INVESTIGATIONS OF SELECTED EUROPEAN CYCLONES BY MEANS OF SERIAL ASCENTS

Case 4: February 15—17, 1935.

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Introduction.

Sounding-balloon ascents closely succeeding each other in time were made technically and financially possible through the invention of light meteorographs. M. Jaumotte, to whom we owe the best solution of the problem of light meteorograph construction, also arranged for the first "swarm ascents" to be carried out with the new instruments at his Institute in Uccle in the years 1928—30.[1] Swarm ascents were for the first time carried out at two places simultaneously: Uccle in Belgium and Ås in Norway, on February 2—3, 1933,[2] however with a rather limited number of ascents.

The aerological undertaking, on which a report will be given in the present paper, was to comprise simultaneous swarm ascents from Norway (Ås), Sweden (Riksgränsen) and Finland (Lauttakylä), but an invitation for collaboration was also sent, through the International Aerological Commission, to all other European observatories equipped for sounding-balloon work. The response was everywhere favourable. All the 18 observatories stated below (and represented in fig. 1) contributed with sounding-balloon ascents in various numbers. The following table gives the net number of soundings (excluding the ascents in which the meteorograph was lost or the record destroyed):

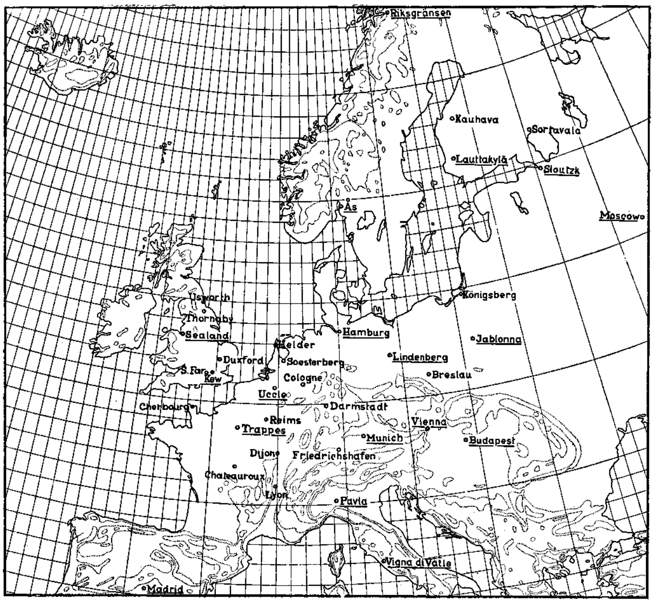

Fig. 1. The network of aerological stations. Sounding balloon stations underlined.

Ås	27	Lindenberg	7
Riksgränsen	1	Munich	1
Lauttakylä	8	Pavia	3
Sealand	9	Vigna di Valle	1
Kew	1	Vienna	3
Uccle	26	Budapest	6
Trappes	8	Jablonna	3
Madrid	6	Sloutzk	6
Hamburg	3	Moscow	1

In all 120 sounding-balloon records were received.

The authors want to express herewith their profound gratitude to all those who shared in the enterprise and made it the most extensive synoptic-aerological research program carried out up to the present time.

It lies in the nature of sounding-balloon work that, in order to get soundings in great numbers, a considerable toll of lost instruments must be paid. An unexpectedly great loss of instruments occurred

[1] The results are published as follows: Case 1: March 28—30, 1928, Beiträge zur Physik der freien Atmosphäre, Band 21, Heft 1, 1933. Case 2: Dec. 26—28, 1928, Geof. Publ. Vol. IX, No. 9. Case 3: Dec. 30—31, 1930, Geof. Publ. Vol. XI, No. 4.

[2] E. Palmén: "Registrierballonaufstiege in einer tiefen Zyklone", Mitteilungen des Meteorologischen Instituts der Universität Helsingfors. No. 26.

at the northernmost outpost, the Swedish observatory Riksgränsen, where only one meteorograph was returned out of the "swarm" of twenty. In future work of the same kind it is to be hoped that the outposts can be equipped with radio-sounding-balloons.

In addition to the sounding-balloon ascents, kite ascents in great number were carried out in Lindenberg. Furthermore, the usual routine ascents, and some extra ascents of aeroplanes, supplement very usefully the aerological data from balloons and kites.

The sounding-balloon observations, which were sent in to the Geophysical Institute in Bergen, have in due course been duplicated and distributed to all the participating observatories. Extra copies are still available on application. Most of the observations have appeared, or will appear, in national publications[1] and in the publications of the International Aerological Commission, so that we have found it superfluous to publish the observational data in this paper.

I. The Synoptic Situation.

The swarm ascents were, according to preliminary agreement, to be launched on telegraphic notice from Bergen some time between the 1st of January and the 31st of March, 1935. Within this period there was a pre-arranged "international week" from the 11th to the 16th of February. It was of course desirable to let the swarm ascents take place during the international week, provided that no situation of outstanding importance presented itself earlier. Fortunately the end of the international week offered disturbed conditions of a kind that would affect not only western and northern Europe but practically all the countries in the aerological network. The starting telegram was sent on the morning of the 15th asking for ascents during the 15th and the 16th. When it turned out that interesting conditions would continue also during the 17th, prolongation of the ascent series was telegraphically asked for on the 16th at noon. The series were thereupon prolonged to include also part of Sunday the 17th, but inevitably the intervals between ascents had to be increased in order to make the prolongation possible.

In the morning of the 15th of February, 1935, (fig. 2) the warm front of a depression, still out in the Atlantic, had reached SW.Ireland and extended in a SE direction along the French W.coast to the Pyrenees. The warm air behind was of genuine tropical origin from W of the subtropical High centred between Portugal and the Azores.

By the evening of the 15th (fig. 3) the Atlantic depression was occluded, maritime polar air masses from its rear having reached W. Scotland and N. Ireland. On the cold front, which extended from N. Ireland to NW of the Azores, a wave cyclone was forming.

In the morning of the 16th (fig. 4) the primary depression was near the Färoes, its occluded front extending across the North Sea to Hamburg. The secondary depression had already reached Scotland, so that only a narrow tongue of maritime polar air remained between the preceding occluded front and the warm sector of the secondary. The frontal structure of the secondary was not so simple as is normally the case in young depressions. The centre W of Scotland is probably to be regarded as the original secondary, but a new one was also forming E of Scotland. The large scale arrangement of air masses had now become favourable for the transformation of potential energy into kinetic energy, because a large portion of tropical air was about to be trapped between the cold continental air masses and the maritime polar air then breaking in behind the Scottish secondaries.

By the evening of the 16th (fig. 5) the secondary from E. Scotland was over the Skagerak and was almost as deep as the primary depression, which was then stagnant off the Norwegian W. coast. The maritime polar air had invaded the British Isles at great speed. The cold front, which had reached SE. England, had no well-marked pressure trough, but another pressure trough followed behind over the Irish Sea. In the northernmost part of that second trough there was a piece of a back-bent occlusion, but the rest of the trough was non-frontal, as can be seen from the fact that it moved slower than the wind component normal to the trough line. As is usually the case, the non-frontal trough did not extend far out towards the high pressure area. The rise of pressure behind the non-frontal trough was so strong as to give a barometric tendency of +11.8 mb at Malin Head.

On the morning of the 17th (fig. 6) the secondary depressions had amalgamated in a centre of 964 mb over the Baltic. The cyclone had just recently occluded while farther S some of the warm sector remained,

[1] The Ås ascents are to be found in Jahrbuch des Norwegischen Meteorologischen Instituts 1935.

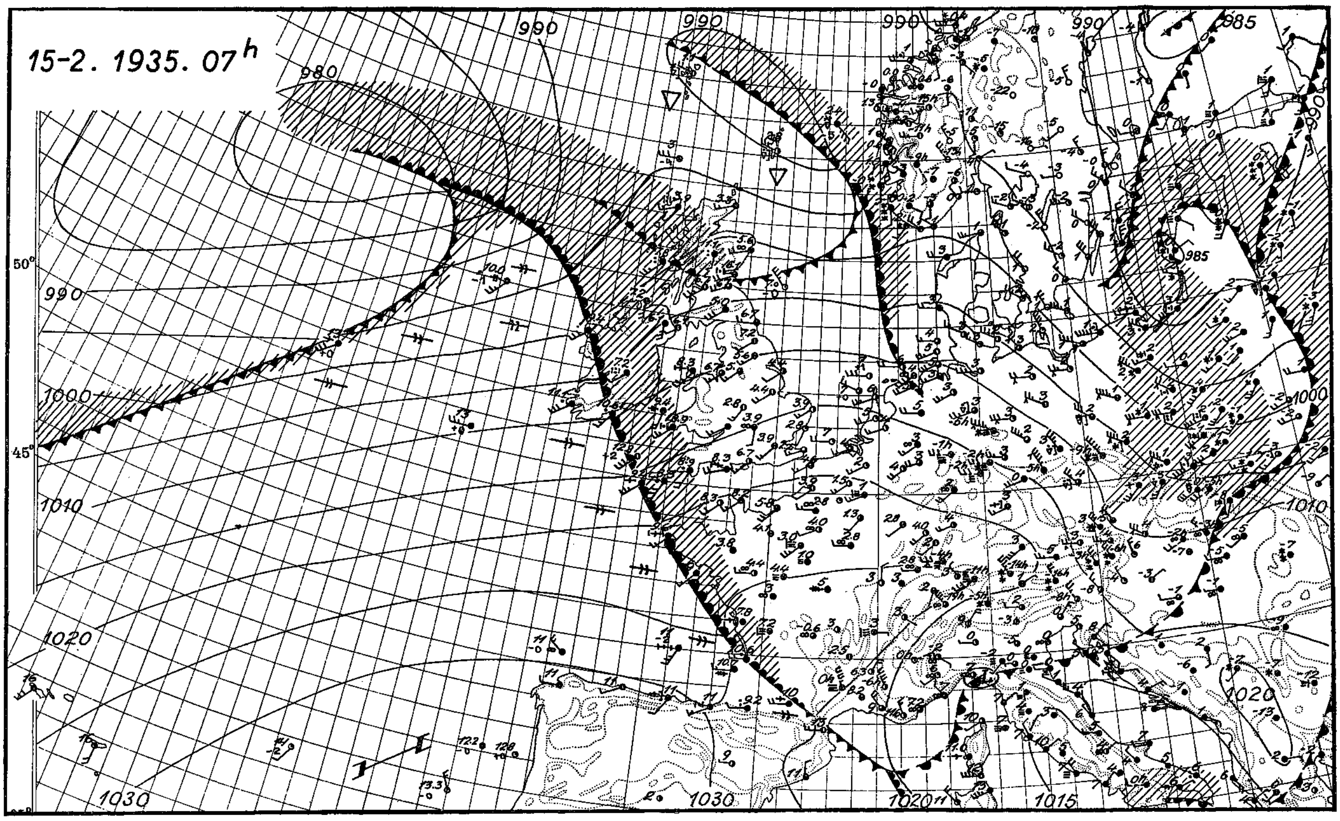
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Fig. 2.

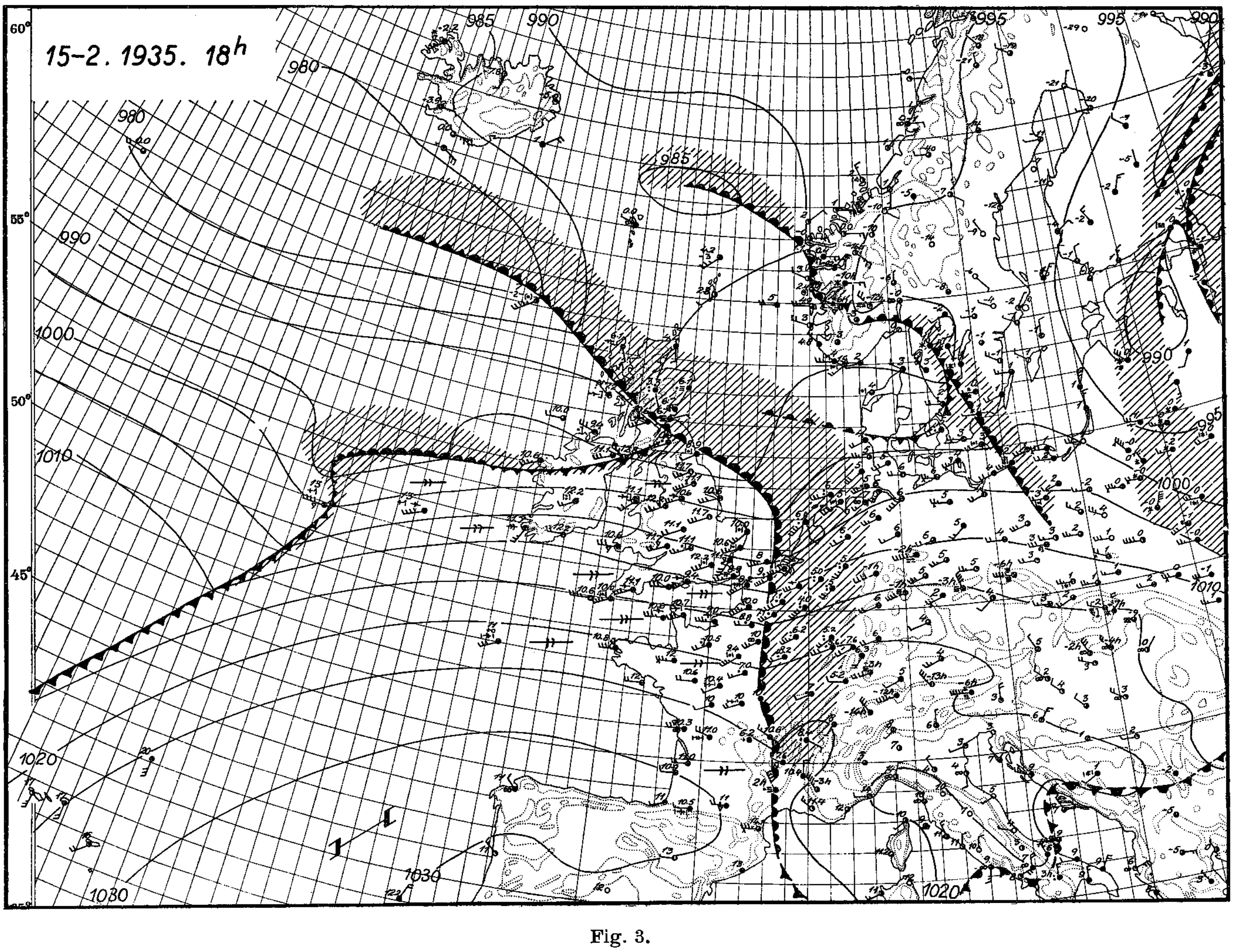
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Fig. 3.

