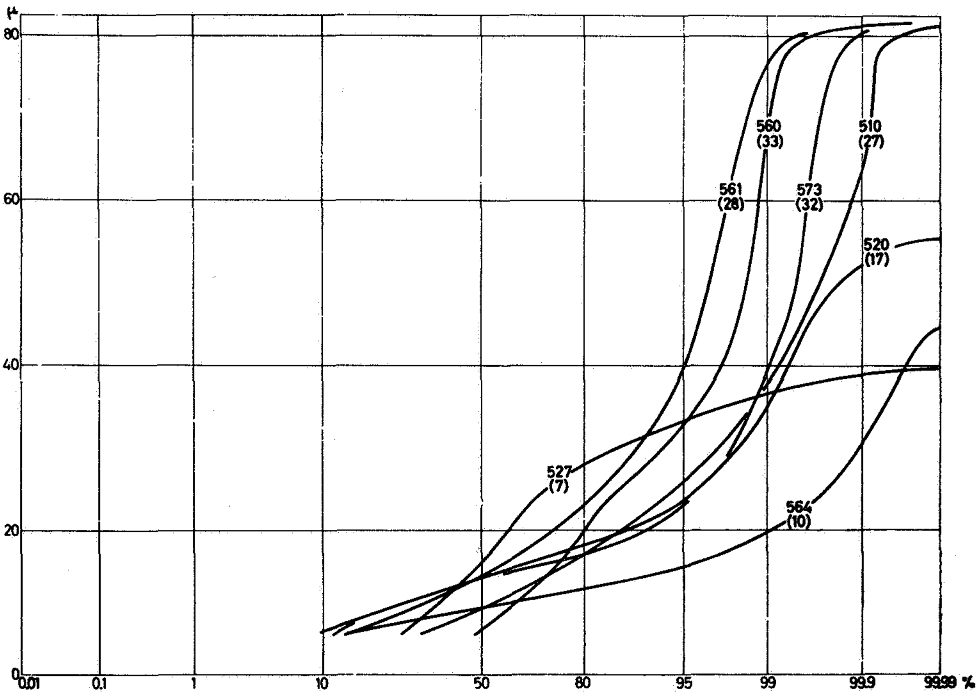
Condensation of water vapour generally takes place in the form of drops, but when these drops have temperatures below zero degrees centigrade, ice-crystals may be formed. Generally speaking, in the free atmosphere this does not happen immediately at a temperature of 0°C but as a rule at lower temperatures. Even at temperatures down to −10 or −20°C or, in extreme cases, even −40°C, waterdrops can still exist as such. They are called super-cooled droplets.

For the forming of ice-crystals it is generally that small crystalline particles are present in the waterdrops; around these crystalline particles, the waterdrops start freezing. So there are clouds consisting of waterdrops only. There are also clouds at higher levels that exclusively consist of ice-crystals. Very important for the genesis of precipitation are those clouds that largely consist of waterdrops but in which ice-crystals are present too.

Clouds are something special in that the presence of particles in the atmosphere is necessary for condensation to take place. The phenomenon of precipitation in itself is also peculiar. For there are clouds that do not give precipitation and others that do. This means that apparently something special must be the matter with a cloud, should it discharge in terms of rain, hail, or snow.

That such a special thing has to happen, becomes clear when the dimensions of cloud particles are compared with those of precipitation drops. A normal cloud drop has a diameter in the order of 10 to 100 micron. On the other hand, a normal rain drop has dimensions of 1 to 6 mm. This means that, as its volume is proportional to its radius to the third power, one thousand to one million cloud drops should coagulate to form just one element of precipitation. It is clearly out of the question that a million or even one thousand cloud drops present in a cloud, should in some way or other, without any obvious cause, unite to one precipitation drop.

It has been proved that the origin of precipitation can only be explained when, in a cloud, particles of various dimensions are present. The larger particles in such a cloud, having a higher rate of fall than the smaller particles, will fall, with regard to the latter, and thus overtake them. It has also been proved that, in this way, such a large falling cloud element can catch so many small cloud elements that the former at last gets dimensions that correspond with the dimensions of a normal raindrop.



1. Distribution of cloud drop dimensions in cumulonimbi, observed during 7 test flights with a Fokker Friendship. Numbers of flights are indicated. Number 564 refers to the flight over Amsterdam. Numbers between brackets give estimated surface windspeed in knots. (From F.H. Schmidt in Geofisica Pura e Applicata, *50*, 176, 1961).

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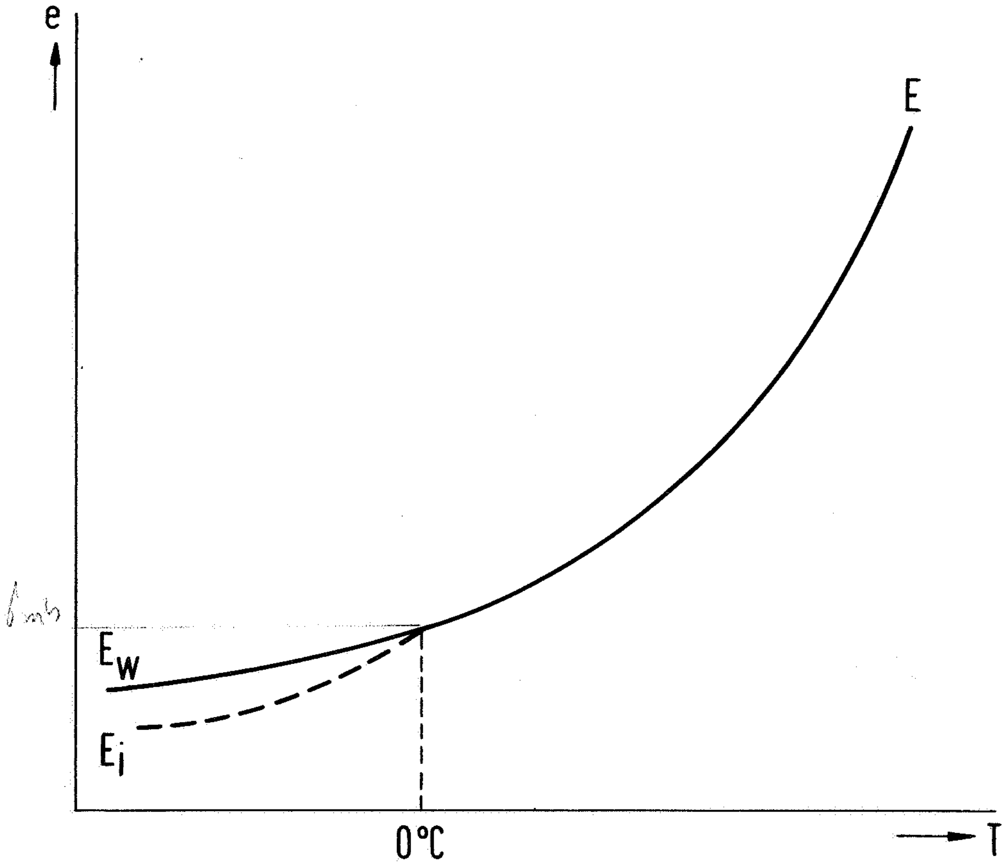
AUTHORS

Prof.dr. F.H. SCHMIDT	Royal Netherlands Meteorological Institute, De Bilt
Drs. H.A.R. DE BRUIN	Royal Netherlands Meteorological Institute, De Bilt
Dr. W. ATTMANNSPACHER	Deutscher Wetterdienst, Meteorological Observatory at Hohenpeissenberg, FRG
Dr. T.W. HARROLD	Meteorological Office, Bracknell, U.K.
Prof.Ir. D.A. KRAIJENHOFF VAN DE LEUR	Agricultural University, Wageningen

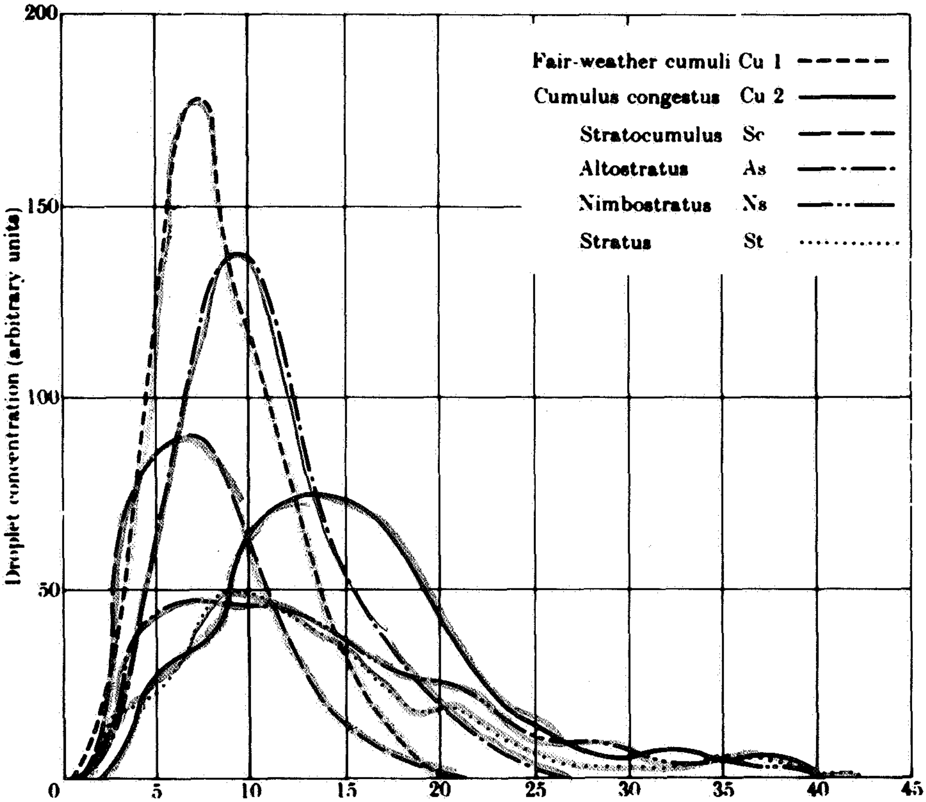
Now another difficulty remains in that the question should be answered: how do such large particles originate in a cloud? In the first place this may be due to coagulation, when there is more or less strong turbulence in the cloud. However, as a rule this process leads to no more than the relatively small droplets that occur in a drizzle.

To come to real precipitation, i.e. real rain or real snow, nature has two important processes at its disposal. It has already been mentioned that condensation nuclei of different dimensions can be present in the atmosphere and that the larger condensation nuclei generally lead to the occurrence of larger drops. Here one finds a first possibility for large particles to be originated in a cloud. These large particles fall, with regard to the smaller ones, coagulation takes place and precipitation drops are formed. This process mainly



2. Water vapour pressure as a function of temperature.
   - e = vapour pressure;
   - E = maximum vapour pressure for temperatures above freezing-point;
   - E_w = maximum vapour pressure for vapour in contact with supercooled water;
   - E_i = maximum vapour pressure for vapour in contact with ice;
   - T = temperature.

takes place when sufficient large condensation nuclei are present, and that is generally the case above seas. For sea-salt nuclei generally have large dimensions and, by their large hygroscopicity, change into large cloud drops. Notably in tropical regions, precipitation due to such large nuclei is a rather normal phenomenon. Also at moderate latitudes do sea-salt nuclei occur. This has been evident from measurements made during testflights of a Fokker Friendship. It was then found that above land, and especially over the city of Amsterdam, a great number of small condensation nuclei and, therefore smaller droplets, were present in the clouds, whereas above the North Sea also large drops were observed to occur in the clouds examined. (Fig. 1).



3. Droplet spectrum in some cloud families.
Precipitating clouds are:
altostratus
nimbostratus
stratus (drizzle)
Non-precipitating are:
fair-weather cumuli
stratocumulus.
Cumulus congestus may develop into precipitating cumulonimbus.
(from Mason, 1971)

Generally, another process is of much greater importance at moderate and high latitudes. It was already mentioned that, besides waterdrops, also ice-crystals appear in many clouds. Now the maximum pressure of water vapour in contact with ice is lower than the one when the water vapour is in contact with a supercooled water surface (Fig. 2). When in a cloud, a so-called mixed cloud, besides waterdrops also ice-crystals are present, the vapour pressure in the immediate surroundings of the crystals will be lower than the vapour pressure in the immediate surroundings of the water-droplets. This results into a vapour transport from places with a high vapour pressure to places with a low pressure, that is to say from the environment of the waterdrops to the environment of the ice-crystals. The vapour pressure around the waterdrops now becomes too low; the waterdrops partly evaporate while the pressure around the ice-crystals becomes too high. Higher, at least, than the maximum pressure that belongs to an ice-surface at the prevailing temperature. This means that the ice-crystals will grow because vapour will condense on them. And the large crystals start falling with respect to still smaller ones and small waterdrops. They will overtake these and grow by coagulation. In due course they will continue their fall as snow-crystals or snowflakes. This process notably takes place at moderate and high latitudes, which means that precipitation in these latitudes then generally originates as snow. Figure 3 shows the distribution of droplet radii in some precipitating and non-precipitating clouds.

When temperatures between the level where the snow is formed and the ground remain below freezing-point everywhere, snow is maintained as such. When the temperature near the surface of the earth, or the lower air-layers lies above the freezing-point, the snow will melt and it will start raining. Everybody knows that, at temperatures close to the freezing-point, e.g. +2° or +3°C, raindrops that fall on a windshield of a car may give the impression of little spurts that can be interpreted as snow-crystals that just melted and which essentially fall apart against that shield.

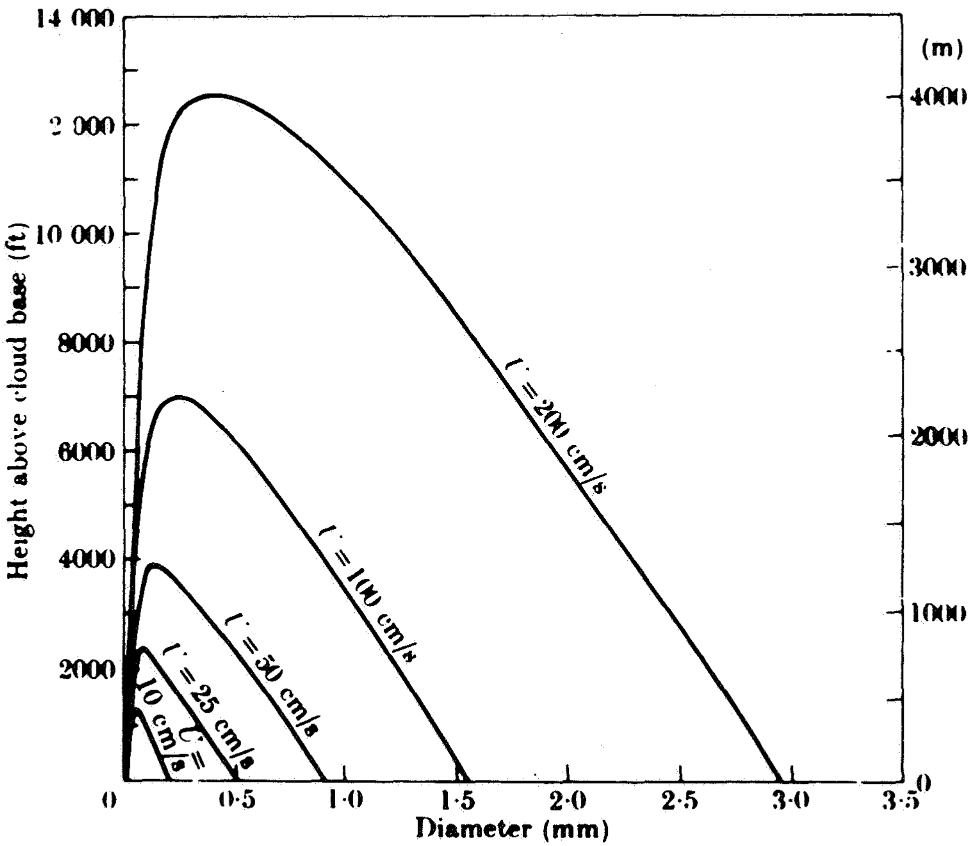
The two processes sketched above both entail on opportunity for rainmaking. The first one is mainly of importance in the tropics, where the following procedure is sometimes applied. When a cloud does not show any tendency of giving precipitation, a quantity of salt is brought into it, for instance by means of a balloon. Explosives next cause that salt to be fragmented into small crystals. Being highly hygroscopical, these salt-crystals, when diffused through the cloud, cause the forming of large drops. These large drops, overtaking the small ones, cause precipitation.

At temperature latitudes, precipitation can be generated when a cloud is available with temperatures below zero degrees centigrade in its upper parts, in which the amount of ice-crystals, necessary to cause appreciable precipitation, is insufficient.

Particles are then brought into that cloud that have a crystal structure so much resembling the structure of ice-crystals that the waterdrops contacting these crystals, freeze. A well-known crystal used in this process is silveriodide. One can also decrease temperatures in a cloud to about –60 or –70°C, by letting liquid CO_2 escape; in such low temperatures waterdrops cannot maintain themselves in the atmosphere and generally a sufficient number of ice-crystals will be formed to cause precipitation in that way.

The vital point in measuring precipitation is the question to what degree the intensity of precipitation will difffer from place to place. It may be clear from the above that one cannot expect precipitation to be equally heavy in all places.

In fact there may be more ice-crystals in one place than in others. Besides, for various reasons, the number as well as the diameters of the waterdrops in a cloud are highly dependent on the upward movements in the cloud. This may be accounted for as follows: due to its upward motion, the air will cool and as the cooling is more intense, starting from the same quantity of vapour in the air, the amount of vapour that will condense will be larger. It means that also more or larger drops will then come into existence than in a situation when the upward velocities in the air are lower. In the tests of the Fokker Friendship, which we mentioned, the number of drops varied between 50 and 350 per

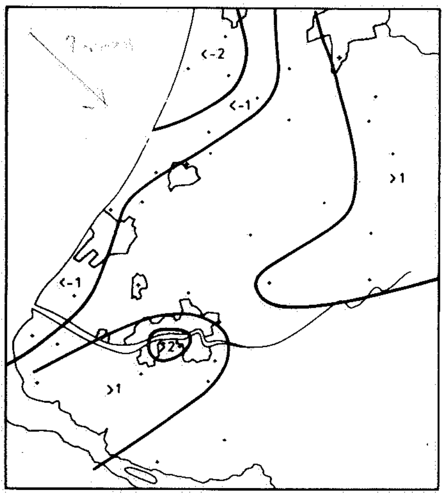


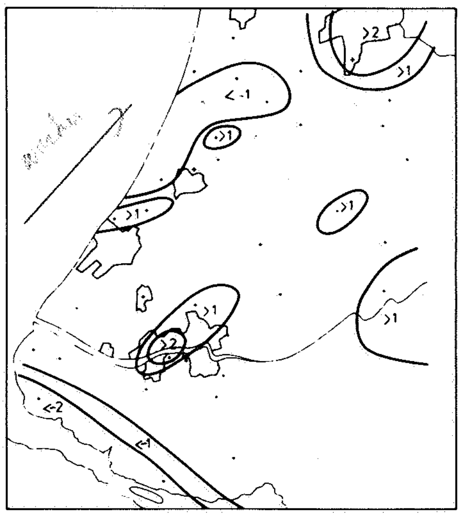
4. The variation of drop diameter for range of updraught velocities according to Bowen. (from Mason, 1971)

cm^3 for the 7 flights, while the number of drops with a diameter of 50 μm varied from 1.79 to 0.03 per cm^3.

A second reason why the amount of precipitation is related to the upward velocities in a cloud, lies in the fact that the stronger the updraught, the longer the larger cloud elements can remain in the cloud, and they then have an opportunity to overtake more small particles (Fig. 4). In this way one can expect that clouds in which a strong upward motion takes place, e.g. cumulonumbi, will generally give more precipitation than clouds with weak upward currents like the layer-clouds that form along warm fronts. As a matter of fact, the velocities at which the precipitating clouds pass the observation place is also of importance. But, also within the cumulonimbus, the upward movement in one place will be stronger than in another place; even in such apparently even clouds as warm-front clouds, a stronger upward velocity will be present in one place than in an other. It means that in both situations more precipitation will fall in one place than in another place, and such precipitation differences may be present at rather short distances. To get a good impression of the precipitation distribution in a certain area, one must therefore use a great number of pluviometers.

Up to now there has been some talk of an accidental distribution of vertical speeds, leading to an accidental distribution of water quantities in a cloud, and it is difficult to find a conclusive explanation for this phenomenon. But there are other causes of precipi-





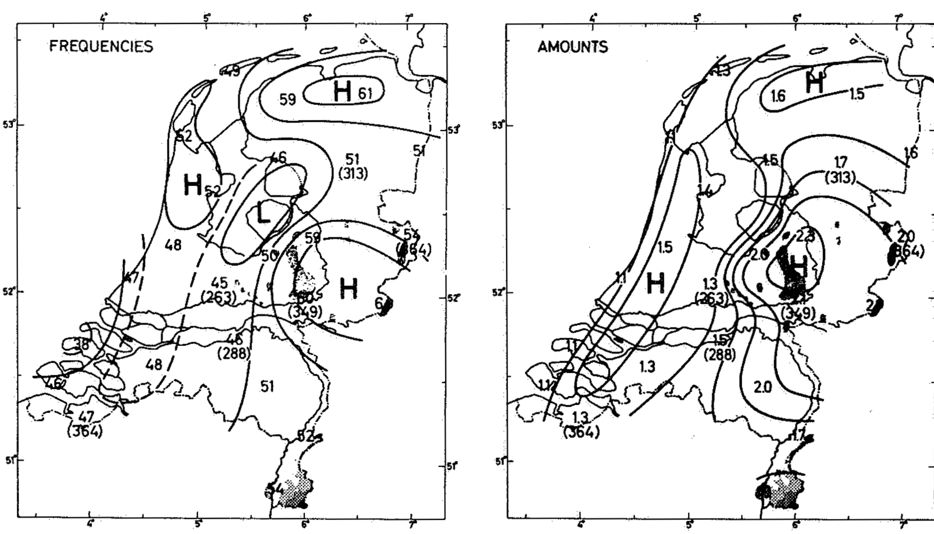
5. The effect of the city of Rotterdam on the amount of precipitation during NE-ly winds in summer. Numbers are normalized as follows: (precipitation - average precipitation over the region): standard deviation. Examples: Rotterdam, station at the Westerkade average 9.4 mm, regional average 8,9 mm, lowest value at Zandvoort, at the coast, west of Amsterdam 5.5. $\frac{x-\bar{x}}{s}$
6. The effect of the city of Rotterdam on the amount of precipitation during SW-ly winds in winter. Normalization as in Figure 5. Precipitation between 5 and 10 mm. Figures 5 and 6 are based on an an investigation by Mr. G. J. Yperlaan.

tation differences that can be understood, as will now be shown with reference to some relevant examples.

A First of all, generally more precipitation falls at the leeward side of big cities than at the windward side; this can be understood with the help of the preceding reasoning. A city is a source of heat and, as a result, an upward movement of the air above the city arises. An upward movement that moreover is intensifed due to the larger roughness in the city compared with the surrounding area, which forces the horizontal air current (the wind) to evade upward. So one might expect that at the leeward side of a city, more water will be present in the clouds than at the windward side; as a rule more precipitation will then fall at the leeward side of a city. One can even find this on some charts of the Climatological Atlas of the Netherlands. Recent investigations have also shown that there are slight differences in the quantities of precipitation; this can partially be explained by the effect of cities and large industrial areas (Figs. 5 and 6).

