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Air mass characteristics above Athens during snowfall

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Summary
A description is given of the structure of the atmosphere above Athens during periods of snowfall, based on surface and upper-air observations at the Hellinikon meteorological station for the period 1956–73. Mean tephigrams are plotted, and mean vector winds at all standard levels calculated. The variation of mean temperature and geopotential for standard levels is studied from two days before the beginning of a snowy period to one day after the end of the period, and also the covariation of temperature and geopotential at 500 and 850 hPa. The performance of various snow predictors defined by previous researchers (mainly British) is checked against the results, proving that they are generally successful in the Athens region.

1. Introduction

This paper is the third to discuss snowfall in Athens, and it deals mainly with the air mass structure above Athens during snowfall. The first paper (Prezerakos and Angouridakis 1979) considered basic characteristics of snowfall in Athens, and the second (Prezerakos and Angouridakis 1984) studied the synoptic evolution of atmospheric circulation systems causing those snowfalls.

Brief reviews of previous papers which dealt with snowfall in Greece (and specifically in Athens) were provided in the earlier papers. Although these two papers describe the data and method used and they give the statistical characteristics of the meteorological elements under consideration, there is a need for some of this material to be briefly repeated here for the sake of continuity.

2. Data and method used

To study the structure of the atmosphere above Athens during snowfall, the surface and upper-air observations at the Hellinikon meteorological station (World Meteorological Organization No. 16716) were used. Hellinikon is about 15 km south of the centre of Athens and about 300 metres from the coastline, at an elevation of 10 m.

Comparison of the snowfall dates at Hellinikon with other stations in the Athens region led to the conclusion that when it snowed at Hellinikon, it also snowed in the city of Athens. Thus 'snow-days' at the Hellinikon meteorological station can be considered as 'snow-days' in Athens.

Dates of snow-days in Athens, selected in the 1956–73 period according to the definition below, are again used for the sake of continuity with the two previous papers. A 'snow-day' in Athens is said to occur when at least one of the eight synoptic observations daily at Hellinikon reports snow in the present weather (ww 70 to 79) or past weather code (W = 7), independently of whether the snow remains lying on the ground or whether it melts.

From the dates found it was realized that occurrences could be divided into groups of one, two and three successive days. Examination of the daily meteorological bulletins for the days selected showed that during the first day ('F-day') of a group of successive snow-days in Athens, the synoptic situation was typical and could be classified into two types, A and B (Prezerakos and Angouridakis 1979, 1984). *Type A* is associated with a surface anticyclone over western Europe with its centre over the Alps, whereas *type B* is associated with a surface anticyclone which extends over central Europe with its centre over south-western Russia.

To study the structure of the atmosphere above Athens during snowfall, mean soundings at 0000 GMT have been calculated and plotted on tephigrams for each synoptic type A and B separately and for the days F–2, F–1, F, and for E-day, where E-day (or END day) is the day immediately following a group of sucessive snow days. Graphs have also been prepared showing the variation of mean temperature and mean geopotential at the 850, 700, 500, 300 and 200 hPa* isobaric surfaces, for both synoptic types A and B, together with mean sea level (MSL) pressure variation for days F–2, F–1, F and E. Other graphs have also been drawn to present the covariation of geopotential and temperature at standard isobaric levels on day F at 0000 GMT. Finally a mean tephigram for Hellinikon closest to the time of snowfall in Athens (independently of the synoptic situation) has been plotted and analysed. Scatter diagrams have been prepared showing the relationship between temperature and geopotential at 850 and 500 hPa levels.

3. Mean structure of the atmosphere above Athens during snowfall

There were 40 snow-days during the period considered; 15 occurrences were attributable to type A and 8 to type B, some pressure systems causing two or three snow-days (Prezerakos and Angouridakis 1984). The mean structure of the atmosphere and particularly of the troposphere for the groups of days F–2, F–1, F and E is presented schematically in Figs 1 to 8 for types A and B separately and together.

3.1 *Type A situations*

Fig. 1(a) shows the mean structure of the atmosphere on day F–2 (that is, two days before snow in Athens) with a synoptic situation of type A. The air mass is relatively moist since the dew-point depression is less than 6 K in the lower layers from the surface up to 750 hPa. The air mass is also relatively cold and conditionally stable so that it does not appear as even a pseudo-latent type of instability. Examination of the wet-bulb curve (dotted line) shows that the wet-bulb potential temperature (θ_w) increases with height. Thus the relation $\partial\theta_w/\partial z > 0$ is valid for every layer of the atmosphere, and so the atmosphere is also potentially stable. The variation of wind direction with height shows some cold advection in the lower layers from the surface up to about 650 hPa, whereas above 650 hPa the advection does not appear with a constant sign.

On day F–1 (Fig. 1(b)) the situation remains almost unchanged with a slight increase in humidity and consequent decrease in potential stability. The cold advection in the lower levels is maintained.

On day F (Fig. 2(a)) the absolute humidity of the lowest part of the troposphere up to 650 hPa has decreased but the relative humidity has increased (since the temperature has decreased). The dew-point depression shows that the cloud amount is probably more than 6 oktas in the layer from lifting condensation level up to 650 hPa (Meteorological Office 1975). The air mass stability remains unchanged and the variation of wind direction with height shows a little warm advection from the surface up to 900 hPa, whereas from 900 hPa up to 600 hPa the advection becomes cold. Above 600 hPa there is no distinct thermal advection. The height of the tropopause has also lowered during these three days.

* The hectopascal (hPa) is now the preferred unit of atmospheric pressure; it is numerically equal to the millibar (mb). Both units will continue to be used in *Meteorological Magazine* at present, the choice being left to individual authors.

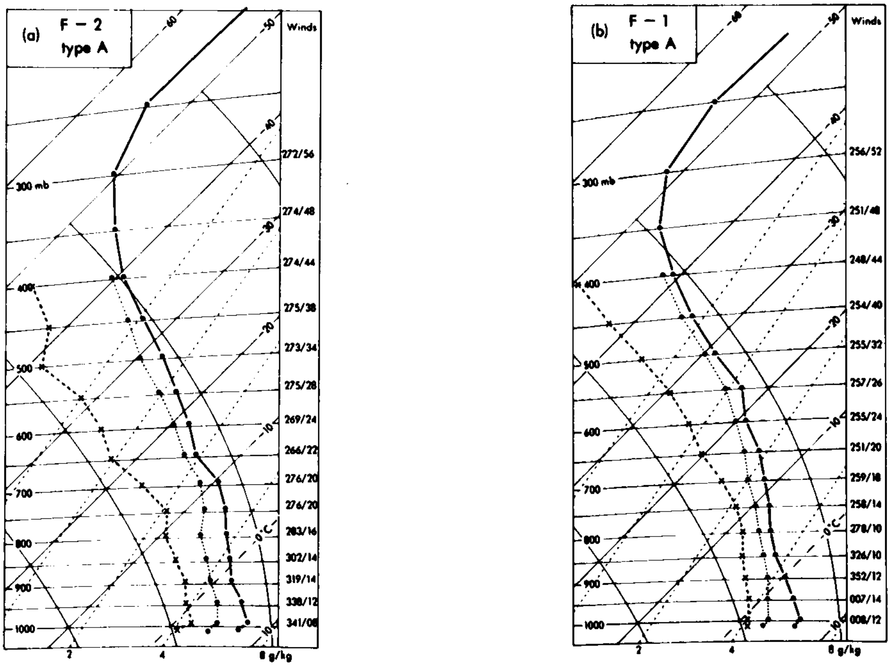


Figure 1. Mean sounding for Hellinikon for type A synoptic pattern at 0000 GMT on day F−2 (a) and F−1 (b). The continuous line is air temperature, the dotted line wet-bulb temperature and the dashed line dew-point temperature. Direction (degrees) and speed (kn) of the vector mean wind are given for standard levels.

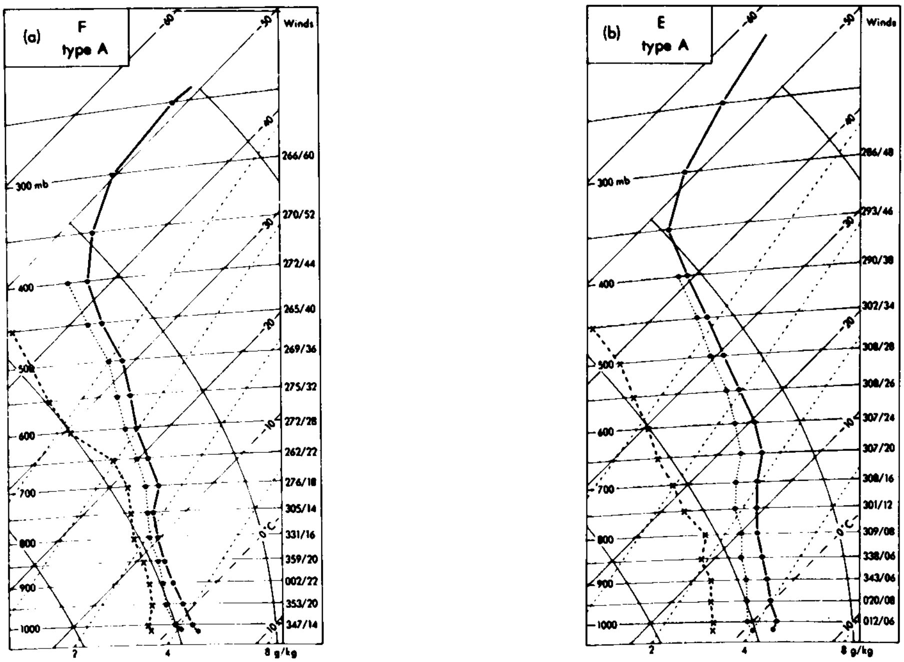


Figure 2. As for Fig. 1 but for day F and E.

The atmospheric structure above Athens on day F (Fig. 2(a)) indicates that the clouds would be basically stratocumulus and nimbostratus, while the cause of the air mass lifting must be dynamical (vorticity advection, convergence) or forced lifting. This is shown more clearly in the evolution of the pressure systems which cause snowfall in Athens (discussed in Prezerakos and Angouridakis 1984).

On day E (Fig. 2(b)), the relative humidity (and therefore the absolute humidity) has decreased, whereas the stability of the atmosphere has increased. The surface winds have veered to north-easterly while the cold advection in the lower part of the atmosphere is unchanged.

3.2 *Type B situations*

Fig. 3(a) shows the mean structure of the atmosphere above Athens two days before snowfall, with a synoptic situation of type B. The structure is stable, with the greatest stability in the layer nearest the ground. The variation of wind direction with height shows cold advection in the layer from the surface up to 650 hPa; there is no clearly-defined thermal advection at higher levels.

On the following day (F−1), Fig. 3(b), the humidity increases in the same way as that of the corresponding day of type A, but only up to 750 hPa; cloudiness in the lower layers is, therefore, increased. The cloud must consist of stratocumulus owing to the general stability of the air mass. Winds have become more northerly; their variation with height clearly shows cold advection in most of the troposphere.

Fig. 4(a) shows that on day F in type B situations the structure of the atmosphere is slightly unstable in the layer from the surface up to 850 hPa (and particularly so up to 950 hPa). This is due to the warming of the air passing above the sea coming from regions which lie to the north-east of Athens. Because of this warming the planetary boundary layer becomes unstable. The structure of the atmosphere above 850 hPa is stable both conditionally and potentially. The winds are north-easterly up to 800 hPa and north-westerly above that. Their variation with height shows some cold advection above 750 hPa.

On day E, Fig. 4(b), the relative humidity decreases and the structure of the atmosphere becomes stable even in the lowest part of the boundary layer (owing to the cold ground). The winds become north-easterly up to 650 hPa, backing slightly with height, showing that some cold advection is still present.

3.3 *Mean ascents during snowfall*

The mean ascent of upper-air observations at Hellinikon nearest to the time of snowfall occurrence (independently of the type of synoptic situation) is shown in Fig. 5.