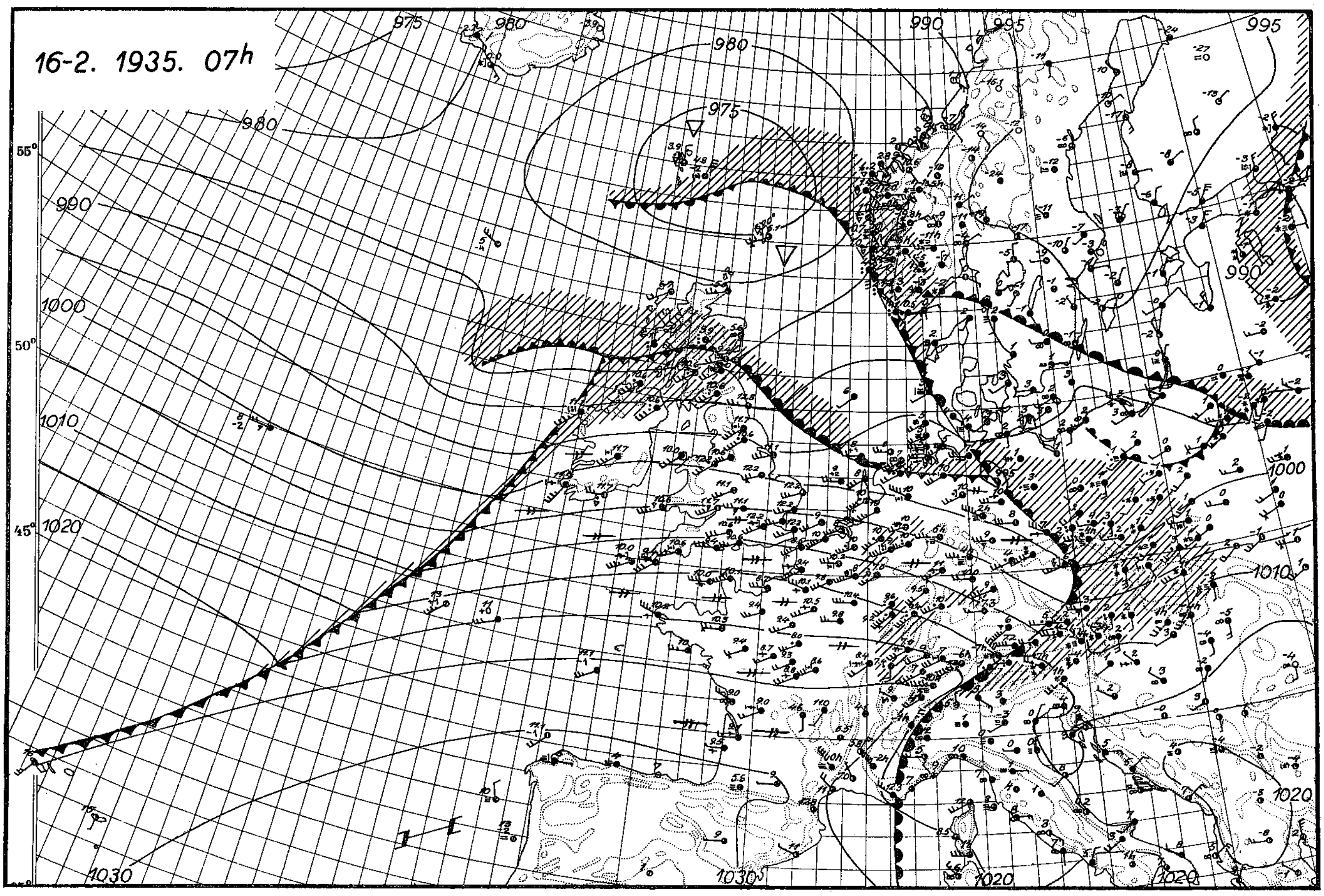

Fig. 4.

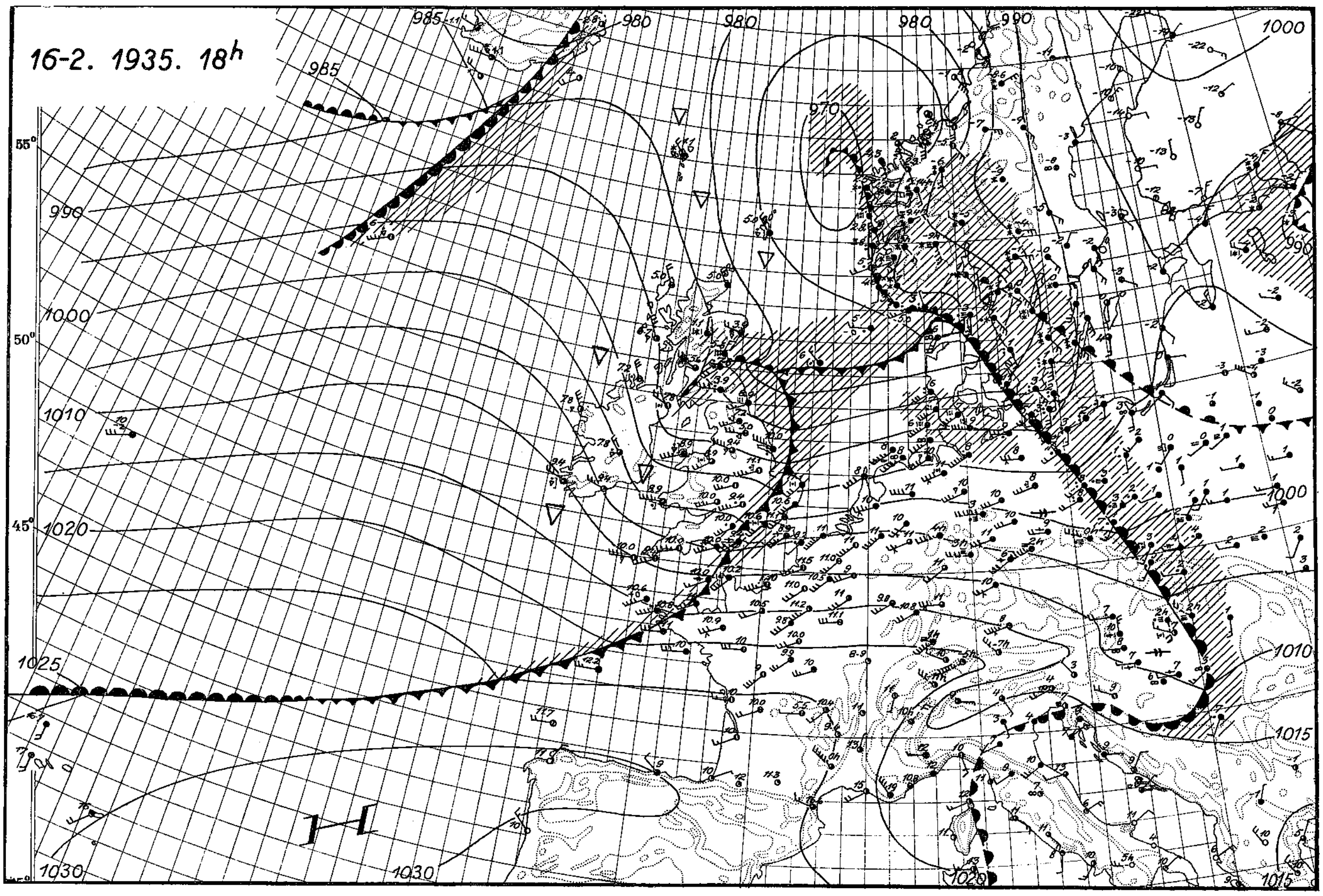

Fig. 5.

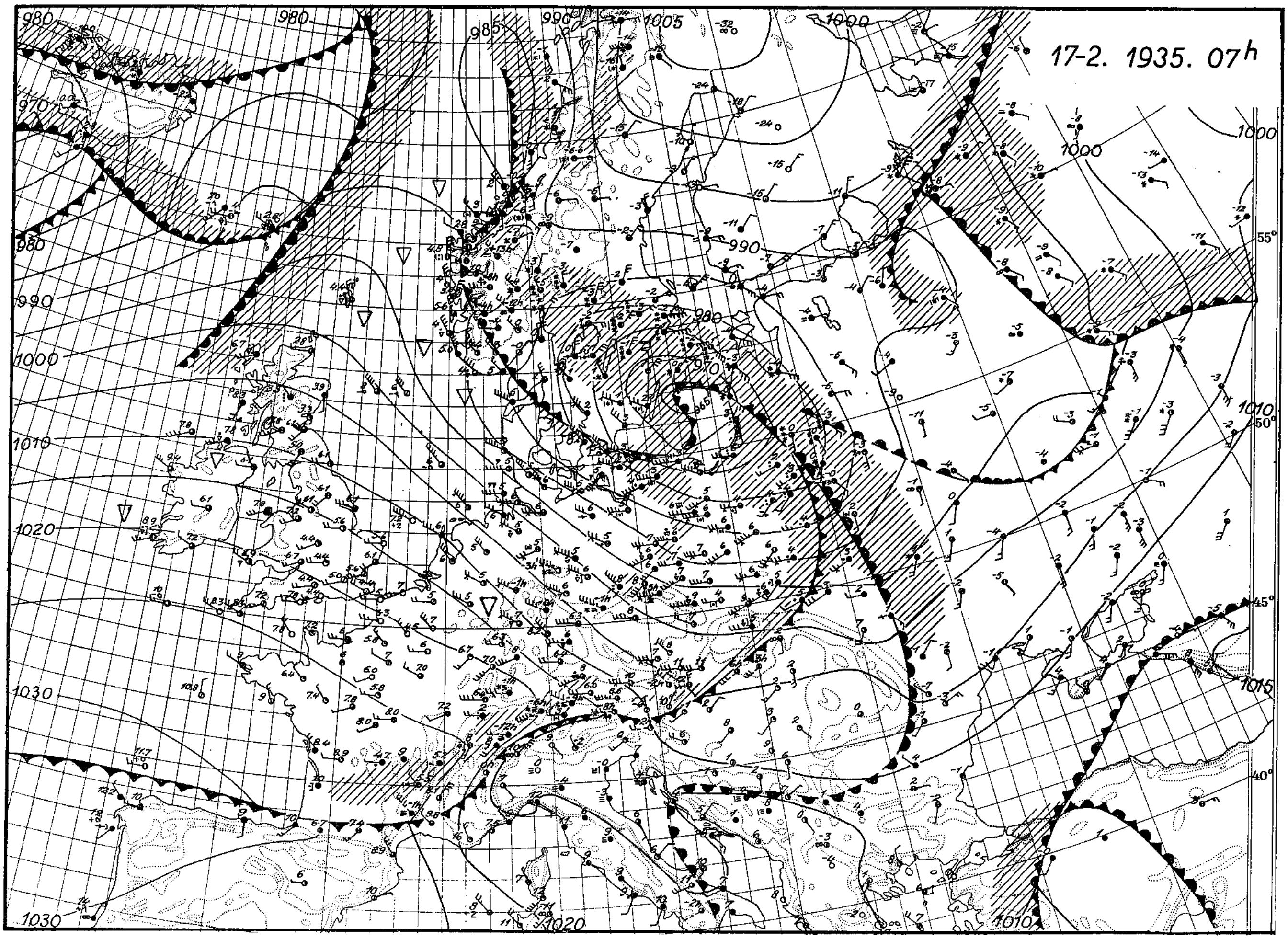

Fig. 6.

concealed, however, by a surface layer of cold air (the Budapest ascent proves the presence of tropical air aloft). The non-frontal trough extended from the centre of the depression towards Breslau and marked the transition from a phase of slightly falling or steady pressure to a phase of rising pressure. S of Breslau the non-frontal trough ended, and there was rising pressure from the very moment of the passage of the cold front. The rise of pressure extended over the whole of Western Europe and the maximum values of barometric tendencies, about +8 mb, are reported at the German North Sea coast.

Along the British western seaboard warm front altostratus was appearing without previous fall of pressure. The air current behind this new warm front was at first (evening of 17th, not reproduced) rather too cold at the surface to be characterized as "tropical", but it warmed up gradually to the tropical standard. In the upper troposphere, as will be seen in the following chapter, it was just as warm as the tropical air encountered over the warm front of the 15th of February.

The fronts over Eastern Europe were more or less indistinct and will not be treated in the present investigation.

II. Analysis of the Tropospheric Field of Temperature.

1. The Properties of the Principal Air Masses.

A general survey of the different air masses is first of all to be obtained through an examination of ascents in the regions undisturbed by the frontal zones. In fig. 7 three such ascents in different air masses are put together: Lauttakylä (Finland) February 16th, 0^{h} 50, Hamburg 15th 7^{h} 30 and Trappes 16th 7^{h}.

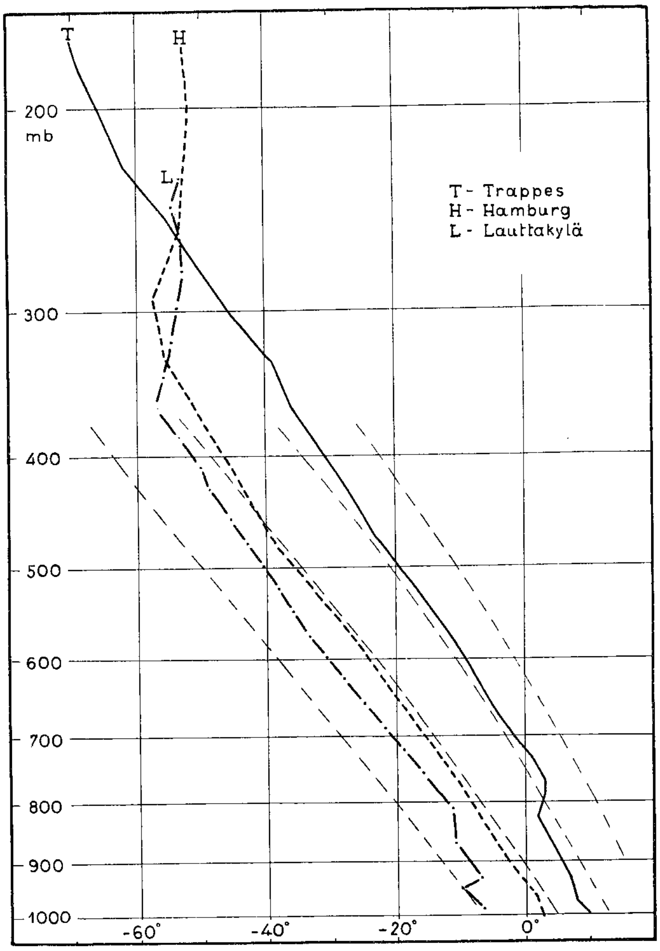

Fig. 7. Sample ascents in arctic (Lauttakylä), maritime polar (Hamburg) and tropical air (Trappes). Saturated adiabatics stippled.