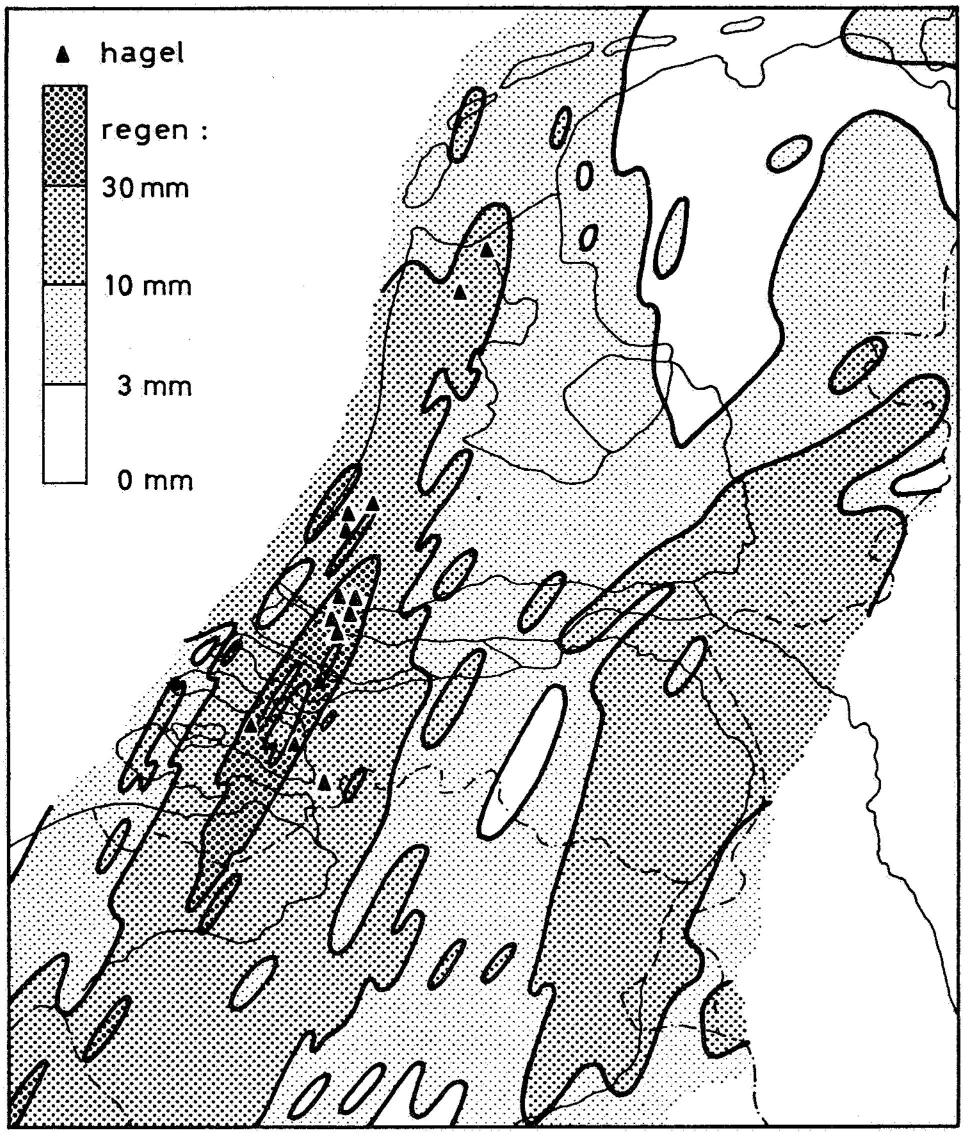
B Natural obstacles like woods and hills also cause vertical rising movements of air and, along with them a strengthening of the precipitation. Some investigators are convinced of the existence of a wave motion perpendicular to lines along which the surface roughness changes more or less abruptly, like coastlines. The effect of such a wave motion, characterized by very weak upward and downward motions, seems to be present in the precipitation pattern in the Netherlands (Fig. 7). How strong precipitation can differ from place to place during showery weather, appears from the total amounts when cold

WINDDIRECTION W-N



7. Possible effect of a weak wave motion, east of the coastline, on precipitation in the Netherlands (from Timmerman, 1963).

air penetrates the Netherlands from the south, putting an end to a hot summer period. It is during such invasions of cold air that most precipitation falls in this country. Figure 8 shows how strongly this precipitation can differ from place to place.



8. Precipitation in mm between 18 July 1964, 7.40 gmt. and 19 July 1964, 7.40 gmt., due to passage of a cold front after a period of hot summer weather in the Netherlands. (hagel = hail; regen = rain). (from H. R. A. Wessels, 1965).

An irregular precipitation pattern is often found along weak warm fronts. In such situations it even happens that no precipitation is falling at all.

All in all it seems that even in a flat country like the Netherlands, with only few natural obstacles, the distribution of precipitation may be quite irregular. Some aspects of this irregularity appear on climatological maps. This means that, certainly for every day or for every period of one hour or a few hours, the differences from place to place will be large. Consequently, if one wants to have, for all kinds of purposes, an accurate impression of the amount of precipitation that falls in small areas, special measurements should be made to obtain sufficiently reliable results.

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THE ACCURACY OF MEASURING AREAL PRECIPITATION WITH A RAIN GAUGE NETWORK

H.A.R. DE BRUIN
Royal Netherlands Meteorological Institute, De Bilt

1. SYNOPSIS

The main purpose of this technical meeting of the Committee for Hydrological Research TNO, is to report on the use of radar for precipitation measurements. In practice, areal precipitation is at the moment mainly determined by rain-gauge measurements at individual points. The aim of this paper is to discuss the accuracy which this method may yield.

Firstly, the instrumental errors of a rain-gauge proper are described. Secondly, the spatial variability of a precipitation field is discussed. Thirdly, a theoretical approach of the problem is given and, lastly, some results are shown of experiments carried out with a relatively dense network of rain-gauges.

2. INTRODUCTION

For many hydrological applications, information on the precipitation depth averaged over a certain area, S, and fallen in a certain time interval, T, is desirable. In practice, this average is mostly determined with rain-gauge measurements taken at a finite number of individual points.

It is perhaps interesting to relate that this method was first described many centuries ago; Biswas (1971) reports that the Chinese scientist Chiu Chiu Shao, who lived about 1200 A.D., discussed it in his book "Shu Shu Chiu Chang".

The accuracy of the method depends on several factors. The instrumental accuracy is, of course, important; the spatial statistical structure of a precipitation field, depending on the climatological characteristics of the area of interest, plays a prominent part, while further the quantities S, T, the number of rain-gauges, N, the spatial distribution of the gauges, and the method of estimation itself may affect the accuracy of the method.

3. INSTRUMENTAL ERRORS OF A RAIN GAUGE

General remarks

A rain gauge is an instrument that has been used for thousands of years (Biswas, 1971). It usually has the form of a collector above a funnel leading into a receiver.

In the Netherlands, the so-called "national standard rain gauge" is installed with its rim 40 cm above ground level, while since 1962 its orifice has an area of 200 cm^2. (Before 1962, its area was 400 cm^2 and before 1946 its height 1.50 m).

In general, the catch of this rain gauge differs from the true precipitation depth because of instrumental and observational errors. Some of these are discussed in the following sections.

Systematic errors

A rain gauge is affected by a number of systematic errors. For instance: evaporation and wetting losses; rain can splash in and out, and wind also acts on the gauge. Although these errors depend on such factors as the nature of the gauge, the meteorological conditions and characteristics of the site, Kurtyka (1953) (see also Rodda (1971)) has made the following approximations (see Table 1) of the magnitude of the most important systematic measuring faults:

Table 1

Evaporation	Adhesion	Colour	Inclination	Splash	Wind and exposure
−1.0%	−0.5%	−0.5%	−0.5%	+1%	−5 to −80%

It will be seen from this table that, except for the error caused by splash, the effects are all negative, while the error due to wind and exposure appears to be the most important. Therefore, this error will now be discussed in more detail.

The process of wind acting on a rain gauge is rather complicated. Qualitatively it can be described as follows:
Wind interacts with the gauge and the feature of the site to produce eddies and turbulence in such a way that, in the region immediately above the gauge, raindrops or snowflakes are accelerated. This results in an alteration of their trajectories by which a fraction of the drops or flakes, which otherwise would have reached the orifice without disturbance, is not caught by the gauge.

In Figure 1, taken from Green and Helliwell (1972), streamlines are drawn as measured above a rain gauge installed in a steady wind flow of about 3 m/s produced in a wind tunnel. It is immediately noticeable that the effect of the rain-gauge is to cause a convergence of streamlines from the windward edge of the gauge. The alteration of raindrop trajectories is due to this convergence. Obviously, the trajectories of small raindrops are

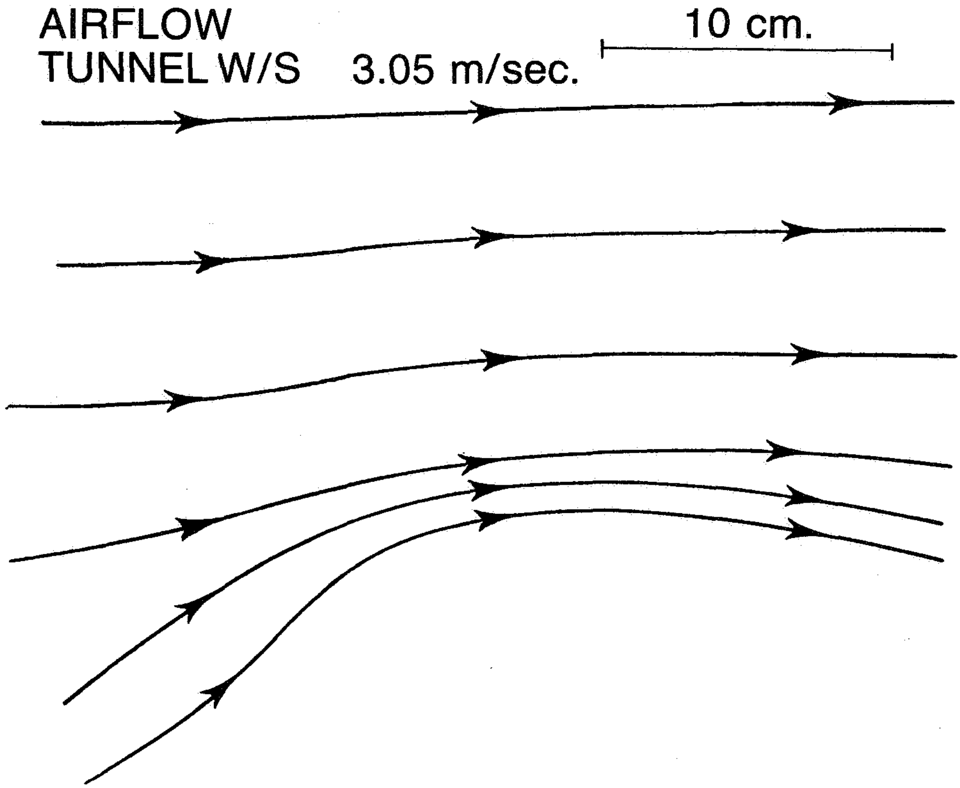


Fig. 1. Streamlines above the Mk 2 rain-gauge funnel with a wind speed of 3.05 m/s, taken from Green et. al. (1972).

altered more than those of bigger ones, while a much greater proportion of snowflakes is being deviated because of their larger surface area and their lower fall-velocities.

The results shown in Figure 1 are valid for steady wind flows. In natural field conditions, the flow will be far from steady: it will be more gusty. Furthermore, the wind flow is influenced by obstructions, such as bushes, in the vicinity of the rain-gauge. All these factors together make that a quantitative description of the process is hardly possible.

In practice, the order of magnitude of the error due to the wind, as far as liquid precipitation is concerned, is obtained by comparing the catch of the rain gauge of interest with that of a so-called "pit gauge". This is a rain gauge installed in such a way that its rim is at ground level. Usually an anti-splash grid is constructed to protect the pit gauge against splash.

It is generally accepted that a pit gauge is not affected by wind. In the Netherlands, a number of comparisons between the national standard rain gauge and a pit gauge was carried out (see e.g. Braak (1945), Dey (1968), Colenbrander and Stol (1970)).

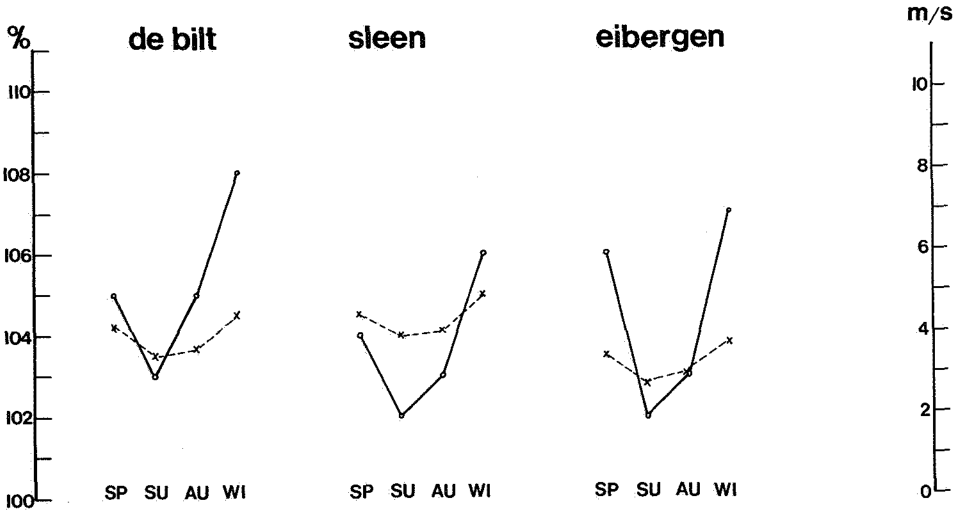


Fig. 2. Seasonal fluctuation of the difference in precipitation depth between the standard rain gauge, h_s, and an international reference pit gauge, h_p (o-o-o), and the seasonal fluctuation of the mean wind speed (x-x-x); $h_p = 100\%$.

In Figure 2, the result of such comparisons made in 1972–1975 within the framework of an international comparison programme, organized by the World Meteorological Organization through its Commission for Instrumentation and Methods of Observation, are given for the Dutch stations De Bilt, Eibergen (Hupsel) and Noord Sleen. It is seen that in winter the pit gauge catches 6–8 per cent and in summer about 3 per cent more than the national standard gauge.

4. RANDOM ERRORS OF A RAIN GAUGE

Random errors due to reading the measuring glass

In the Netherlands, the measuring glass is read to 0.1 mm. The average daily rainfall amount is about 4 mm, if only wet days are taken into account. So the random error due to reading the measuring glass is about 2.5 per cent of the average daily precipitation depth. As about 50 per cent of the days are dry, the random error of monthly totals due to this effect can be estimated with an accuracy better then 1 per cent.

Errors due to the "random character of rain"

Let us imagine a cloud which procedures raindrops, so that the dropsize distribution and the rainfall intensity do not vary with time. During fixed time intervals, T, we take samples of the rainfall depth, h. If N_i is the number of raindrops caught during the i^{th} sample, we can write:

$$h_i = \frac{\pi}{6} \frac{N_i(\widetilde{D^3})_i}{0} \tag{1}$$

where $(\widetilde{D}^3)_i$ is the average value of the third power of the diameter of the raindrops caught during the i^{th} sample, and 0 the area of the rain gauge. Doing so, it will be observed that in general the values of h_1, h_2, h_3, ... are not equal. This variation is caused by random variation in N and $\widetilde{D}^3$, due to sampling effects. Appendix I shows, that the relative value of the random error caused by this effect is given by:

$$\frac{\sigma(h)}{\bar{h}} \approx \sqrt{\frac{\widetilde{D}^6}{N(\widetilde{D}^3)^2}} \tag{2}$$

It should be noted that N is proportional to 0, so $\sigma(h)/\bar{h} \backsim \sqrt{1/0}$.

During 1968 and the first few months of 1969, raindrop-size measurements were carried out at De Bilt, using a filter-paper technique (Wessels, 1972). These data (one purpose to collect them was to determine the relationship between rainfall rate and radar reflectivity) were used to get an impression of the order of magnitude of $\sigma(h)/\bar{h}$. It should be remarked that the measurements could only be carried out in working time. Table 2 shows some results for "daily" rainfall amounts and for two values of 0 (20 and 200 cm^2).

Table 2

		$0 = 20\ cm^2$	$0 = 200\ cm^2$	
date	$\bar{h}$ mm	$\sigma(h)/\bar{h}.100$	$\sigma(h)/\bar{h}.100$	$N(0=20\ cm^2)$
5-1-68	0.16	6.7	2.1	1355
9-2-68	0.60	4.3	1.4	4651
14-2-68	0.06	4.9	1.5	3402
21-2-68	0.48	2.5	0.8	5898
21-3-68	1.9	2.4	0.8	11216
2-4-68	1.3	2.5	0.8	13820
4-4-68	2.7	2.8	0.9	12462
23-4-68	3.0	1.5	0.5	23892
29-4-68	0.2	5.2	1.6	1727
7-5-68	1.6	3.4	1.1	8705
13-5-68	0.3	6.5	2.1	2903
6-6-68	4.0	1.7	0.5	34815
17-6-68	4.3	2.9	0.9	18364
10-7-68	8.3	1.5	0.5	39856
26-9-68	6.3	1.0	0.3	50025
20-2-69	1.4	3.9	1.2	6916
13-3-69	11.2	1.0	0.3	48142

Table 3 shows some examples of short time intervals:

Table 3

		$0 = 20\ cm^2$	$0 = 200\ cm^2$	
ΔT min	$\bar{h}$ mm	$\sigma(h)/\bar{h}.100$	$\sigma(h)/\bar{h}.100$	$N(0=20\ cm^2)$
5	0.427	7.3	2.3	1271
5	0.096	8.8	2.4	1691
1	0.177	13.0	4.1	463
1	0.1732	14.6	4.6	512
1	0.329	8.9	2.8	799
1	0.052	21.0	6.7	175
1	0.241	17.6	5.6	1099
5	0.245	10.3	3.3	422

From these tables it can be concluded that the error due to random character of rain is negligible for rain gauges having an area of 200 cm^2.

In the literature (see for instance Kalma et al. 1969), rain gauges with smaller areas are described. It is seen from Tables 2 and 3 that for these types of gauges the values of $\sigma(h)/\bar{h}$ are not negligible.

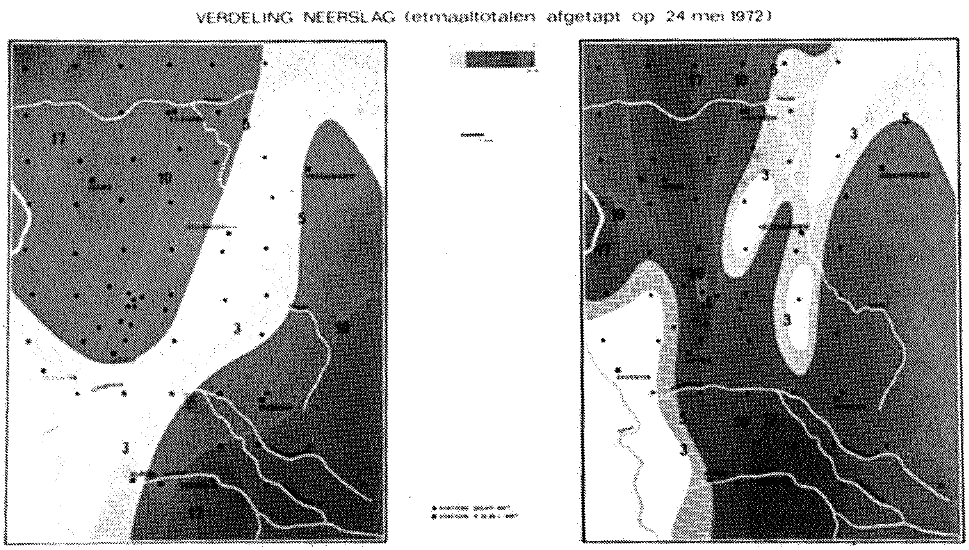


Fig. 3. Spatial distribution of daily precipitation amounts, as measured in Salland on May 24, 1972, with the national rain gauge network (Fig. 3a) and a dense network (Fig. 3b).

5. SPATIAL VARIABILITY WITHIN A PRECIPITATION FIELD

From our own experiences we know that the spatial distribution of precipitation can be highly irregular. The meteorological processes responsible for this variability were reviewed this morning by Schmidt (1976).

To illustrate the possible differences over a short distance, the spatial distribution of the daily precipitation amounts is shown in Figure 3, as measured on May 24, 1972, with a dense network installed in Salland, a district in the eastern part of the Netherlands, in 1970–1972. This was done within the framework of a hydrological research programme (Santing et al., 1974).

In this figure the same precipitation field is drawn as "seen" by the (less denser) national rain gauge network of the Royal Netherlands Meteorological Institute.

This example shows that, over distances of 1 kilometre, differences of 20 mm/day can occur.

Generally, these differences within areas such as Salland are not systematic. For most of the practical applications, areas in the Netherlands of, say, 1000 km^2 can largely be considered to be climatologically homogeneous and isotrope.

It will be pointed out in the next section that a convenient quantity to describe the statistical spatial structure of a precipitation field is the correlation coefficient, ρ, defined by:

$$\rho_{ik} = \frac{\overline{h_i h_k} - \overline{h}_i \overline{h}_k}{\sigma_i \sigma_k} \tag{3}$$

in which the bars indicate an average over a long time series; h_i and h_k are the precipitation depth measured at the i^{th} and k^{th} station, while σ_i and σ_k are the standard deviations (of time series) of these stations.

If the distance between the two stations is denoted by r_{ik}, it is to be expected that ρ_{ik} decreases with increasing r_{ik}, if a homogeneous and isotrope area is considered.

It has been found (see e.g. Stol (1972) and De Bruin, (1975) that the "correlation function", $\rho(r)$, can be (more or less) satisfactorily described by:

$$\rho(r) = \rho_0 \, e^{-r/r_0} \tag{4}$$

which relationship can be written for $r << r_0$ in the form:

$$\rho(r) = \rho_0 - r/r_0 \tag{5}$$

Stol fitted the daily observations of a 10-year period from 9 selected stations situated in the eastern part of the Netherlands to equation (4). He did so for each month. The ("smoothed") results are given in Figure 4 (according to Stol, 1967), while the values of ρ_0 and r_0, as found by Stol, are listed in Table 4. It should be noted that he only took those days when at least 0.5 mm had been recorded at each station.

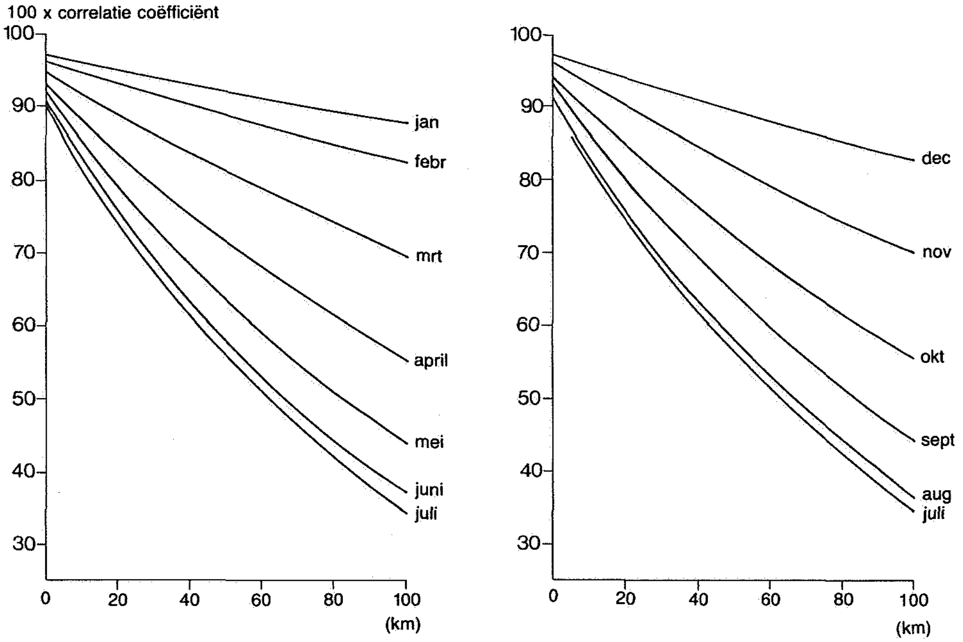


Fig. 4. The correlation coefficient, ρ, as a function of the distance, r, for each month (daily totals), taken from Stol (1967).

Table 4

month	ρ_0	r_0 km
1	.97	962
2	.96	630
3	.95	320
4	.94	190
5	.93	135
6	.91	111
7	.91	104
8	.92	110
9	.93	133
10	.95	186
11	.96	309
12	.97	609
Year	.94	188

Ten days total from 36 stations in the same area and of the period 1966–1972, were processed by De Bruin (1975). He concluded that for r, up to about 30 km, equation (5) can be used. However, he found a big scatter plotting ρ against r (a scatter which, after all, is also present in the data of Stol). De Bruin pointed out that this scatter is partly due to random observational errors, because these lower the value of ρ and, therefore, increase its variability.

Table 5 shows the values of ρ_0 and r_0 from (5), as found by De Bruin for 14 selected stations. Furthermore, the corresponding values of all 36 stations are listed.

Table 5

period	14 selected stations		all 36 stations	
	ρ_0	r_0 km	ρ_0	r_0 km
Nov.–March	.99	1000	.95	1000
Sept., Oct., Apr., May	.99	270	.98	270
July–Aug.	.99	190	.99	190

It is seen from Tables 4 and 5 that a seasonal effect exists: in winter, r_0 has a maximum and in summer a minumum value. This is so because in winter mainly frontal (that is large-scale) precipitation occurs, whereas in summer also convective storms (small-scale precipitation) are likely to occur.