This plotted tephigram is the most representative of the structure of the atmosphere during snowfall. The diagram is based on the average of 39 soundings (40 snow-days during the period 1956–73, less one missing sounding). The basic characteristics of the tephigram also reflect the stable stratification, the cold air mass ($T_{850} = -7.5$ °C, $T_{500} = -31.5$ °C) and the increased moisture in the lower layers. The latter fact implies a considerable amount of stratiform clouds, and specifically that the cloudiness in the layer from 950 hPa up to almost 800 hPa is 8 oktas because the dew-point depression is less than 2 K (Air Weather Office 1959, Meteorological Office 1975).

The winds are mainly from the north. They are the vector means of both types A and B but the winds of the type A synoptic situation are north-westerly in the lower levels of the troposphere during snowfall, whereas the winds of the type B synoptic situation are north-easterly, producing a rather complicated resultant. Thus north-westerly winds occur near the surface, north-easterly winds are found above them and finally north-westerly winds return again in the region above 750 hPa.

4. The variation of temperature and geopotential of standard isobaric surfaces from day F−2 to E

The variation of surface temperature and that of the standard isobaric levels was examined from day F−2 to E. The data considered were 0000 GMT radiosonde observations, and inevitably there were cases

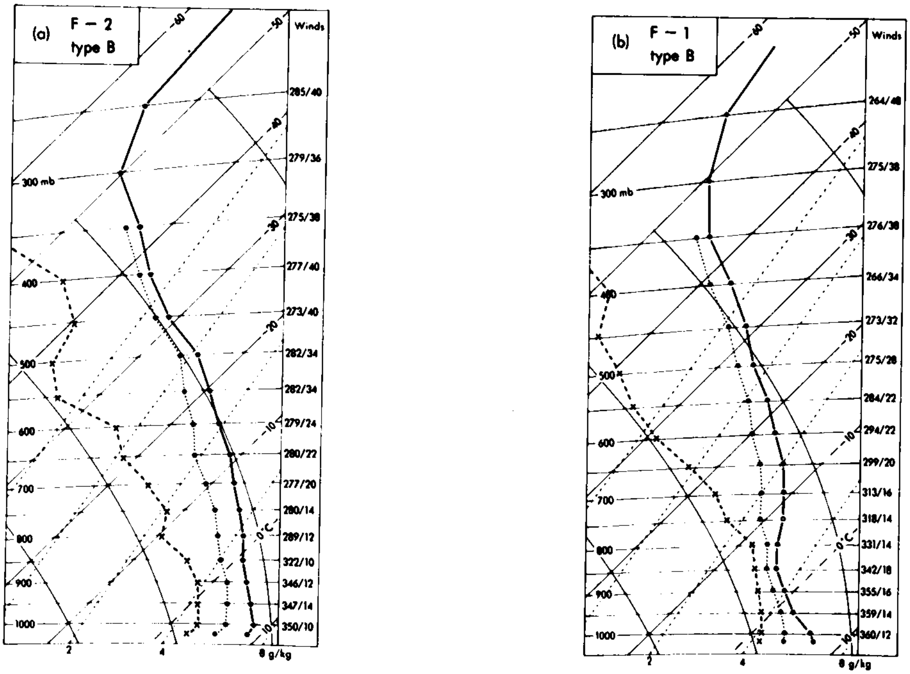


Figure 3. As for Fig. 1 but for type B synoptic pattern.

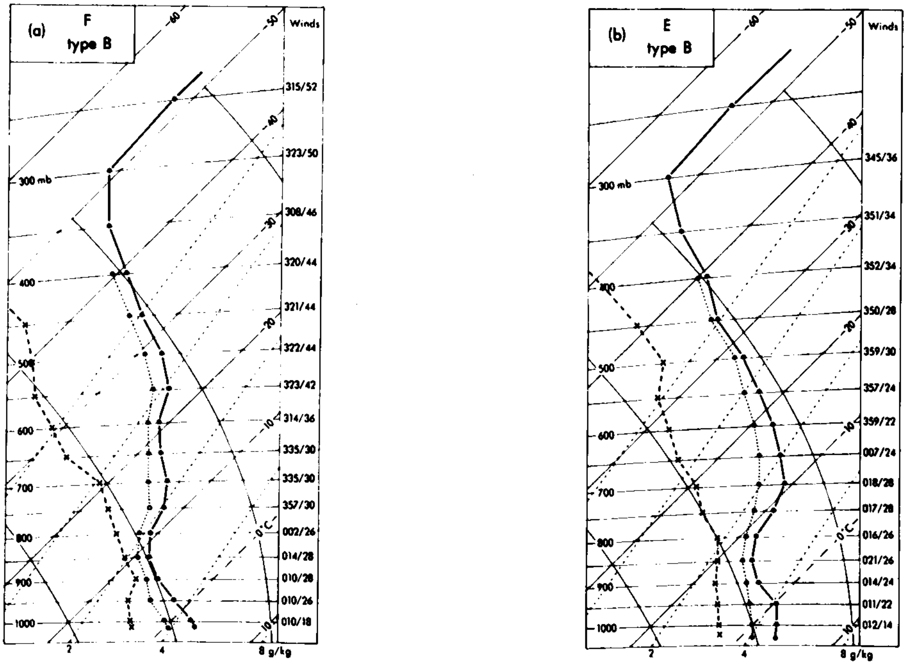


Figure 4. As for Fig. 2 but for type B synoptic pattern.

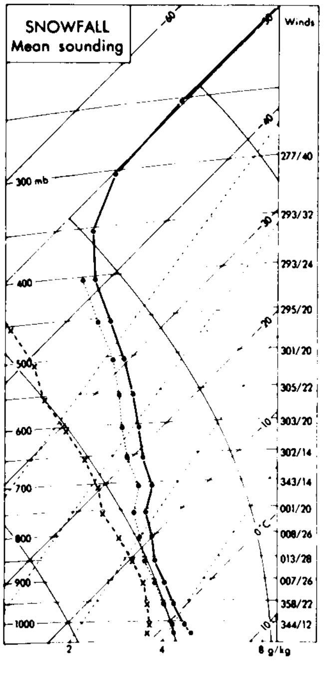


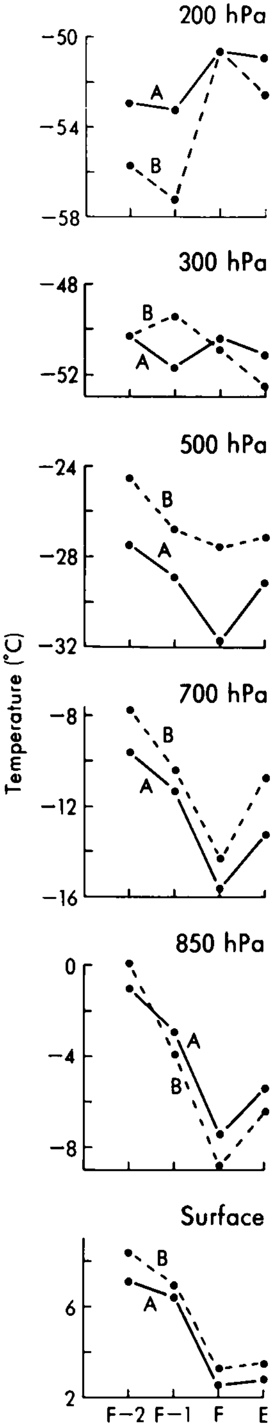
Figure 5. Mean tephigram for Hellinikon for the upper-air sounding nearest to the time of snowfall (independently of synoptic pattern). Direction (degrees) and speed (kn) of the vector mean wind are given for standard levels.

where the radiosonde launch was more than 12 hours from the time of snowfall. This departure was, on average, five hours for synoptic type A and nine hours for type B on day F (Prezerakos and Angouridakis 1979).

Day F was, on average, the coldest day at all levels up to 500 hPa, whereas it appeared slightly warmer at the 300 and 200 hPa levels (Fig. 6, and Table I). Day F−1 was the coldest day at 300 and 200 hPa because the tropopause lay beneath the 300 hPa level. Also worthy of attention is the fact that type A temperatures appear generally to be colder than type B and only at 850 hPa on day F−1, day F and E does the latter appear to be colder than the corresponding day of type A. At 300 hPa day E of type B appears to be colder than the corresponding day of type A, whereas both types on day F have almost the same temperature. Day F of both types also appeared to have the same temperature at 200 hPa. The values of the mean temperature by day and by type given in Fig. 6 have been a significant help in forecasting snowfall in Athens.

In Fig. 7 and Table II the variation of barometric pressure at station level and the variation of the geopotential of the standard isobaric surfaces of both types A and B from day F−2 to day E are shown. From this it is clear that the barometric pressure at station level reaches its minimum value on day F−1 and that the pressure values of type A are less than the corresponding values of type B. It is also apparent that the variation in pressure under type B from day F−2 to day F−1 is negligible, whilst its increase from day F−1 to day F is significant, as in type A. These data suggest that snowfall in type A synoptic situations occurs when some perturbation, which causes the pressure to fall in the Athens area, has passed, whereas snowfall of type B is due to strengthening of the anticyclonic circulation (which increases the barometric pressure) and the mass uplifting is of the forced type. Such cases arise when an anticyclonic ridge extends towards Greece from the north.

The 850 hPa geopotential under type B appears unchanged from day F−1 to day F, whereas at all other levels the geopotential falls to a minimum value on day F.



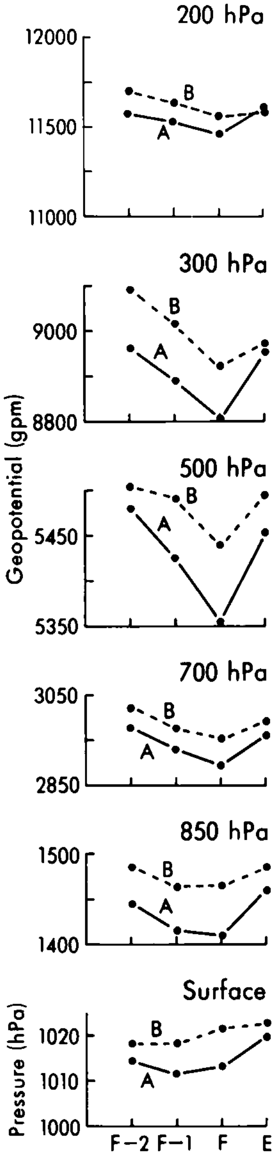


Figure 7. Variation of geopotential of standard isobaric surfaces and surface pressure for type A and B synoptic patterns from day F-2 to E.

Figure 6 (*left*). Mean temperature for the surface and standard isobaric levels for type A and B synoptic patterns from day F-2 to E.

Table I. *Mean temperatures (°C) and standard deviations (SD, in K) at standard isobaric surfaces and at the surface for days F–2 to E at Hellinikon near Athens (see text for further details). These values are presented graphically in Fig. 6*

	Day F–2	Day F–1	Day F	Day E
200 hPa				
Mean — type A	-53.1	-53.2	-50.6	-51.0
— type B	-55.7	-57.2	-50.6	-52.6
SD — type A	4.8	4.7	4.2	6.8
— type B	5.5	5.5	2.6	3.3
300 hPa				
Mean — type A	-50.3	-51.7	-50.4	-51.1
— type B	-50.2	-49.4	-50.9	-52.5
SD — type A	2.9	2.5	3.3	2.0
— type B	4.4	2.8	2.4	3.6
500 hPa				
Mean — type A	-27.5	-28.9	-31.7	-29.1
— type B	-24.5	-26.8	-27.5	-27.1
SD — type A	4.9	4.9	3.3	5.4
— type B	2.9	3.8	4.1	2.8
700 hPa				
Mean — type A	-9.6	-11.3	-15.5	-13.2
— type B	-7.7	-10.4	-14.3	-10.7
SD — type A	5.4	3.0	2.0	5.5
— type B	4.1	3.7	3.2	3.4
850 hPa				
Mean — type A	-1.0	-2.9	-7.2	-5.4
— type B	0.9	-3.9	-8.8	-6.4
SD — type A	5.7	3.6	1.8	5.6
— type B	4.8	4.2	2.5	2.1
Surface				
Mean — type A	7.1	6.4	2.4	2.6
— type B	8.4	7.0	3.3	3.6
SD — type A	4.1	3.2	1.4	3.4
— type B	3.0	3.0	2.1	2.9

The average geopotentials of the isobaric surfaces and the values of the barometric pressure at station level of type A synoptic situations were less than those of type B. These facts support the above hypothesis, namely that snowfall of type A is due to perturbations causing positive advection of vorticity towards the Athens region, while snowfall of type B is caused basically by the forced uplifting of the very cold air above orographic obstacles. This cold air is directed towards the Athens region by a north-easterly wind associated with the elongated ridge of high pressure.