The ascent at Lauttakylä took place in cold air of typically arctic origin, the ascent at Hamburg exemplifies maritime polar air and the ascent at Trappes tropical air. In the arctic type of air normal convective lapse-rate does not prevail till above the 800 mb surface (1.5 dyn km). The maritime polar air, however, has as usual a lapse-rate close to the saturated adiabatic all through the troposphere. At the ground the temperature difference between the maritime polar and the arctic air is about 9°, but above 800 mb the difference amounts to only 4°—6°. The corresponding temperature difference between the pure tropical air (Trappes) and the maritime polar air (Hamburg) is 14°—17° (except for the surface layers where the difference goes down to 7°). Thus the solenoid field in the layers between the tropical air and the maritime polar air is much more powerful than the solenoid field of the "arctic front", which is rather hypothetical in this case. The importance of the "arctic front" for the development of the cyclone from the 15th to the 17th of February is likely to be rather negligible.

Representative temperatures for the tropical and arctic air masses, at different heights, are given in Table 1, the first column of which gives the average of all Uccle ascents in the undisturbed part of the tropical current, while the second contains corresponding average temperatures from arctic air ascents in Lauttakylä.

Table 1.

Average temperature distribution in the tropical air (Uccle) and in the arctic air (Lauttakylä).

Height dyn km	Mean temperature		Temperature difference
	Uccle	Lauttakylä	
Ground	10.1	— 8.0	18.1
1	5 2	— 8.3	13.5
2	3.3	—14.5	17.8
3	— 1.9	—23.5	21.6
4	— 8.2	—32.2	24.0
5	—14.6	—39.7	25.1
6	—21.6	—47.4	25.8
7	—29.3	—54.7	25.4
8	—37.4	—53.7	16.3
9	—45.7	—52.8	7.1
10	—53.7	—52.0	— 1.7
11	—62.3	—50.9	—11.4

The Uccle tropical air temperatures are, from about 1.5 dyn km upward, somewhat higher than the corresponding representative winter temperatures of the tropical air, according to the statistics of G. Schinze.[1] On the whole, table 1 presents about the highest tropical air temperatures that can be obtained during the winter.[2] The corresponding temperature values in Lauttakylä are somewhat higher than the representative temperatures of the pure arctic air according to Schinze, but the difference is small enough to allow the air mass in question to be characterized as arctic.

The difference in temperature between Uccle and Lauttakylä is remarkably great, and attains, as is usual in analogous situations, its maximum value (about 26°) at 6 dyn km. At that level the observed temperature over Uccle is as high as the average

1 G. Schinze, Die praktische Wetteranalyse. Archiv der Deutschen Seewarte, Hamburg 1932.

2 See for instance Deutsche Seewarte: Tägl. Wetterber. 15—17 Febr. 1935.

temperature for February in latitude 35° N,[1] and that of Lauttakylä is as low as the extrapolated mean temperature over the polar region. Roughly stated, those isotherms which normally are distributed over a meridional distance of about 50 degrees in our case were for some time concentrated in a region of less than 10 degrees (the distance from the polar front to Lauttakylä on the morning of the 16th). The greatest horizontal temperature gradient over those 10° of latitude is to be found at the polar front surface, but a considerable temperature gradient prevails also outside the real frontal zone.

Fig. 8 presents four tropical air ascents on the 16th of February: Sealand (53° N), Trappes (49° N), Madrid (40° N) and Rabat (34° N). Apart from the layers below 800 mb the temperature lapse-rate is very uniform and the horizontal temperature differences small. Approximate barotropy thus characterizes the tropical air above 800 mb over a zone of about 19° of latitude.[2] Above 450 mb (about 6000 dyn m) barocline conditions are, however, to be found north of Trappes, Sealand being colder than Trappes in these upper tropospheric layers. This is the beginning of the strongly barocline conditions prevailing in the tropical air at the polar front, which was at that time about 300 km north of Sealand.

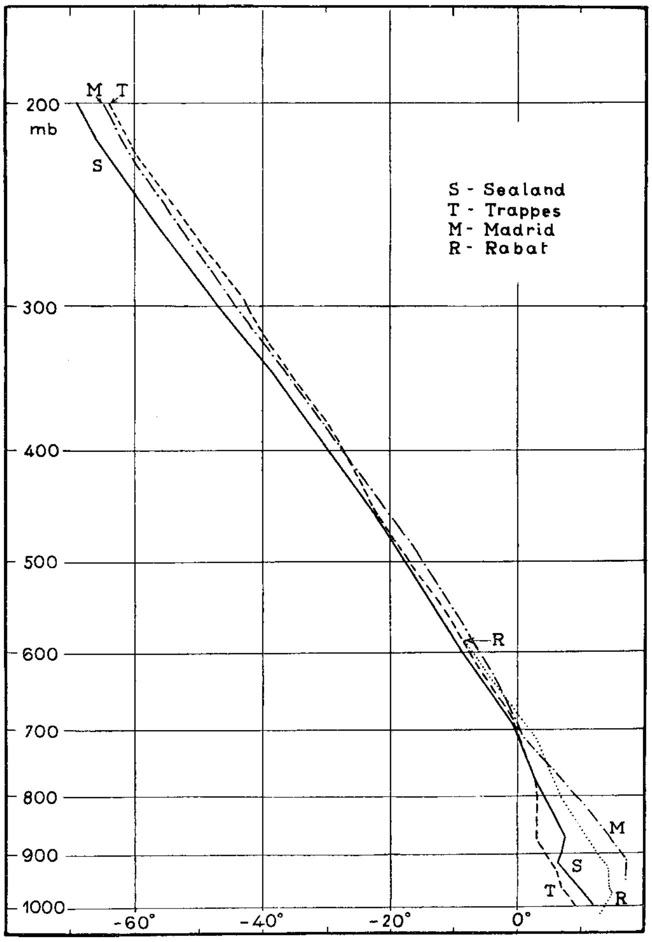

Fig. 8. Four tropical air ascents, 16th of February 1937, showing quasi-barotrope conditions.

In fig. 9 the Hamburg ascents which took place on the 15th of February at 7h 39, 13h 27 and 19h 24 and those of Sealand the same day at 12h 50 and 18h 00 are put together to illustrate the transition from the maritime polar to the tropical air. The temperature over Hamburg at 7h 39 shows all the characteristic properties of freshly arriving maritime polar air (great lapse-rate of temperature in the troposphere, low tropopause and relatively high stratosphere temperature). The following ascent, at 13h 27, shows tropical air masses above 7.4 dyn km (368 mb). In the polar air below 7.4 dyn km an important rise of temperature in comparison with the previous ascent can be observed within the whole column, although the frontal zone does not as yet assert a direct influence here. Very characteristic is the appearance of several inversions in this part of the polar air (1.35, 3.3 and 4,8 dyn km). At 19h 24 there is tropical air over Hamburg down to 3.7 dyn km (622 mb), but that air is by far not as warm as the tropical air over Sealand at 12h 50, the time of the warm front passage. The final, representative temperature of the undisturbed tropical air is not observed at Sealand till the 18h 00 ascent, that is well into the warm sector.

The temperature difference between Sealand 18h 00 and Hamburg 7h 39 is thus only partly concentrated in the warm front surface, both the cold wedge and the tropical air above it being barocline. The prefrontal increase of temperature may be due in part

[1] See E. Palmén, Über die Temperaturverteilung in der Stratosphäre und ihren Einfluß auf die Dynamik des Wetters. Met. Zeitschr. 1934, p. 17.

[2] Barotropy of course also implies that the surfaces of equal potential temperature (isentropic surfaces) coincide with isobaric surfaces. If the isobaric surfaces are horizontal, barotropy becomes synonomous with "horizontal isentropy" introduced by T. Bergeron and G. Swoboda in "Wellen und Wirbel . . ." Veröff. d. Geoph. Inst. d. Univ. Leipzig, Bd. III, Heft 2, p. 137. Also with inclined isobaric surfaces barotropy becomes for all practical purposes equivalent to "horizontal isentropy". We prefer to use the conception barotropy because of its intimate connection with dynamics (see Hydr. Phys. § 24).

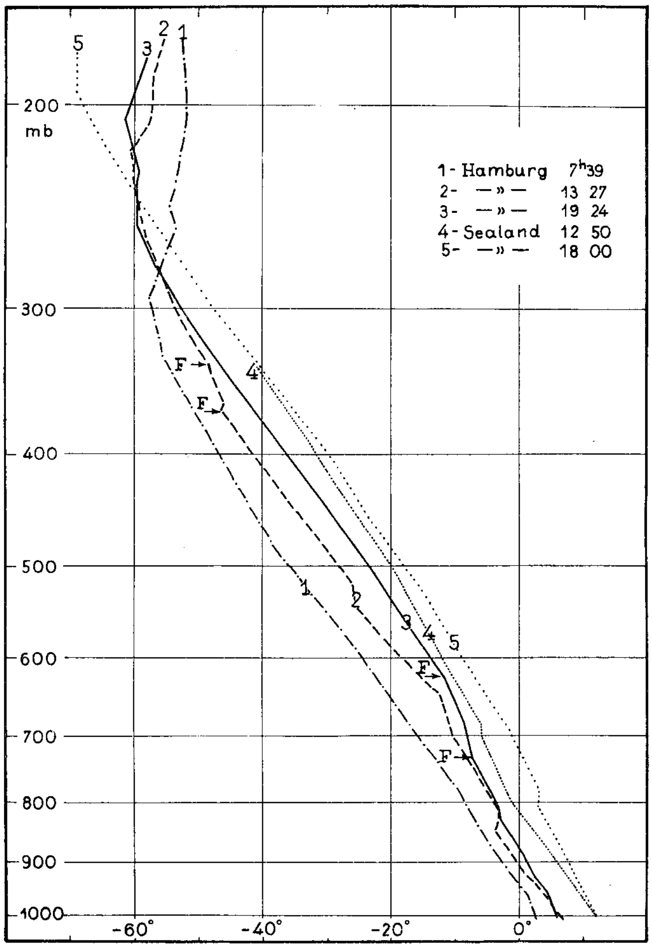

Fig. 9. Ascents showing the transition from maritime polar to tropical air.

to subsidence, but more important is probably the different life history of the maritime polar air masses at different distances from the front.

During and after the 15th of February, to which fig. 9 refers, a process of frontogenesis concentrated the temperature contrast at the warm front surface proper into a more narrow space (shown in fig. 10). Above Trappes and Uccle the frontal layer is still very indistinct; later it appears over Hamburg as an almost isothermal zone and still later over Lindenberg the frontal layer is characterized by a pronounced inversion of 1°.

2. The Field of Temperature in the Vicinity of the Warm Front Surface.

The air that ascends along the warm front surface must have come from the low layers of the warm sector. If the ascent has been adiabatic (usually at first dry-adiabatic then saturated-adiabatic), the wet-bulb potential temperature,[1] at the warm front surface will be the same as that in the bottom layer of the warm sector. In the case before us the warm sector had the usual small meridional temperature gradient (quasi-barotropy) acquired over the Eastern Atlantic with its small meridional gradient of water temperature. The wet-bulb potential temperature has a still smaller meridional

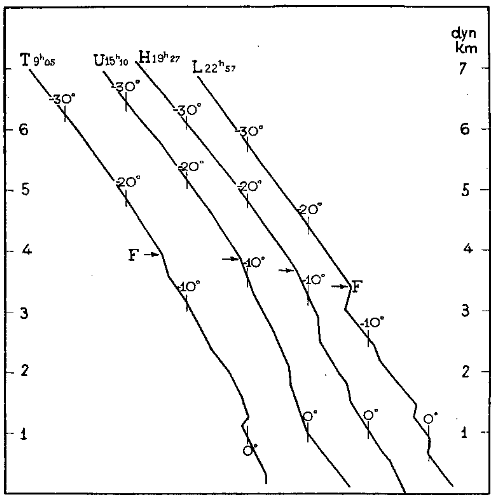

Fig. 10. Ascents showing the progressive sharpening of the warm front boundary layer.

gradient than the temperature itself, because there is in the warm sector a northward increasing relative humidity and a northward decreasing pressure, which both tend to obliterate the effect of the northward decreasing temperature.

In fig. 11 the surface air in the warm sector has been assumed to have a pressure of 1000 mb, a temperature of + 12° and a relative humidity of 79 %, which were the readings at Sealand on the 15th of February at 18^h. Its wet-bulb potential temperature is then 10°, and approximately the same wet-bulb potential temperature ought to be found at the warm front surface. In other words, the readings at the warm front surface ought to arrange themselves along the "characteristic saturated adiabatic" intersecting 1000 mb at 10°.

In tables 2 and 3 the warm front readings have been tabulated as obtained from the shape of the various ascents through the warm front surface. By definition the warm front surface has always been placed at the top of the layer of feeble or inverted lapse-rate that represents the transition from the cold

[1] See for instance: D. Brunt, Phys. and Dyn. Meteorology, p. 86.

wedge to the overlying tropical air. The values from Ås and Lauttakylä are tabulated separately because of the decidedly lower temperature of the tropical air in high latitudes.

Table 2.

Temperature and pressure at the warm front surface. (W and Central Europe).

Ascent			Geopotential dyn m	Temperature °C	Pressure mb
Hamburg	15.	13 27	8000	—49.5	335
Lindenberg	15	15 44	7900	—50.5	335
Sloutzk	17	9 32	6800	—41.5	380
Sealand	14	17 20	6000	—33.0	460
Trappes	15	9 05	5430	—24.0	501
Uccle	15	11 00	5360	—24.6	504
Uccle	15	11 55	4970	—20.1	532
Budapest	16	6 58	4670	—16.7	555
Uccle	15	15 10	3850	—11.1	617
Hamburg	15	19 24	3710	—12.0	622
Lindenberg	15	22 57	3410	—13.0	645
Breslau	16	8 00	2750	— 5.7	703
Uccle	15	16 05	2460	— 3.1	740
Cherbourg	15	7 45	2300	— 2.5	762
Vienna	16	7 09	2180	— 1.3	763
Lindenberg	16	5 37	2120	— 2.2	758
Reims	15	15 00	2000	— 0.5	792
Lindenberg	16	6 52	1970	— 1 2	776
Budapest	16	13 24	1890	— 0.6	799
Lindenberg	16	8 51	1870	— 0 5	790
Kew	15	12 33	1660	1.0	821
Uccle	15	18 05	1350	3.4	852
Lindenberg	16	10 50	833	4.2	899
Berlin	16	8 00	600	5.3	929
Sealand	15	12 50	5	12.0	1002

Table 3.

Temperature and pressure at the warm front surface. (Norway and Finland.)