Table 6 lists the values of ρ_0 and r_0 from equation (5) for 5-day periods and different "seasons", as found in Salland by means of the records of the dense raingauge network mentioned before. The values found in this network for 1 and 10-days periods were similar to those listed in Tables 4 and 5.

Table 6

period	selected stations	
	ρ_0	r_0 km
"year"	.97	230
"summer"	.97	150

6. ACCURACY OF THE ESTIMATES OF AREAL PRECIPITATION

Estimate with one rain gauge

A homogeneous, isotrope and square area, S, with edge ℓ is considered. In the centre. P, of this area, a rain gauge is installed.

The precipitation depth averaged over the area is denoted by h_G, while that measured at P is called h_0. As the area is supposed to be homogeneous, an estimate of h_G is:

$$\hat{h}_G = h_0 \tag{6}$$

Appendix II shows that the standard error of this estimate, E, defined by $E^2 = \overline{(h_G - \hat{h}_G)^2}$, can be written as:

$$E^2 = \sigma^2 \left[(1-\rho_0) + 0.23 \frac{\sqrt{S}}{r_0} \right] \tag{7}$$

where σ is the standard deviation of a long time series taken at an arbitrary point in S.

This relationship is valid, if equations (4) or (5) are supposed to be true.

Introducing $Z = E/\overline{h}$ and $C_v = \sigma/\overline{h}$, we obtain:

$$Z = C_v \sqrt{(1-\rho_0) + 0.23 \frac{\sqrt{S}}{r_0}} \tag{8}$$

Table 7 shows some characteristic values of C_v for different time intervals.

Table 7

T days	$C_v = \sigma/h$
1	1.4
5	1.1
10	0.9
30	0.5

Accuracy obtained with N evenly distributed rain gauges

Next, the case is considered that N rain gauges are evenly distributed over area S. The recorded precipitation depths at these stations are denoted by $h_1, h_2, ..., h_N$.

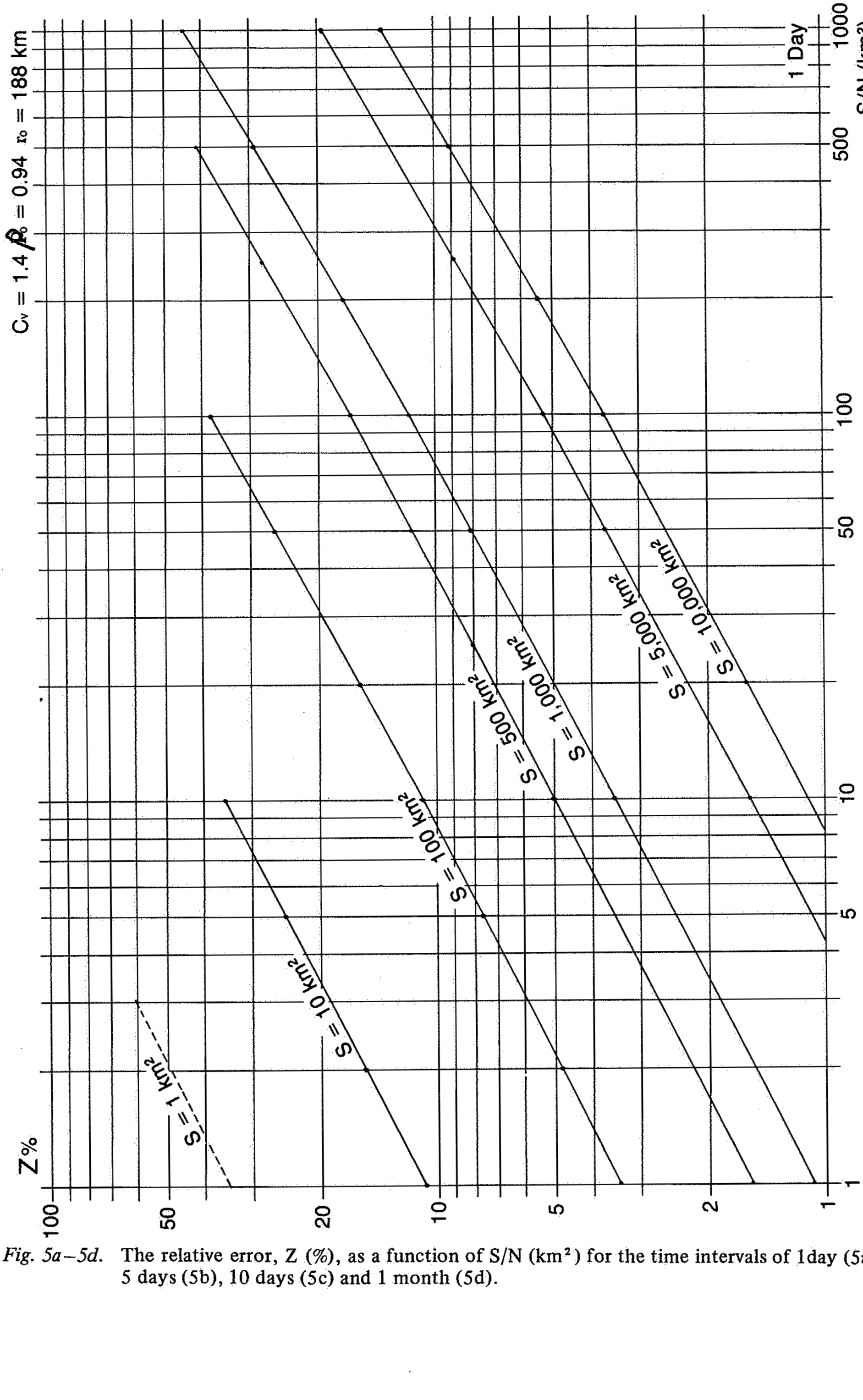


Fig. 5a–5d. The relative error, Z (%), as a function of S/N (km²) for the time intervals of 1day (5a), 5 days (5b), 10 days (5c) and 1 month (5d).

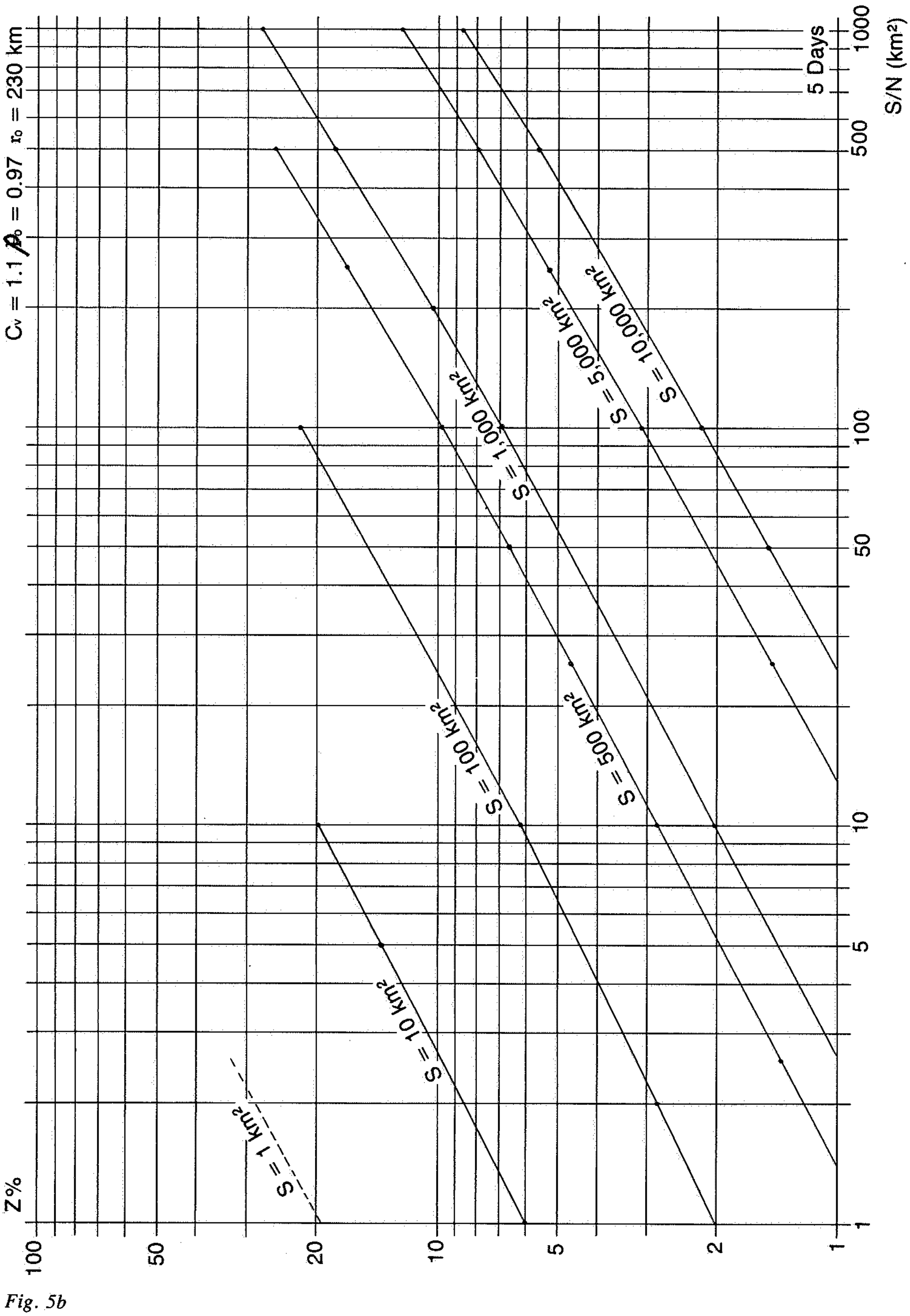
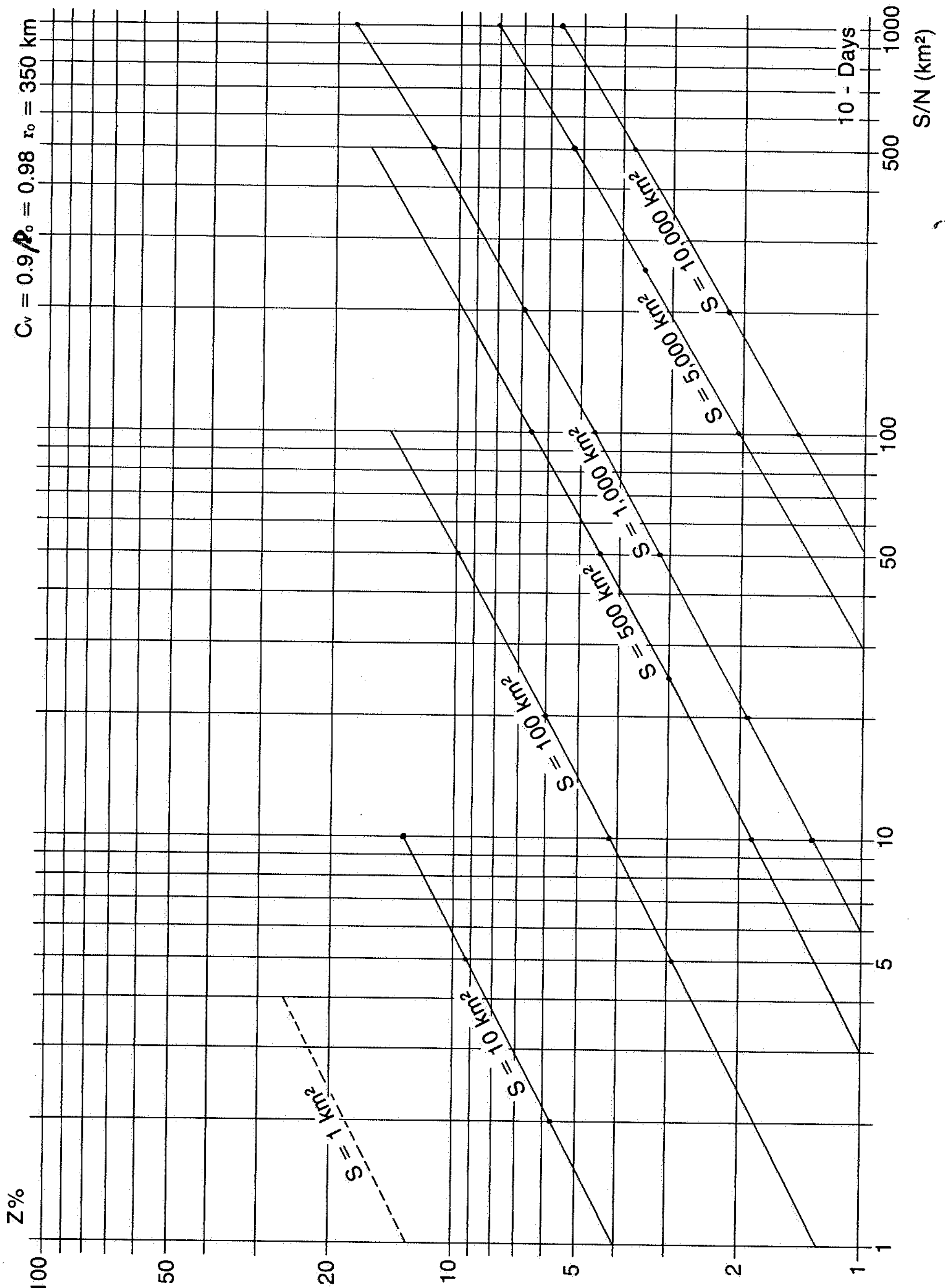

Cᵥ = 1.1 ρ₀ = 0.97 r₀ = 230 km
Z%
S/N (km²)
5 Days
S = 1 km²
S = 10 km²
S = 100 km²
S = 500 km²
S = 1,000 km²
S = 5,000 km²
S = 10,000 km²
100
50
20
10
5
2
1
1
5
10
50
100
500
1000

Fig. 5b



$C_v = 0.9$ $\rho_0 = 0.98$ $r_0 = 350$ km
10 - Days
S/N (km²)
Z%
S = 10,000 km²
S = 5,000 km²
S = 1,000 km²
S = 500 km²
S = 100 km²
S = 10 km²
S = 1 km²
1000
500
100
50
10
5
1
100
50
20
10
5
2
1

Fig. 5c

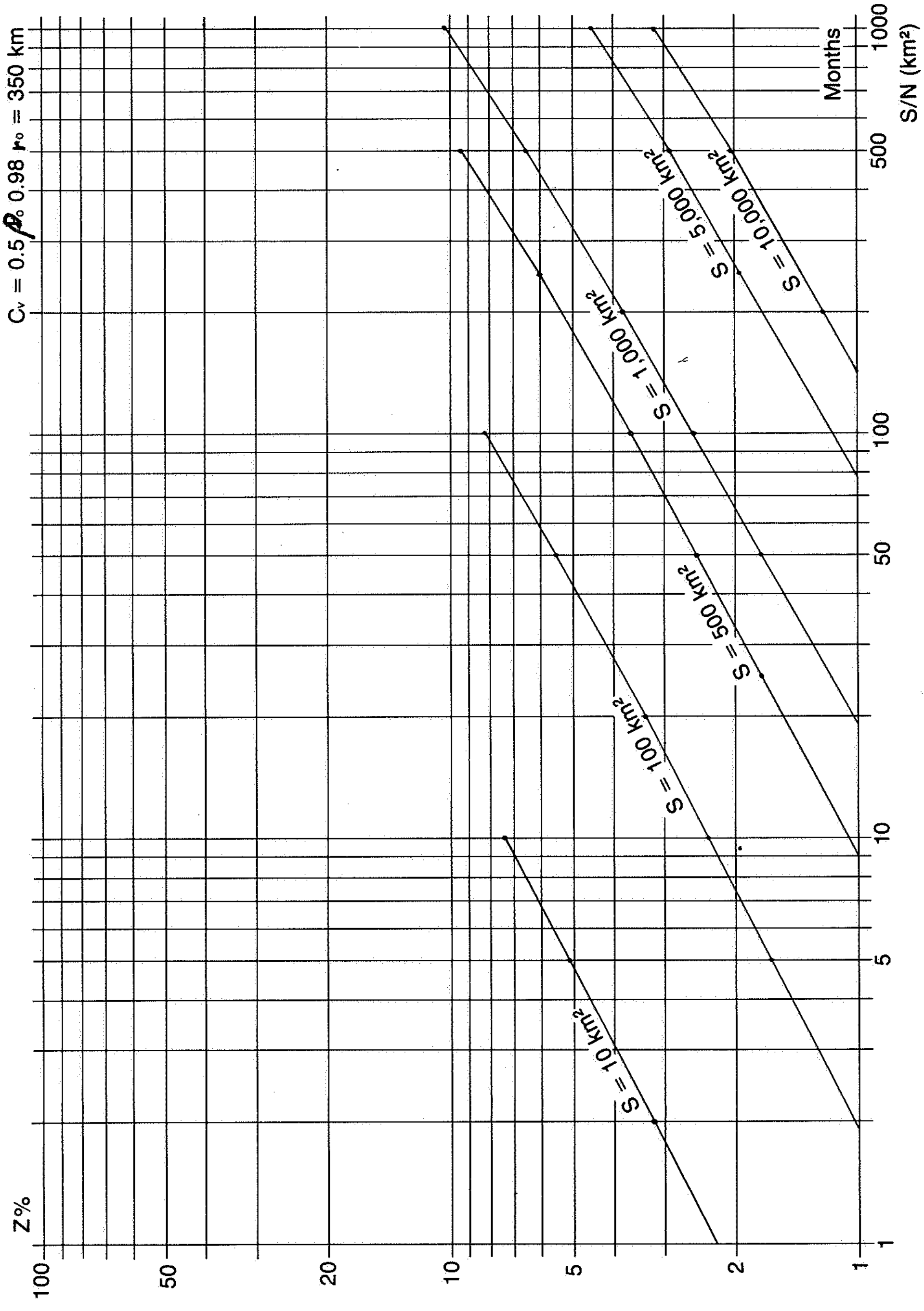

Cv = 0.5 ρo 0.98 ro = 350 km
Months
S/N (km²)
Z%
S = 10,000 km²
S = 5,000 km²
S = 1,000 km²
S = 500 km²
S = 100 km²
S = 10 km²
1000
500
100
50
10
5
1
100
50
20
10
5
2
1

Fig. 5d

An estimate of h_G is now given by:

$$\hat{h}_G = \frac{1}{N} \sum_{i=1}^{N} h_i \tag{9}$$

while the relative error Z can be written as

$$Z = C_v \sqrt{\frac{1}{N} \left[(1 - \rho_0) + \frac{0.23}{r_0} \sqrt{\frac{S}{N}} \right]} \tag{10}$$

(See Appendix II, and also Kagan, 1972).

Figures 5a, b, c and d present Z drawn as a function of S/N for various values of S and for different time intervals.

Some remarks on equation (10)

The reliability of equation (10) depends on the quantities ρ_0 and r_0, found with the empirical correlation function given by (4) and (5). But, because of the rather big scatter observed when correlation coefficient ρ is plotted against distance r, the accuracy of these quantities (notably that of ρ_0) is not very high. This is one of the main problems of this "theoretical" approach.

Quantity ρ_0 has been introduced as a result of the random errors of observation (see Appendix II). Although ρ_0 can be determined only with a relatively low accuracy, it is interesting to see what happens when the value of ρ_0 is altered.

In Figure 6, Z is drawn as a function of S/N for 1000 km^2 and 10-day totals, while the following values of ρ_0 were chosen: 0.96, 0.99 and 0.9975, corresponding with $\Delta = 0.20$, 0.10 and 0.05 respectively (Δ is the random error of observation divided by σ). Suppose we have a rain-gauge network with S/N = 100 km^2 and $\Delta = 0.20$. It is then seen from Figure 6 that in a network with S/N $\approx$ 230 km^2 and $\Delta = 0.10$, or S/N $\approx$ 320 km^2 and $\Delta = 0.05$, the same value of Z is obtained. This example is given to illustrate that under certain circumstances it can be more economical to decrease the number of rain gauges, if at the same time the random measuring error is lowered (for instance, by means of automation of the network).

7. COMPARISONS WITH A DENSE NETWORK IN SALLAND AND IN ZUID-HOLLAND

It is in itself a problem to test the reliability of equation (10). The only thing one can do is compare the records of a dense and a less dense network of rain-gauges with one another, assuming that the densest network gives the true areal precipitation depth.

In this manner, the data of a network of rain gauges installed in Salland during 1970–1972 were used.

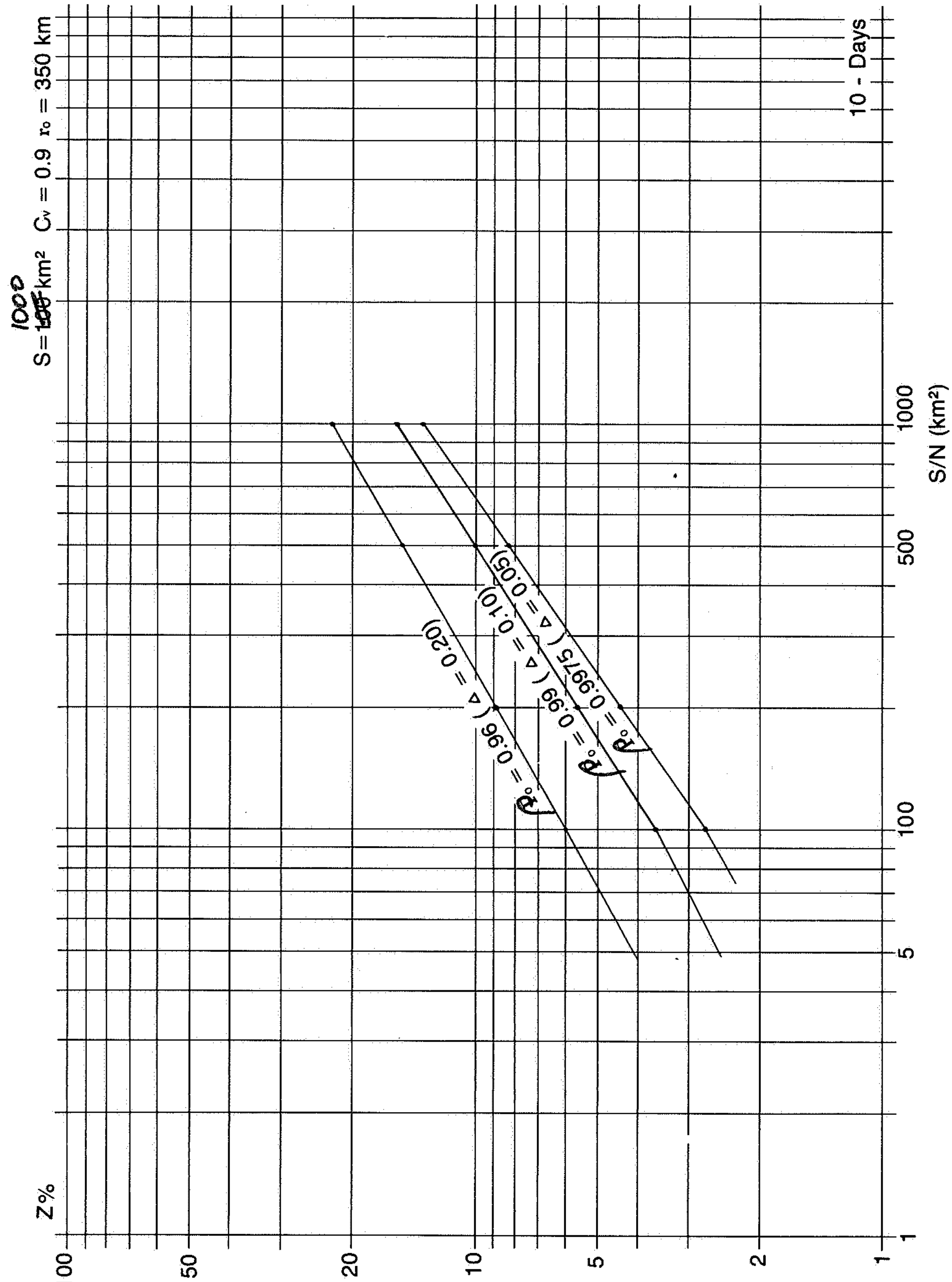


Fig. 6. The relative error, Z (%), as a function of S/N (km^2) for different values of ρ_0 (10-days period and S = 1000 km^2).

Within an area of about 1100 km², rain gauges were erected at 65 points. These are plotted in Figure 7. It is seen that in principle the gauges are evenly distributed over the area. They form a 5 x 5 km grid. In the centre of the area (in the neighbourhood of Heeten), the network had a special density.