Table II. *Mean and standard deviation geopotentials (gpm) for standard isobaric surfaces and mean and standard deviation of surface pressure (hPa) for days F–2 to E at Hellinikon near Athens (see text for further details). These values are presented graphically in Fig. 7*

	Day			
	F–2	F–1	F	E
200 hPa geopotential				
Mean — type A	11 583	11 518	11 460	11 597
— type B	11 694	11 642	11 557	11 592
SD — type A	109	83	82	106
— type B	102	144	122	107
300 hPa geopotential				
Mean — type A	8 961	8 889	8 805	8 954
— type B	9 095	9 016	8 924	8 976
SD — type A	146	129	111	166
— type B	110	126	79	86
500 hPa geopotential				
Mean — type A	5 482	5 427	5 355	5 453
— type B	5 506	5 490	5 441	5 499
SD — type A	99	82	73	108
— type B	78	83	50	49
700 hPa geopotential				
Mean — type A	2 974	2 931	2 898	2 962
— type B	3 021	2 978	2 953	2 992
SD — type A	58	52	56	61
— type B	53	58	29	32
850 hPa geopotential				
Mean — type A	1 448	1 416	1 410	1 462
— type B	1 485	1 464	1 467	1 487
SD — type A	36	44	54	52
— type B	41	58	35	33
Surface pressure				
Mean — type A	1 014.5	1 011.5	1 013.7	1 019.5
— type B	1 018.3	1 018.1	1 021.6	1 022.7
SD — type A	3.9	5.4	6.8	7.7
— type B	6.1	8.1	5.4	4.1

5. Covariation of temperature and geopotential of the 850 and 500 hPa isobaric surfaces on snow-days

Forecasters' experience suggests that knowledge of the temperature and geopotential at 850 and 500 hPa and their covariation when it is snowing in Athens are a significant help in forecasting snowfall. For these reasons an attempt has been made to correlate the 0000 GMT geopotential and temperature at 850 and 500 hPa on day F for each synoptic type situation separately. This correlation has been done graphically because there are few cases (15 snow-days of type A with 14 available 0000 GMT upper-air observations, and 8 of type B with 7 available 0000 GMT observations), they are very easy to plot and the regression curve can be easily drawn (Meteorological Office 1975). Linear correlation equations have been derived only for the (39) observations nearest in time to the snowfall, independently of synoptic type.

5.1 *850 hPa*

The values of 850 hPa temperature and geopotential on day F at 0000 GMT are illustrated in Fig. 8(a) for synoptic type A, and Fig. 8(b) for type B. Temperatures ranged from –5.5 °C to –11.5 °C for type A and from –7 °C to –13 °C for type B, i.e. in both cases there was a range of 6 K. The geopotential values for type A ranged from 1330 gpm to 1485 gpm (a range of 155 gpm) and from 1390 gpm to 1505 gpm (a range of 115 gpm) for type B. It is clear from Table III that north-westerly winds dominate in type A and north to north-easterly in type B. The type B air mass at 850 hPa appears to have higher relative humidity (Table III) because it has been advected by north-easterly winds passing over the Black Sea and the Aegean Sea thus picking up moisture. The type A air mass is advected by north-westerly winds passing over land (central Europe). Figs 8(a) and 8(b) also show that on snow-days in Athens the 850 hPa geopotential usually increases with decreasing temperature. Finally Figs 8(a) and 8(b) show that snow is unlikely in Athens unless the 850 hPa temperature is below –5.5 °C and the associated geopotential less than 1505 gpm.

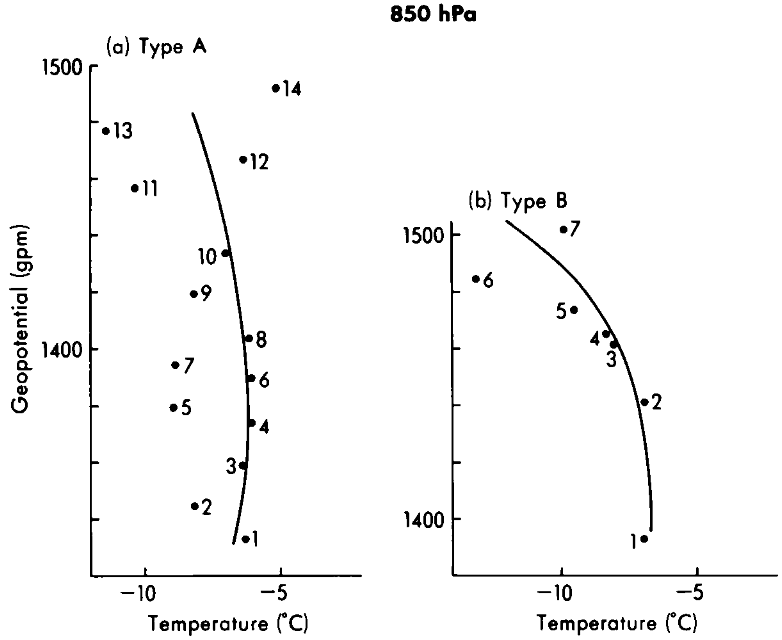


Figure 8. Scatter plot of temperature and geopotential of the 850 hPa surface for (a) synoptic type A and (b) synoptic type B on day F, at 0000 GMT. The numbered points refer to the values in Table III.

Table III. *Wind direction (degrees) and speed (m s^{-1}) and humidity (%) for the numbered points on the scatterplot of 850 hPa temperature against geopotential (Figs 8(a), 8(b)). The points are numbered consecutively from the lower axis and do not necessarily correspond to similarly-numbered points in Tables IV or V or Figs 9 and 10*

Fig. 8(a) (Type A)

Point No.	Wind direction	Wind speed	Humidity
1	350	08	75
2	040	09	77
3	340	10	80
4	330	04	85
5	350	12	90
6	290	08	57
7	350	19	99
8	040	15	99
9	350	08	95
10	290	04	87
11	350	11	90
12	310	09	89
13	350	18	81
14	330	08	95

Fig. 8(b) (Type B)

Point No.	Wind direction	Wind speed	Humidity
1	020	18	87
2	020	15	87
3	350	15	86
4	020	16	92
5	350	12	81
6	020	12	90
7	020	22	98

5.2 *500 hPa*

Figs 9(a) and 9(b) show the correlation of geopotential and temperature at the 500 hPa isobaric surface. In these figures temperature usually increases with geopotential as do most isobaric surfaces (Angouridakis 1977), but in the lower part of the type A curve (Fig. 9(a)) the variation is reversed. Furthermore it can be seen that days with snow in Athens are characterized by temperatures at 500 hPa in the range from −24 °C to −35.5 °C and geopotentials ranging from 5270 gpm to 5510 gpm; westerly winds predominate in synoptic type A and north-westerly winds in type B (Table IV). Relative humidity appears to be higher in type A than in type B. This seems to be due to the dynamical large-scale upward velocity resulting in adiabatic cooling which occurs on snow-days in Athens with type A synoptic situations, whilst the upward velocities are forced and limited to the lower levels with type B synoptic situations (Prezerakos and Angouridakis 1984). Type B, as mentioned above, appears to be warmer than type A at 500 hPa with a temperature range from −35 °C to −24 °C and a geopotential range from 5380 gpm to 5510 gpm.

5.3 *All snow-days*

The general argument in the discussion of Figs 8 and 9 is based on a rather small sample which may limit the validity of the derived conclusions. Therefore, Figs 10(a) and 10(b) have been drawn based on the data for all snow-days (40 days with 39 upper-air observations available). The temperature and geopotential values used in Figs 10(a) and 10(b) are more representative of the snowfall time than those used in Figs 8 and 9.

Figs 10(a) and 10(b) are more interesting because they provide, to a good approximation, the true relation between temperature and geopotential at the 850 hPa and 500 hPa isobaric surfaces during snowfall in Athens. On these figures are shown two kinds of regression lines, the continuous one drawn empirically by hand and the pecked one which has been derived mathematically.

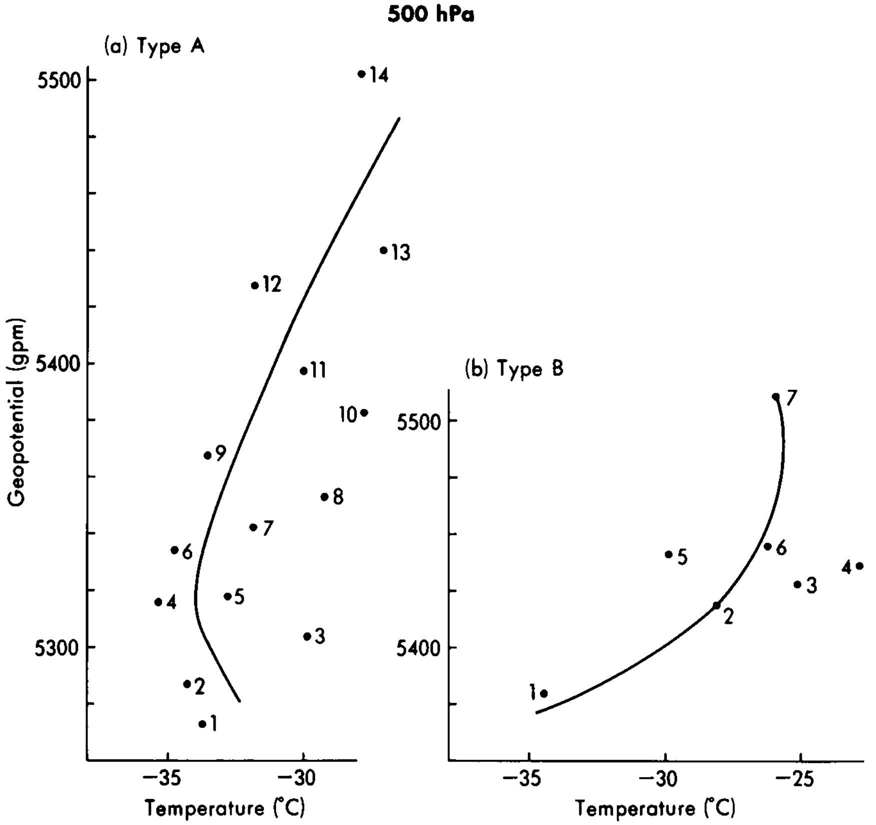


Figure 9. As for Fig. 8 but for the 500 hPa surface, the numbered points referring to Table IV.