Ascent			Geopotential dyn m	Temperature °C	Pressure mb
Lauttakylä..	16.	21 55	7000	—50.0	367
Ås.........	15	23 00	5820	—38.5	447
Ås.........	16	0 00	5080	—30.8	500
Ås.........	16	2 00	4940	—30.0	511
Ås.........	16	3 00	4760	—30.0	523
Ås.........	16	4 00	4100	—24.7	575
Ås.........	16	7 07	3870	—20.5	593

The values from tables 2 and 3 are entered in fig. 11, those of Ås and Lauttakylä as rings, and those of other stations as crosses. The crosses very nearly coincide with the saturated adiabatic line which

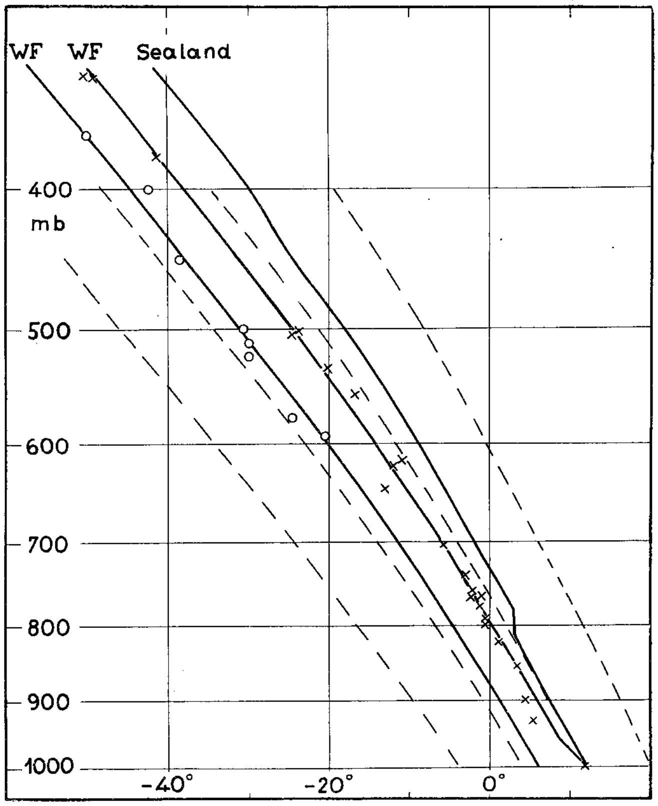

Fig. 11. The Sealand ascent 15th of February 18h (tropical air) and the curves of adiabatic ascent of a tropical air element from the ground. Crosses represent readings over Central Europe at the warm front surface, rings corresponding readings in Norway and Finland. Interrupted lines represent saturated adiabatics.

characterizes the temperature decrease of the ascending tropical air. The wet-bulb potential temperature is thus practically the same along the warm front surface over all stations in W and Central Europe. This rule applies even to the warm front surface over Sloutzk on the 17th and to the ascents in Hamburg at 13^h 27 and Lindenberg at 15^h 44 on the 15th, although these ascents probably go through the warm front surface above the maximum height of the up-gliding motion. The greatest deviations from the characteristic saturated adiabatic occur at Lindenberg and Berlin in the evening of the 15th and the morning of the 16th. The reason for that will appear on p. 23.

Fig. 11 shows furthermore that the readings at the warm front surface over N Europe (Ås and Lauttakylä) group around a saturated adiabatic line with a wet-bulb potential temperature of 6°, thus 4° lower than the corresponding wet-bulb potential temperature for W and Central Europe (including

Sloutzk). The trajectories of the tropical air over S Scandinavia lead back north of the British Isles to colder parts of the Atlantic than that over which the rest of the tropical air acquired its uniformly high temperature. Perhaps also the whole column of tropical air, which has attained its characteristic properties in the radiation climate of low latitudes, perceptibly loses heat when staying a day or so in latitude 60°.

Fig. 11 also contains the Sealand ascent on the 15th at 18h, which represents the undisturbed tropical air. It shows throughout a stable lapse-rate[1] and is therefore in all levels warmer than the ascending air at the warm front surface. At the 500 mb surface, for example, the temperature over Sealand is —18°, at the warm front surface over Central Europe —24°.5 and at the warm front surface over Ås —31°. At the same isobaric surface over Finland in the cold air mass of arctic origin, the temperature is —40°, that is, only 9° lower than that at the front surface over Ås. At 500 mb the temperature difference between Sealand and Ås within the tropical air is thus greater than the temperature difference between Ås and Lauttakylä, although the polar front surface lies between these last mentioned stations.

Figs. 29 and 30 (p 42 and 43) show vertical cross-sections which demonstrate many of the air mass and front properties described in this paragraph.

3. The Field of Temperature in the Vicinity of the Cold Front Surface.

For the determination of the position and structure of the cold front surface, only the following numbers of ascents are available: Sealand 3, Trappes 1, Uccle 4, Munich 1, Lindenberg 1, Vienna 1 and Jablonna 1. In addition there are aeroplane ascents from Darmstadt and Königsberg. From Ås several ascents were undertaken in the rear of the cyclone, but Ås was N of the cyclone track, so that there was no cold front passage in the lowest layers.

[1] The great vertical stability of the tropical air is typical for all winter situations. In the summer the stability of the tropical air often changes into lability over the continents. The marked stability of the great winter cyclones shows that the hypothesis of A. Refsdal concerning the importance of the "Feuchtlabilität" is not applicable to winter cyclones.

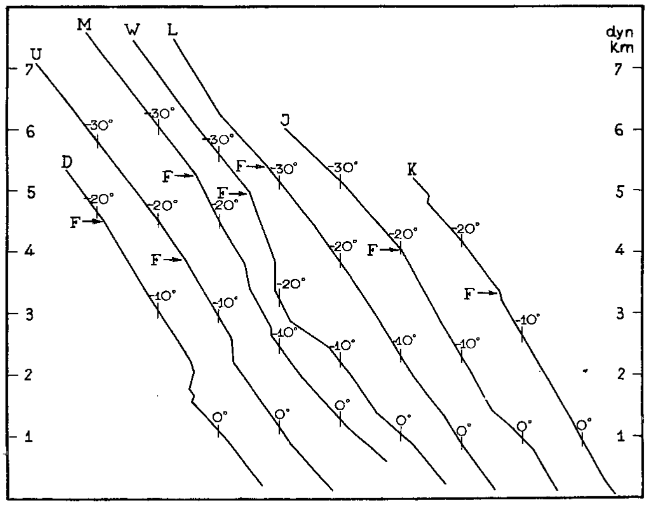

Fig. 12. Ascents through the cold front surface.

In fig. 12 some of the ascents that go through the cold front surface are given. The arrows indicate the cold front surface, which is always located at the top of the transitional layer, just where the uniform tropical air lapse-rate begins. Certain soundings, as for example the ascents from Vienna, Darmstadt and Uccle, show distinct inversions farther down, probably indicating subsidence in the polar air. They occur only rather far from the centre of the cyclone. The soundings at Sealand show complete frontolysis, so that it is impossible to determine the frontal surface in this region by the usual methods; these very interesting ascents will be treated in a later connection.

In tables 4 and 5 the temperature and pressure values recorded at the cold front surface are given.

Table 4.

Temperature and pressure at the cold front surface in W and Central Europe.

Ascent			Geopotential dyn m	Temperature °C	Pressure mb
Lindenberg	17.	6 54	5420	—32.0	481
Munich	17	13 41	5240	—23.6	513
Uccle	17	7 15	5200	—24.4	511
Sealand	16	18 00	5150	—29.0	500
Vienna	17	16 38	4940	—24.9	525
Darmstadt	17	7 00	4500	—19.0	565
Jablonna	17	8 48	4020	—19.8	580
Uccle	17	1 05	3870	—15.4	609
Königsberg	17	7 00	3300	—13.5	632
Sealand	16	16 05	1700	— 1.0	805

Table 5.

Temperature and pressure at the cold front surface in Norway and Finland.

Ascent	Geopotential dyn m	Temperature °C	Pressure mb
Lauttakylä . 17. 10^{02}	7550	—52.5	340
Ås .... 17 10^{00}	6620	—40.7	400
Ås .... 17 6^{00}	6150	—39.0	423
Ås .... 17 4^{00}	5090	—30.0	493
Ås .... 17 2^{00}	4450	—23.2	539
Ås .... 16 22^{00}	3240	—13.1	637

Fig. 13 shows these same readings at the cold front surface arranged in a diagram analogous to that of fig. 11. The values for W and Central Europe (crosses) and those for Norway and Finland (rings) all arrange themselves approximately along a saturated adiabatic intersecting 1000 mb at 8°.5. This is different from what was found at the warm front surface, where the temperature data grouped around two different saturated adiabatics with wet-bulb potential temperatures respectively 10° and 6°.

The temperature along the warm front surface is thus, in our case, over W and Central Europe considerably higher than farther north over Scandinavia and Finland, while on the contrary the temperature at the cold front surface is practically independent of the latitude. Above the stations in W and Central Europe the temperature is higher at the warm front surface than at the cold front surface. In Scandinavia and Finland the temperature at the cold front surface is higher than at the warm front surface.

The internal solenoid field in the tropical air is thus over W and Central Europe more strongly developed in the vicinity of the cold front than in the vicinity of the warm front. This difference is apparently connected with the life history of the different parts of the tropical air. At the cold front, "tropical" air masses of very different origin and life history are brought together in a relatively narrow zone. Thereby a strong internal solenoid field is formed outside the real front. The origin of this internal solenoid field will be more closely illustrated in the following detailed analysis of the isopleths for Sealand and Uccle (Chapter IV).

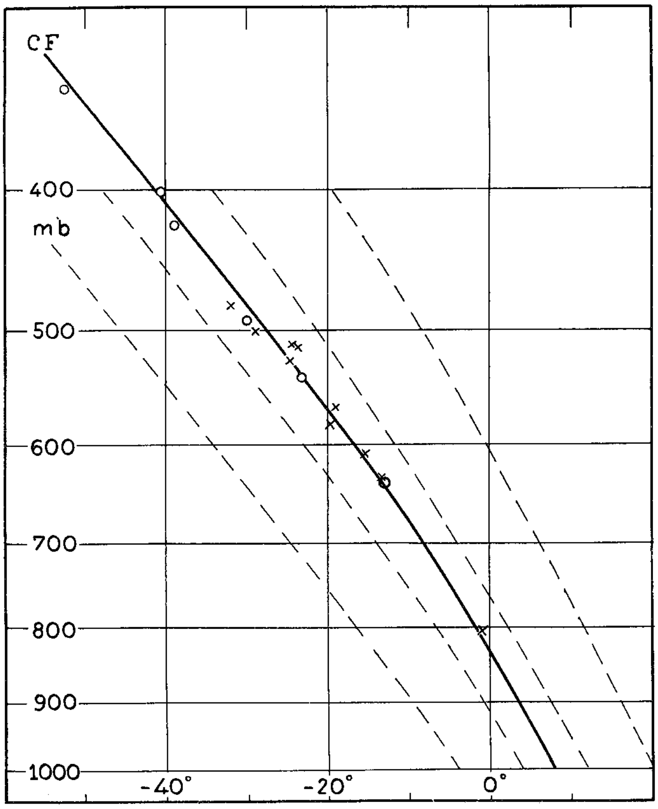

Fig. 13. Readings at the cold front surface (+ Central Europe, ○ Norway and Finland). Interrupted lines represent saturated adiabatics.

III. Analysis of the Tropopause and the Stratospheric Field of Temperature.

1. Definition of the Tropopause.

The ideal boundary between the troposphere and the stratosphere is a single surface of discontinuity of the first order, with discontinuous change of temperature lapse-rate.[1] Rather often, however, several points of the ascent curve in the tropopause region may show that kind of discontinuity, so that the lapse-rate changes from its tropospheric to its stratospheric value by successive steps. We then have a "multiple tropopause" (double, triple etc.).

The soundings at Uccle from the 15th of February at 11^h to the 16th at 5^h show good examples of a double tropopause. In all the ascents two levels with distinct decrese of lapse-rate can be found. In fig. 14 they are indicated with arrows and the corresponding potential temperatures are inserted. In table 6 the geopotentials of these levels (characteristic

[1] Hydr. Phys. § 37.

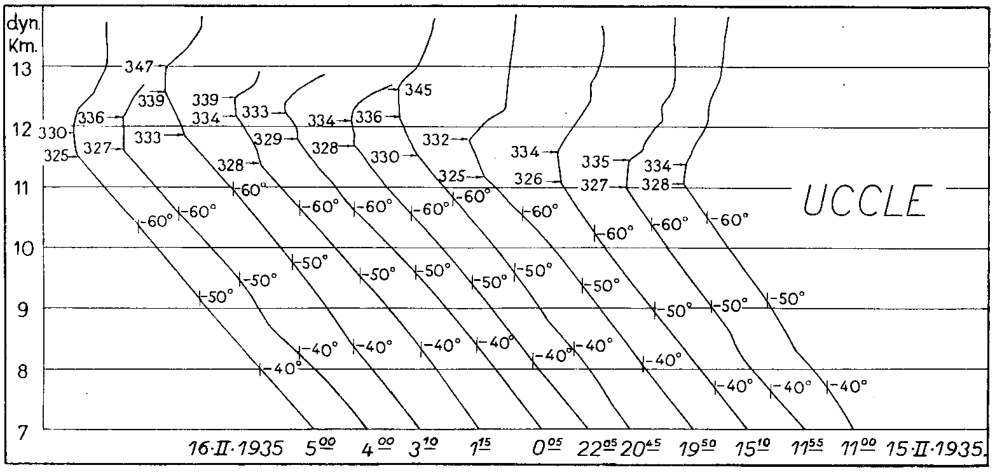

Fig. 14. The multiple tropopause, as shown by part of the Uccle series of ascents.

points) are given, as well as the difference of temperature and the mean lapse-rate between them. The last column contains the potential temperatures of the two characteristic points.

Table 6.

The double tropopause in the ascents at Uccle February 15th 11^h — 16th 5^h.

Ascent	Geopotential of the characteristic points in dyn m		Temp. diff. °C	Lapse-rate °C per 100 dyn. m	Pot. temp.	
15. 11^{00}	11067	11359	0.3	—0.10	328	334
11^{55}	10991	11457	0.3	—0.06	327	335
15^{10}	11100	11604	—0.5	0.10	326	334
19^{50}	11178	11802	—2.6	0.42	325	332
20^{45}	11524	12180	—3.1	0.47	330	336
22^{05}	11680	12086	—0.7	0,17	328	334
16. 0^{05}	11817	12229	—2.1	0.51	329	333
1^{15}	11405	12210	—4.0	0.50	328	334
3^{10}	11864	12587	—3.4	0.47	333	339
4^{00}	11632	12174	—0.2	0.04	327	336
5^{00}	11499	11863	—0.7	0.19	325	330

If the tropopause were defined as the level where the temperature gradient definitely sinks below a certain limit, for instance 2° per 1000 dyn m, the lower characteristic point in table 6 would be the tropopause in the ascents at 11^{00}, 11^{55}, 15^{10}, 22^{05}, 4^{00} and 5^{00}, whereas the upper characteristic point would be the tropopause in the ascents at 19^{50}, 20^{45}, 0^{05}, 1^{15} and 3^{10}. This seems unnatural, and we therefore prefer to define such a system of successive changes in lapse-rate as a multiple tropopause (double, triple etc.).