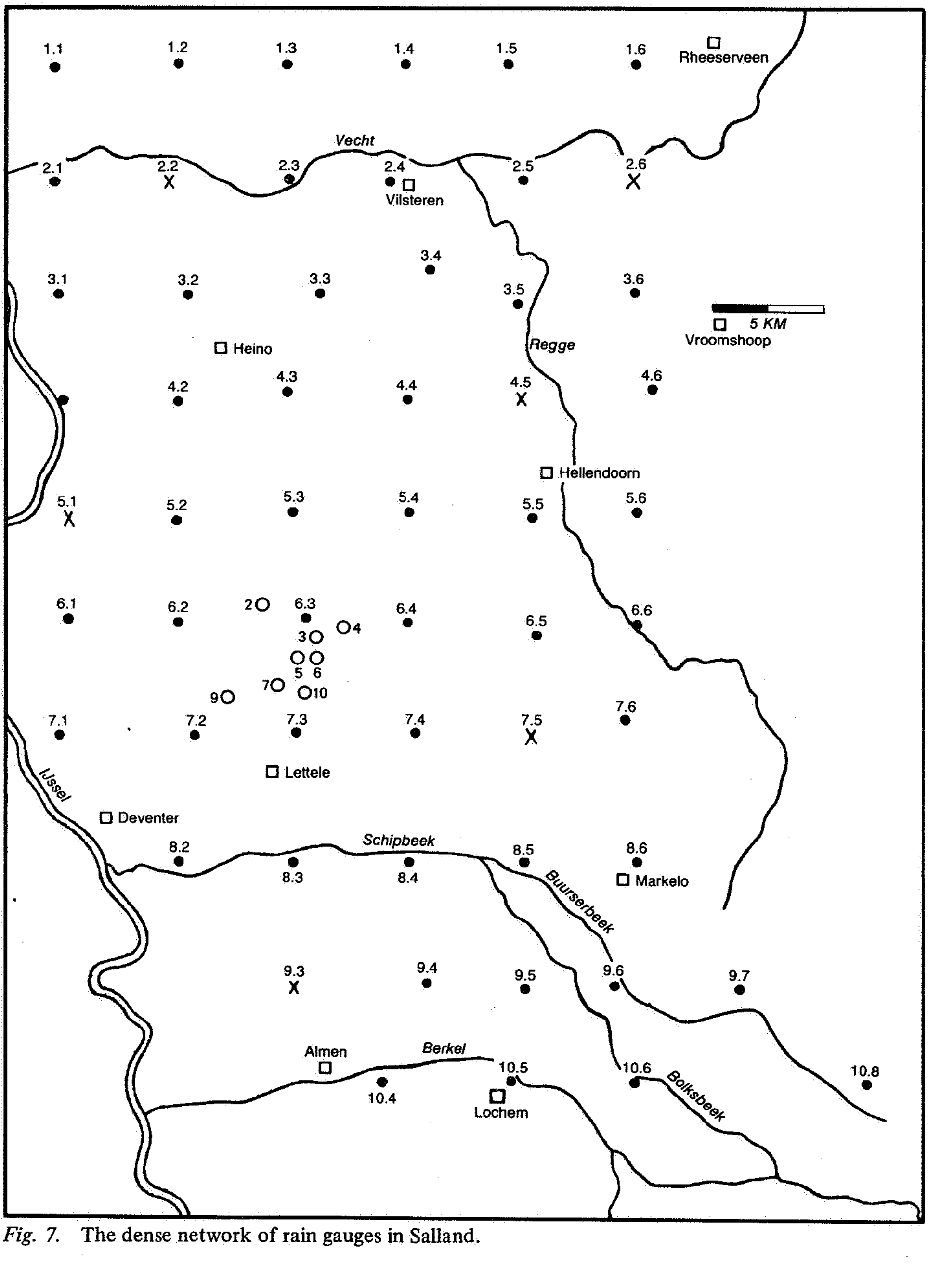


Fig. 7. The dense network of rain gauges in Salland.

Here the distances between two stations were up to about 750 m. In order to avoid the systematic measuring error due to wind, pit gauges were used. For budgetary and organizing reasons, a special, cheap type was developed; see Figure 8.

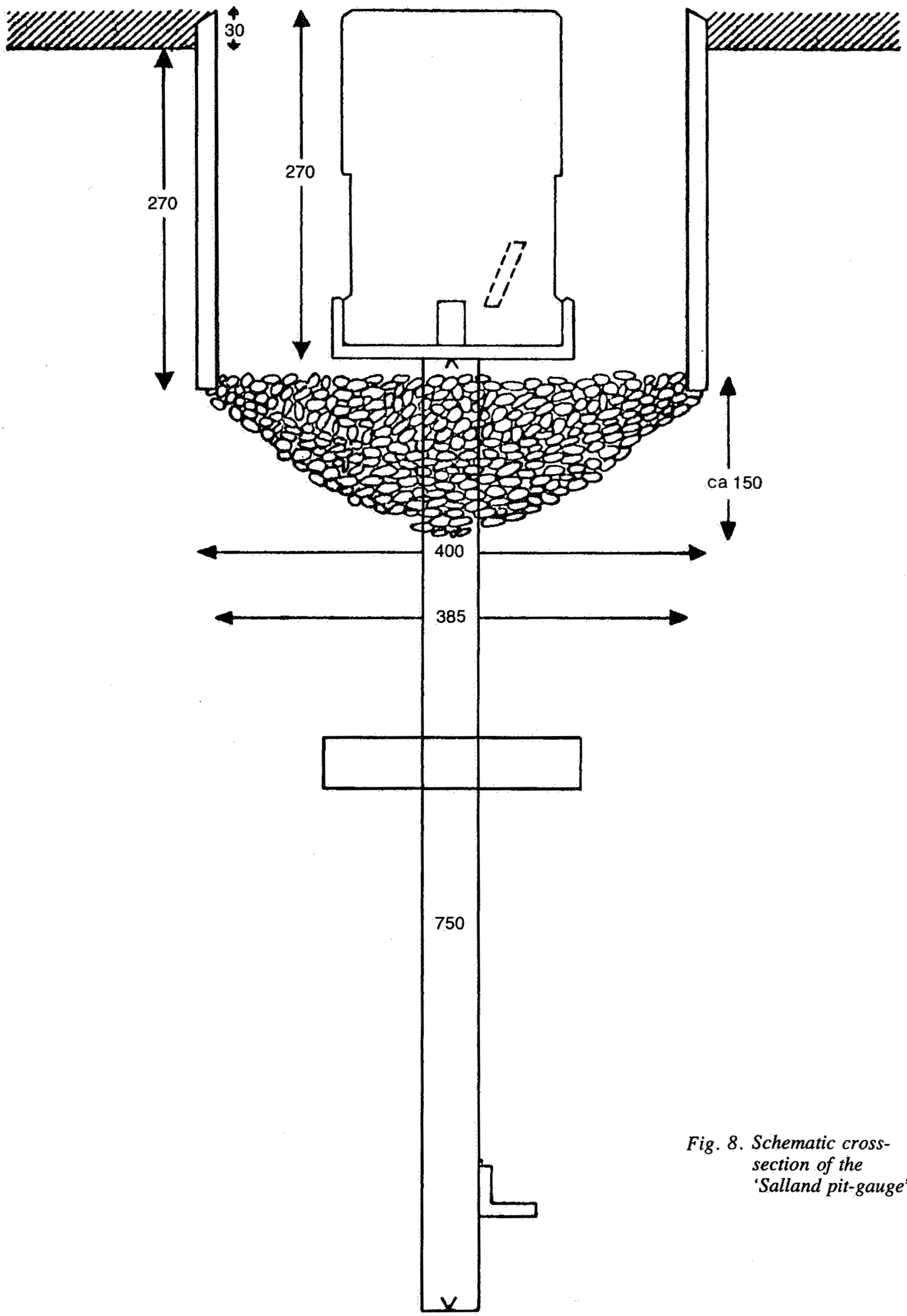


Fig. 8. Schematic cross-section of the 'Salland pit-gauge'

Comparisons between this type of rain gauge and an international standard pit gauge, mentioned in one of the previous sections, were carried out at De Bilt in 1972–1975.

Two "international" pit gauges were installed next to three "Salland" pit gauges.

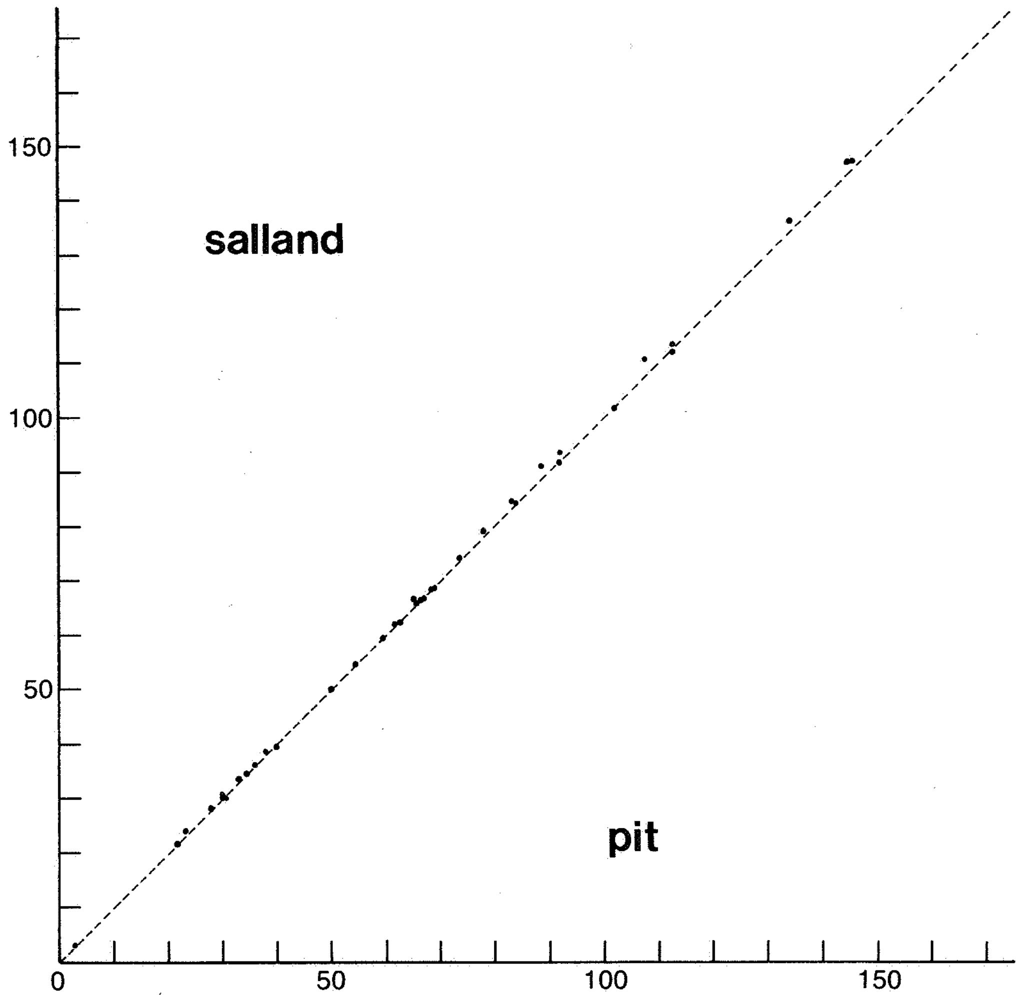
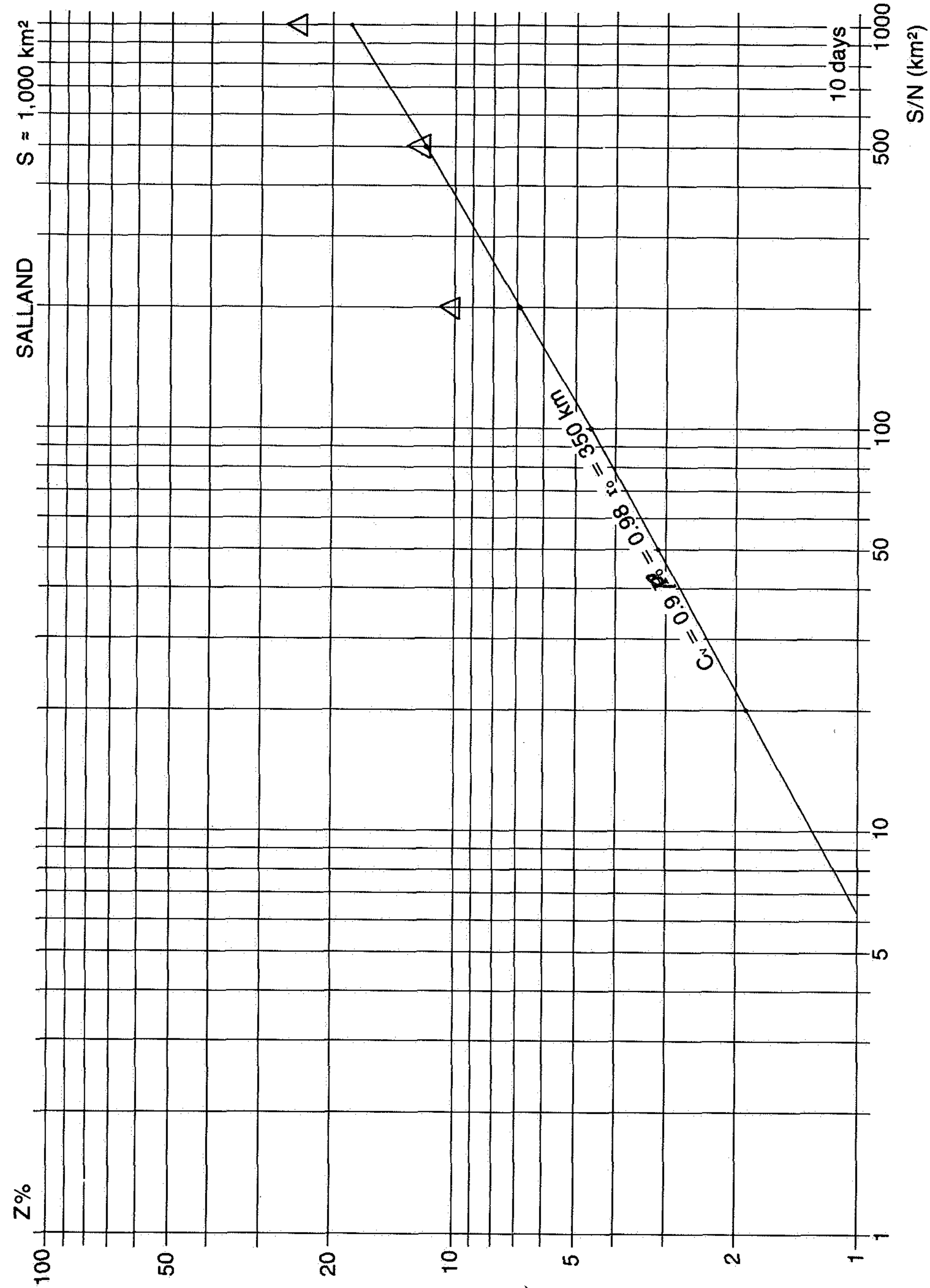


Fig. 9. Comparison between the monthly precipitation depth measured at De Bilt with the "Salland pit-gauge" and the international reference pit-gauge (mm).

In Figure 9, the means of the monthly totals measured with the three "Salland" gauges are plotted against those of the two "international" gauges. It is seen that there are hardly any differences between the two types of pit gauges. (De Bruin, 1973).

Unfortunately, it appeared that the daily records of about 30% of the stations were inadequate (Yperlaan, 1974), so only the data of 42 stations could be used for testing equation (10) as far as daily totals were considered. For 10-day totals all 65 stations were



Figs. 10–12. Comparison between theoretical and experimental values of Z (%) as found, with a dense network of rain gauges in Salland for:
an area of ≈ 1000 km² and a time interval of 10 days (Fig. 10);
an area of ≈ 1000 km² and 1 day (Fig. 11);
and ≈ 1 and ≈ 100 km² and 1 day (Fig. 12).

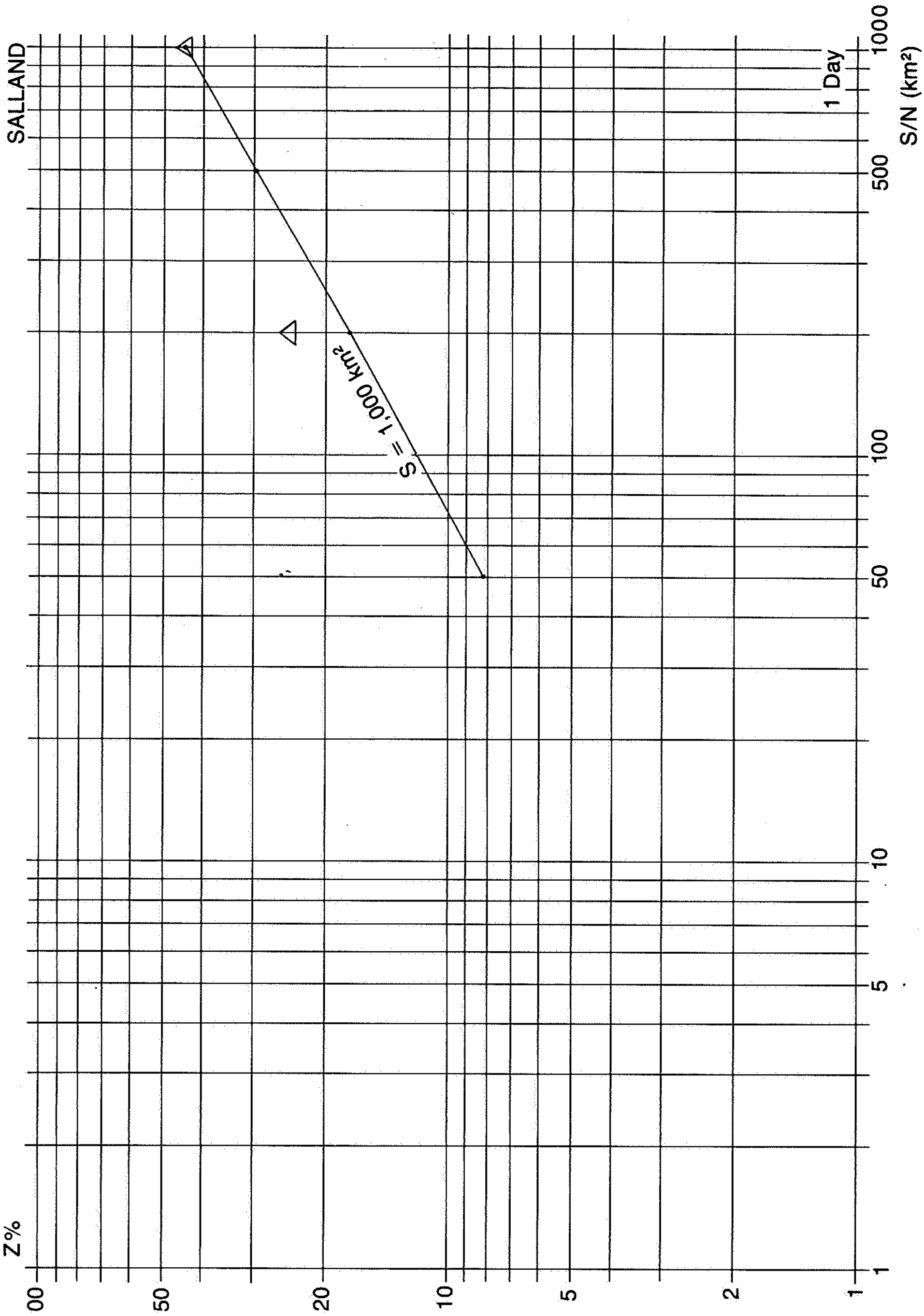

SALLAND
Z%
100
50
20
10
5
2
1
S/N (km²)
1000
500
100
50
10
5
1
1 Day
S = 1,000 km²

Fig. 11

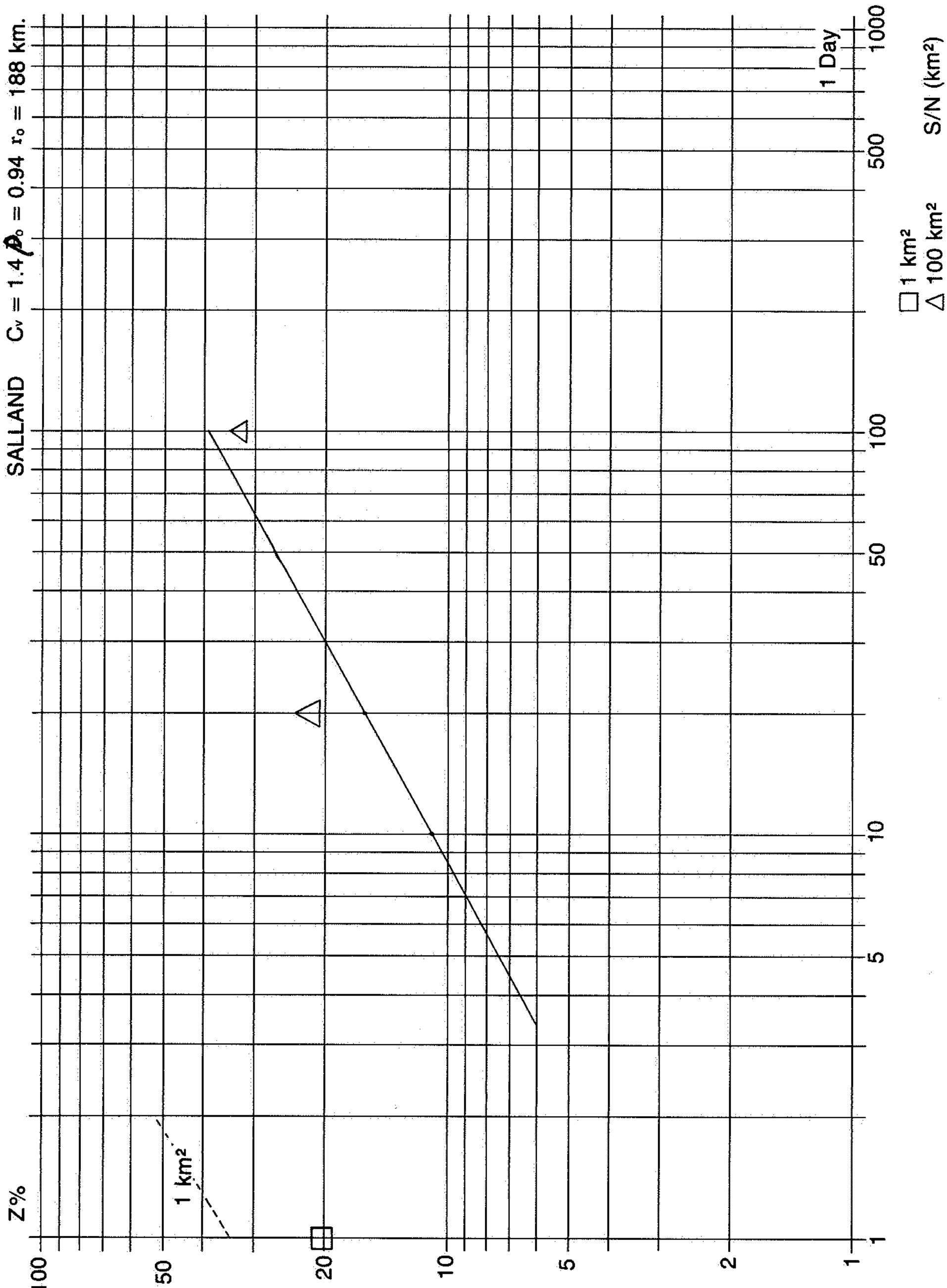

SALLAND
$C_v = 1.4$ $\rho_o = 0.94$ $r_o = 188$ km.
Z%
1 km²
1 Day
S/N (km²)
□ 1 km²
△ 100 km²
100
50
20
10
5
2
1
1
5
10
50
100
500
1000

Fig. 12

used. The arithmetic means of the records of these data were taken as the true areal precipitation depth.

The areal precipitation depths of 72 10-day periods were estimated with 1, 2 and 5 stations of the national rain gauge network, respectively, each covering about 1000, 500 and 200 km^2, respectively. As the five stations were not evenly distributed over the area, the "Thiessen average" was taken as an estimate. These estimates were compared with the "true" precipitation depth obtained with the dense network. The respective values of Z were calculated; they are plotted in Figure 10. In this figure, also the theoretical line given by equation (10) is drawn. It is seen that there is fairly good agreement between experimental and theoretical values of Z. It can be shown that (10) underestimates Z, if the stations are not evenly distributed. This explains the rather big deviation of the experimental Z-value for $S/N = 200$ km^2.