Table IV. *Wind direction (degrees) and speed ($m\,s^{-1}$) and humidity (%) for the numbered points on the scatterplot of 500 hPa temperature against geopotential (Figs 9(a), 9(b)). The points are numbered consecutively from the lower axis and do not necessarily correspond to similarly-numbered points in Tables III or V or Figs 8 and 10*

Fig. 9(a) (Type A)

Point No.	Wind direction	Wind speed	Humidity
1	220	22	67
2	220	14	78
3	260	36	36
4	240	11	34
5	240	20	43
6	020	13	61
7	020	20	40
8	260	30	55
9	260	14	56
10	330	45	19
11	260	35	36
12	270	25	27
13	260	22	19
14	260	39	48

Fig. 9(b) (Type B)

Point No.	Wind direction	Wind speed	Humidity
1	290	25	56
2	320	30	33
3	040	30	20
4	330	41	18
5	290	22	42
6	310	31	19
7	020	28	40

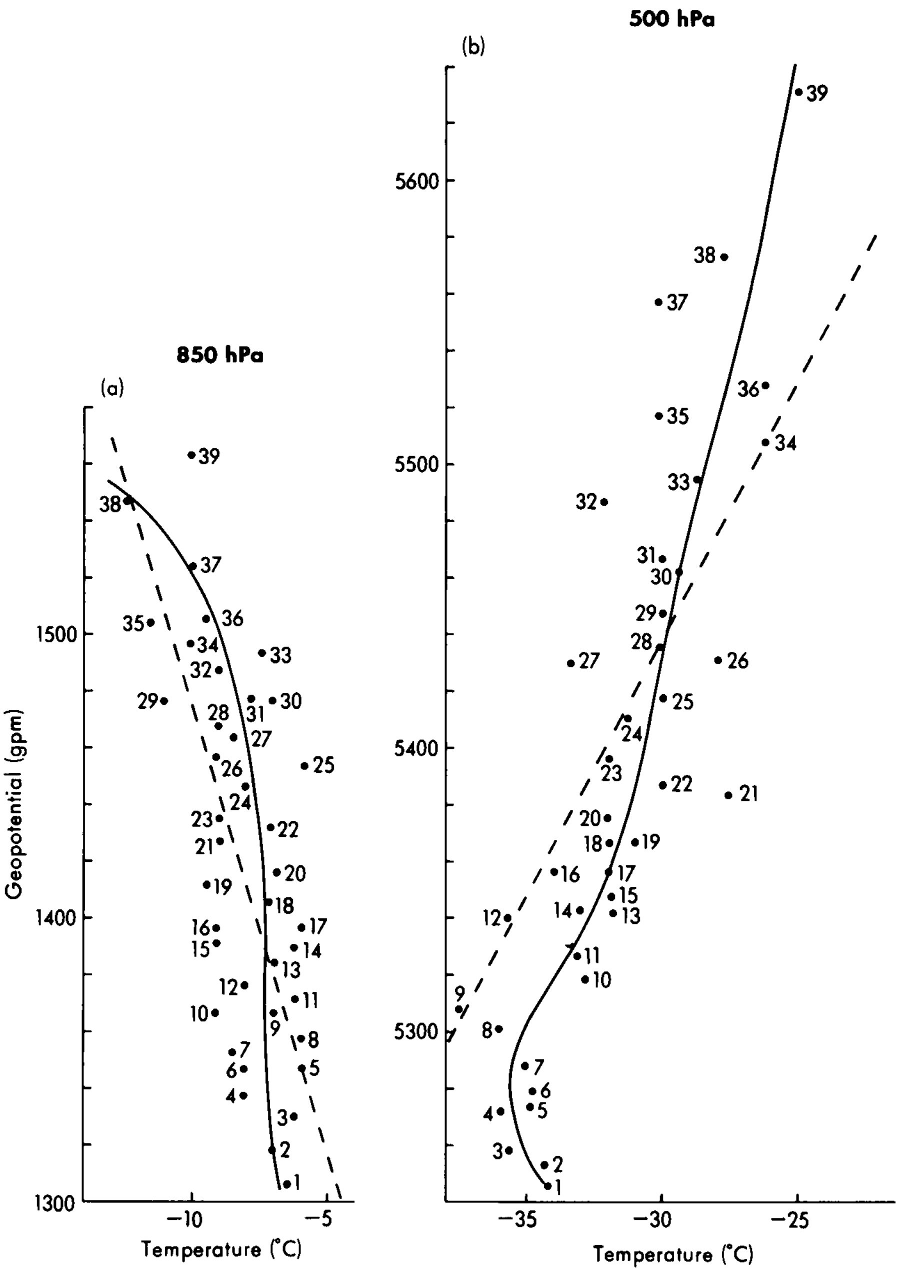


Figure 10. Regression lines of temperature and geopotential of (a) the 850 hPa surface and (b) the 500 hPa surface for both synoptic types A and B together for upper-air observations closest to the time of snowfall. The pecked line is the regression line derived mathematically, the continuous line is the regression line drawn empirically. The numbered points refer to Table V.

At 850 hPa (Fig. 10(a)) temperatures range from −6 °C to −12 °C. The mean temperature is −7.7 °C with standard deviation 3.2 K. The mean vector wind (Table V) is north-easterly, speed 14 m s^{-1} (about 28 kn) and the mean relative humidity 83%, implying 7–8 oktas cloud at this level. There is a considerable range in geopotential (250 gpm, from 1310 gpm to 1560 gpm). Furthermore it is important to mention that, while the temperature and geopotential vary in the same sense (Austin 1953, Angouridakis 1977), the curves of Fig. 10(a) and especially their upper parts show an inverse correlation, i.e. the temperature decreases as the geopotential increases. This is explained by the fact that the 850 hPa isobaric surface is influenced by the cold mass of the anticyclonic ridge which accompanied snowfall. In other words the 850 hPa geopotential in this case presents the same correlation with the MSL pressure as shown (Holton 1973), i.e. cold advection results in an increase in surface barometric pressure.

Table V. *Wind direction (degrees) and speed (m s^{-1}) and humidity (%) for the numbered points on the scatterplot of 850 hPa and 500 hPa temperature against geopotential (Figs 10(a), 10(b)). Identically-numbered points do not necessarily relate to the same ascent*

Fig. 10(a) (850 hPa)				*Fig. 10(b) (500 hPa)*			
Point No.	Wind direction	Wind speed	Humidity	Point No.	Wind direction	Wind speed	Humidity
1	020	27	72	1	040	04	73
2	360	10	81	2	350	25	60
3	350	08	81	3	360	30	50
4	360	15	83	4	260	10	46
5	020	13	83	5	350	25	48
6	030	09	84	6	220	22	67
7	020	12	77	7	350	20	60
8	020	14	83	8	200	21	79
9	020	15	85	9	340	40	55
10	350	12	87	10	240	02	43
11	150	04	91	11	290	20	45
12	020	20	90	12	350	38	45
13	020	19	85	13	040	05	43
14	040	21	85	14	300	08	52
15	350	19	99	15	300	10	40
16	010	15	85	16	260	12	68
17	030	15	90	17	280	10	40
18	340	22	80	18	290	10	55
19	350	16	70	19	240	15	79
20	340	20	70	20	020	09	32
21	350	17	75	21	330	45	19
22	350	24	69	22	300	10	60
23	340	22	70	23	040	05	40
24	020	15	84	24	260	13	77
25	200	03	81	25	260	30	45
26	030	14	86	26	290	38	19
27	020	10	92	27	080	29	21
28	020	10	85	28	040	10	45
29	020	15	82	29	260	20	45
30	020	16	83	30	350	16	41
31	040	15	81	31	330	40	30
32	020	18	85	32	260	30	30
33	020	15	82	33	330	23	19
34	360	19	84	34	260	30	30
35	350	17	87	35	330	10	30
36	110	05	91	36	330	10	35
37	020	23	80	37	330	10	40
38	350	12	91	38	330	22	19
39	020	15	96	39	110	04	36

The empirical regression line of 500 hPa geopotential with temperature (continuous line in Fig. 10(b)) is almost straight (with the exception of its lowest part) indicating that 500 hPa temperature and geopotential are linearly correlated during snowfall in Athens. However, in the area of the lower geopotentials there are four observations in which the trend is in the opposite direction. The mean temperature is −31.1 °C and the mean relative humidity 45%. The range of temperature is from −25 °C to −36 °C, whilst the range of the 500 hPa geopotential is from 5250 gpm to 5680 gpm. The 500 hPa mean vector wind (Table V), as mentioned above, is north-westerly (301°) at 10 m s^{-1}.

In Figs 10(a) and 10(b) the pecked lines are the calculated regression lines which show the linear correlation between geopotentials (H, gpm) and temperatures (T, °C). The equations of these lines are:

$$H = 21.98T + 6088.10$$

at 500 hPa with correlation coefficient $R_{500} = 0.77$ and standard error of estimation = 59.28, and

$$H = -27.86T + 1198.19$$

at 850 hPa with correlation coefficient $R_{850} = -0.70$ and standard error of estimation = 47.87. Thus conclusions drawn from the empirical regression lines are also supported by the mathematical regression line since there are no significant differences between them with the exception of the inverse relationship shown at the bottom of Fig. 10(b) which is ignored by the calculated equation. However, the F-test (Brooks and Carruthers 1953) shows that both correlation coefficients are significant at 0.01 level of significance.

6. Comparison of the results with some snow predictors

Various meteorologists have attempted to contribute to the improvement of snow forecasting by devising indices. Most of these are a combination of various meteorological parameters of the free atmosphere and the boundary layer. For this study a check of the performance of these indices, using upper-air data very close to the time of snowfall, has proved very useful. (Only the snow predictors and meteorological parameters which can be used easily and quickly in a forecasting office are considered in this context.)

The interest in the verification of the various snow predictors comes from the fact that the upper-air data used in this study differed from the time of snowfall occurrence at Hellinikon by an average of only one hour (Figs 10(a) and 10(b)). A similar study has been made by Lioki-Livada-Tselepidaki (1979) but most of the upper-air data used differed by up to 12 hours from the time of snowfall. Furthermore, snow-days in Athens in that study were even considered to include days on which it snowed only at Tatoi (a meteorological station located north of Athens, close to Parnitha Mountain, at an elevation of 237 m).

Many research meteorologists have dealt with the relation between snowfall and the 1000–500 hPa thickness (Murray 1952, Boyden 1964, Lamb 1955, Smith 1970, Lioki-Livada-Tselepidaki 1979). In this paper it has been found that the mean thickness value at Hellinikon for snow-days in Athens, and for the radiosonde observations nearest in time to the snowfall occurrence, is 5230 gpm with a standard deviation of 47 gpm (ranging from 5379 gpm to 5189 gpm). These results are in agreement with other researchers' results: Murray (1952) found values which ranged from 5160 gpm to 5230 gpm, but he also found cases with values as high as 5420 gpm. Lamb (1955) found a mean value of 5280 gpm (with a standard deviation 45 gpm) but some of his values reached 5440 gpm. Boyden (1964) found a mean value of 5260 gpm. The above researchers used British data from different stations, which may have caused small differences. It could be concluded that the results of the present research are essentially in agreement with the above-mentioned researches, since they do not show deviations much greater than those referred to.

The comparison of the results of this paper with those in Lioki-Livada-Tselepidaki's thesis (who found values in the range 5150 gpm to 5510 gpm, with a mean value of 5291 gpm) confirms what was expected, namely lower thickness values (because the upper-air observations are those nearest to the time of snowfall). Otherwise the results are in agreement with those of Lamb (1955), who found that the mean 1000–500 hPa thickness at coastal sites is, at the time of snowfall, about 60 gpm less than the corresponding value coming from other regions. (This is because the surface air at coastal stations will, in most cases, have been warmed by advection over a much warmer sea; the mean 1000–500 hPa temperature (i.e. thickness) thus needs to be considerably lower than it does at inland stations for the surface air to remain cold enough for snow to reach the ground.)

A more important factor than the 1000–500 hPa thickness is the 1000–850 hPa thickness, because this has a greater influence on whether the snow will reach the ground. Boyden (1964) found that the 1000–850 hPa thickness on snow-days ranged from 1280 to 1300 gpm; furthermore, he calculated a 50% probability of snow if the 1000–850 hPa thickness was less than 1295 gpm. Similarly, Lioki-Livada-Tselepidaki (1979) found that 1000–850 hPa thickness values of 1260–1350 gpm (with a mean value of 1300 gpm) occurred on snow-days, while Sahsamanoglou and Makrogiannis (1978), working on data from northern Greece, found a mean 1000–850 hPa thickness value of 1310 gpm, with a minimum value of 1280 gpm and a maximum of 1350 gpm.

In this research a mean 1000–850 hPa thickness of 1282 gpm (with a standard deviation of 34 gpm and a range from 1263 to 1309 gpm) was found. These values appear less than those found by Lioki-Livada-Tselepidaki (1979) for the same reasons as mentioned above. They are also less than the values derived by Sahsamanoglou and Makrogiannis (1978) using surface observations covering the whole of northern Greece, often some hours away from the 1200 GMT radiosonde ascent at Thessaloniki/Mikra (WMO No. 16622). In the present case, surface and upper-air data are from the same station and are almost simultaneous.