It is to be kept in mind, that an error of the magnitude of ± 2° in the temperature records corresponds to an error of ± 3—4° in the potential temperature, and errors of pressure may further increase the errors in potential temperature. It is therefore possible that the irregular variations in the potential temperature of the characteristic points are mostly due to instrumental errors. This justifies the adoption of the working hypothesis that the potential temperature at each one of the tropopauses in a bundle is relatively constant although its height varies.

Furthermore we find from fig. 14 and from other similar cases, that the upper surface of the double tropopause is in general better developed if the whole system is temporarily lifted, whereas the lower surface becomes dominant when the whole system is temporarily lowered.

In the following we have taken into consideration all characteristic points of the temperature records which might indicate a step in the transition between troposphere and stratosphere. If such a characteristic point, according to the usual definition of the tropopause, is to be considered as the upper limit of the troposphere, the surface in question has been indicated with a thick full line in all isopleth-diagrams and vertical cross-sections. Where the characteristic point is less distinct, the corresponding tropopause line is stippled. This method signifies only that every surface which marks a distinct change in the lapse-rate within the so-called sub-stratosphere is taken into consideration, and that no preference is given a priori to any surface according to purely schematic rules.

An earlier attempt at an analogous analysis of the tropopause has been undertaken by E. Palmén.[1] Also J. Reger has earlier given expression to similar ideas.[2]

[1] See E. Palmén, Aerologische Untersuchungen der Atmosphärischen Störungen. Mitteilungen des Meteorologischen Instituts der Universität Helsingfors, No. 25.

[2] J. Reger, Die Grenze zwischen Troposphäre und Stratosphäre. Arb. des Aerol. Obs. bei Lindenberg, No. 16, 1930.

2. Analysis of the Tropopause in the Soundings at Ås.

In applying the given definition of the multiple tropopause, another interesting fact appears in the analysis of the soundings at Ås. In fig. 15 the temperature records between 5 and 12 dyn km are given for the ascents on the 15th of February at $7^h 09$, $15^h 04$, $16^h 00$, $18^h 05$ and $20^h 01$. The characteristic points on the temperature curves are again indicated with arrows. The lowest characteristic points in the five soundings are found at 7.4, 6.9, 6.45, 6.55 and 6.2 dyn km respectively. In the three first cases the characteristic points mentioned represent the tropopause unequivocally. The ascents went through maritime polar air masses (probably with arctic air in the lowest strata), and the unusually low tropopause (between 7.4 and 6.45 dyn km) is characteristic for old occluded systems like that which covered Scandinavia. After 16^h the tropical air begins to arrive at great altitude. The sounding at $18^h 05$ gives a very interesting record of this invasion of the tropical air masses above the old tropopause of the polar air. The invading tropical air masses first of all cause a considerable temperature decrease in part of the space initially occupied by the polar air stratosphere, and a new tropopause appears above the pre-existing one. The ascent at $18^h 05$ thus shows two distinct tropopauses, the old polar air tropopause at 6.55 dyn km and a new tropopause of the tropical air at 8.25 dyn km. At the following ascent, $20^h 01$, only one tropopause is found, at the altitude of 9.35 dyn km. The lower characteristic point in fig. 15 now represents the commencing warm front surface. The earlier low tropopause surface, characteristic for the polar air occlusion, has thus wholly disappeared.

In fig. 16 the process is represented schematically. The pure polar air conditions (low temperature in the troposphere, low tropopause and high stratosphere temperature) and the pure tropical air conditions (high temperature in the troposphere, high tropopause and low stratosphere temperature) are schematically represented by the curves in full. The dotted lines, AB and A'B' respectively, represent the temperature distribution in the transitional air in two different cases. If the transitional layer lies between the points A and B, that is to say in the troposhere, a typical record is obtained through a front surface. If, on

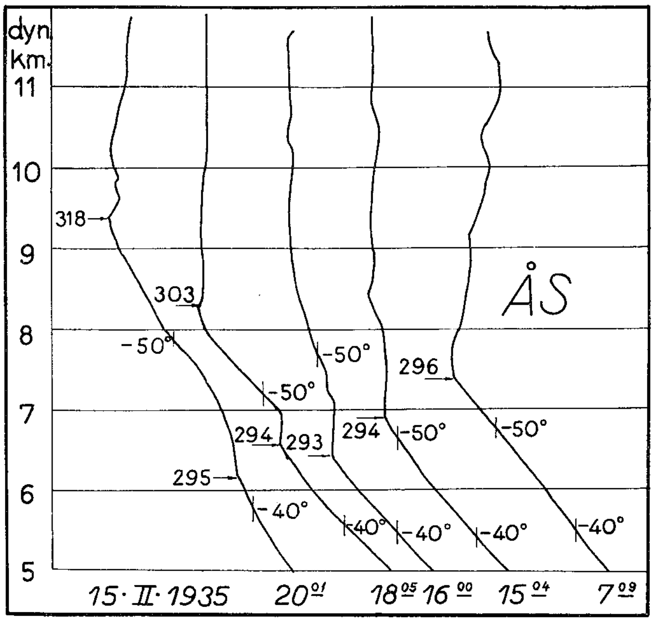

Fig. 15. The tropopause over Ås during the transition from polar air to tropical air in the upper atmosphere.

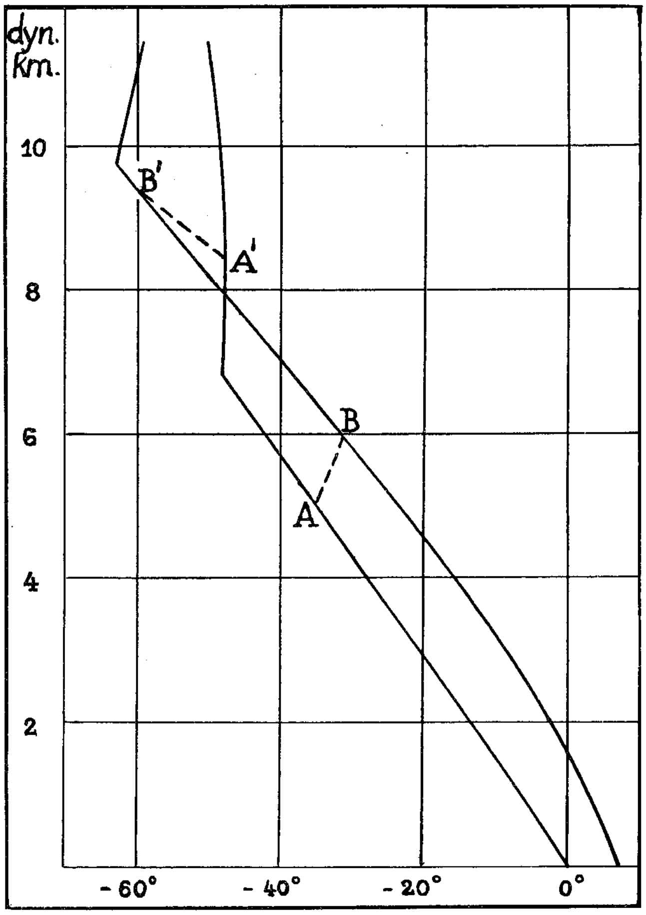

Fig. 16. Schematic deduction of ascent curves for the region where the tropical air overruns the polar air.

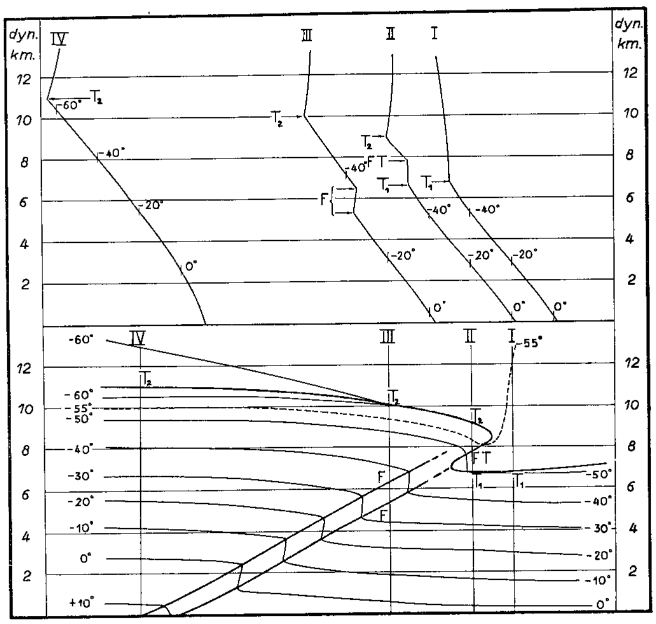

Fig. 17. Schematic vertical cross-section through a warm front surface that reaches the tropopause.

the contrary, the transitional layer between the polar air and the tropical air lies above the polar air tropopause, the dotted line A′B′ shows the resulting temperature distribution. The tropical air lapse-rate at that level generally being nearly adiabatic, the transition layer between A′ and B′ initially tends to become superadiabatic, but the vertical "Austausch" of course sets a limit to that. According to the $18^h\,05$ ascent the lapse-rate in the transitional layer was 1.01° C/100 dyn m between 7000 and 7660 dyn m.

In a schematic vertical cross-section, or in an isopleth diagram, the front and the tropopause should be drawn as in fig. 17. Ascent I corresponds to the temperature curve for polar air in fig. 16, ascent IV to the curve for tropical air. Ascent II shows the polar air tropopause in point T_1 (the same as in ascent I). Between T_1 and FT the instrument rises through the stratosphere of the polar air. In point FT the instrument enters the troposphere of the tropical air, that is colder than the polar stratosphere. In point T_2, the instrument at last, reaches the tropopause of the tropical air (the same as in the ascents III and IV). In fig. 17 the tropopause has therefore been drawn as a folded surface, intersecting the vertical II in the three points T_1, FT and T_2.

The first invasion of the tropical air masses at great altitude thus has the character of a cold front phenomenon in the lower strata of the previous polar stratosphere. At the lower limit (FT) of this cold air wedge, that penetrates into the relatively warm polar stratosphere, vertical convection must arise. The foremost prefrontal Ci may perhaps in part owe their existence to that convection, but that is of course a question for further inquiry.

At a later series of ascents, Oct. 17—19, 1935, a similar folding of the tropopause appeared at the time of the first invasion of the tropical air in the region of the polar air stratosphere. The phenomenon in question is likely to be a general one, but in series of less frequent ascents the phenomenon of the folded tropopause will often be missed, because it moves rapidly past the fixed point of observation.

3. General Analysis of the Tropopause.

We now propose to assemble all the various "characteristic points" of single and multiple tropopauses (see p. 16) in tabular form, using the data from all places of ascents. As an instance of the method of analysis we begin with the series at Ås. Tab. 7 contains the most remarkable points in the upper atmosphere at which the lapse-rate of temperature suddenly decreases from a high value underneath to a lower one above. For each such point the geopotential, temperature and potential temperature are tabulated. The numerals that refer to the polar air tropopause are printed in italics.

The values of the potential temperature suggest a grouping of the characteristic points. Group I, with the lowest potential temperatures, is observed only in the occluded cyclone with polar air in the whole troposphere. Group II, with somewhat higher potential temperatures, represents in our case the lowest possible tropopause surface of the tropical air. The following groups, with higher potential temperature, represent higher tropopause surfaces in the same tropical air. The potential temperature is roughly constant within each tropopause group, and might have shown a closer approximation to constancy if all instrumental errors had been eliminated. Only the sounding of February 15th at $22^h\,00$ disagrees with the scheme. According to the potential temperature the tropopause in that ascent ought to be placed in group I, but since it occurs in tropical air it has been transferred to group II. Group IV, which is represented only in one ascent, corresponds apparently to the lowest tropopause group of the soundings at Uccle (tab. 6).

Table 7.

Analysis of the tropopause in the ascents at Ås.

Ascents	Group I			Group II			Group III			Group IV		
	dyn km	t° C	T′	dyn km	t° C	T′	dyn km	t° C	T′	dyn km	t° C	T′
15. 7^{09}	*7.39*	*—55*	*296*	-	-	-	-	-	-	-	-	-
15^{07}	*6.92*	*—51*	*294*	-	-	-	-	-	-	-	-	-
16^{00}	*6.45*	*—48*	*293*	-	-	-	-	-	-	-	-	-
18^{05}	*6.59*	*—48*	*294*	8.25	—58	303	-	-	-	-	-	-
20^{01}	-	-	-	-	-	-	9.35	—58	318	-	-	-
22^{00}	-	-	-	8.32	—63	297	-	-	-	-	-	-
24^{00}	-	-	-	8.25	—58	303	-	-	-	-	-	-
16. 2^{00}	-	-	-	8.20	—56	304	-	-	-	-	-	-
3^{00}	-	-	-	8.44	—56	308	-	-	-	-	-	-
4^{00}	-	-	-	8.47	—59	305	-	-	-	-	-	-
6^{00}	-	-	-	8.24	—58	303	-	-	-	-	-	-
7^{07}	-	-	-	8.36	—56	306	-	-	-	-	-	-
16^{00}	-	-	-	8.70	—58	309	-	-	-	10.24	—60	328
20^{00}	-	-	-	8.15	—54	307	8.85	—56	314	-	-	-
21^{00}	-	-	-	8.11	—54	306	8.87	—56	314	-	-	-
22^{00}	-	-	-	8.00	—52	307	8.59	—55	311	-	-	-
23^{00}	-	-	-	8.09	—55	305	-	-	-	-	-	-
24^{00}	-	-	-	8.00	—51	309	-	-	-	-	-	-
17. 2^{00}	-	-	-	7.97	—54	305	-	-	-	-	-	-
4^{00}	-	-	-	8.22	—55	307	-	-	-	-	-	-
6^{00}	-	-	-	8.53	—56	310	-	-	-	-	-	-
10^{00}	-	-	-	8.38	—54	309	-	-	-	-	-	-
16^{00}	-	-	-	-	-	-	9.70	—62	316	-	-	-
18^{00}	-	-	-	-	-	-	9.82	—65	314	-	-	-

The mean values of potential temperature in the different tropopause groups over Ås are:

I	II	III	IV
294°	306°	315°	328°

The above mentioned sounding on the 15th 22^h was neglected in the computation of the average potential temperature for group II.

Table 8 brings the result of the analogous tropopause analysis over Uccle.

Group I in the material from Ås is missing in the ascents at Uccle. Group I was, in the Ås material, represented only in the region where the polar air filled the whole troposphere; therefore it is quite natural that that group is not represented over Uccle.

The tropopause group VI appears only sporadically, and even where tabulated the point is hardly to be considered as a tropopause in the established sense. Only at the ascent of the 16th at 20^h 45 is this highest tropopause well developed. It is not always easy to distinguish between groups IV and V, because the potential temperature does not remain sufficiently constant in each of them. A certain arbitrariness is therefore hardly to be avoided.