The same procedure was followed for daily totals. The results for 1000, 100 and 1 km^2 are given in Figures 11 and 12. Again, there is agreement although with a tendency that (10) underestimates Z.

Unfortunately, the procedure under consideration only gives a few examples for testing equation (10); it is, therefore, dangerous to draw definite conclusions.

Finally, similar experiments are being considered, carried out by Kruizinga and Yperlaan, 1976 with the records of 35 stations situated in the province of Zuid Holland, in the western part of the Netherlands. The total area was about 4000 km^2.

The mean daily precipitation depth over this area was estimated with the daily records of 4 stations, and the average over a sub-area of about 400 km^2 was estimated with the measurements at 1, 3 and 4 points, respectively. In the case of four points, the stations were situated outside the area.

Kruizinga and Yperlaan did not use quantity Z as a measure of error of estimate. They took the mean of the absolute value of the differences between the estimate and the true areal rainfall depth. So, in the first instance, their results are not comparable with equation (10). However, experiments with the records of the "Salland network" reveal that the "mean absolute difference" error is about 0.5 times the "root mean square" error, from which Z is derived. This is partly due to the sensitiveness of Z to outliers.

Using this conversion factor, we plotted the results of Kruizinga and Yperlaan in Figure 13, where also the theoretical lines for 400 and 4000 km^2

It should be noted that Kruizinga and Yperlaan described the error of estimates as a function of the "true" areal rainfall depth and the four seasons, while dry days were also concerned. The values given in Figure 13 are recalculated for rainy days. No distinction is made between the rainfall depth and the seasons.

Again, it can be concluded that equation (10) fits fairly well to the experimental values of Z. However, we should bear in mind that the experimental values of Z for this example were obtained in a rather crude way.

ACKNOWLEDGEMENT

The author is greatly indebted to Mr. S. Kruizinga for valuable advice.

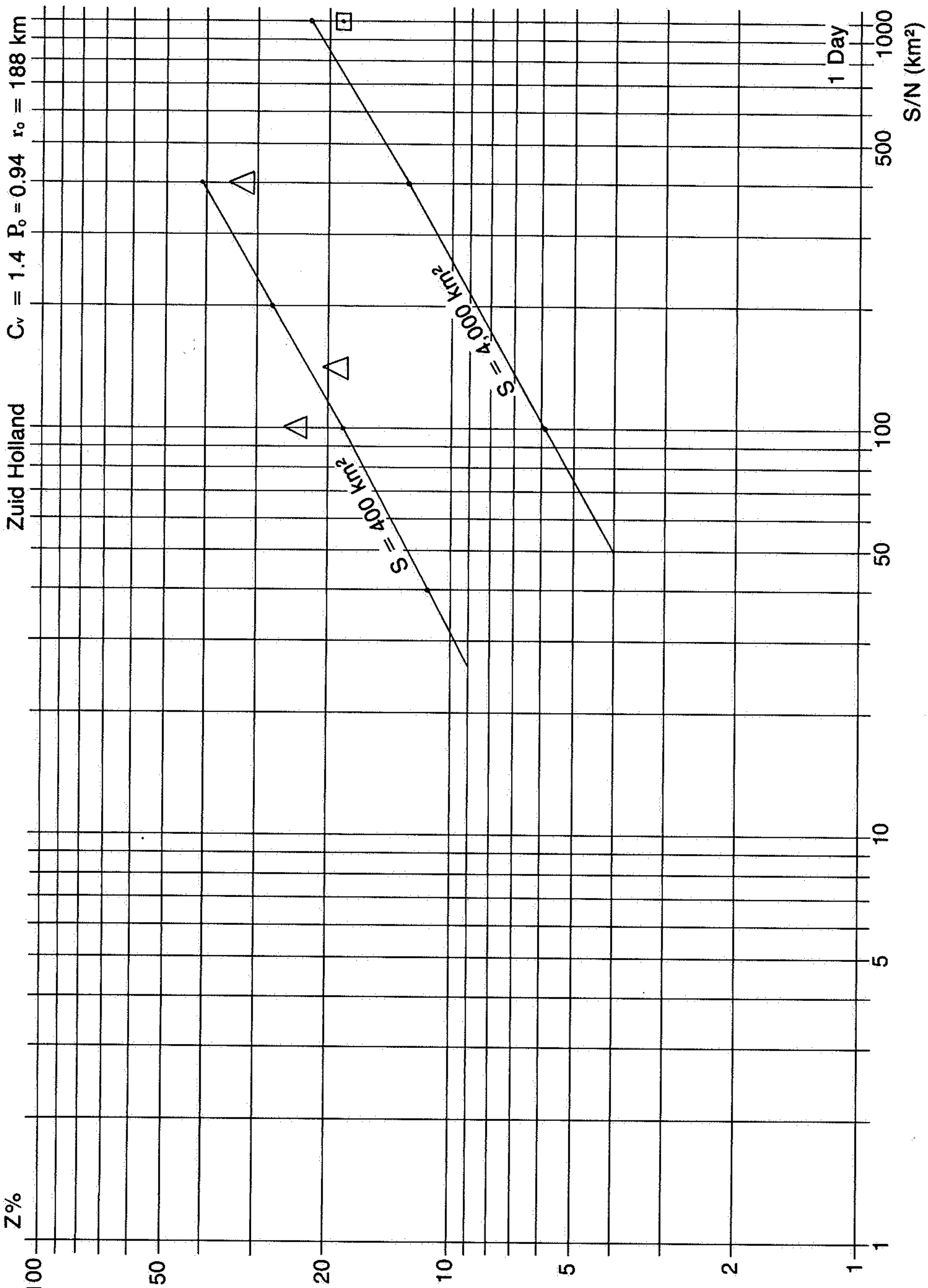


Fig. 13. Comparison between theoretical and the experimental values of Z (%) as found by Kruizinga and Yperlaan in Zuid-Holland ($\approx$ 400 and $\approx$ 4000 km^2, 1 day).

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APPENDIX I

ERRORS DUE TO THE "RANDOM CHARACTER OF RAIN"

When the precipitation depth, h, is measured with a rain gauge, a number of raindrops, N, with different diameters is caught during a certain time, T.

If we were able to repeat the experiment, it would be expected that we would not find the same number of drops and the same dropsize distribution. This indefiniteness in both the number of drops and the size distribution springs from the random nature of the rain-producing process.

We can write:

$$h = \frac{\pi}{6} \frac{N \widetilde{D}^3}{0} \tag{I.1}$$

where $\widetilde{D}^3$ is the average value of the third power of the dropdiameters, and O is the area of the orifice of the rain gauge. If we could repeat the measurement of h many times, we would get the average value of h, N and $\widetilde{D}^3$, denoted by respectively $\overline{h}$, $\overline{N}$ and $\overline{\widetilde{D}^3}$.

Our problem is to find the random fluctuations of h, N and $\widetilde{D}^3$ round their means.

A measure for these functions is the standard deviation σ defined by:

$$\sigma^2(x) = \overline{(x - \overline{x})^2} = \overline{x^2} - \overline{x}^2$$

First, we remark that it is reasonable to assume that the random fluctuations of N and $\widetilde{D}^3$ are independent, so we can write for small σ (N) and σ ($\widetilde{D}^3$):

$$\frac{\sigma(h)}{\overline{h}} = \sqrt{\left(\frac{\sigma(N)}{\overline{N}}\right)^2 + \left(\frac{\sigma(\widetilde{D}^3)}{\overline{\widetilde{D}^3}}\right)^2} \tag{I.2}$$

Further, we note that the standard deviation of the mean of P samples of a stochastic variable is equal to the standard deviation of the variable itself, divided by $\sqrt{P}$. So we can estimate $\sigma(\widetilde{D}^3)$ with

$$\sigma^2(\widetilde{D}^3) = \frac{\sigma^2(D^3)}{\overline{N}} \approx \frac{\overline{\widetilde{D}^6} - \overline{(\widetilde{D}^3)^2}}{\overline{N}}$$

This results into:

$$\left(\frac{\sigma(\widetilde{D}^3)}{\overline{\widetilde{D}^3}}\right)^2 = \frac{1}{\overline{N}} \frac{\overline{\widetilde{D}^6}}{(\overline{\widetilde{D}^3})^2} - \frac{1}{\overline{N}} \tag{I.3}$$

The problem to determine σ (N) remains. We, therefore, divide the time interval, T, into M equal time intervals, where $M >> \overline{N}$. If it is assumed that two raindrops never reach the orifice simultaneously, which makes it possible to choose M in such a manner, that never more than one drop is caught during an interval; the chance that during any interval a raindrop is caught is $\overline{N}/M$. The chance that during any given interval no drop is caught is $(1-\overline{N}/M)$. It can then be proved that the probability that during N intervals a raindrop is caught can be written as:

$$P(N) = \frac{M!}{N!\,(M-N)!}\left(\frac{\overline{N}}{M}\right)^{N}\left(1-\frac{\overline{N}}{M}\right)^{M-N} \tag{I.4}$$

For large N, this equation transforms into:

$$P(N) = \frac{\overline{N}^{N}\,e^{-\overline{N}}}{N!} \tag{I.5}$$

This is the Poisson distribution, for which

$$\sigma(N) = \sqrt{\overline{N}} \tag{I.6}$$

From (I.2), (I.3) and (I.6) we obtain:

$$\frac{\sigma(h)}{\overline{h}} = \sqrt{\frac{1}{\overline{N}}\,\frac{\overline{\widetilde{D}^6}}{(\overline{\widetilde{D}^3})^2}} \tag{I.7}$$

If we have only 1 sample $\frac{\sigma(h)}{\overline{h}}$ is of course estimated with $\sqrt{\frac{1}{N}\,\frac{\widetilde{D}^6}{(\widetilde{D}^3)^2}}$

APPENDIX II

DERIVATION OF EQUATIONS (7) AND (10)

It is assumed that the region of interest is square-shaped and has an area S, while at its centre, P, a rain gauge is installed. Let h_0 be the precipitation depth measured at P during a certain time interval, T. The mean precipitation depth over S is denoted by h_G.

Further, it is assumed that the area is climatologically homogeneous and isotrope.

Because of these assumptions, a reasonable estimate of h_G will be:

$$\hat{h}_G = h_0 \tag{II.1}$$

Now let us imagine that S is covered by a dense and regular network of ideal rain gauges. The precipitation depths obtained at these stations are denoted by $h_1^*, h_2^* \ldots, h_N^*$.

We can write:

$$h_G = \frac{1}{N} \sum_{i=1}^{N} h_i^*$$

In fact:

$$h_G = \lim_{N \to \infty} \frac{1}{N} \sum_{i=1}^{N} h_i^* \tag{II.2}$$

A measure of accuracy of the estimate given by equation (II.1) is the quantity defined by

$$E^2 = \overline{(h_G - \hat{h}_G)^2} \tag{II.3}$$

The bar denotes an average over a large series of h_G and $\hat{h}_G$.
From equations (II.1), (II.2) and (II.3), we obtain:

$$E^2 = \lim_{N \to \infty} \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \left(\overline{h_i^* h_j^*} - \overline{h_i^* h_0} - \overline{h_j^* h_0} + \overline{h_0^2} \right) \tag{II.4}$$

The measured precipitation depth, h_0, can be written as

$$h_0 = h_0^* + \delta \tag{II.5}$$

where h_0^* is the true precipitation depth at P, and δ is a random measuring error with $\overline{\delta} = 0$. (For the sake of simplicity it is assumed that h_0 has been corrected for systematic errors such as those due to wind).

Further, it is reasonable to assume that:

$$\overline{h_i^* \delta} = 0 \quad (i = 0, 1, 2 \ldots N) \tag{II.6}$$

The Quantity $\overline{\delta^2}$ is the standard error of observations at P.
From equations (II.4), (II.5) and (II.6) we get:

$$E^2 = \lim_{N \to \infty} \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \left(\overline{h_i^* h_j^*} - \overline{h_i^* h_0^*} - \overline{h_i^* h_0^*} + \overline{h_0^{*2}} + \overline{\delta^2} \right) \tag{II.7}$$

As the homogeneity of S has been assumed, we can write:

$$\overline{h_0^*} = \overline{h_i^*} \equiv x \quad \text{and} \quad \left(\overline{h_0^*}\right)^2 - \overline{h_0^{*2}} = \left(\overline{h_i^*}\right)^2 - \overline{h_i^{*2}} \equiv \sigma^2 \tag{II.8}$$

$(i = 1, \ldots N)$.

Thus the correlation coefficient

$$\rho_{ij}^* = \frac{\overline{h_i^* h_j^*} - \overline{h_i^*}\,\overline{h_j^*}}{\sigma_i \sigma_j} \tag{II.9}$$

is equal to

$$\rho^*_{ij} = \frac{\overline{h^*_i h^*_j} - x^2}{\sigma^2} \tag{II.10}$$

So we obtain:

$$E^2 = \lim_{N\to\infty} \frac{\sigma^2}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \left(\rho^*_{ij} - \rho^*_{i0} - \rho^*_{j0} + 1 + \Delta^2 \right) \tag{II.11}$$

where

$$\Delta^2 = \frac{\overline{\delta^2}}{\sigma^2} \tag{II.12}$$

As we have seen, the correlation coefficient can be described by equation (4):

$$\rho\,(r) = \rho_0\, e^{-r/r_0}$$

It is noted that this relationship has been found for the correlation coefficients computed with the series of measured precipitation depths, while ρ^*_{ij} stands for the correlation between two series of true precipitation.

If, analogously to (II.5), we write:

$$h_i = h^*_i + \delta_i, \quad h_i = h^*_j + \delta_j \tag{II.13}$$

then it can easily be shown that:

$$\rho_{ij} = \rho^*_{ij} \frac{1}{\Delta^2 + 1} \; (i \neq j)$$

If it is assumed that:

$$\overline{\delta_i \delta_j} = 0, \; \overline{\delta_i} = \overline{\delta_j} = 0, \; \overline{\delta_i h^*_j} = 0, \; \overline{\delta^2_i} = \overline{\delta^2_j} = \overline{\delta^2}, \text{ while } \overline{\delta^2}/\sigma^2 = \Delta^2$$

the "true" correlation function, ρ^*_{ij}, will reach the value 1 for $r \to 0$.*) Thus, combining equations (4) and (II.13), we find:

$$\rho_0 = \frac{1}{\Delta^2 + 1} \tag{II.14}$$

and

$$\rho^*_{ij} = e^{-r_{ij}/r_0} \tag{II.15}$$

If $\Delta \lesssim 0.25$, then $\frac{1}{\Delta^2 + 1} \approx 1 - \Delta^2$, so $\Delta^2 \approx 1 - \rho_0$ (II.16)

*) assuming that ρ_0 is independent of r_{ij}.

Substituting (II.15) and (II.16) into (II.11) we get:

$$E^2 = \sigma^2 \left[(1-\rho_0) + \lim_{N\to\infty} \frac{1}{N^2} \sum_{i=1}^{N} \sum_{i=1}^{N} \quad e^{-r_{ij}/r_0} - e^{-r_{io}/r_0} - e^{-r_{jo}/r_0} + 1 \right] \tag{II.17}$$

With the aid of a computer, the term

$$\lim_{N\to\infty} \frac{1}{N^2} \sum_{i=1}^{N} \sum_{j=1}^{N} \left(e^{-r_{ij}/r_0} - e^{-r_{io}/r_0} - e^{-r_{jo}/r_0} + 1 \right)$$

was determined for a square area S (see Kagan, 1972).

In this way

$$E^2 = \sigma^2 \left[(1-\rho_0) + \frac{0.23}{r_0} \sqrt{S} \right] \tag{II.18}$$

is obtained.

Next, we consider the situation that n rain gauges are evenly distributed over S. The precipitation depths recorded at these points are denoted by $h_1, h_2, ..., h_n$.

Each rain gauge covers a subarea S/n.

The areal precipitation depths of this subarea is now estimated with the precipitation depth recorded at the station belonging to it.

That means that:

$$\hat{h}_G = \frac{1}{n} \sum_{i=1}^{n} h_i \tag{II.19}$$

The standard error of estimation for the i^{th} subarea is given by (II.18):

$$E_i = \sqrt{\sigma^2 \left[(1-\rho_0) + \frac{0.23}{r_0} \sqrt{\frac{S}{n}} \right]} \tag{II.20}$$

Assuming that the errors of estimate for the various subareas are independent, we can write for the standard error of the estimate given by (II.19):

$$E = \sqrt{\frac{1}{n} E_i^2} = \sqrt{\frac{1}{n} \sigma^2 \left[(1-\rho_0) + \frac{0.23}{r_0} \sqrt{\frac{S}{n}} \right]} \tag{II.21}$$

THE PRINCIPLES OF MEASUREMENT OF PRECIPITATION WITH RADAR: A SHORT INTRODUCTION

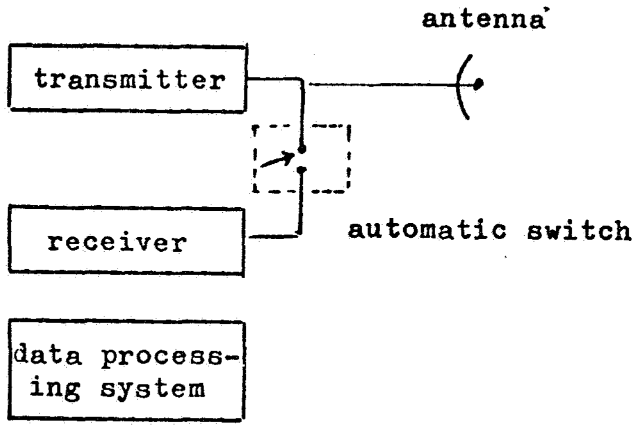
H.A.R. DE BRUIN
Royal Netherlands Meteorological Institute, De Bilt

1. SYNOPSIS

It is to be expected that the technical meeting on radar measurements of precipitation organized by the Committee for Hydrological Research TNO will be attended by hydrologists and others concerned with water management and water supply, who are generally not acquainted with the principles of radar and its application for precipitation measurements. Therefore, the basic physical formulae and their background are briefly reviewed.

2. QUALITATIVE OUTLINE

Radar is based on the principle that an electromagnetic wave propagages along a straight line through the atmosphere with a velocity equal to light speed ($\approx 3.10^8$ m/s). Objects such as raindrops will scatter and reflect electromagnetic energy. When an instrument is constructed which is able to transmit an electromagnetic pulse and which can also intercept reflected signals, one can measure the time interval between the transmission of the pulse and the reception of the reflected signal and thus determine the distance to the object. Measuring the intensity of the reflected pulse will yield information of the object itself. Such an instrument is called 'radar', when it transmits radio waves (with wavelengths of the order of cm – m).



Radar (see Fig. 1) essentially consists of a transmitter which produces the power at the radio frequency, a parabolic antenna which radiates the power in a narrow beam, an antenna which intercepts the reflected signals, a receiver which detects and amplifies the intercepted signals, and an automatic switch to close the receiver during the short interval when the transmitter is operating.

Finally, a radar is equiped with a system which processes the returned signals. Often a cathode ray tube is used, together with an oscilloscope, to visualize the received power. Today also computer systems are "interfaced" to the radar set.

It is noted that in practice a single antenna is used for both transmission and reception.

3. BACKSCATTERING OF A SINGLE SMALL SPHERICAL PRECIPITATION PARTICLE

Let W_0, U_0 and λ respectively be the power per unit area, the amplitude and the wavelength of the transmitted radio wave. This wave is scattered in all directions by a spherical precipitation particle at a distance r from the radar set and with a diameter $D \ll \lambda$. The amplitude, U_b, of the backscattered wave intercepted by the radar antenna can qualitatively be derived by the following dimensional analysis.

The incident wave will induce an electric polarization of all molecules of the particle, which varies periodically in time with the same frequency $\nu (= c/\lambda$, c = light speed) as the incident wave. All these induced vibrating dipoles in turn emit secundary waves in all directions. As $D \ll \lambda$, the differences in phase of the secundary waves intercepted by the antenna are very small. Thus the amplitudes intensify one another. This means that U_b is proportional to the number of molecules and thus to volume V of the particle; so $U_b \sim D^3$.

The remaining quantities involved with the problem are U_0, λ, r and the optical properties of the particle described by the (complex) refraction index, N. Obviously, $U_b \sim U_0/r$ (the intensity $I_b \sim U_b^2$ is proportional to $1/r^2$).

These arguments give:

$$U_b \sim \frac{U_0}{r} \cdot V \cdot f(N) \cdot g(\lambda) \tag{1}$$

where f(N) and g(λ) are, respectively, functions of N and λ.

The dimension of $\frac{V}{r}$ is [length^2]. As N is dimensionless, g(λ) must have the dimension of [length^{-2}], so we obtain:

$$g(\lambda) \sim \frac{1}{\lambda^2} \tag{2}$$

This gives:

$$U_b \sim \frac{U_0 V}{r \lambda^2} f(N) \sim \frac{U_0 D^3}{r \lambda^2} f(N) \tag{3}$$

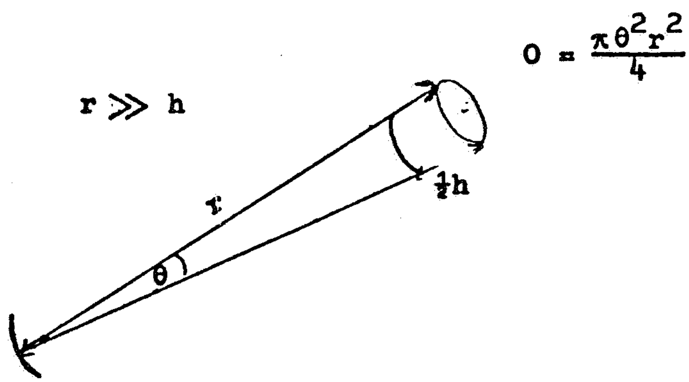
The intensity of the scattered wave intercepted by the antenna is proportional to U_b^2, so we can write (as $U_0^2 \sim W_0$):

$$W_b \sim \frac{W_0 D^6}{r^2 \lambda^4} f^2 (N) \tag{4}$$

where W_b is the power per unit of area intercepted by the antenna. Equation (4) is the well-known scattering law of Rayleigh.

4. BACKSCATTERING OF A NUMBER OF PARTICLES

When one uses radar for precipitation measurements, the radar beam illuminates a large group of raindrops (snowflakes, hailstones, etc.). So the backscattering of a large number of particles has to be considered. A radar pulse covers, during "pulse time" T, a distance h (the "pulse length"). The power backscattered by particles at a range (r + h/2) from the front of the pulse of length h arrives at the antenna, at the same time as the power backscattered by particles at a range r from the rear of the pulse. At a certain time t, the antenna receives signals from particles within a volume that is equal to about $\frac{\pi\theta^2 r^2 h}{8}$ (see Fig. 2) where θ is the "beam-width".



Assuming that the particles are distributed at random, the total power per unit of area received by the antenna can be written as:

$$W_b \sim \frac{W_0}{r^2 \lambda^4} f^2 (N) \, \theta^2 r^2 h \sum_{vol} D_i^6 \tag{5}$$

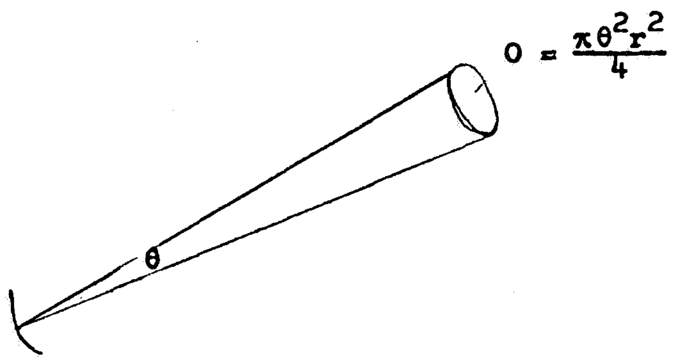
where $\sum_{vol} D_i^6$ is the sum of the 6th power of the diameter of the particle within a unit of volume. $Z = \sum_{vol} D_i^6$ is often called the 'reflectivity factor'.

It has been found that the backscattered signal varies from one reflected pulse to the next. This is caused by movements of the particles with respect to one another. Therefore, in practice the averaged power per unit of area $\overline{W}_b$ of a number of reflected pulses is detected.

5. THE RADAR EQUATION

Until now the averaged reflected power per unit area $\overline{W}_b$ intercepted by the antenna has been described as a function of W_0, the power per unit of area intercepted by the scattering precipitation particles.

The power, P_t, transmitted by the radar set is concentrated in a narrow beam of width θ. Particles at a range r along the axis of this beam will receive a power per unit of area W_0, which is proportional to P_t/O (see Fig. 3).