It is commonly accepted that the freezing level plays the most important role in determining whether the precipitation that reaches the earth's surface will be snow, sleet or rain. Lowndes *et al.* (1974) found that the probability of precipitation falling as snow is greater than 50% when the freezing level is 38 to 41 hPa above the earth's surface, and that this probability rose to 90% when the freezing level was within $\pm$ 4 hPa of the surface. In this work the freezing level averaged 980 hPa, some 38.6 hPa above the surface (since the mean barometric pressure at station level at the time of radiosonde release was 1018.6 hPa Prezerakos and Angouridakis (1979)). This value is in agreement with the results of Lowndes *et al.* (1974) wherein an index I_w (defined as surface pressure minus pressure at wet-bulb freezing level) was introduced to define the precipitation type. It was found that when $I_w \leqslant 10$ hPa the probability of the precipitation falling as snow was 50%, but when $I_w \leqslant 0$ rain or sleet never occurred. In this work the mean value of I_w was found to be 8.6 hPa (standard deviation 1.4 hPa), a fact which suggests that the index could be used in deciding whether the precipitation will be snow or not.

Another index which seems very useful in determining likely precipitation type was proposed by Booth (1973). His conclusion was that when the sum of the surface temperature T and dew-point T_d is $\leqslant 1$ °C precipitation was always snow. Here care is needed in the application of the index, for it does not in itself help forecast precipitation but only helps in the determination of precipitation *type* (the likelihood or otherwise of precipitation being forecast by other methods; obviously there will be cases when the sum $T + T_d \leqslant 1$ °C, under clear skies for example). In all cases of snow in Athens the sum $T + T_d$ appeared to be less than 1 °C, with a mean value of −0.6 °C.

Another useful meteorological parameter in determining precipitation type is θ_w at the 850 hPa level (Bradbury 1977) and/or the wet-bulb temperature T_w. Bradbury found that when θ_w at 850 hPa was less than 3 °C then the probability that precipitation will fall as snow was 60% or more. Lioki-Livada-Tselepidaki (1979) found that values of θ_w at 850 hPa ranged between −4 °C and +10 °C, with a mean value of +2.4 °C; in this work the mean value of θ_w at 850 hPa was +0.5 °C (with values ranging between −6 °C and +3 °C). The average wet-bulb temperature at 850 hPa was found to be −9 °C in this research, whereas Lioki-Livada-Tselepidaki found values ranging between −13.3 °C and +8 °C. These latter boundary values, compared with the results obtained by Booth (1973) and during the current investigation, indicate an extension of the 850 hPa wet-bulb temperatures to the positive end point. One final point arising from the observations used in this work is that they show the lower part of the troposphere to be much colder than was shown by Lioki-Livada-Tselepidaki's corresponding values.

7. Conclusions

The main conclusions derived from this study are briefly as follows:

(i) There were 23 synoptic systems associated in the snowfall in Athens in the period under consideration (1956–73) — fifteen cases belong to type A, and eight cases to type B. Thus the number of F days are 15 for type A and 8 for type B, whereas the total number of snow-days is 40.

(ii) The atmospheric structure above Athens during F−2, F−1, F and E days has been found to be potentially stable ($\partial\theta_w/\partial z > 0$). However, for type A on F day the stability decreases in the lower levels and the relative humidity increases, the same also happens for the type B and even on E day the layer from surface up to 850 hPa appears to be potentially unstable.

(iii) The characteristics of the atmospheric structure during snowfall are:

(a) stable stratification

(b) cold air mass (mean temperature at 850 hPa −7.5 °C, mean temperature at 500 hPa −31.5 °C)

(c) increased relative humidity ($T - T_d < 3$ K in the layer from 950 up to 800 hPa)

(d) northerly winds from 950 to 750 hPa, backing higher up, to north-westerly. At the surface type A is characterized by north-westerly winds, and type B by north-easterly winds.

(iv) Temperature values derived from Hellinikon 0000 GMT upper-air observations showed that temperature decreases from F−1 day to F day and increases on END day at all standard levels. The warmest day at 200 hPa is F day. Also type A appears to be colder than type B at almost all standard levels and warmer on F−1, F, E days at 850 hPa level.

(v) The standard isobaric surface geopotentials decrease from day F−2 to F and they increase on E day apart from the geopotential of 200 hPa isobaric surface which continues decreasing on the E day.

(vi) Various indices defining the precipitation types which have been introduced over a long period of meteorological practice have been compared with the results of this research. Characteristic values of various meteorological parameters associated with these indices for snowfall time are:

(a) mean value of 1000–500 hPa thickness = 5230 gpm with a standard deviation 47 gpm

(b) mean value of 1000–850 hPa thickness = 1282 gpm with standard deviation 34 gpm

(c) mean distance of the freezing level from the ground = 38.6 hPa

(d) mean value of index I_w = 8.6 hPa with a standard deviation 1.4 hPa

(e) $T + T_d \leqslant 1$ °C on average

(f) θ_{w850} = 0.5 °C on average

(g) T_{w850} = −9 °C on average.

These values when compared to the corresponding values found by other researchers showed that the lower part of the troposphere was much colder than the observations used by previous researchers had shown.

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An analysis of extreme rainfalls observed in Jersey

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Summary

A depth–duration–frequency analysis of extreme rainfalls observed on the island of Jersey has been made using the methods described in the Natural Environment Research Council's *Flood Studies Report*, Volume II. Results of a comparison of the Jersey observations and the Flood Studies model (which was based on rainfall observations on the mainland of the United Kingdom) are given. It is concluded that for most practical purposes the Flood Studies model is applicable to extreme rainfalls on Jersey.

1. Introduction

On 31 May 1983 a violent shower occurring between 20 and 21 GMT produced an hourly rainfall total of 43.3 mm at Jersey Airport, with a 09 to 09 GMT total of 52.2 mm at the nearby St Peter's Rectory. Serious flooding resulted in the west of the island with consequent insurance claims and questions regarding the effectiveness of drainage systems and practices. On the basis of records dating from the last century both the Jersey Meteorological Department and the news media described this event as a rare occurrence. However, less than a week later, another thunderstorm approached the island from the south-west in an unstable airstream across France. Again, flooding occurred but this time the worst-affected area was in the north-west of the island, although some unfortunate farmers and householders suffered damage to their properties on both occasions. Soon after midnight on 5 June the rain-gauge at the airport recorded 28 mm in less than an hour and other daily gauges recorded higher totals (such as the 53.3 mm at an unofficial site at St John's Rectory).

Although surprising, and of considerable concern to enquirers who suffered damage or loss, this repetition of events is understandable to the meteorological statistician with a clear idea of the 'return period' concept. However, these two storms did highlight the need to produce extreme rainfall data for design or planning purposes for Jersey and it was decided to study the application of Volume II of the *Flood Studies Report* (Natural Environment Research Council 1975b), hereinafter referred to as the FSR, to the island. The FSR provides design rainfall estimates for the United Kingdom† based on plentiful and comprehensive precipitation data. Applicability of the findings of the report to Jersey could not be assumed without investigation, since the proximity of Jersey to France means that the island often receives heavy rainfalls in spring and summer originating from thundery activities over Brittany and Normandy where temperatures can be much higher than those recorded over southern England. The precipitation regime of Jersey might thus differ sufficiently from that of coastal areas of southern England to invalidate the FSR approach.

After discussions between the Jersey Meteorological Department and the Agriculture and Hydrometeorology Branch of the Meteorological Office at Bracknell, it was decided to carry out a co-operative study of Jersey rainfall data to assess whether rainfall regimes were sufficiently alike to those

* Formerly of the Meteorological Office, Bracknell.

† The United Kingdom comprises England, Wales, Scotland and Northern Ireland, but not the Channel Islands or the Isle of Man.

in the United Kingdom to apply the FSR methods, and to produce estimates of extreme-event occurrence. These are of potential interest not only to climatologists but also more practically to engineers, farmers and insurance companies, and those individuals who have to cope with flooding on a local scale. It was intended that the benefits of a statistical extreme-value analysis, using records of considerable length, would be shared by both the organizations.

2. The data and their extraction

The climate of Jersey, located as shown in Fig. 1, has been described by Blench (1967) who points out that the major factor influencing rainfall is the occurrence of moist westerly winds from the Atlantic Ocean. Average monthly rainfall amounts are fairly constant throughout the year but the average annual rainfall (AAR) is greater in the north of the island where the higher ground is situated (see Fig. 2). The island is also subject to heavy rainfalls from convective storms which form over France and move northwards.

Long rainfall records are available from a number of sites on Jersey. The rain-gauge sites, types of data and record lengths are given in Table I.

For each month the maximum rainfalls for the following durations were extracted from the records: 1, 2, 4, 6 and 12 hours from Jersey Airport and St Louis Observatory, 1 and 2 days from all sites; 3, 4, 8 and 25 days from Millbrook Reservoir.

Durations of from 1 to 12 hours start on integral GMT hours ('clock hours'), days start at 09 GMT and all durations are allotted to the month in which they start. The analysis involves the series of maximum rainfall amounts (for a specified duration) in each year, or in a particular month or season of that year, and these are referred to as the annual, monthly and seasonal maxima respectively (as in the FSR).

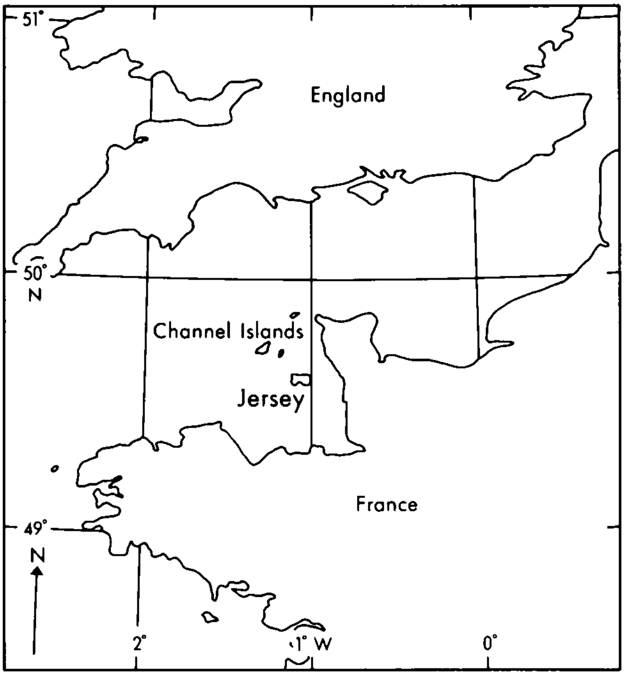

Figure 1. Location of Jersey in relation to England and France.

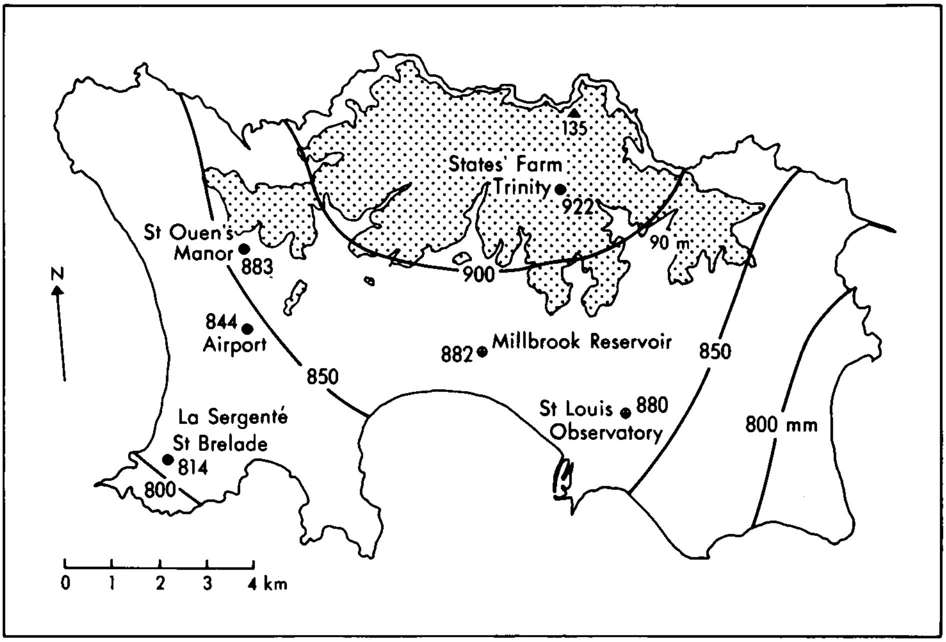

Figure 2. Location of rain-gauges on Jersey with average annual rainfalls in millimetres (1941–70 averages). Dotted area indicates ground over 90 m above mean sea level.