The ascent of February 17th at 1^h 05 shows at 7.58 dyn km a relatively distinct low tropopause surface with the same potential temperature as that of group II in the series from Ås. This surface of discontinuity in the vertical temperature distribution has the character of a "new-formed tropopause", which later becomes more distinct over Germany. The following ascent, at 7^h 15, shows the same tropopause group II at the height of 8.53 dyn km. In this case, however, the lapse-rate above the surface in question is so great that it is hardly to be considered as a real tropopause.

In table 9 a survey is made of the height, temperature and potential temperature of the tropopause surfaces for all other participating stations. The grouping of the different tropopauses is made according to the same scheme as for Ås and Uccle. Brackets indicate indistinct tropopause surfaces. The values, referring to tropopause surfaces in the polar air regime, are printed in italics as in table 7.

The essential content in tables 7—9 may be summed up as follows:

The tropopause group I, with a potential temperature varying between 288° and 301°, is found only

Table 8.

Analysis of the tropopause in the ascents at Uccle.

Ascents	Group II			Group III			Group IV			Group V			Group VI		
	dyn km	t° C	T'	dyn km	t° C	T'	dyn km	t° C	T'	dyn km	t° C	T'	dyn km	t° C	T'
15. 11^{00}	-	-	-	-	-	-	11.07	—64	328	-	-	-	-	-	-
11^{55}	-	-	-	-	-	-	10.99	—64	327	-	-	-	-	-	-
15^{10}	-	-	-	-	-	-	11.10	—66	326	11.60	—66	334	-	-	-
19^{50}	-	-	-	-	-	-	11.18	—66	325	11.80	—69	332	-	-	-
20^{45}	-	-	-	-	-	-	11.52	—66	330	12.18	—69	336	12.70	—69	345
22^{05}	-	-	-	-	-	-	11.68	—69	328	12.09	—70	334	-	-	-
16. 0^{05}	-	-	-	-	-	-	11.82	—70	329	12.23	—72	333	-	-	-
1^{15}	-	-	-	-	-	-	11.42	—67	328	12.21	—71	334	-	-	-
3^{10}	-	-	-	-	-	-	11.86	—68	333	12.59	—71	339	13.02	—71	347
4^{00}	-	-	-	-	-	-	11.63	—69	327	12.17	—69	336	-	-	-
5^{00}	-	-	-	-	-	-	11.50	—70	325	11.86	—71	330	-	-	-
6^{05}	-	-	-	-	-	-	12.03	—70	332	-	-	-	-	-	-
7^{20}	-	-	-	-	-	-	11.61	—66	331	-	-	-	12.40	—66	345
8^{05}	-	-	-	-	-	-	11.63	—68	329	-	-	-	12.58	—68	345
10^{25}	-	-	-	-	-	-	12.02	—69	332	-	-	-	-	-	-
11^{30}	-	-	-	-	-	-	-	-	-	12.23	—68	336	-	-	-
15^{25}	-	-	-	-	-	-	11.82	—68	331	-	-	-	-	-	-
16^{50}	-	-	-	-	-	-	11.91	—68	333	12.40	—70	338	-	-	-
20^{45}	-	-	-	-	-	-	10.71	—58	331	-	-	-	12.06	—62	345
17. 1^{05}	7.58	—44	309	9.40	—54	321	10.27	—58	328	-	-	-	-	-	-
7^{15}	8.53	—54	308	9.50	—57	317	-	-	-	-	-	-	-	-	-
8^{45}	-	-	-	9.63	—58	316	-	-	-	-	-	-	-	-	-
10^{05}	-	-	-	9.79	—59	317	-	-	-	-	-	-	-	-	-

in the region of the old occluded cyclone. The group is distinctly represented in the material from Ås and Lauttakylä and less pronounced in some soundings at Riksgränsen, Jablonna, Lindenberg and Budapest.

The tropopause group II, with a potential temperature of 303—311°, is observed in Central Europe and over Scandinavia (especially at Ås) at soundings that are made in the peripheric parts of the tropical current. The tropical current appears in these cases as an upper warm current above the polar front surface. According to table 9 a similar group is more or less distinctly represented in Central Europe before the commencing invasion of the tropical air. This tropopause group (printed in italics in table 9), however, is probably not to be identified with the corresponding tropopause group in the tropical air. After the passage of the cold front at the earth's surface we can find the same group represented over Uccle and better developed at Lindenberg, nearer to the centre of the occluded cyclone.

Group III, with a potential temperature of 314—321°, is well represented over the whole of Central Europe above the warm front surface and after the passage of the cold front in the W and S parts of the cyclone (Sealand, Uccle and Munich). In the central parts of the cyclone a trace of this tropopause group is visible over Lindenberg in the morning of the 17th of February. At Ås the tropopause of group III appears only sporadically in the area of the deep cyclone, but seems to become the principal one at the two last soundings in the region of the following wedge of high pressure.

Groups IV and V, with potential temperatures of 325—333° and 330—339° respectively, are represented in fully developed shape only in the region covered with tropical air at the ground. However, the former group is also represented, at least in one ascent, at Ås, and that happens approximately at the time for the greatest northern elongation of the tropical air. Both groups IV and V appear clearly in the material from Uccle, as shown in tab. 6 and fig. 14, and they are also well represented in the ascents at Sealand, Trappes, Madrid, Vienna and Budapest. It is sometimes impossible to decide to which of these two groups every individual "characteristic point" ought to be referred. However, it appeared clearly from the detailed analysis of the soundings at Uccle that groups IV and V are separate.

Group VI, with a potential temperature of 343—348°, is sporadically represented in the ascents at

Table 9.

The multiple tropopause Febr. 15—17, 1935.

Ascents			I			II			III			IV			V			VI		
			dyn km	t° C	T′	dyn km	t° C	T′	dyn km	t° C	T′	dyn km	t° C	T′	dyn km	t° C	T′	dyn km	t° C	T′
Riksgränsen	15	16^{53}	*(6.07*	*—49*	*288)*	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
Lauttakylä	15	22^{08}	*7.25*	*—58*	*291*	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
—»—	16	0^{50}	*7.03*	*—57*	*290*	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
—»—	16	12^{00}	*6.57*	*—50*	*292*	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
—»—	17	10^{02}	-	-	-	8.59	—60	306	-	-	-	-	-	-	-	-	-	-	-	-
—»—	17	15^{55}	-	-	-	8.69	—63	303	-	-	-	-	-	-	-	-	-	-	-	-
Jablonna	15	19^{21}	*(7.50*	*—51*	*301)*	8.32	—54	307	-	-	-	-	-	-	-	-	-	-	-	-
Hamburg	15	7^{39}	*(7.92*	*—55*	*300)*	8.65	—57	307	-	-	-	-	-	-	-	-	-	-	-	-
—»—	15	13^{27}	-	-	-	-	-	-	9.60	—58	317	10.35	—60	325	-	-	-	-	-	-
—»—	15	19^{24}	-	-	-	-	-	-	9.81	—60	318	-	-	-	(11.09	—62	335)	-	-	-
Lindenberg	15	6^{45}	*7.03*	*—49*	*296*	*8.31*	*—54*	*307*	-	-	-	-	-	-	-	-	-	-	-	-
—»—	15	15^{44}	-	-	-	-	-	-	9.48	—60	315	-	-	-	-	-	-	-	-	-
—»—	15	22^{57}	-	-	-	-	-	-	9.38	—60	312	-	-	-	-	-	-	-	-	-
—»—	16	6^{52}	-	-	-	-	-	-	10.07	—62	319	-	-	-	-	-	-	-	-	-
—»—	17	6^{54}	(6.28	—40	299)	7.48	—48	305	(8.54	—49	318)	-	-	-	-	-	-	-	-	-
Budapest	15	6^{58}	-	-	-	*8.68*	*—53*	*311*	-	-	-	-	-	-	-	-	-	-	-	-
—»—	15	13^{06}	*6.76*	*—47*	*295*	*(8.20*	*—52*	*307)*	9.42	—55	321	-	-	-	-	-	-	-	-	-
—»—	16	6^{58}	-	-	-	-	-	-	9.37	—54	317	-	-	-	10.88	—59	333	-	-	-
—»—	16	13^{24}	-	-	-	-	-	-	-	-	-	-	-	-	11.45	—66	331	-	-	-
Vienna	15	17^{05}	-	-	-	*(8.00*	*—52*	*305)*	9.38	—66	319	-	-	-	-	-	-	-	-	-
—»—	16	7^{14}	-	-	-	-	-	-	-	-	-	(10.86	—60	328)	11.24	—61	334	-	-	-
—»—	17	16^{44}	-	-	-	(8.52	—53	310)	-	-	-	-	-	-	11.17	—63	334	-	-	-
Munich	17	13^{41}	-	-	-	-	-	-	8.95	—52	314	-	-	-	-	-	-	-	-	-
Sealand	14	17^{20}	-	-	-	-	-	-	9.79	—62	315	-	-	-	-	-	-	-	-	-
—»—	15	16^{20}	-	-	-	-	-	-	-	-	-	11.48	—67	329	-	-	-	-	-	-
—»—	15	18^{00}	-	-	-	-	-	-	-	-	-	11.58	—69	328	-	-	-	-	-	-
—»—	16	1^{00}	-	-	-	-	-	-	-	-	-	11.38	—68	326	-	-	-	-	-	-
—»—	16	4^{30}	-	-	-	-	-	-	-	-	-	11.90	—69	333	-	-	-	-	-	-
—»—	16	7^{00}	-	-	-	-	-	-	-	-	-	11.98	—72	329	-	-	-	-	-	-
—»—	16	18^{00}	-	-	-	-	-	-	(8.00	—43	317)	9.91	—52	332	-	-	-	-	-	-
—»—	17	7^{00}	-	-	-	-	-	-	10.42	—65	318	-	-	-	-	-	-	-	-	-
Trappes	15	9^{05}	-	-	-	-	-	-	-	-	-	11.04	—68	321	-	-	-	-	-	-
—»—	15	18^{00}	-	-	-	-	-	-	-	-	-	-	-	-	12.35	—70	337	-	-	-
—»—	16	1^{05}	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	12.98	—72	343
—»—	16	7^{00}	-	-	-	-	-	-	-	-	-	-	-	-	12.75	—72	338	-	-	-
—»—	16	14^{00}	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	13.29	—70	348
—»—	17	7^{00}	-	-	-	-	-	-	-	-	-	10.73	—63	325	-	-	-	-	-	-
Madrid	15	13^{00}	-	-	-	-	-	-	-	-	-	12.33	—72	329	-	-	-	-	-	-
—»—	16	18^{00}	-	-	-	-	-	-	-	-	-	11.24	—68	320	-	-	-	-	-	-
—»—	17	13^{00}	-	-	-	-	-	-	-	-	-	-	-	-	12.12	—67	334	(12.94	—68	346)

The brackets indicate that the surface in question is indistinct. The observations in Madrid do not permit any satisfactory grouping. The radio-sound observations at Sloutzk are not used in this table because of their lack of details. The ascent at Trappes on the 15th 9^{05} probably gives a too low temperature in the upper atmosphere. The soundings at Sealand are difficult to analyse because of their lack of details. The difference between the tropopause groups IV and V is rather indistinct.

Trappes and Madrid and appears more or less distinctly at several soundings at Uccle. This tropopause group with the highest potential temperature is thus distinctly developed only within the region of the subtropical anticyclone over SW Europe. The group disappears wholly in the regions nearer the polar front. The tropopause in question is situated high up, between 12.1 and 13.3 dyn km.

In all tropopause groups the potential temperature seems to decrease slightly polewards (see for instance the meridional cross-section fig. 29, p. 42).

It will be understood that many uncertainties attach to the above classification of tropopauses in groups. Already the small, but frequent, errors in the records of temperature and pressure make, in certain cases, a definite grouping on the basis of the potential temperature impossible. Particularly unsuited for a detailed analysis are those soundings which are published in whole degrees only. For future investigations of the structure of the tropopause it is desirable to increase the accuracy both of observation and of evaluation.

It is not to be expected that the tropopause groups established above will be identified from case to case with unaltered graduation of potential temperature. It is, however, very probable that multiple tropopause systems of some kind will occur in other similar situations. The same phenomen was earlier established in the cyclones of April 14—16, 1925 and February 2—3, 1933. These earlier cases differ, however, from the present one in that they concern the formation of a tropopause depression ("Tropopausentrichter"), at which the advective processes were of less importance than in the case before us. Our detailed analysis of the situation during February 15—17 results in the working hypothesis, that the tropopause in every synoptic situation in the temperate zone is likely to show multiple structure over some part of the map. This multiple structure naturally disappears in the distribution of the average temperature and can only be studied synoptically or with isopleths from ascents at short intervals.

Rather significant for the formation of any theory on the multiple tropopause is the fact that such a structure is limited to the region just over the sloping polar front surface and the region nearest to the south of that. This is also the region of maximum meridional tropopause inclination and the region of barocline tropical air in the upper troposphere, in other words a region within which a meridional contraction has distorted the normal meridional cross-section of the atmosphere. This meridional contraction is probably just the frontogenetical process, which in the lower atmosphere is known to exist in the barocline wedge of polar air and which is likely to exist as well in the barocline wedge of tropical air above.

To the question of how the multiple structure of the tropopause originates only vague and tentative answers can be given.

One possibility is that surfaces of discontinuity in the upper troposphere in their mode of formation may be "mirror pictures" of the surfaces of subsidence in the lower troposphere. The normal surface of subsidence is the lower limitation of downward air currents, and their counterpart in the upper troposphere may be the upper limitation of upward currents. Upward currents within the tropical air, both local and en masse, are to be found in the equatorial easterlies,[1] and that may be the region where the multiple structure of the tropopause is first created. Certain parts of that equatorial air find their way W of the subtropical anticyclones to extratropical latitudes, and may bring along a multiple tropopause. During the latter part of that transportation towards increasing latitudes the upper tropospheric air descends, since otherwise it would bring along a higher tropopause than has ever been observed in temperate latitudes. If, at the same time, the tropical air in mid-troposphere descends less, or even rises, the vertical contraction of the upper tropospheric air continues and favours the maintenance of the multiple tropopause system.

The outlined life history of the upper tropospheric air of equatorial origin must of course not be assumed to be adiabatic. That would lead to higher tropopause temperatures at the end of the trajectory than those observed. A systematic loss of heat by radiation must take place along the path of the upper tropospheric air, which is just what the radiation theories also claim (emission layer of Albrecht).[2]

More purely radiational theories of the multiple tropopause might also be attempted, but the authors have not succeeded in establishing any plausible solution.

[1] See Hydr. Phys. § 173.

[2] Met. Zeitschrift 1931, Heft 2 and 12.