Obviously, $O \sim \theta^2 r^2$. So we obtain, if the radar beam is entirely filled with precipitation particles:

$$W_0 \sim P_t/\theta^2 r^2 \tag{6}$$

Substituting (6) in (5), we can write:

$$\overline{W}_b \sim \frac{P_t h}{r^2 \lambda^4} f^2(N)Z \tag{7}$$

The total average power, $\overline{P}_r$, intercepted by the antenna is equal to $\overline{W}_b \cdot A_e$, where A_e is the effective area of the antenna. It is found that $A_e \approx \frac{2}{3} A$ (A is the actual area of the antenna), if the antenna is circular.

This gives:

$$\overline{P}_r \sim \frac{P_t A h}{r^2 \lambda^4} f^2(N) Z \tag{8}$$

It can be shown that $f^2(N) = \left| \frac{N^2 - 1}{N^2 + 2} \right|^2$ (9)

It should be noted that $f^2(N)$ may be taken as 0.93 and 0.20 for water and ice particles, respectively.

From (8) and (9) we obtain the "radar equation":

$$\overline{P}_r = C \frac{P_t A h}{r^2 \lambda^4} \left| \frac{N^2 - 1}{N^2 + 2} \right|^2 Z \tag{10}$$

where C is a constant (which is about 50).

This equation can be written as:

$$\overline{P}_r = C_1 \cdot C_2 \frac{Z}{r^2} \tag{11}$$

where C_1 is a parameter that depends on the characteristics of the radar set, and C_2 is related to the optical properties of the precipitation particles.

In the derivation of (11) no provision is made for attenuation caused by absorption and scattering of precipitation present between radar and observed volume.

Some authors (e.g. Collier) introduce an attenuation factor K; (11) can then be written as:

$$\overline{P}_r = C_1 C_2 K \frac{Z}{r^2} \tag{12}$$

This factor is introduced to describe the effect of attenuation of precipitation that occurs between the radar set and the observed volume.

6. SOME NUMERICAL EXAMPLES

In order to get an impression of the magnitude of the various quantities, Table 1 lists some characteristic values.

Table 1

Quantity	
transmitted power P_t	$10^4 - 10^6$ W
minimum detectable power P_{min}	$10^{-14} - 10^{-13}$ W
pulse time T	$0.2 - 4\ 10^{-6}$ s
pulse length h	60–1200 m
beam width θ	0.5–2 deg.
wavelength λ	0.03–0.1 m

Now it is supposed that we have a radar set with $P_t = 2.10^5$ W, $P_{min} = 2.10^{-13}$ W, $A = 1.5\ m^2$, $\lambda = 0.05$ m and $h = 900$ m, while we observe spherical waterdrops at a distance of 100 km.

Equation (10) can then be written as:

$$\overline{P}_r \approx 2.10^4\, Z \qquad \text{(W)} \tag{13}$$

The backscattered power intercepted by the radar antenna will be detected only if threshold value P_{min} is crossed.

So the waterdrops are "seen", if:

$$Z > 10^{-17} \qquad (m^6/m^3) \tag{14}$$

Typical values of drop sizes are 1 mm and 0.15 mm during rain and drizzle, respectively, whereas cloud droplets have diameters in the order of 3.10^{-6} m.

So condition (14) is fulfilled if within 1 m^3 at least 10 raindrops, 9.10^5 "drizzle drops" or about 10^{10} cloud droplets are present. For rain and drizzle these values are realistic, but the maximum concentration of cloud droplets is about 10^8 per m^3. So this example shows that with radar rain and drizzle are detectable, whereas clouds are not "seen".

7. SPECIAL PROBLEMS

Some factors limit the use of radar for precipitation measurements. We would mention four:

A. *Melting layer*

In one of the previous sections it has been seen that water and ice have different optical properties, resulting in different values of $\left|\frac{N^2-1}{N^2+2}\right|^2$ (0.93 and 0.20, respectively).

When a snowflake is melting, first a water-skin arises. At this stage, the flake is "seen" by the radar as a waterdrop with the size of the snowflake. This causes an increase of the radar echo. So, when the radar beam intersects the 0°C layer, a "bright band" is observed. This effect limits the accuracy of rainfall measurements.

B. *Attenuation*

It was noted before that rain present between the radar set and the volume of interest will attenuate the power. Especially with heavy rainfall, this problem is serious.

C. *Alteration of rainfall rate with height*

Radar detects rain above ground level; so when the rainfall rate alters between the radar beam and the ground, a wrong rainfall rate is observed.

D. *Anomalous propagation*

Under some atmospheric conditions, radio waves do not propagate along a straight line.

8. RELATION BETWEEN Z AND PRECIPITATION RATE, R

The last step missing is the relation between Z and precipitation rate R. This relationship depends on the shape of the distribution of the sizes of particles. This follows directly from:

$$R = \sum_{D=0}^{D_{max}} N_D M_D u_D \Delta D \tag{15}$$

where N_D is the number of drops per unit size range per unit volume of air in the class ΔD whose average diameter is D, M_D is the mass of the individual particles and u_D is the fall speed. (Obviously, $M_D = \frac{\pi}{6}\rho D^3$, ρ is the density).

In the same way, Z can be written as:

$$Z = \sum_{D=0}^{D_{max}} N_D D^6 \Delta D \tag{16}$$

Marshall and Palmer (1948) gave an approximation of N_D:

$$N_D = N_0 e^{-\Lambda D} \tag{17}$$

where Λ is a constant.

With the approximation of Gunn and Kinzer (1949) for the fall speed:

$$u_D \approx 130\sqrt{D} \tag{18}$$

(D in metres, u_D in m/s) we obtain, after elimination of Λ, if we take a typical value for $N_0 = 10^7 m^{-4}$:

$$Z \approx 210 R^{14/9}\ mm^6 \cdot m^{-3} \tag{19}$$

(R in mm/hr).

This analysis is too crude. Therefore, in practice, the empirical relationship

$$Z = A \cdot R^B \tag{20}$$

is used.

The problem of this equation is the large variability of its parameters A and B.

Several authors (see e.g. Hitschfeld and Bordan (1954) and Collier (1975) have shown that the accuracy of radar precipitation measurements will be improved if rain gauges are used to calibrate parameters A and B. This is, for instance, achieved in the Dee Project and in the experiments at Hohenpeissenberg. Both will be described during this meeting.

For more detailed information on the principles of radar, and its application for precipitation measuring, readers are referred to Battan (1973) and Kessler and Wilk (1968).

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AREAL PRECIPITATION MEASURED BY RADAR

W. ATTMANNSPACHER
Deutscher Wetterdienst, Meteorological Observatory at Hohenpeissenberg, FRG

1. INTRODUCTION

The measuring of precipitation seems to be the oldest meteorological facility on earth; from the literature it is known that about 400 B.C. the first rain measurements were done in India (Kautilya, 1915). About 2000 years later Wren built the first rain gauge in England (PLeiss, 1969). Most of the rain gauges used in Europe today were developed in the 1880's and 1890's (Hellmann, 1885 and Fuess, 1897). The necessity of higher accuracy has led to new designs and constructions of rain gauges in recent years, for example the Obrometer of Joss (1967) and that of the Observatory at Hohenpeissenberg (Attmann-spacher, 1973). Such an instrument provides suitable knowledge on local precipitation in terms of duration, amount and intensity at one point.

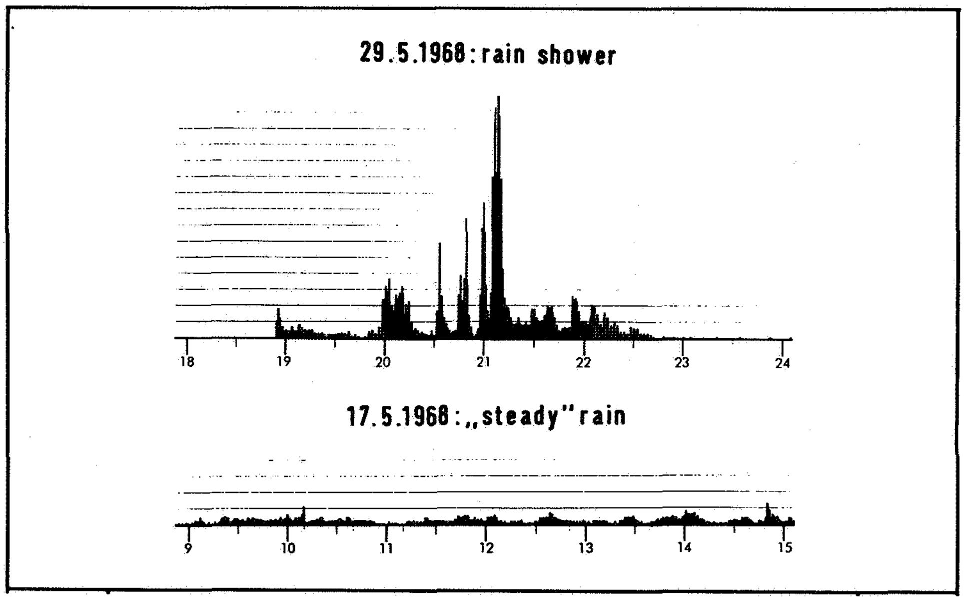

Fig. 1: Minute precipitation intensities during a rainshower (upper part) and during "steady" rain; Hohenpeissenberg, 1968.

An example of variation in precipitation intensities from minute to minute is shown in Fig. 1. Within a shower, the precipitation intensity changes from minute to minute by a factor of 10. For steady rain, the minute-to-minute differences reach a factor of 4. The areal variations are at least of the same order as those of time. Single measurements at one point, over a period of some years, will smooth the differences in time and area. But for many specific questions in hydrology, knowledge of the true areal precipitation for single cases is doubtless necessary. To be able to get sufficient knowledge of these areal precipitations, a dense rain-gauge network may be useful. Else an other facility for areal precipitation measurement will have to be found.

Fig. 2 shows the amount of precipitation in a dense rain-gauge network, which is installed north of the Hohenpeissenberg, and the results of areal precipitation measurement using radar. Even with the data of this dense network of about 50 rain-gauges operating in an area of 750 km^2, it is impossible to draw realistic isohyets; but they can be drawn easily if radar measurement is practised.

As early as 1904, the German Patent Office granted a patent to an engineer, Huelsmeyer (Brockhaus Lexikon, 1954) for a navigation instrument which received radio waves being reflected from a ship. About 30 years later, this idea was used in systems for radio navigation developed in several countries. Those systems were soon found to be disturbed in situations of bad weather. Meteorologists next studied the unwanted echos and, finally, radar became an important tool in meteorological science. For example, for measuring areal precipitation. As the literature reveals, more than 20 years ago the first steps in this field were taken by Americans (Battan, 1973) and Russians (Borovikov, 1967). Over the last 10 years scientists in other countries also started much work on radar used for hydrological measuring.

2. PRINCIPLE OF AREAL PRECIPITATION MEASUREMENT USING RADAR

The measuring of areal precipitation is a simple technique: a pulse of the radar equipment is sent out and the echos coming back from the precipitation, that is from the raindrops, are received and their intensity is measured. As long as the rain-drop diameters are much smaller than the radar wavelength (Ralleigh Approximation) based in the Mie Theory, the radar equation can be written as:

$$\overline{P}_r = \frac{P_t \cdot h \cdot A_e}{8\pi \cdot r^2} \cdot F \cdot K \cdot \Sigma\sigma \qquad (1)$$

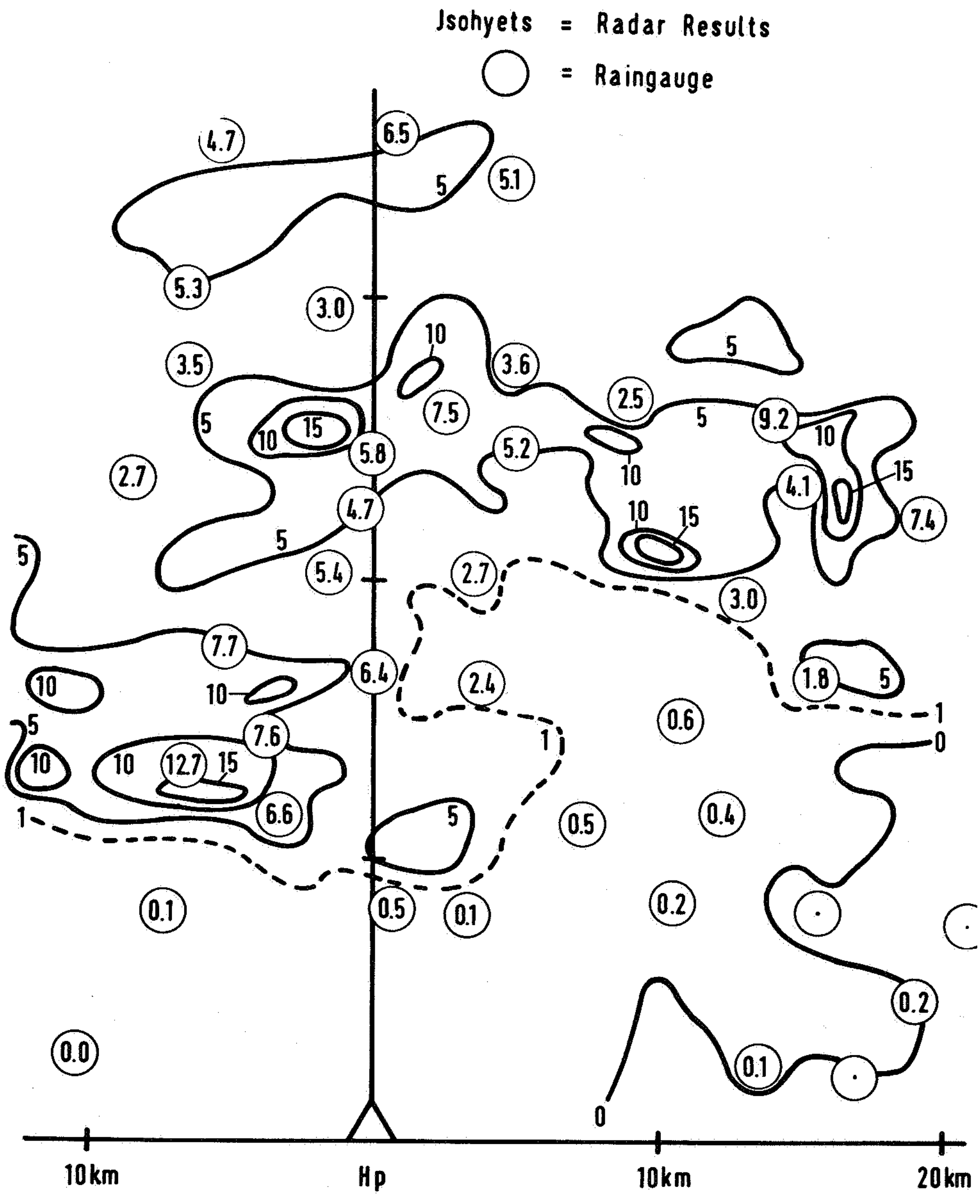

Fig. 2: Radar- and rain gauge results; Hohenpeissenberg, 1.9.1970, 13.35–17.30 MET.

$\overline{P}_r$ = mean received power
P_t = transmitted peak power
h = pulse length
A_e = effective aerial aperture
r = range
F = correction factor for partly filled radar volume
K = Correction factor for attenuation of radar power by the rain itself
σ = equivalent back scattering cross-section of all drops in unit volume of the beam.

The factors in this equation are functions of the radar equipment; or of the objects causing the echo, in other words: the raindrops.

To be able to get realistic measurements of areal precipitation, the electronic data of the radar device have to be as steady as possible. This can hardly be realised with old radars using electronic tubes only; normally it is much easier with fully transistorized radars with integrated circuits. Three more conditions must be satisfied:

- the radar must have a very sharp radiation beam, to make sure that the radar volume, that is the volume from which the echo is coming back, is always filled with precipitation;
- the areal precipitation should be measured close to the earth surface, so that the precipitation measured reaches the surface in the very spot where it is measured;
- the height of the freezing level, at which water-drops and ice particles co-exist, should be avoided; echos from mixed originators can hardly be used for precipitation measurements.

These three points call for a sharp vertical radar beam and a horizontal one. Fig. 3 shows a radar beam of one degree and another of two degrees, as a function of distance and altitude. It can be seen that the beam of one degree transmitted at a vertical angle of one degree already reaches an altitude of three kilometers at a distance of 100 km. Due to the earth curvature, the radar beam, which is nearly straight, will reach higher and higher areas with increasing distance. This means there exists a limit for realistic quantitative measurements of areal precipitation; it seems to be at about 100 to 120 km radius from the radar. However, qualitative measurements are possible beyond these distances. In practice, a radar beam of 1 degree can easily be realised for short waves. For the wavelength of 3 cm, today an aerial of 2.1 m is necessary; for 5 cm one of 3.5 m and for 10 cm one of 7 m diameter.

Another very important point that has to be considered is the attenuation of the radar energy by the precipitation itself; factor K in equation 1. As can be seen in Fig. 4, this attenuation reaches the highest values at 3 cm and the smallest at 10 cm wavelength. Theoretically it is possible to correct this attenuation but correction factor "K" is a function of the precipitation amount which is to be measured.

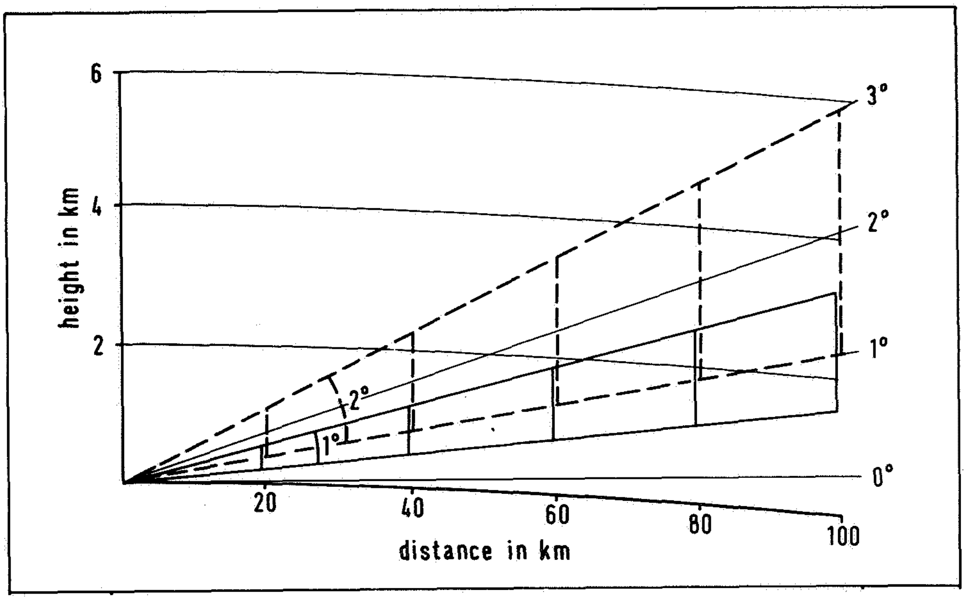

Fig. 3: Radar beam width of one and two degrees and distance.

$$K = 10^{-2} \int_0^r \frac{K}{10}\, dr : K = K_0 \cdot R^{\gamma} \tag{2}$$

$R = \text{mm/h}$

Factor K_0 respectively exponent γ, is mainly dependent on the wavelength of the radar device.

This precipitation amount also contains the electronic data of the instrument itself; cf equation (1). Theoretically, one may obtain a good correction factor as long as an echo is coming back; but an error of 1 dB may lead to useless correction (W.M.O., 1968). Based on these attenuations, the best equipment would be the 10 cm radar. But, as it was shown before, a radar beam of 1 degree can only be realised for a 10 cm wavelength through specialized engineering and at high financial cost. A 5 cm radar has an attenuation value of one sixth to one seventh of that of the 3 cm radar (see Fig. 4). Therefore this radar, the so-called C-band radar, seems to be a good compromise as regards meteorological needs and economic features.

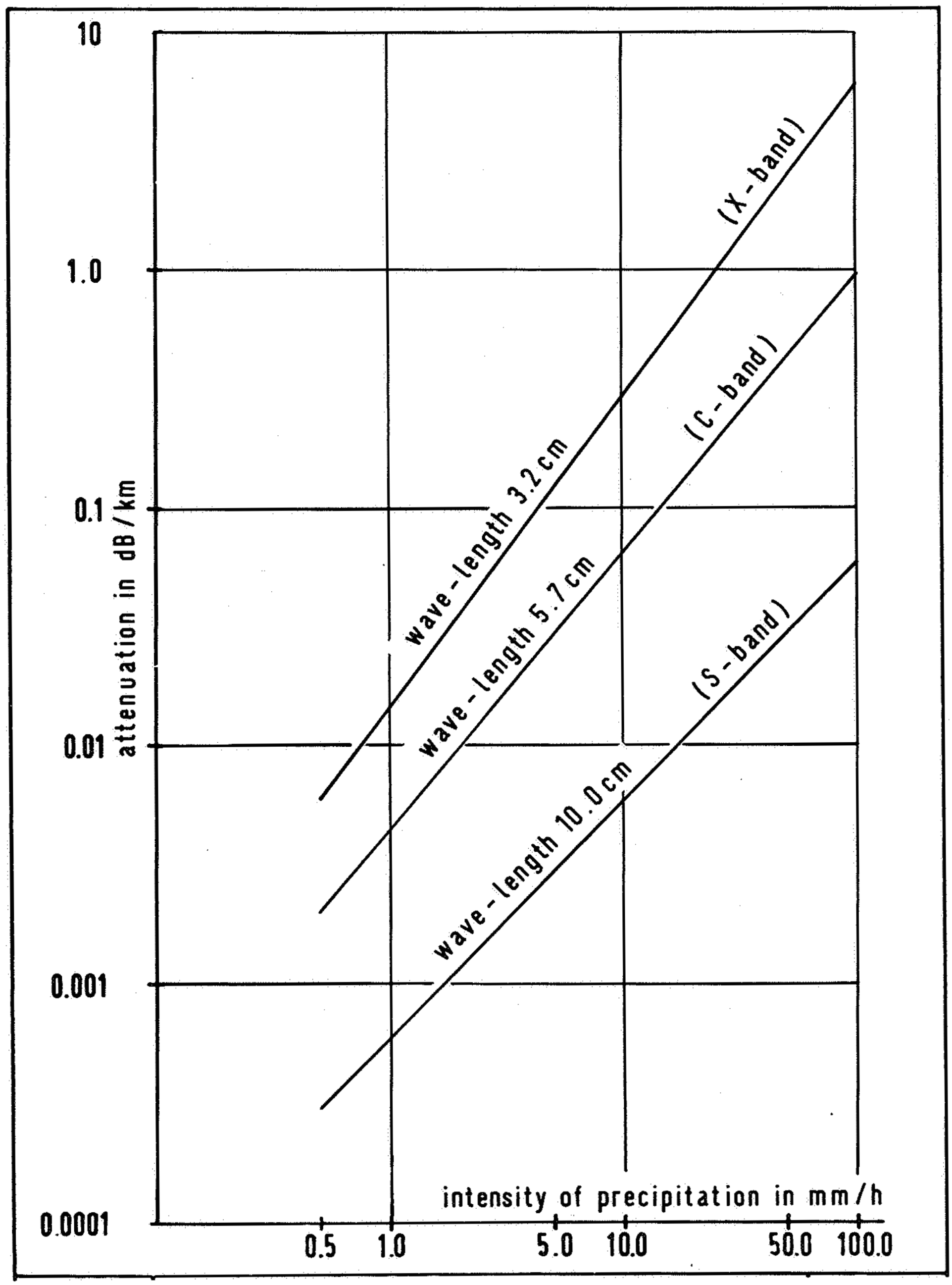

Fig. 4: Attenuation of radar energy by precipitation for 3 wave length.

The equivalent back-scattering cross section "σ", can be used to calculate the precipitation; it can be written as:

$$\Sigma\sigma = \frac{\pi^5}{\lambda^4}\left(\left|\frac{\epsilon - 1}{\epsilon + 2}\right|\right)^2 \Sigma n_i D_i^6 \tag{3}$$

The sum, $n_i D_i^6$, which is the so-called radar reflectivity (Z), is a function of the precipitation amount:

$$Z = \Sigma n_i D_i^6 = AR^b \tag{4}$$

A and b are factors, respectively exponents; R represents the rain amount in mm per hour. A and b are largely functions of the precipitation itself, that is of the meteorlogical processes.

The main meteorological problem in measuring areal precipitation amount by radar is shown in equation (4). In it, the 6th power of the diameter of the rain-drop sizes has to be known, but normally the rain-drop size distribution is unknown during the measurement operation. To avoid this difficulty, at the Hohenpeissenberg Observatory the so-called "Aneichmethode" (calibration method) was developed. First, a "rough precipitation amount" using a mean dropsize distribution is calculated. These rough data added up over the whole period of time that percipitation occurs at one rain-gauge point, are compared with the results of the rain-gauge measurements at this station. Dividing the rain-gauge amount by the rough radar value, we obtained a factor to be used for calibrating the whole radar measurement. Authors in other countries, for example J.W. Wilson (1970), also successfully calibrate their data in a similar way.