Table I. *Rainfall records for Jersey. The sites referred to are shown in Fig. 2*

Site	Universal Transverse Mercator (UTM) grid ref. Zone 30U WV (554)	Altitude *metres*	Daily (D) or hourly (H) values	Period of record
Jersey Airport	583513	84	H and D	1951–82
St Louis Observatory	662493	54	H	1940–43, 1947–82
			D	1894–1920, 1925–83
St Brelade	566486	58	D	1917–41, 1948–83
St Ouen's Manor	583530	79	D	1917–34, 1948–74
Trinity	650540	104	D	1936–41, 1948–83
Millbrook Reservoir	632507	16	D	1912–82

3. Analysis of extreme-value data

A good introduction to the use of statistics of extremes in meteorology is provided by Tabony (1983); only an outline is given here.

Consider a series of observations of the maximum values of some variable occurring in equal time intervals arranged in ascending order 1 to n, for instance, the maximum 1-hour rainfalls in each year R_1 to R_n. With the mth value R_m we need to associate a probability F that it will not be exceeded in any year. The median value of this probability can be estimated by Chegodayev's formula (see FSR):

$$F = (m - 0.31)/(n + 0.38). \tag{1}$$

The probability of R_m being exceeded is $1 - F$ and the average duration in years between exceedances is

$$T = 1/(1 - F) \tag{2}$$

where T is the return period. For instance, if $n = 10$ and $m = 5$, $F = 0.452$ and $T = 1.82$ (years) for the associated value R_5. F is also called the 'plotting position' and the plotting positions given by equation (1) were used in the analysis in the FSR.

Annual, monthly and seasonal maxima of rainfall amounts for specified durations are frequently used in extreme-value analysis because they are usually readily retrievable from convenient data sets. However, their use overestimates the return period of exceedance for a specified amount and the effect is important, especially for return periods of less than 5 years. This arises because in some years the second (or third etc.) largest rainfall, which is ignored in the annual maximum series, can be greater than the maximum for another year which is included in the series. As a consequence a more accurate estimate of the return period can be obtained by selecting and ordering all exceedances of the specified amount into the 'partial duration' series. Fortunately a return period T_E from the partial duration series can be related to its corresponding T_M in the annual maximum series by the formula (Langbein 1949)

$$T_E = 1/\{\ln T_M - \ln(T_M - 1)\}$$

so that the convenience of the analysis of maxima in fixed intervals can be retained. For $T_M \geqslant 5$ years, $T_E \approx T_M - 0.5$ years and the rainfalls corresponding to the two periods are usually negligibly different.

For a particular duration, to estimate the rainfall amount for a specified T, interpolation within or extrapolation beyond the T values for each member of an ordered series given by equations (1) and (2) is required. The theory of extreme values indicates that the ordered maxima can be related to the 'reduced variate' $y = -\ln\{\ln(1/F)\}$.

As an example, Fig. 3 shows the annual maximum rainfall of 1-hour duration at Jersey Airport (32 points) plotted against y_M, where $y_M = -\ln\{\ln(1/F_M)\}$, and $F_M = (T_M - 1)/T_M$, on linear scales. The values lie on a fairly smooth curve with an indication, mainly from the two uppermost points, of a slope increasing with y_M. Denoting the general extreme-value variable by x, a curve of the form

$$x = x_0 + (\alpha/k)\{1 - \exp(-ky_M)\} \tag{3}$$

can be fitted to the extreme values where α is the scaling factor and k is the slope curvature factor. The slope $dx/dy_M = \alpha \exp(-ky_M)$ so that for small k, $dx/dy_M \approx \alpha$ while for the straight-line case $k = 0$, $dx/dy_M = \alpha$ and hence $x = x_0 + \alpha y_M$. For T less than 10 years, maximum rainfalls as in Fig. 3 can usually be represented by a line of small or zero k.

Results in this paper are quoted for standard return periods of $T_E = 0.5$, 1 and 2 years and T_E or $T_M = 5$, 10, 20 and 50 years. With the record lengths available, rainfalls for T up to 50 years can be estimated well because they are covered by the range of individual T values given by equations (1) and (2); beyond 50 years extrapolation of the extreme-value distribution is involved with increasing uncertainty. For Millbrook Reservoir with its longer record of 71 years, the upper limit of T can be increased to 100 years.

From individual maxima plotted as in Fig. 3 reliable observed values for the standard return periods are calculated using one of three methods:

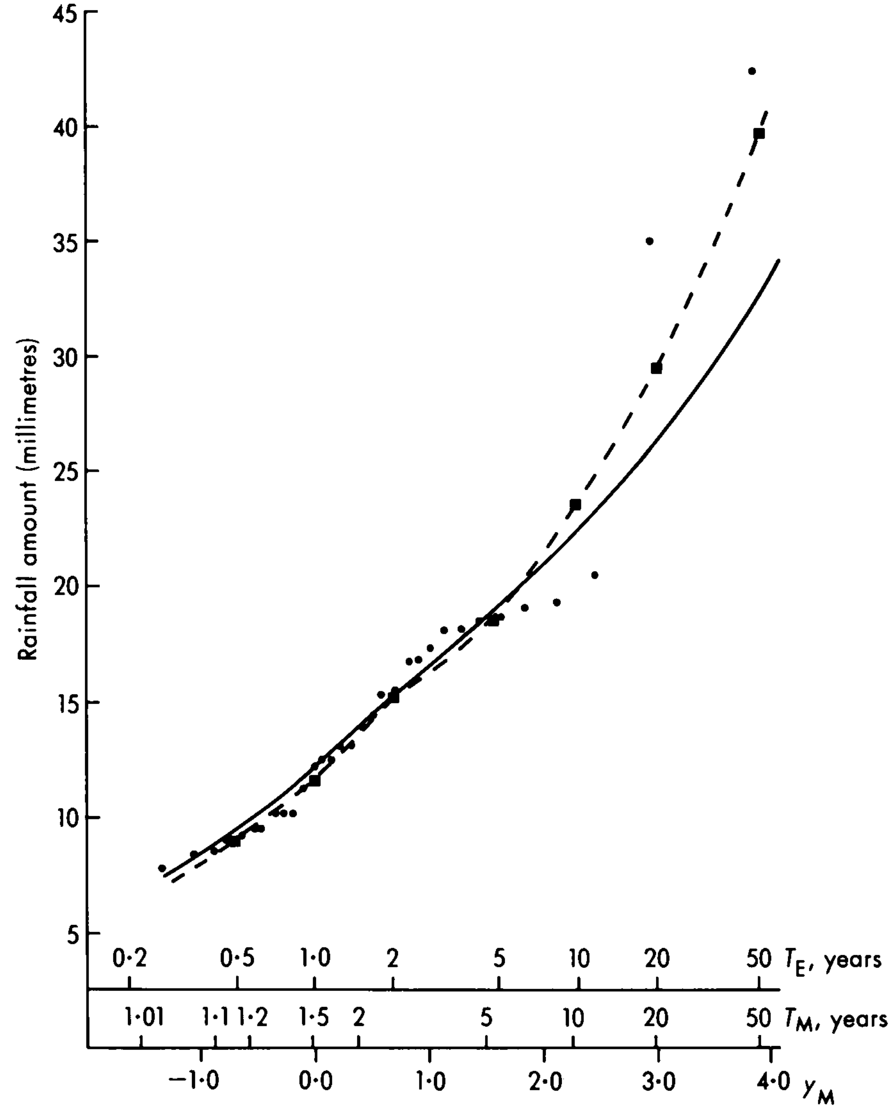

Figure 3. Analysis of annual maximum 1-hour rainfalls at Jersey Airport (32 years).
● observed annual maximum 1-hour rainfalls
■--■ adopted observed values for standard return periods
—— FSR model

(i) *For T_E = 0.5 and 1 year (2M and 1M values)*

Jenkinson (1975) showed that if a set of maxima in ascending order was divided by quartiles then the geometric mean of the values below the lower quartile is a good estimate of the twice-yearly rainfall (denoted by 2M), i.e. for T_E = 0.5 years, and the geometric mean of the values lying between the lower quartile and the median gives the 1M value (T_E = 1 year).

(ii) *For T_E = 2 years (M2 value)*

The geometric mean of the values in the interquartile range is a good estimate of the value for $T_M = 2$ years which is equivalent to $T_E = 1.44$ years. The value for $T_E = 2$ years is calculated by adding to the T_M = 2-year value a correction of $+0.325\alpha$. The factor 0.325 is the difference between the values of y_M for $T_E = 2$ years (or $T_M = 2.54$ years) i.e. $y_M = 0.692$ and for $T_M = 2$ years i.e. $y_M = 0.367$; strictly this factor applies only to linear extreme-value distributions but in practice the error in the correction is negligible.

(iii) *For $T_M \geqslant 5$ years (M5, M10, M20, M50, M100)*

Curves of the form of equation (3) are fitted to the values over the whole range of y_M using the maximum likelihood technique (Jenkinson 1969). Interpolated values appropriate to the standard return periods taken from the curves are regarded as the best estimates based on these observations.

4. Errors of 2M, 1M ... estimates

For a linear distribution of extreme values an estimate of the standard error (S.E.) of a point on the line is given by the FSR (Natural Environment Research Council 1975a):

$$\text{S.E.} = (\alpha/n^{1/2})(1.11 + 0.52y_M + 0.16y_M^2)^{1/2}$$

where n is the number of observations in the sample. This expression is also adopted here where the distribution of extreme values is close to being linear.

5. The Flood Studies model

In the early 1970s a large number of rainfall records for sites in the United Kingdom were analysed by the Meteorological Office as a part of the background work for the FSR. It was later shown that the depth–duration–frequency relationship for all locations could be represented by a simple computer-based model described by Keers and Wescott (1977). To estimate the rainfall amount for a particular duration (in the range 5 minutes to 25 days) and return period (greater than 0.5 year) the model requires four input parameters to be specified. These are:

(i) 60-minute M5 rainfall;
(ii) 2-day M5 rainfall;
(iii) 25-day M5 rainfall; and
(iv) average annual rainfall.

The 60-minute M5 rainfall (which is not constrained to start on a clock hour) has been found to be greater than the 'constrained' clock-hour value by a factor of 1.15.

The first step in the application of this model is to estimate (essentially by interpolation within the above values) the M5 value for the duration of interest. This does not necessarily require all the input items (i) to (iv) above to be specified — for instance, to estimate the 1-day M5 rainfall requires only items (i) and (ii) to be known. The second step is to calculate the value for the required T_M from the M5 value from step 1. For $T_M \geqslant 5$ years the relationship

$$\text{M}_t = \text{M5}\cdot\exp\{c(y_M - 1.5)\}$$

is used, in which c is a constant which depends upon the M5 value, and y_M is the reduced variate corresponding to T_M. For $T_E = 0.5$, 1 or 2 years the corresponding 2M, 1M and M2 values are available as a tabulated percentage of the M5 value, for different ranges of M5.

6. Results

(i) *General*

Comparisons have been made between rainfalls for various durations from 1 hour to 25 days and return periods from 0.5 to 100 years as determined from the Jersey observations (denoted by O) and those from the FSR model (F) as calculated by the Keers and Wescott computerized method.

As an example, the results for annual maximum 1-hour rainfalls at Jersey Airport are tabulated against return period in Table II. Differences ($O - F$) are given in millimetres, as percentage relative difference ($\{(O - F)/F\} \times 100\%$) and as a fraction of the estimated S.E.