IV. Discussion of Selected Isopleth Diagrams.

The next step towards establishing the complete synoptic aerological analysis will be the study of some isopleth diagrams. Several of the participating aerological observatories had a sufficient frequency of ascents to make it possible to construct reliable isopleths. Amongst these we have selected the Lindenberg, Uccle, Sealand, Ås and Lauttakylä isopleths for publication in this chapter. In all the isopleth diagrams time increases from right to left, so that they are comparable (although of course not identical) to vertical cross sections W—E seen from the south.

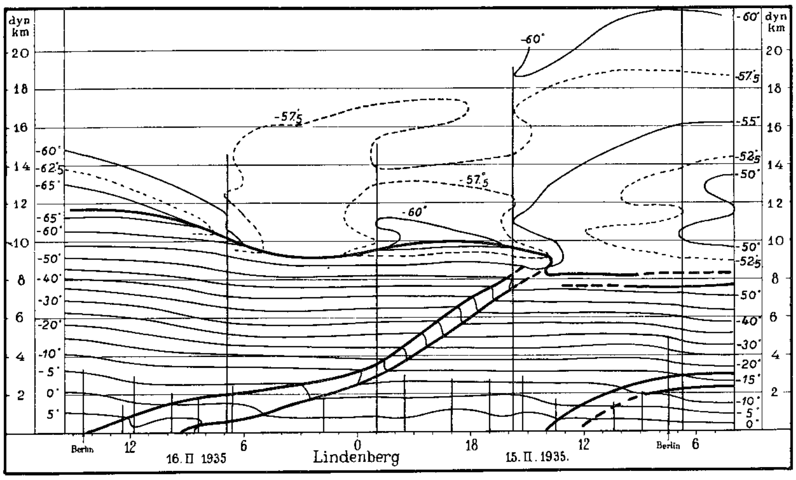

Fig. 18. Lindenberg isopleths.

1. Lindenberg Isopleths (fig. 18).

A series of frequent kite ascents, together with two Berlin aeroplane ascents, provide a good base for the analysis of the lower troposphere. Four sounding balloon ascents add thereto an outline of the temperature field higher up. There was one more high reaching balloon sounding in the morning of the 17th, but no reliable interpolation can be carried out through the preceding 24 hours, which were without high ascents (ice deposits spoiled two intervening ascents).

The important phenomenon presented by the Lindenberg diagram is the warm front surface, that can be followed from the ground up to the tropopause. The junction of the warm front surface with the tropopause has been drawn in accordance with the experience from Ås (see fig. 17), where the high ascents were more frequent than in Lindenberg. The ascents of the 15th at $15^h 44$ and $22^h 57$ (fig. 10) both show a sharply defined warm front surface, the frontal zone being only 0.4 dyn km thick. Where the warm front surface approaches the ground the frontal zone is, however, about 1.5 dyn km thick and has a complicated structure. The slope of the warm front surface appears smaller near the ground than higher up, but that is due only to the small speed of the front in the region just W of Lindenberg.

The isotherms in the tropical air run very nearly horizontally until the morning of the 16th. Later on, at the approach of the second disturbance, the tropical air temperature goes up. The tropopause height, which is rather closely correlated to the upper troposphere temperature, also remains relatively low during the first phase of the tropical air, but rises later. These isopleth features resemble those found at Ås (p. 31) where there was a relative minimum of tropical air temperature and of tropopause height in connection with the first occluded disturbance. Lindenberg was in the southern outskirts of that disturbance and shows the same upper air features only feebly (early in the morning of 16th). Uccle and Sealand isopleths show still less influence of that first disturbance. This is also the reason why the Lindenberg wet-bulb potential temperature at the warm front surface is lower than that in W. Europe (see table 2 and fig. 11).

2. Uccle Isopleths (fig. 19).

The Uccle sounding balloon ascents were made with great frequency, so that detailed data is available for the drawing of isopleths at all heights (except for the night between the 16th and the 17th). The ordinates, representing the various ascents, have been replaced by short vertical lines at the ground level. Under the thermo-isopleths will be found the following data from the Uccle self-recording instruments: the wind direction and Beaufort strength at each whole hour, the rainfall registration and the temperature registration. Next follows a set of "aerological barograms" from standard levels 2 dyn km apart plus the barogram at station level. And finally, at the bottom of the diagram, the directions and distances in km to the landing points of the balloons are inserted.

In the thermo-isopleths we see the lowest part of the warm front surface at the right. It is not so sharp as later on at Lindenberg (fig. 18), but has more the character of a frontal zone almost 2 dyn km

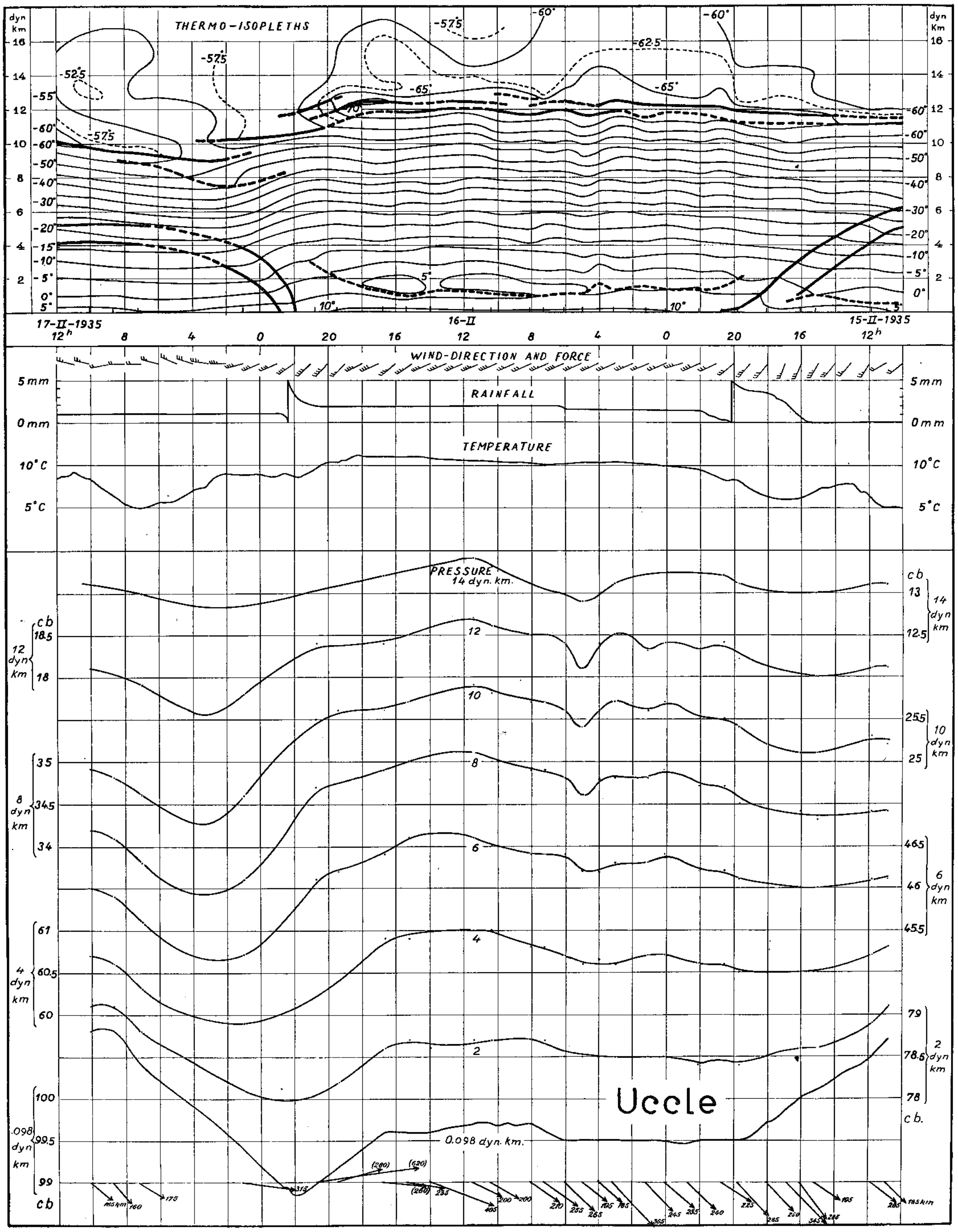

Fig. 19. Uccle isopleths.

thick in the vertical and 200 km broad in the horizontal direction. The upper limit of this frontal zone is by definition the warm front surface. The tropical air temper ature above the warm front slope goes on rising gradually until the warm front has also passed at the ground. (This fact was found and discussed already in Chapter II, p. 11). Then, in the warm sector proper, the temperature of upper levels remains practically constant with superimposed quick fluctuations of magnitude $\pm 2^\circ$ C, which are probably due to instrumental errors. To these, probably unreal, temperature fluctuations correspond unreal variations in the height and potential temperature of the tropopause (see Chapter III).

A well defined temperature inversion is found over the warm sector Stratus, at roughly 1.5 dyn km. Towards the warm front, and also towards the cold front, the inversion is lifted, and through the same process of rising motion it fades away. The saturated adiabatic cooling of the Stratus air (100 %), together with the stronger dry-adiabatic cooling of the air above (50—60 %), satisfactorily explains the vanishing of the thermal inversion in the rising parts of the warm sector air.[1] Moderate rain is also experienced behind the warm front and ahead of the cold front, probably falling not only from the Stratus but also from cloud layers higher up, that are also intensified by the general ascending motion. During the rains in question the temperature at the ground is slightly depressed under its normal value in the warm sector.

The cold front is of a frontolyzed type, but the barogram and the sharp maximum of rain intensity locate it quite well. Fig. 20 contains the Uccle ascents nearest before and after the cold front passage. The first one, at $16^h 50$ on the 16th, shows undisturbed tropical air conditions with a well-marked inversion (I) over the Stratus. At $19^h 30$ the same inversion is lifted and is on the point of vanishing. Simultaneously the inversion and the cloud underneath it has reached up to the isotherm of 0°, so that snow and ice particles can form in the upper part of the cloud. A little after the prefrontal precipitation reaches the ground, which can be taken as an indication of the importance of the solid particles in the upper part of the cloud in starting the rain.[2]

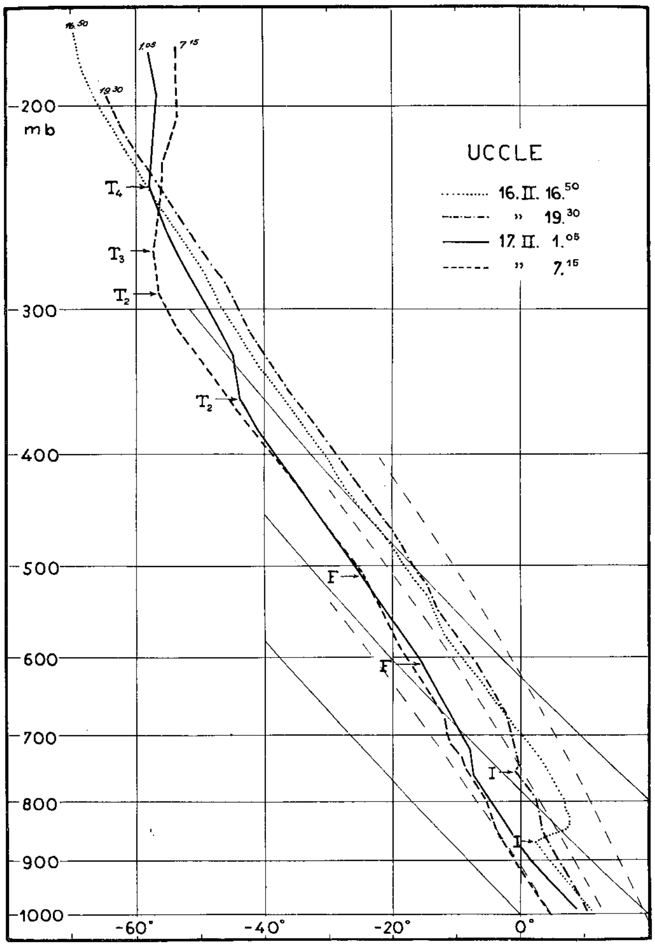

Fig. 20. Ascent before and after the cold front passage in Uccle.

Rather remarkable is the rise in temperature from $16^h 50$ to $19^h 30$ above the level of 680 mb. The same rise of temperature was also found in the Sealand series of ascents and is therefore probably real. As for interpretations see the discussion of the Sealand isopleths p. 27.

The two ascents behind the cold front ($1^h 05$ and $7^h 15$ of the 17th) show an indistinct cold front surface and above that tropical air of a colder kind than the prefrontal tropical air. The ascent at $1^h 05$ shows at 609 mb (3.87 dyn km) a transition from the small lapse-rate of the frontal zone to the uniform great lapse-rate of the tropical air above. That point being accepted as indicating the cold front surface, the connection with an analogous point at 511 mb (5.20 dyn km) on the 17th at 7^h seems appropriate, and from that time there is again a downward trend of the frontal surface. The moderate vertical thickness of the polar air tongue at Uccle is natural, considering the fact that the cold front

[1] See for instance H. v. Ficker: Über die Entstabilisierung eines aufsteigenden Luftmassensystems, Met. Zeitschr. 1936, Heft 12, p. 472.

[2] T. Bergeron: On the Physics of Clouds and Precipitation. Mem. de l'Association de Météorologie de l'U. G. G. I. Lisbonne, Sept. 1933, Paris 1935. W. Peppler: Zur Aerologie der Wolken, besonders des Nimbus. Beitr. Phys. fr. Atm. Bd. 23, Heft 4, p. 275, 1936.

slowed down over S-France and that a warm front approached over the British Isles by the time the ascent series came to its end.

The tropical air temperature drops considerably (at places as much as 10° C) from the time when the cold front passes at the ground and reaches a minimum about 5 hours later (fig. 19). Towards the end of the series the temperature above the polar front surface is again rising towards normal tropical air values. This phenomenon might already have been expected from the evidence presented in chapter II, that over Western and Central Europe the temperatures at the cold front surface are 3° to 4° lower than those at the warm front surface, and that the temperatures of the warm front surface are in their turn 4° to 8° lower than those of the undisturbed tropical air away from fronts. The isopleths now present the change from the warmest, undisturbed phase of tropical air to the coldest phase of tropical air, over the cold front surface, within about 7 hours, and a considerable slope of the isotherms thus results. We do not, however, introduce any frontal surface separating the cold and warm phases of the tropical air, because they appear to be simply the manifestations of: a) meridional elongations of the barocline tropical air over the polar front surface and b) vertical displacements of the same tropical air. These questions will be reconsidered later in connection with the Sealand- and Ås-isopleths and also in the chapter on aerological synoptics.