From all these results, an important statement can be derived: Normally the rain-drop size distribution in a single shower does not change much; otherwise these good results would have been impossible.

3. SOME EXAMPLES AF AREAL PRECIPITATION MEASURING

At the Hohenpeissenberg Observatory research was started in 1968 with an X-band radar. The project was sponsored by the Deutsche Forschungsgemeinschaft. As was shown in the previous chapter, a radar of 3.2 cm wavelength can in special cases be used only as long as there is some weak or no attenuation of the radar energy by the rain itself. Fig. 5 presents data on a situation of weak precipitation. The circles represent the precipitation amount measured by single rain-gauges, the isohyets are based on radar measurement; they were calculated by the "Aneichmethode". The results of the two measurement techniques are in good agreement.

With the results of our precipitation measurements for 1970, a correlation coefficient was calculated between the radar measurement at each of the 50 stations within the rain-gauge network over an area of about 750 km² and the rain-gauge measurement amount

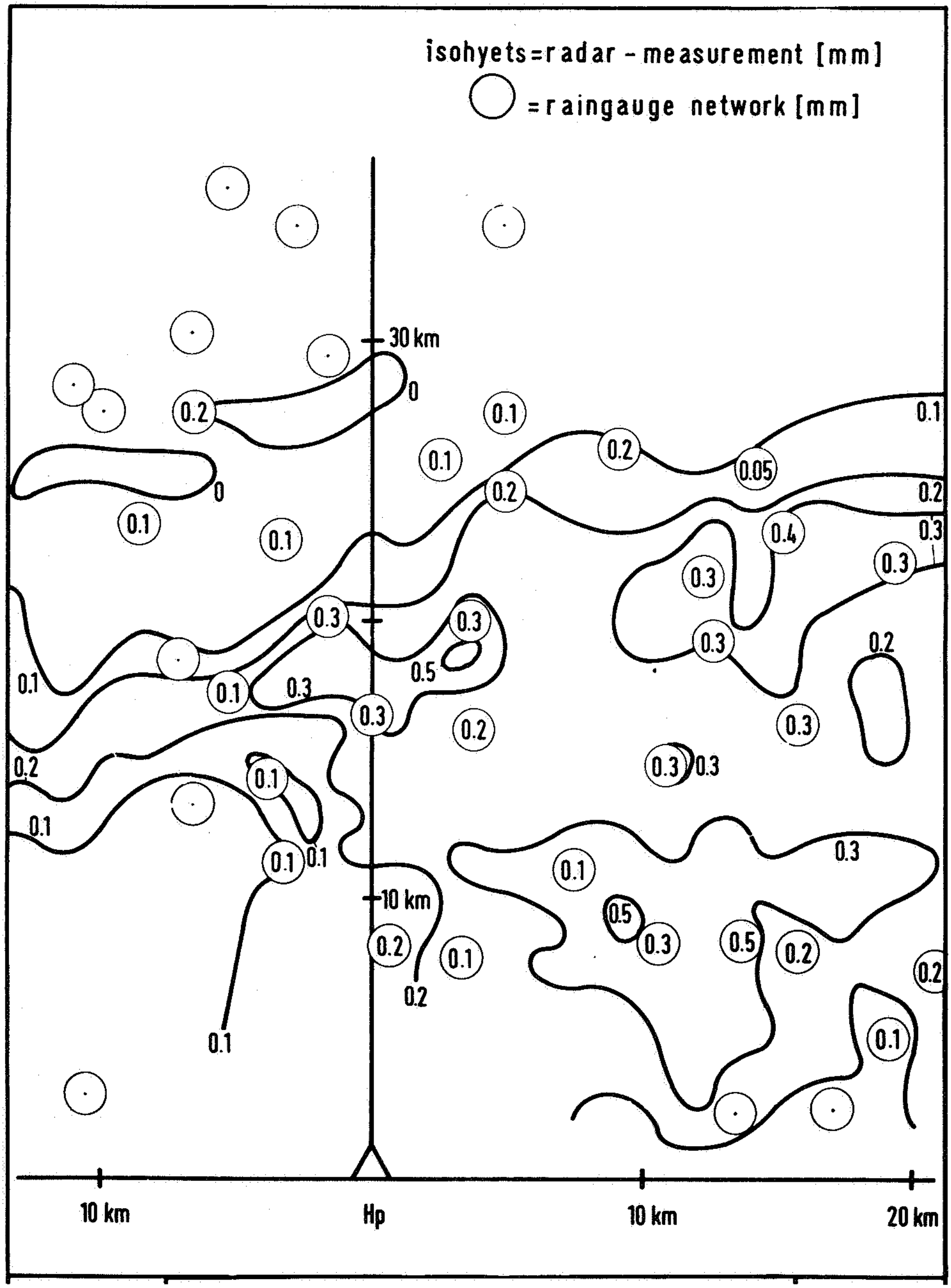

Fig. 5: Radar and rain gauge results; Hohenpeissenberg, 20.7.1970, 13.40–14.20 MET.

there. The correlation was better than 0.99. This coefficient may seem unexpectedly high. However, only separate cases with weak or no attenuation of precipitation itself were measured. Moreover, the Hohenpeissenberg is an ideal place for radar measurement; it is a single mountain overlooking a relatively flat area. Which means it is possible to measure between cloud level and surface. And thirdly the correlation coefficient does not give an answer to the problem of correspondence of absolute values of measurements. To check this correspondence, another statistical test had to be made. The precipitation amounts were classified by a logarithmic scale: smaller than 0.6, 0.10, 0.16, 0.25, 0.40 mm and so on; next, the amount of rain at each station based on radar measuring was compared with that of the rain-gauge. The result is given in Table 1.

Table 1. Differences of classified amount of rain measured by radar (Ra) and rain-gauges (Rg), as a function of distance (Hohenpeissenberg, 1970).

		Ra – Rg					
		2	1	0	–1	–2	–3
Distance from radar station (km)	> 30			2	3	4	5
	25–30			14	15	2	
	20–25		2	20	9		
	15–20	1	1	41	5		
	< 15	1	3	26			
	Σ	2	6	103	32	6	5

Up to 20 km, the two measurements are in good agreement. At stations more than 20 km away from the radar, the amounts of the rain-gauges are higher than those of the radar measurements. These differences are caused by attenuation, and we tried to correct it.

The results of the measurements shown in Fig. 6 could thus be improved. However the overall results of measurements in the last few years have confirmed our tentative conclusion that for a 3-cm radar attenuation corrections can be made that only display limited usefulness.

About one year ago, a new fully automatic 5.4-cm radar device was installed; it has a digitalised videa signal and direct connection to a computer. Fig. 7 compares the results measured by an X-band radar, a C-band radar and the rain-gauge network. The 3.2-cm radar data are corrected for attenuations, the 5.4-cm radar data are not. This figure confirms that measurements made by a C-band radar, at least in many cases, do not need an

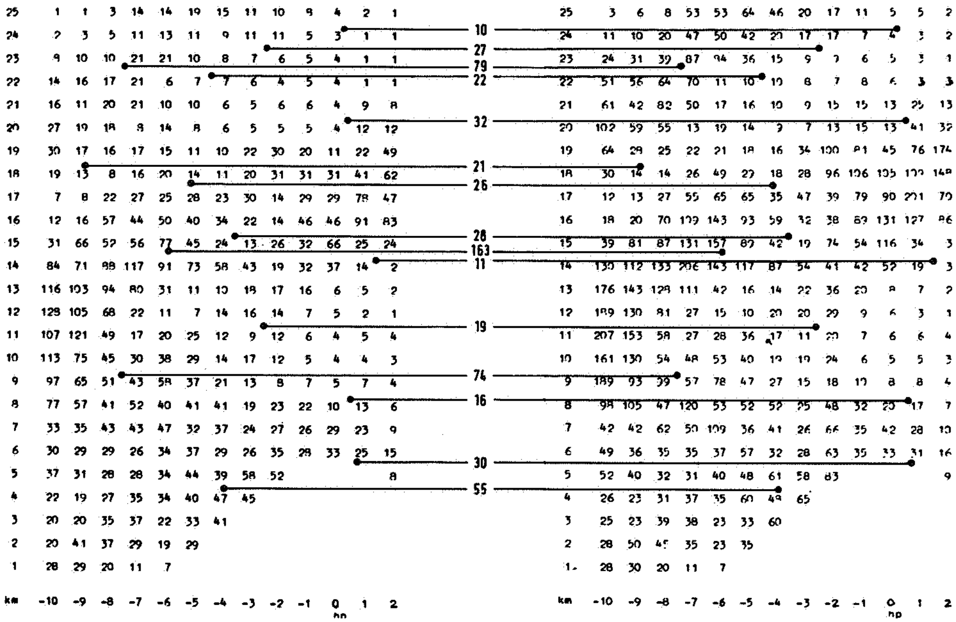

Fig. 6: Comparison of corrected (right) and uncorrected (left) measurement and rain gauge results; Hohenpeisenberg, 24.7.1971, 16.50–17.50 MET.

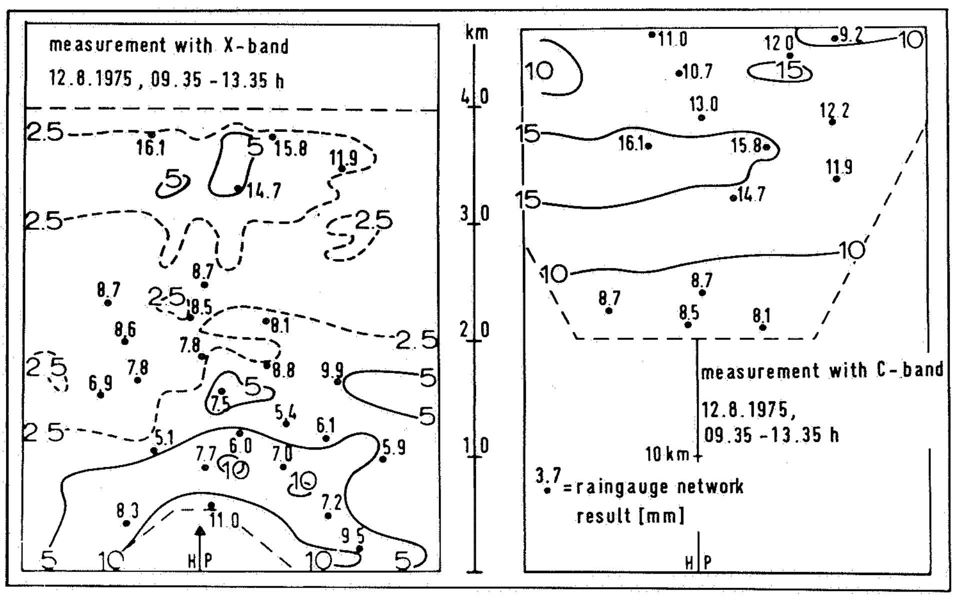

Fig. 7: Comparison of corrected X-band-radar- (left), uncorrected C-band-radar-measurements (right) and rain gauge results in mm; Hohenpeissenberg, 12.8.1975, 09.35–13.35 MET.

attenuation correction. For technical reasons, our C-band measuring starts further off the radar station than that of the X-band. Calculation of areal precipitation, using the latter facility, was found inadequate at distances more than 38 km.

In co-operation with the Hydrological Institute "Wasserbau III" of Karlsruhe University, some radar areal precipitation measurements were used as input for inflow forecasts on water reservoirs. One example is shown in Fig. 8 (Anderl and Schultz, 1975). The

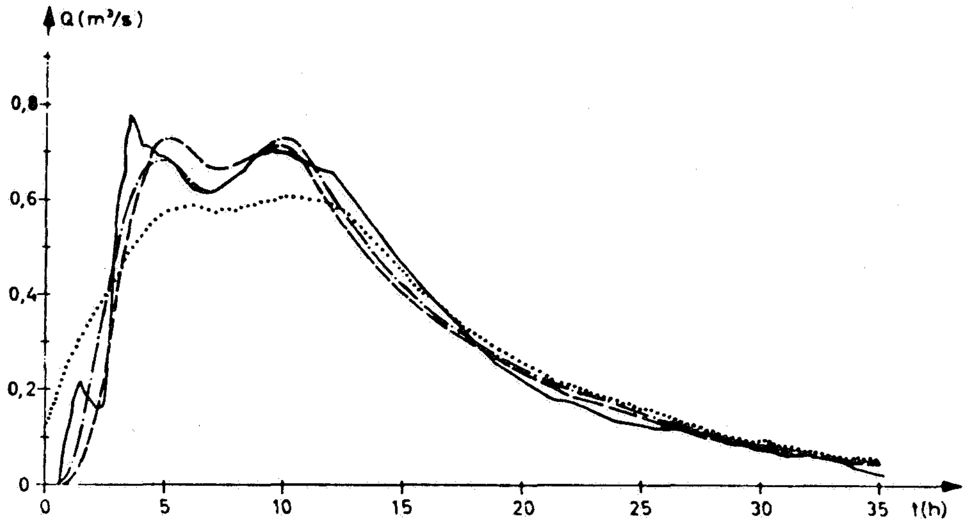

Fig. 8: Windach river flood of 28./29.7.1972.
observed hydrograph
computed hydrographs based on rainfall measurements by:
——— weather service rain gauge network
-.-.-.- special rain gauge network Hohenpeissenberg
.... radar

computed hydrography based on rainfall measuring by radar, fits best the observed hydrography. The computed hydrography based on the special network may fit the needs of hydrologists, too. The isohyets within the catchment area and the results of the rain-gauges are compared in Fig. 9, where it will be seen that in this special case the results of these rain-gauges fairly represent the mean value of precipitation within this small area. A bigger catchment area with local, high-density rain, outside a station, would have caused a poor result. The last curve, which is based on the normal official weather service network, is useless for this hydrological purpose.

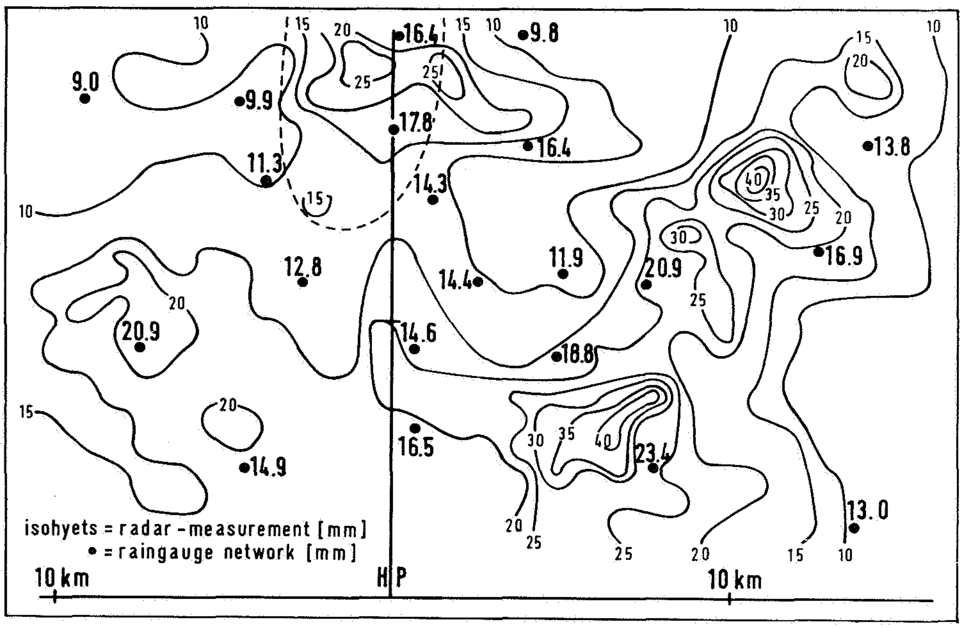

Fig. 9: Radar- and rain gauge network results; Hohenpeissenberg, 28./29.7.1972, 20.20–03.20 MET. (Windach catchment area inside of dashed line.)

4. USE OF RADAR FOR AREAL PRECIPITATION MEASURING IN HYDROLOGICAL PRACTICE

All results shown above have been obtained in a research program. At first measurements were mainly manual, (Attmannspacher and Aniol, 1971), later semi-automatic (Attmannspacher, Hartmannsgruber and Riedl, 1974) and finally fully automatic and there was a direct connection of the radar device to a computer.

Generally speaking radar should be used for measurement only if its economic benefits are higher than its cost. Modern radar devices are electronically stable enough for these measurements. Calculation of radar results, and of the rainfall run-off model, has to be made during the measurement operations and with the aid of a computer. Using the "Aneichmethode" it is generally not necessary to know the dropsize distribution, though some cases may show up where these distributions should be known. It may happen that the "Aneich-Ombrometer" value does not reach the computer station, or that really strong showers outside the "Aneich Ombrometer" place are measured; or that one has to correct the measurement because of attenuation of the radar energy. This correction has to be made with rough data for the first period of time. In these situations it will be useful to have the so-called rough precipitation measurement value as close as possible to the real value. There are several instruments (mechanical, optical and electronic systems) that measure the raindrop size distribution. The resulting data are highly different, and today there are still meteorologists who think there is no possibility to connect the rain-drop size distribution to meteorological conditions. In view of the statement at the end of chapter 2, it seemed useful to look for such relationships. Therefore, the rain-drop size distributions of the precipitations at Hohenpeissenberg over a period of 2 years were classified by Aniol (1975), using four simple meteorological classes:

- precipitation due to warm air advection;
- precipitation due to cold air advection and "Stau" situations;
- precipitation based on weak pressure gradients, and
- thunderstorms.

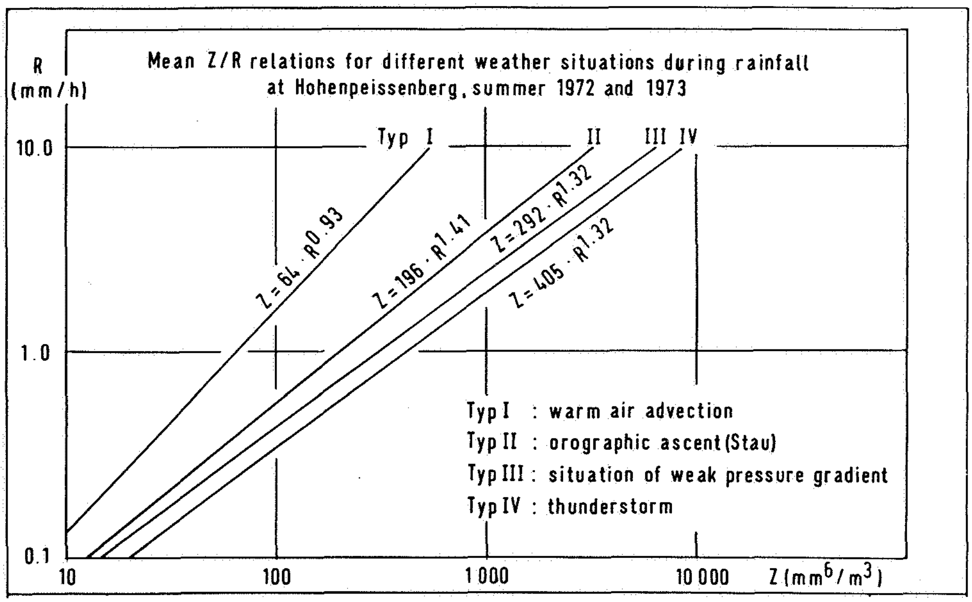

Fig. 10: Radarreflectivity Z and precipitation R for special weather types.

These weather types show significant differences in the A and b values of the Z/R relation (Fig. 10). The measurements of these rain-drop size distributions were made by an electronic instrument: the "Distrometer", as described by Joss and Waldvogel (1967). Such measurements only represent the situation at one point, or in the area were they have been made. However the rough data for this area can be improved by using typical Z/R relations for different weather situations.

The differences between calculated areal precipitation measurement data using a mean weighted Z/R relation, or weather type Z/R relations, are shown in Fig. 11; it is copied from an article published by Aniol and Riedl (1976). The mean weighted relation only

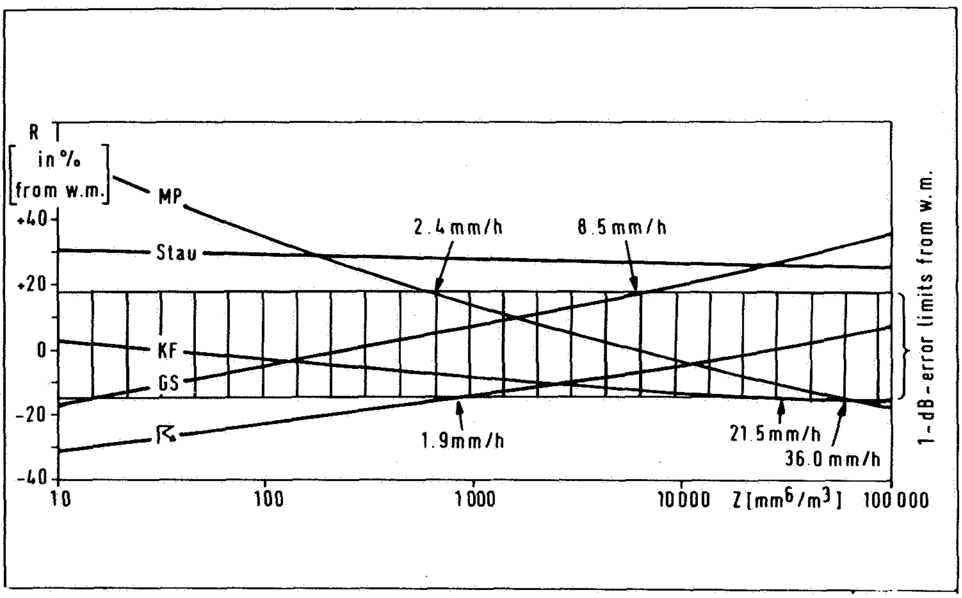

Fig. 11: Precipitation amount for weather type Z/R-relation in % of a weighted mean Z/R relation (w.m.).

may be "off" 40% with regard to the real value. These calculations are based on an electronic error of ± 1 dB; in practice this is the absolute minumum which may be reached.

Another tool to improve the radar areal precipitation measurement should be some type of areal precipitation statistics. Such statistics will be a matter for the hopefully near future; sufficient areal precipitation measurements are not available yet. Future statistics should help meteorologists to a sound basis for a forecast of the precipitation amount, in terms of at least some hours. This, anyway, seems to be very important for hydrological purposes.

I would emphasize, finally, that areal precipitation measuring cannot and should not replace precipitation measuring at one point. The two types of measuring together can help meteorologists and hydrologists in attaining improved knowledge of the real precipitation falling on the surface of our earth.

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MEASUREMENT OF RAINFALL BY RADAR IN THE UNITED KINGDOM

T.W. HARROLD
Meteorological Office, Bracknell, United Kingdom

1. INTRODUCTION

Ever since it was realized that radar could be used to detect the presence of precipitation, attempts have also been made to measure the quantity reaching the ground. However, although the qualitative use of radar is now well established, progress in the quantitative measurements has been slow. This has been largely because of (i) uncertainties in the accuracy of the measurements and (ii) the technical difficulties involved in processing the large amount of data obtained by using radar. During the last four years considerable progress has been made in the United Kingdom towards the solution of these problems.

It is worthwhile summarizing the advantages of the radar method of measuring precipitation compared with the only practical alternative of using rain-gauges. The advantages are that:

- the measurements can be made over an area (most users, if not all, require measurements over an area);
- the measurements can be processed, transmitted and displayed in near real time;
- the measurements are made from a single location, which simplifies maintenance in comparison with that associated with the operation of a network of rain-gauges;
- snowfall can be measured; and that
- the availability of measurements over an extensive area enables short term quantitative forecasts of precipitation to be made.

Possible disadvantages of a radar-based system of measuring rainfall, compared with the use of a rain-gauge network, are that accuracy may be less and cost may be greater. This is referred to briefly in Section 6.

2. THEORY

The basic principle of the technique of using radar to measure precipitation is that the power (P) back-scattered to the radar from the precipitation particles of diameter D at range r is given by

$$P = \frac{C \, \Sigma \, D^6}{r^2} \tag{1}$$

where the summation is over unit volume and the constant (C) consists of various measureable characteristics of the radar and of the dielectric constant of the particles. The rate of rainfall (R) is related empirically to D by an expression of the form

$$\Sigma D^6 = AR^B \tag{2}$$

so that

$$P = \frac{CA}{r^2} R^B \tag{3}$$

Hence R can be estimated from P by using predetermined values of C, A and B. A and B are variable because there is no unique drop-size distribution for each rate of rainfall, and this variability is a potential source of error in the radar technique. The error can be reduced if the radar estimates are calibrated at a single point within the field by matching the radar estimate over that point with an independent estimate for the point, obtained for instance by using a rain-gauge (see Harrold, English and Nicholass (1974), Wilson (1970) and section 3 of this paper).