Table II. *Comparison of Jersey Airport and FSR rainfalls of 1-hour duration*

Return period *years*	Observed rainfall (O) and S.E. *millimetres*	FSR rainfall (F) *millimetres*	$(O-F)$ *millimetres*	$(O-F)/F$ *per cent*	$(O-F)/$S.E.
0.5	9.0 (0.8)	9.5	-0.5	-5.3	-0.63
1	11.8 (0.8)	12.1	-0.3	-2.5	-0.38
2	15.3 (1.0)	15.3	0.0	0.0	0.00
5	18.6 (1.4)	18.9	-0.3	-1.6	-0.21
10	23.6 (1.8)	22.4	1.2	5.3	0.67
20	29.7 (2.2)	26.4	3.3	12.5	1.49
50	39.8 (2.7)	32.6	7.2	22.1	2.67

The observed and FSR rainfalls in Table II are also shown in Fig. 3 described previously. It can be seen that the estimated observed values for the standard return periods obtained using the methods described in section 3 are a good representation of the basic values over the range $T = 0.5$ to 10 years and particularly for $T \leqslant 5$ years. For $T \leqslant 10$ years the relative difference is less than 6% and the difference relative to the S.E. is in the range -0.7 to +0.7. For $T > 10$ years larger differences appear which are discussed below.

(ii) *Annual maxima*

In Fig. 4 are plotted the percentage relative differences for durations of 1, 4 and 12 hours for Jersey Airport and St Louis Observatory for the range of return periods. For $y = 5$ years the differences are small or zero (because of the method of application of the FSR); for $T < 5$ years they vary systematically from negative values for 1 hour to positive values for 12 hours for both sites, but all differences are within 10%. For $T > 5$ years the differences again vary systematically and in the same way for both sites but this time are positive for 1 hour and negative for 12 hours duration. This indicates that Jersey experiences larger rainfalls of short duration and smaller rainfalls of longer duration than the United Kingdom for the rarer events. These rainfalls are associated with summer thunderstorms ahead of upper troughs west of France, and with cold fronts weakening and becoming quasi-stationary near the Channel Islands. Nevertheless, the FSR model fits these sub-daily rainfall observations within -20% to + 30% for T up to 50 years.

Fig. 5 shows results in the same style for 1-day duration rainfalls from all the six sites. For $T < 5$ years the differences tend to be positive but are less than 11%, while for $T = 50$ years the errors are larger, ranging between ±15%. There is no obvious relationship between sites with differences of the same sign and so these ranges of differences can be regarded as being typical for a selection of sites randomly situated over this island.

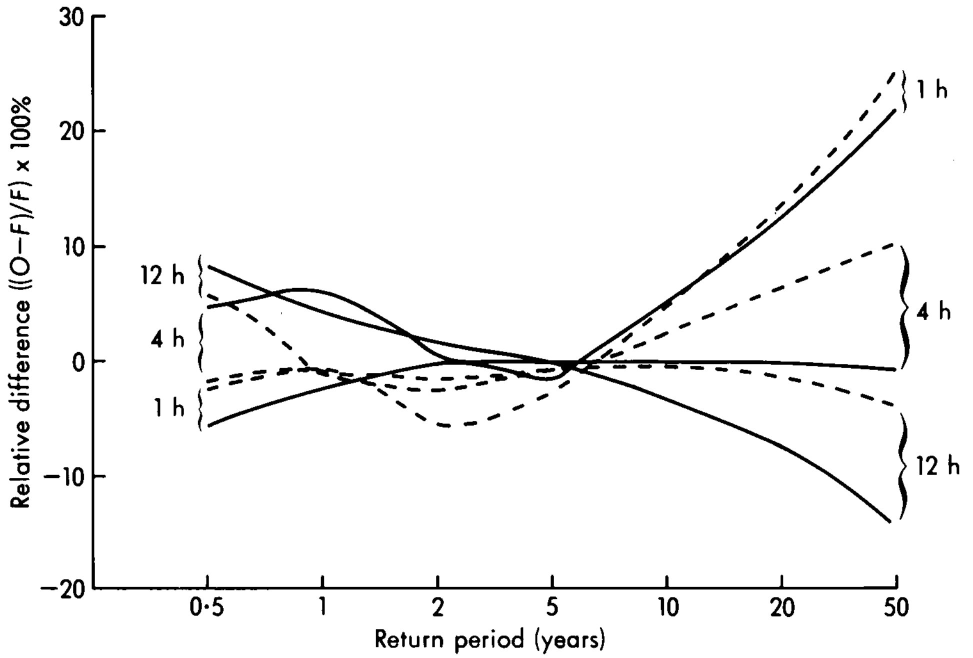

Figure 4. Variation of percentage relative differences (observed – FSR) for sub-daily duration rainfalls at Jersey Airport (——) and St Louis Observatory (- - -).

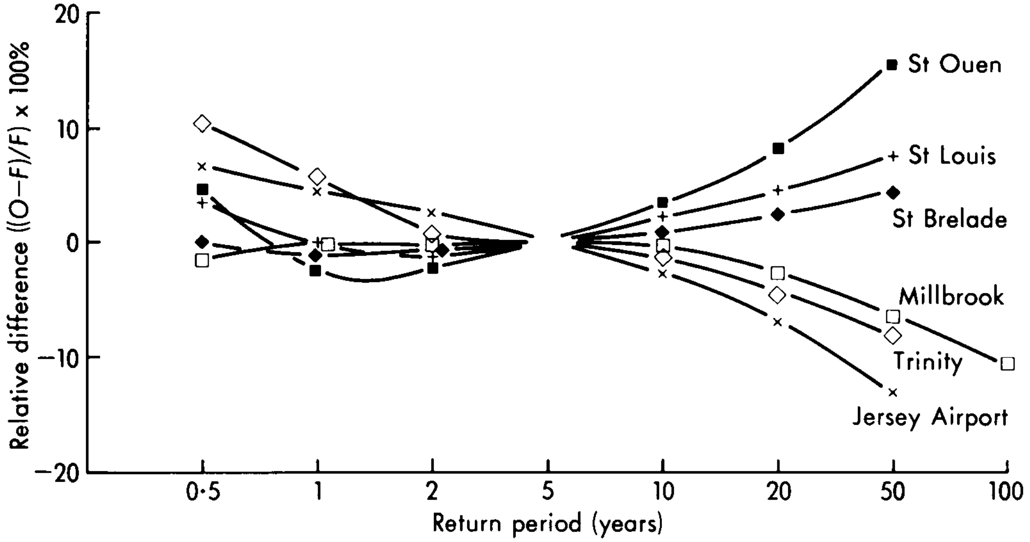

Figure 5. Variation of percentage relative difference (observed – FSR) for 1-day duration rainfalls.

Results for 2-, 4-, 8- and 25-day durations for Millbrook Reservoir are shown in Fig. 6. The differences here are mainly negative, increasingly so for the larger T, but even so for the whole range $0.5 \leqslant T \leqslant$ 100 years the differences are all smaller than 16%. From the similarity of the 1-day duration (Fig. 5) and the 2-day to 25-day duration (Fig. 6) curves for Millbrook it is suggested that the 2-day to 25-day duration curves for the other sites will show approximately the same spread about their 1-day duration curves.

The largest individual difference relative to the S.E. is for St Louis Observatory, 1-hour duration, T = 50 years, at +3.3; otherwise the differences are usually in the range ±2.0 S.E.

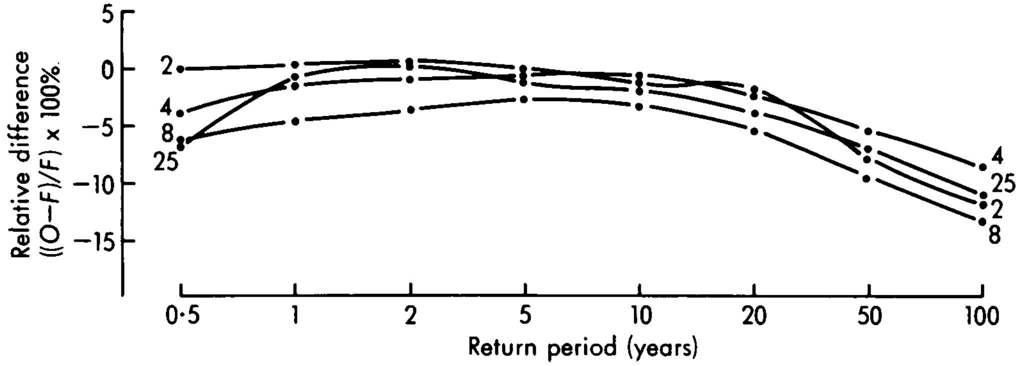

Figure 6. Variation of percentage relative difference (observed − FSR) for 2- to 25-day duration rainfalls at Millbrook Reservoir.

(iii) *Monthly and seasonal maxima*

Results of depth-duration-frequency analyses of annual maxima have been described, but for some purposes similar analyses of rainfalls in particular months or seasons need to be considered. Such analyses of the Jersey data proceeded exactly as for annual maxima except that the maximum rainfalls extracted were for each month and season (winter — November to April, summer — May to October, to agree with the FSR).

In the FSR the variations of M5 for monthly and seasonal maxima are expressed as a percentage of the M5 for annual maxima, for durations from 1-hour to 25 days and for a range of average annual rainfalls; the Jersey results are presented here in a similar way. The variation of percentage with month from the FSR model (AAR range 800–1000 mm) is shown in Fig. 7 in comparison with the percentages from the Jersey data for durations of 1 hour, 1 day and 25 days only for clarity. For each duration the form of the curves from the two sources is very similar but there are small differences of up to 16%, mainly in February and March, and October to December. Differences are also apparent, but to a lesser extent, in the seasonal results also shown in Fig. 7.

In general then it appears that the FSR model of the depth-duration-frequency relationship applies well to the Jersey extreme rainfalls within the percentage relative differences indicated, but adjustments based on Figs 3–7 can be applied if required.

7. Representative depth-duration-frequency curves for Jersey

(i) *Annual maxima*

The differences quoted in section 6 indicate the extent to which the FSR model can represent depth-duration-frequency relationships for both United Kingdom and Jersey extreme rainfalls, for specified input parameters (i) to (iv) in section 5. They are not necessarily representative of the errors of predicted

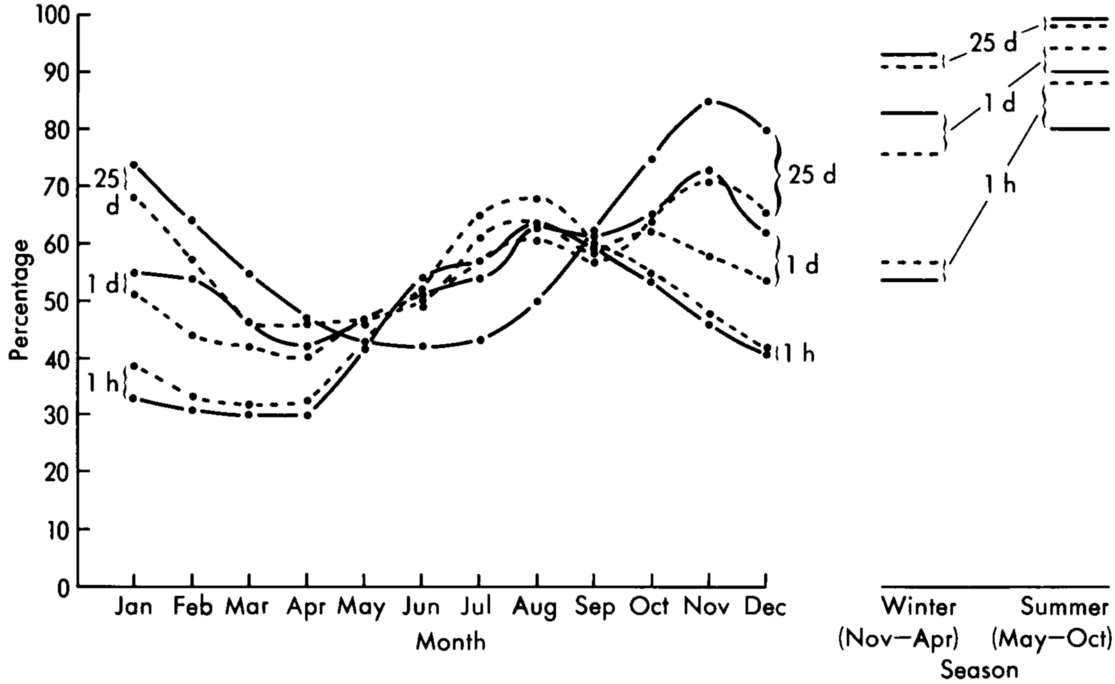

Figure 7. Comparison of variations of M5 for Jersey (——) and FSR (- - -) for monthly and seasonal maxima as a percentage of M5 for annual maxima of 1-hour, 1-day and 25-day durations.

extreme rainfalls or return periods for an ungauged location, which are dependent upon the accuracy with which the input parameters can be estimated. The available values of these parameters are given in Table III.

Table III. *Values of FSR model input parameters for Jersey sites*

Location	M5 rainfalls 60-minute	2-day	25-day	Average annual rainfalls 1941–70 averages
			millimetres	
Jersey Airport	21.4	48.7	—	844
St Louis Observatory	21.3	53.9	—	880
Trinity	—	55.3	—	922
St Ouen's Manor	—	50.1	—	883
St Brelade	—	50.6	—	814
Millbrook Reservoir	—	55.8	196.6	882

From the entries in Table III it can be seen that there are no large variations of the M5 values of 60-minute, 2-day or the AAR over this small area. We can therefore adopt the following representative set of values for Jersey:

60-minute M5	21.9 mm
2-day M5	52.4 mm
25-day M5	200.0 mm
AAR	880.0 mm

— which differ by no more than 7% from the individual values in Table III (in the FSR model an x% change in an M5 value for a specified duration changes the M5 value for a similar duration by about the same amount). These representative values have been used with the FSR model to produce the depth–duration–frequency curves (also known as design rainfall curves) in Fig. 8 generally applicable to all locations on the island. Although the record lengths only cover the range of T up to 50–100 years the curves in Fig. 8 have been extended to $T = 1000$ years for guidance.

(ii) *Seasonal maxima*

From the observed percentage curves in Fig. 7 (and for other intermediate durations) and the M5 values in Fig. 8 the generally applicable curves for monthly and seasonal M5 rainfalls for Jersey have been constructed and are shown in Fig. 9. The curves for duration of less than 60 minutes were derived solely from the methods in the FSR as no sub-hourly rainfalls for Jersey sites were extracted from the records. The curves show well the contrary seasonal changes in short- and long-duration M5 rainfalls

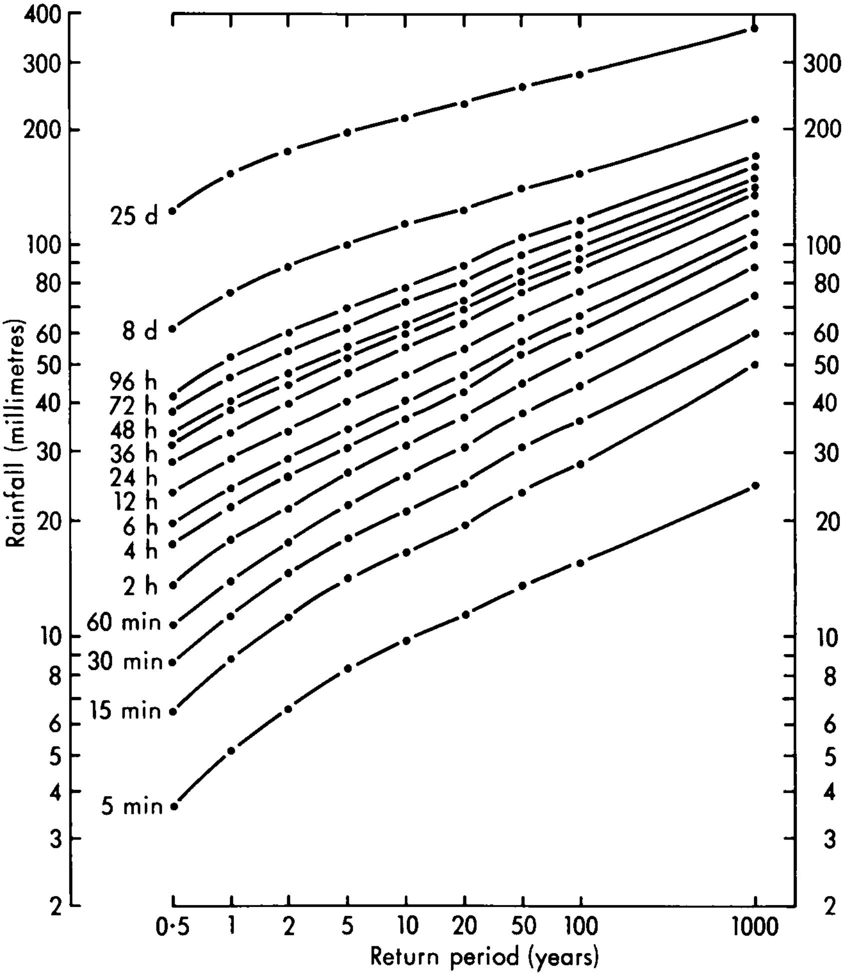

Figure 8. Depth-duration-frequency relationship for Jersey.

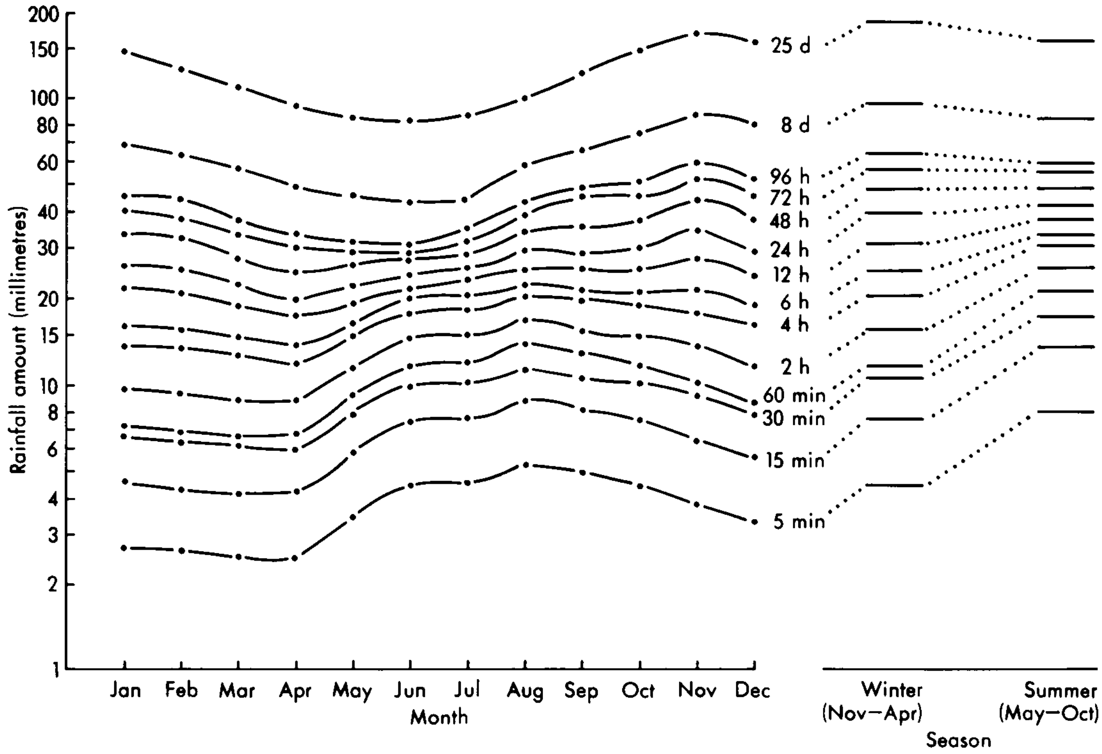

Figure 9. Monthly and seasonal variation of M5 rainfalls for Jersey.

with rainfalls of duration less than 60 minutes being much smaller in winter than in summer (owing to the influence of summer convective storms), and 1-day rainfalls being larger in winter than in summer owing to the greater influence of prolonged rainfalls from winter depressions.

8. Conclusions

Extreme rainfalls for durations from 1 hour to 25 days from six sites on Jersey with records from 32 to 71 years have been analysed. The rainfalls for return periods from 0.5 to 50 years are usually within ±15% of the amounts inferred from the *Flood Studies Report* model, for model input parameters (M5 values of the 60-minute, 2-day and 25-day duration rainfalls and the annual average rainfall) observed at gauged sites. Further errors of up to ±7% can result from errors in estimating the input parameters from ungauged sites.

It is concluded that for most practical purposes the FSR model is applicable to extreme rainfalls in Jersey and the neighbouring Channel Islands.

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Notes and news

Pressure to change

The use of hectopascals (hPa) in the paper by Dr Prezerakos published in this issue has recalled to us the following short poem by Mr Ernest Gold CB, DSO, OBE, FRS published over the initials E.G. in *Symons's Meteorological Magazine* for December 1919 (Vol. 54, p. 136). The occasion was the decision by a majority vote of the Congress of Directors of Independent Meteorological Services, held in Paris, to adopt the millibar as the unit of atmospheric pressure in preference to the millimetre of mercury. (Mr Gold, Deputy Director of the Office during and immediately after the Second World War, played an outstanding role in organizing international meteorological co-operation.)

The Directors of Science in Congress assembled,
Agreed that in future no discord should mar
The values of pressure so often dissembled
By units derived from a platinum bar.
The inch and the metre and gravity trembled,
As into the Congress there tripped lightly skipping
An innocent damsel who'd just 'scaped a whipping;
Her name in plain English was Miss Milly Barr.
Dear Milly Barr
Bjerknesian star;
A thousand of you
Shall be ever our cue,
As our standard of pressure wherever we are.

We look forward to receiving a similar poetic salute to Hecto(r) Pascal.

Books received

The listing of books under this heading does not preclude a review in the Meteorological Magazine at a later date.

Atmospheric ozone, edited by C. S. Zerefos and A. Ghazi (Dordrecht, Boston, Lancaster, D. Reidel Publishing Company, 1985. £69.95) contains progress papers in atmospheric ozone research which were presented at the Quadrennial Ozone Symposium held in Halkidiki, Greece, from 3 to 7 September 1985. The papers are grouped into nine chapters corresponding to the nine sessions of the symposium, the titles of which are as follows: chemical–radiative–dynamical model calculations, ozone–climate interaction, observations of relevant trace constituents and their budgets, analysis of ozone observations, recent developments in observational techniques, interaction of ozone and circulation, laboratory measurements of absorption cross-sections and of chemical rate constants, radiation topics relevant to atmospheric ozone, and non-urban tropospheric ozone.

The global climate, edited by John T. Houghton (Cambridge University Press, 1984; first paperback edition 1985. £11.95) is a paperback edition of this volume issued in hardback last year. The aim of the book is to introduce scientists and students working in climate research or related fields to the global climate problem, and the complex processes and interactions which play a part in climatic change. The text is centred around the World Climate Research Programme, whose aims are to determine, firstly, to what extent changes in climate can be predicted, and secondly the extent of man's influence on climate.

Climatic hazards in Scotland, edited by S. John Harrison (Norwich, Geo Books, 1985. £8.00) covers the proceedings of the joint Royal Scottish Geographical Society and Royal Meteorological Society symposium held at the University of Stirling in June 1984. The Scottish experience of excess weather conditions is reported. Extremes in rainfall and resultant flooding conditions, and windthrow damage in the uplands, are discussed. Also dealt with is the general influence of climatic extremes, with regard to insurance aspects and research, and also as regards the city of Glasgow.

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NOTICE

It is requested that all books for review and communications for the Editor be addressed to the Director-General, Meteorological Office, London Road, Bracknell, Berkshire RG12 2SZ and marked 'For Meteorological Magazine'.

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