3. ACCURACY OF ESTIMATES OF SURFACE RAINFALL

The Meteorological Office, together with the former Water Resources Board, the Dee and Clwyd River Authority and Plessey Radar Limited, have made an intensive study of the accuracy of radar estimates of rain falling on hilly subcatchments of the River Dee in North Wales. This study has been a phase of the Dee Weather Radar Project (Harding, 1972), the ultimate aim of which is to develop a radar system to measure areal precipitation in real time on a space and time-scale appropriate to the hydrological requirements for water management and river regulation.

A standard Plessey 43S weather radar was installed at Llandegla in North Wales during 1971 specifically for this project. This radar operated on a wavelength of 10 cm and had a beam width of 2°. In 1973 the radar was modified to operate on a beam width of 1°, this being accomplished by changing the operating wavelength to 5.6 cm. The change was made in order to secure a reduction in the extent of ground clutter (Harrold, 1974) and in the number of occasions when the radar beam intersected the metting layer.

The radar measurements were evaluated by comparing them with estimates based on a 1000-km^2 network of up to 70 battery-operated modified Plessey MM 37 tipping-bucket rain-gauges incorporating a magnetic-tape event-recorder. All but a few of the 740-cm^2 gauges were mounted in pits, with the rims at ground level and surrounded by a 5 x 5 x 5 cm plastic non-splash lattice.

On occasions when large rainfall gradients occurred the rain-gauge network estimates of the areal rainfall may have contained significant errors. In these circumstances the radar-derived pattern of rainfall was used to interpolate between gauge readings, each reading being assumed correct so as to derive an 'optimum' rainfall field which was then used for comparison with the radar estimate (see Harrold, English and Nicholass (1974) for more detail).

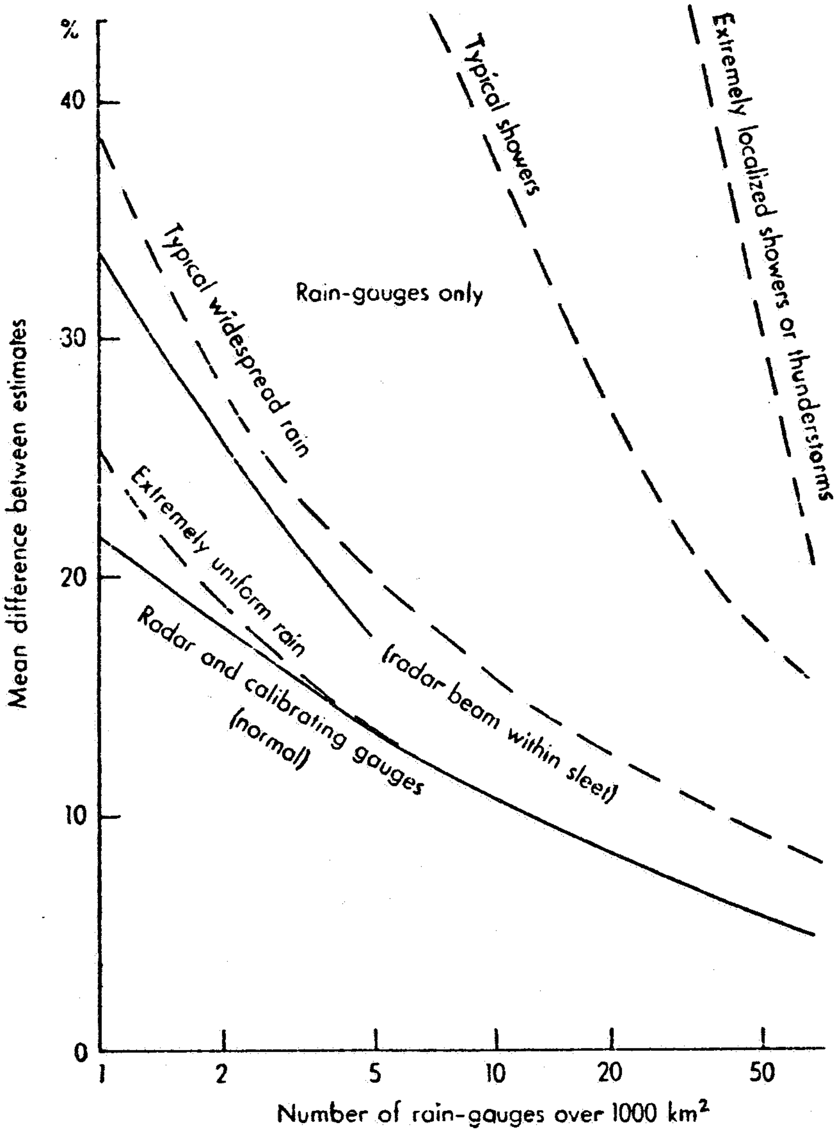

Fig. 1. Difference between estimates of hourly subcatchment rainfall related to rain gauge density (after Collier et. al. 1975).

Figure 1 (full lines) shows the difference between the radar and rain-gauge (or optimum) estimates of hourly rainfall over subcatchments of area typically 60 km^2. Three years of data are summarized. The radar estimates were calibrated by using various numbers of calibrating sites. The lower (greater-accuracy) line is for occasions when the radar beam was entirely within rain. The upper line is for occasions when a mixture of rain and snow was present within the beam. The figure shows, for example, that when the beam was within rain, if the radar was calibrated using two rain-gauges sites per 1000 km^2 then the estimates over subcatchments of 60 km^2 were on average within 18 per cent of those based on the entire rain-gauge network. Since these latter estimates contained some errors the true error of the radar estimate of surface rainfall was less than this value.

Also shown on the figure, as dashed lines, is the average difference between the full rain-gauge network (or optimum) estimates and estimates obtained by using various numbers of rain-gauges without a radar. The lower dashed lines are for widespread rain and the upper lines are for showery conditions. These dashed lines show, for example, that in typical showers estimates of rainfall over catchments of area 60 km^2 obtained by using a network of 20 gauges per 1000 km^2 differ from the optimum estimate by 26 per cent on average.

The figure can be used to derive the number of rain-gauges required to measure rainfall with the same accuracy as a radar system calibrated at a specified number of rain-gauge sites. For example, in typical showers a radar calibrated by using two rain-gauge sites gives the same accuracy over subcatchments of area 60 km^2 as a rain-gauge network of about 50 gauges per 1000 km^2.

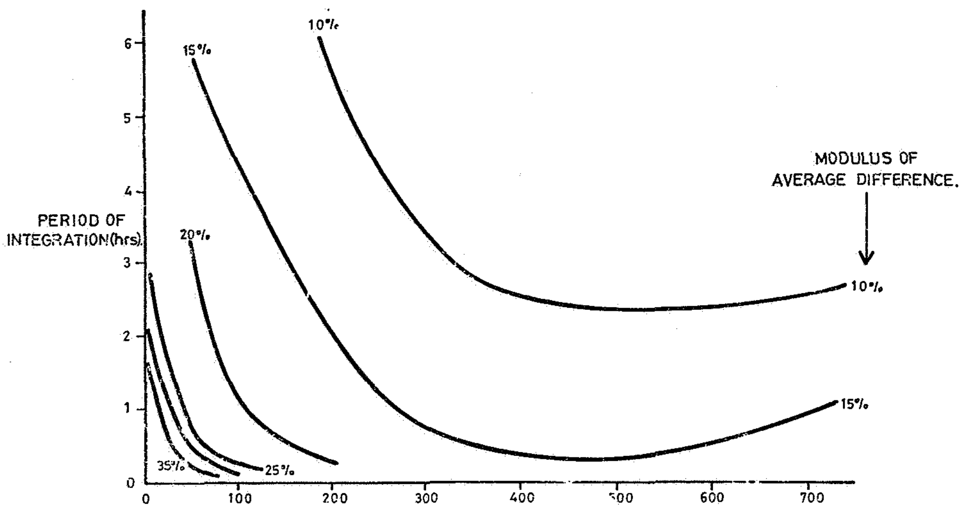

Fig. 2. Variation of the accuracy of radar measurements of areal rainfall in relation to period of integration and catchment, size.

The accuracy of radar estimates of areal rainfall depends on the area of measurement and on the period of integration – see Fig. 2, after Collier (1975). For example, over the Dee, average differences between the radar (calibrated at a single site) and the rain-gauge estimates of hourly rainfall decreased from 37 per cent for point comparisons to 20 per cent for comparisons over 100 km^2 and to less than 15 per cent over areas between 300 and 700 km^2 (Harrold et. al., 1974). This is because the drift of the rain in the wind between the volume sampled by the radar and the ground becomes progressively less important as the area is increased.

The accuracy of the radar estimates of surface rainfall obtained over the Dee is greater than that reported by most other authors, even though the measurements have been made over the type of terrain which would be expected to increase the difficulties of the technique because of ground clutter and the low-level changes which occur in the orographic precipitation of North Wales. The most important single contributory factor to this improved accuracy was probably the use of rain-gauges to calibrate the radar system. Other significant factors were that the measurements were assessed within a range of 50 km, which is much less than in some studies, and that the actual rainfall (the standard against which the radar is evaluated) may have been more accurately estimated than in some studies.

Factors which limit the accuracy of the radar estimates include the following:

- Changes in the drop size distribution relation (equation (2)) over the area.
- Increase of the rain between the radar beam and the ground. In North Wales this growth often exceeds a factor of two in the lowest kilometre of the atmosphere, so the beam has to be as close to the ground as possible in order to minimize the error. This growth is much more pronounced in the hilly terrain of North Wales than over flatter terrain.
- Ground clutter. In the area of the experiment, measurements of precipitation were not made in areas where ground clutter occurred in dry weather. Rainfall was estimated by interpolating from around the area of clutter. The changing of the beam width from 2° to 1° significantly reduced the amount of clutter observed.
- Blocking of beam. At Llandegla the radar view to the south, outside the area of main hydrological interest, is blocked by a nearby hill. In order to estimate rainfall the beam-elevation angle has to be increased, and this seriously reduces the accuracy of estimates of the rainfall reaching the ground. In siting a weather radar with the intention of making quantitative measurements of surface rainfall it is essential that prior consideration be given to he blocking and ground clutter which will occur. A computer technique has been developed which enables this to be done in hilly terrain (Moores and Harrold, 1975).
- The melting layer. When the radar beam intersects the melting layer the accuracy of the estimates of surface rainfall can be in error by at least a factor of two (Collier 1974). A method has been developed (Harrold and Kitchingman, 1975) which enables

this error to be reduced, but even then errors are greater than in the absence of the melting layer – cf the two full lines in Figure 1. The melting layer sets a limit to the accuracy of radar estimates of rainfall in temperate latitudes in winter.

– Anomalous propagation. The distribution of ground clutter did vary occasionally, particularly when the atmosphere was such that anomalous propagation of that radar beam occurred. These occurrences were objectively identified over North Wales once the radar has been converted to operate with a 1° beam, by comparing echo intensities at two beam elevations (0.5° and 1.5°). If these echoes above a single location differed by more than 60 per cent they were attributed to anomalous ground clutter. It should be noted that this technique was successful primarily because quantitative measurements were only attempted within a rainge of 50 km; it would be less successful at greater ranges, or if adopted in connection with radars with a wider beam width.

4. REAL-TIME PROCESSING, TRANSMISSION AND DISPLAY OF WEATHER RADAR DATA

Concurrent with the studies over North Wales the Royal Radar Establishment at Malvern have been developing for the Meteorological Office a system the purpose of which is to enable quantitative radar data to be processed, transmitted and displayed remotely in real time.

Broadly speaking two kinds of data-processing are involved, for convenience termed primary and secondary. The primary processing entails the conversion of radar signals into meaningfull estimates of surface precipitation in cells with a space-time resolution matched to potential users' requirements. The secondary processing entails the synthesis and rearrangment of the primary data into a more convenient format.

The primary processing is carried out at a radar site by a dedicated PDP 11-40 mini-computer. There are several stages in this processing. First the radar signals from a number of pulses are averaged so as to provide a reliable estimate of the average power return in cells, each of which is defined by 600 m in range and 1° in azimuth. (Only this average power, not the noise-like signal returned from single pulses, can be related to rate of rainfall). The averaging is carried out in a specially developed hardware component. There then follow several processes within the computer aimed at obtaining quantitative rainfall measurements. At the present stage of development these processes include allowance for partial blocking of the radar beam by high ground and for ground clutter. Other procedures being developed include automatic calibration of the radar-derived estimates of rainfall using measurement from a few telemetering rain-gauges and an attempt to correct objectively for at least some of the errors introduced when the beam is within the melting layer (Harrold and Kitchingman, 1975).

The last stage in the on-site primary data processing is that of transformation from polar to Cartesian co-ordinates. This dispenses with the needlessly high resolution provided by polar co-ordinates at close ranges and thereby diminishes the amount of data, which can then be transmitted by standard telephone lines.

There are several possible ways of handling and displaying the data at the receiving terminal. To date the work has concentrated on developing a pictorial display on an inexpensive commercial colour-television tube. Some form of data storage is required before the display can be generated and this has been accomplished by using semiconductor storage elements. The entire system is described more fully by Taylor and Browing (1974).

At present the system is used in conjunction with two weather radars, including that at Llandegla in North Wales. Data from the latter are transmitted routinely to the River Control Centre at Bala in North Wales. Data were also transmitted to the Meteorological Office Central Forecasting Office at Bracknell for a 9 month trial, and are now (April 1976) transmitted to the forecasting office at Preston, in north west England.

Current investigations include assessing that the accuracy of the real-time estimates of rainfall over North Wales are similar to those obtained off-line during the Dee Weather Radar Project (see previous Section), using the radar data in a real-time hydrological model being developed for the Dee Catchment, and determining methods of compositing data from a number of radars into a single display.

5. FORECASTING OF PRECIPITATION USING RADAR DATA

When forecasting precipitation using radar data, account must be taken of

- the initial distribution of precipitation;
- the movement of the pattern;
- the development or decay of existing and new areas of precipitation;
- orographic efffects.

The extent to which these various factors are adequately forecast determines the period of usefulness of the forecast. A very simple method of forecasting is to assume that the pattern does not change, so that the existing field can be advected at its observed velocity. This type of forecast can be made using objective techniques developed in Pattern Recognition studies. In widespread rain the forecast can be useful for several hours ahead, the period often being limited by precipitation from beyond the area observed by the radar moving into the area of interest. (A network of radars with overlapping coverage minimises this limitation). In convective conditions the precipitation field changes quite rapidly, and on some occasions detailed quantitative forecasts may not be possible to more than about an hour ahead, see Fig. 3 for an example. However, even in these circumstances it is sometimes observed that, although individual cells change rapidly, they recur within large clusters which persist for several hours and which can be forecast.

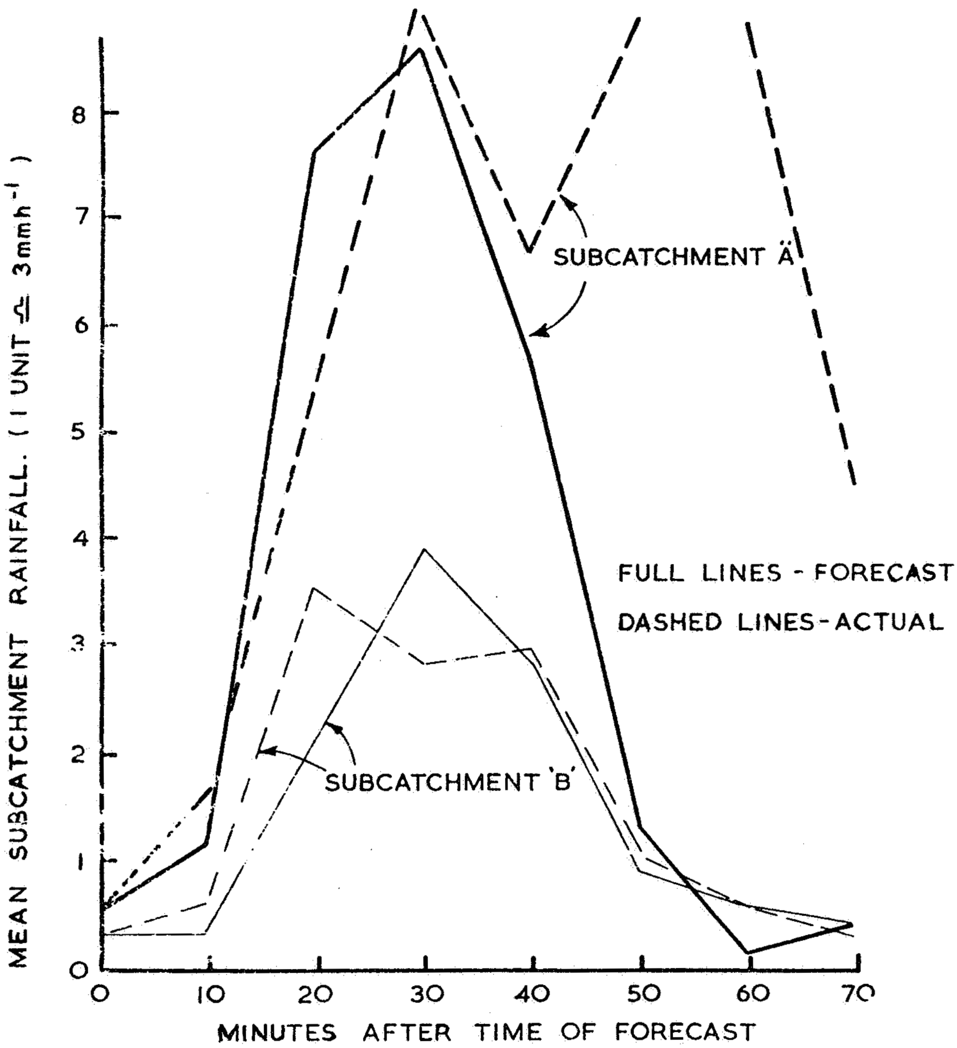

Fig. 3 Examples of objective forecasts of rain over two subcatchments of the River Dee in North Wales on 31 July 1972.

As the period of a forecast is extended, probability forecasts can be made, or the development of the pattern can perhaps be predicted adequately by objective techniques. The latter could be attempted by observing the time changes of the system and extrapolating these forward in time. An alternative means of forecasting some hours ahead is to incorporate the radar data into a numerical (computer-based) prediction of the behaviour of the atmosphere. Work has commenced on these type of forecasts.

6. ECONOMIC CONSIDERATIONS

During the Dee Weather Radar Project a study was carried out into the use of a network of radars in measuring and forecasting precipitation for a number of potential user applications in the United Kingdom (Water Resources Board, 1973). The report included a chapter and detailed appendix on economic considerations. Costs of six alternative radar systems were estimated, the options being beams widths of 1.0° and 0.5° and three degrees of manning: full-time, normal working hours and once a fortnight. The manning was assumed to be required for maintenance purposes only, routine operations being automatic. It was concluded that although equipment costs would be less with a fully manned system, the partially manned systems were to be preferred economically because of the much smaller running costs.

In the report the cost of a radar network covering the United Kingdom, and manned during working hours only, was compared with alternative methods of measuring areal rainfall. It was concluded that 'a rain-gauge network is the preferred means of measuring rainfall over an area limited to less than about 3000 km^2 if forecasts are required for no longer than 30 minutes ahead. For large areas, or if forecasts over longer periods are required, a radar-based system is to be preferred. However, a network of radars covering the entire country is the only means by which the user requirements ... can even be approached'.

7. CONCLUSION

The accuracy of radar-derived measurements of rainfall is now known, and a method of processing transmitting and displaying the quantitative data in real time has been developed. Studies into possible methods of utilising these data in forecasting have commenced. However, the radars generally available for operational use in Europe at present are such that all the potentialities of radar cannot be met. These potentialities can be realised only be developing a network of radars with overlapping coverage, utilising a narrow beam width (1° or less, at least in the vertical) and operating on a non-attenuating wavelength (5 cm or more).

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IS WEATHER RADAR RELEVANT TO DUTCH CIRCUMSTANCES?

D.A. KRAIJENHOFF VAN DE LEUR

For a first attempt at an appraisal of benefits that could be derived from the application of radar techniques for rainfall measurements in the Netherlands, a recapitulation of the specific characteristics of this method may be helpful:

For a considerable area, covering both water and land, radar provides us with a detailed image of moving precipitation in that such pictures are practically continuous in space and time. A high precision of rainfall depths and rainfall rates can be obtained through simultaneous operation of a sparse network of telemetering raingauges for calibration of the radar image.

The amount of detail of such images seems to suffice for determination of rainfall depths on large and small catchment areas within appropriate time intervals. The space-time resolution for the presentation and storage of discretized data from this continuous image can be chosen in accordance with the potential user's requirements.

Dr. Harrold's closing sentence: "... a network of radars covering the entire country is the only means by which the user's requirements ... can even be approached" clearly indicates the unlimited range of questions to which the supplier of precipitation data is expected to provide the answers. Let us try and formulate some questions which a Dutch hydrologist might ask.

Every year, hundreds of millions of guilders are spent on sewer systems designed either to drain new urbanized areas or to reduce the number of overflows from existing sewer systems into surface waters. What is the critical sequence of areal precipitation rates that would produce such overflows? For the Netherlands, there is no satisfactory answer to this question. Nevertheless it is being asked with growing urgency, since the water authorities try to protect the quality of their surface waters with increasingly stringent constraints.

This design problem of urban drainage systems is being met by a world-wide effort to interest hydrometeorological analysts in finding ways for determining the probabilities of major rainstorms over various sizes of areas.

Larger storms can be analyzed with the aid of terrestrial networks of raingauges. Here again the question arises whether the results of such expensive local studies can be transposed to other regions. Obviously, there are climatological limits; they may relate to the classes of atmospheric mechanisms that have produced such areal rainfalls.

High intensities of small areal rainfall are nearly exclusively due to convective local rainstorms. The precipitation field changes quite rapidly so that short-interval information from terrestrial networks calls for a large number of telemetering raingauges. Such a network has been installed elsewhere: 225 recording raingauges were installed in and around St. Louis, Missouri. This "Metromex" programme has definitely shown that, locally, the urban environment substantially modifies the rainfall from a portion of the raincells having already moderate to heavy intensity due to natural atmospheric processes. Among 300 heavy cells that occurred within the study period, the maximum 5-minute volumes showed a medium increase of 43% water yield. It was reported last year that "a major hydrologic implication from the Metromex studies to date is that rainfall rate frequency distributions may vary significantly in large urban industrial regions".

Such combinations of heavy industries and metropolitan areas have also developed in the Netherlands, and the first indications of urban weather modificiation have already been reported.

Would this imply the possibility of significant increases of storm cell precipitation?

Would radar be a helpful tool to study these high-intensity but local and short-lived phenomena that should determine the capacity and layout of our expensive sewer systems?

Referring briefly to all intermediate cases of rainfall runoff relationships in medium size and larger catchment areas, one must suspect that the lack of reliable real rainfall series of sufficient length has led to sub-optimal though generally acceptable drainage projects.

Finally, some hesitant questions can be asked on the feasible role of weather radar as an intermediate tool between a terrestrial network of raingauges and the Global Atmospheric Research Programme (GARP).

A recent WMO-report tells us that the main objective of GARP is to extend abilities to predict the motion of the atmosphere from one day to several weeks. Such extension should permit improvement of meteorological forecasts used as inputs to hydrological forecasting. We are also told that radar measurements may assist in the general analysis of the current weather situation by identifying mesoscale patterns of organization which otherwise would escape detection. Drs. Harrold and Attmannspacher have shown the high accuracy of one to three hourly integrated rainfall measurements over areas up to 700 km^2 by means of a calibrated radar. Could such radar measurements supply the "ground truth" for the calibration of the satellite images that are used in the World Weather Watch (WWW) and in GARP?

WMO is already arranging the preparation of case studies and pilot projects for testing and developing the capability of WWW systems to meet the needs of operational hydrology, especially hydrological forecasting in large international basins such as the Rhine and the Danube.

Piecing these bits of information together does provide a tantalizing vision for a country at the lower end of the River Rhine that would like to operate her water management system of weirs, inlets, sluices and reservoirs on the basis of an optimal flow forecasting system.

How much real promise does this picture hold?

Summarizing the results of this brief reconnaissance, one can assume that weather radar techniques clearly offer interesting opportunities in the field of hydrologic forecasting and prediction in the Netherlands. Therefore, close participation in this new field should seriously be considered.

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