The tropopause over Uccle can be investigated with much detail by aid of the rich aerological data at disposal. It presents a multiple structure nearly all the time, which has been analyzed according to the principles set forth on p. 20. The most distinct of the several tropopauses is always indicated in full and the others in interrupted lines. With the arrival of the cold front the uppermost tropopauses disappear and lower ones dominate instead (see fig. 20). The lowest one (belonging to group II in table 9) never acquires the role of a real tropopause in the conventional meaning and fades away again towards the end of the series.

The barograms show the following phenomena. The pressure fall observed at the ground during the warm front approach disappears with height. A maximum of pressure is found in all the barograms on the 16th at noon, which marks the time of highest temperature in the warm air mass. A pronounced upper minimum of pressure coincides in time with the coldest phase of the tropical air, about 5 hours after the cold front passage at the ground. The pressure difference between the warm sector maximum and the following minimum reaches its greatest value, 17 mb, at 8 dyn km. The corresponding pressure fall goes on at a good rate also at the ground until the arrival of the wedge of polar air. From that moment the pressure effect of the advancing cold wedge overcompensates the pressure fall that goes on in the upper layers.

3. Sealand Isopleths (fig. 21).

The ascents have less frequency than in Uccle, but they suffice for the analysis of large scale features. A balloon ascent from Kew on the 15th at 12^h 33, and an aeroplane ascent from Duxford on the 15th at 8^h, have been inserted in order to obtain the structure of the warm front surface. When introducing these two ascents in the diagram, allowance was made for the time difference between the passage of the warm front at the three stations Sealand, Kew and Duxford.

The Sealand isopleths are rather like those of Uccle. The chief difference lies in the greater intensity of the cold front phenomenon, which will here be studied more in detail.

The cold front passage is best shown in the rain record, which indicates a 6 mm downpour from 16 to 17^h. The temperature drops at the same time (although gradually) and the barometer stops falling. One ascent was made just at the cold front passage at 16^h 05 and another at 18^h. The position of the cold front surface at the latter ascent is ill-defined, but the "characteristic curve" of the cold front surface determinations (p. 15) indicates 5.15 dyn km as the most probable level. The next morning the thickness of the polar air is again about 5 dyn km, after having had a maximum during the night. The resulting profile of the polar air tongue is very steep eastwards and gently sloping westwards.

The cold phase of the tropical air, which was also met with at Uccle, borders so abruptly towards the genuine tropical air that a fall of temperature of up to 15° C occurs during the two hour interval between the ascents. This phenomenon will be analyzed in fig. 22.

Fig. 22 contains the surroundings of the cold front at Sealand with isopleths for the absolute potential temperature. In that same figure the usual kinematical cross-section through a fast moving cold front has also been drawn. The lowest part of the warm air is moving more slowly eastwards than the cold

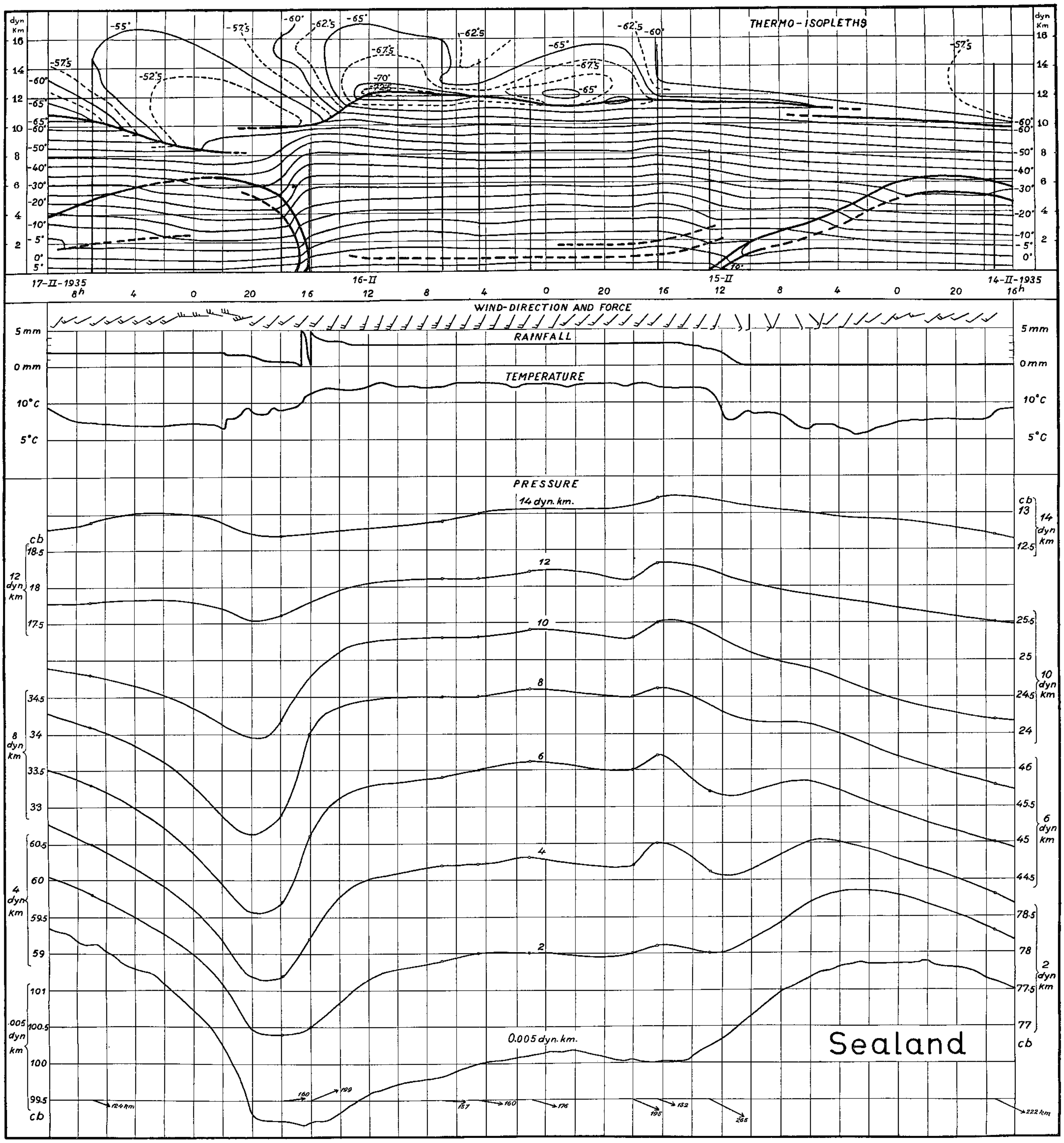

Fig. 21. Sealand isopleths.

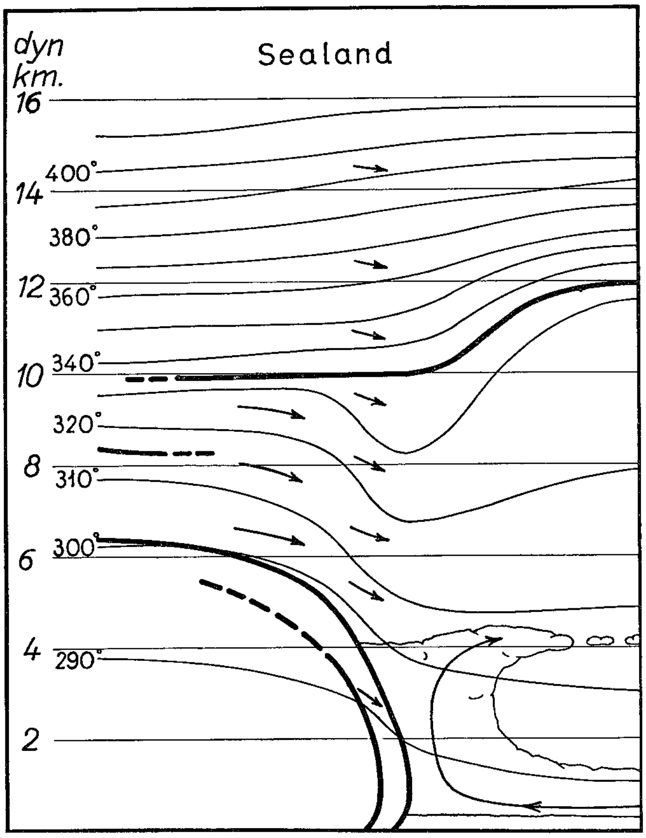

Fig. 22. Absolute potential temperature in the surroundings of the cold front at Sealand. The arrows indicate the projected relative air motion.

front and has thus a relative motion towards the cold wedge. A rapid ascent of the warm air must consequently take place at the cold front, and will continue up to a level where the warm air has a component forwards relative to the moving cold front. (Thermodynamically the saturated adiabatic ascent is in this case possible up to about 4 dyn km.) Still higher up on the cold front slope the warm air glides downwards and warms up dry-adiabatically.

We suppose, for a moment, that the cold front system undergoes no change of structure during its eastward displacement, so that the isopleth representation becomes equivalent to a synoptic W—E cross section. Furthermore, we suppose that the air has no component normal to that vertical cross-section. If dry-adiabatic changes of state prevail in all the down-gliding air, the stream lines of that air (relative to the moving front) would be parallel to the isentropes. The isentropes of the tropical air actually descend approximately parallel to the cold front surface and give a sort of thermodynamical picture of the down-gliding motion, that is assumed to take place there. The maximum descent of a particle, following the isentropes of the W—E cross-section, would be about 2 dyn km.

We shall later see that the descent of the tropical air on the cold front slope takes place along a cyclonic track, so that the air particles do not remain in the W—E cross-section but have a component from the south at the end of the downward displacement. This component from the south explains part of the horizontal gradient of tropical air temperature in the isopleth diagram, but probably most of that gradient comes from the descending motion that is forced upon the tropical air. In other words the interpretation probably still holds true, that the tropical air, after crossing the crest of the tongue of polar air in a relatively cold state, quickly warms up again in the free-air-foehn on the lee-side of the polar air mass.

In fig. 22 we have accordingly indicated the descent of the tropical air by stream lines that slope only a little less than the isentropes. Almost certainly the descent extends up as far as the lower stratosphere and contributes to the warming up of the stratosphere and the lowering of the tropopause in the same region.

When treating these questions on the basis of isopleth diagrams one should of course always remember that only a succession of three-dimensional synoptic pictures of the fields of pressure, temperature and motion can give the necessary complete survey of the problem. We shall therefore return to the same questions in the "Aerological Synoptics" (Chapter V).

The upper pressure minimum, that follows 3 to 4 hours after the cold front passage at the ground, has a steeper profile than that found farther south at Uccle. For instance in the 8 dyn km level the maximum values of $\frac{\partial p}{\partial t}$, respectively in front of and behind the pressure minimum, were roughly:

Sealand	—6 mb/hour	+ 2 mb/hour
Uccle	—2.5 »	+ 2 » .

These values of $\frac{\partial p}{\partial t}$ would be rather extreme even for the conditions at the ground, where the absolute pressure is about 3 times that of the 8 dyn km level. A value of $\frac{\partial p}{\partial t}$ of 6 mb/hour is very seldom met with at all in extratropical cyclones.

The negative values of $\frac{\partial p}{\partial t}$ in front of the upper pressure minimum are damped out downwards, firstly by the temperature distribution inside the tropical air mass (the pressure minimum coincides with the temperature minimum and therefore flattens out downwards), and secondly by the influence of the sloping cold front surface. In that way there is in the ground barogram of Sealand a $\frac{\partial p}{\partial t} = 0$ under the region of greatest pressure fall in 8 dyn km, and in that of Uccle there is even a $\frac{\partial p}{\partial t} = +2$ mb/hour at the corresponding phase of development.

The positive values of $\frac{\partial p}{\partial t}$ behind the upper pressure minimum are both in Sealand and Uccle transmitted all the way down to the ground. The first part of the rise in the Sealand ground barogram, which amounts to as much as 6 mb/hour, very likely coincides in time with the first rise of pressure behind the upper pressure minimum, but unfortunately there are no ascents available to ascertain it. If there was no appreciable change in the average temperature of the air column up to 8 dyn km during that phase, $\frac{\partial p}{\partial t}$ at 8 dyn km would be about 2 mb/hour. Actually there was probably still a little fall of temperature going on in the column, at any rate near the ground, so that $\frac{\partial p}{\partial t}$ at 8 dyn km must have been even less than 2 mb/hour. The sudden rise in pressure starting at 20^h at Sealand would thus be the rise behind the upper pressure minimum transmitted to the ground and also slightly increased by thermal advection. Very soon afterwards the thermal advection is such as to raise the average temperature of the column 0—8 dyn km, and the transmission of pressure rise from above is then counteracted by the thermal change. Towards the end of the diagram $\frac{\partial p}{\partial t}$ is transmitted downwards with unaltered magnitude, instead of being tripled as it would be if the column 0—8 dyn km had kept its temperature unchanged. At Uccle the rise of pressure aloft seems to be compensated in that same way and is accompanied by a rise of pressure at the ground of the same magnitude as in 8 dyn km.

To avoid misunderstanding we want to emphasize that this description of simultaneous pressure changes at different levels does not pretend to unravel the complicated causality at work. For instance the "transmission of a positive pressure change $\frac{\partial p}{\partial t}$ from 8 dyn km to the ground with simultaneous tripling of its amount" means that the air is accumulating in the column 0—8 dyn km (mainly by horizontal convergence), so as to make the rise of pressure three times greater at the ground than at 8 dyn km. That accumulation of air is a process in which all layers from 0—8 dyn km have their share, and is of course not attributable to any particular layer.

The Uccle and Sealand barograms for the ground level are quite typical exponents of the following well established rule of synoptic meteorology, already mentioned in Chapter I (p. 6).

The cold front is rather well defined in the pressure field at some distance (say 500—1000 km) away from the centre, but nearer to the centre the cold front trough becomes more indistinct and is followed by another trough (Sealand the 16th at 20^h) which is "non-frontal".[1])

The barograms from successive levels over Sealand now give the indication that the non-frontal trough in question is simply the projection to low levels of the upper air pressure minimum. The axis of the upper air pressure minimum is so nearly vertical that its projected image in low levels becomes quite well-defined. If its axis were tilting, the projection of the upper air minimum to low levels would of course have no clear meaning.

In the opinion of the authors the upper air pressure trough is to be explained as a cyclonic bend of the westerly tropical current, produced at the lee side of the polar air tongue.[2] The polar air tongue, that moves more slowly eastward than the upper part of the tropical current, acts as an obstacle and forces the tropical air to descend on its leeward (eastern) side. The air that converges, in order to replace the descending masses, necessarily comes into cyclonic rotation. Hence the formation of the cyclonic bend (and later also the complete cyclonic circulation) imbedded in the upper westerly current.

The described lee phenomenon ought to acquire its greatest intensity where the polar air tongue

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1 An analogous non-frontal trough (Febr. 2—3, 1933) was examined aerologically by E. Palmén in Mitteilungen d. Met. Inst. d. Univ. Helsingfors, No. 26, 1934.

2 Hydr. Phys. § 183, p. 810—816.

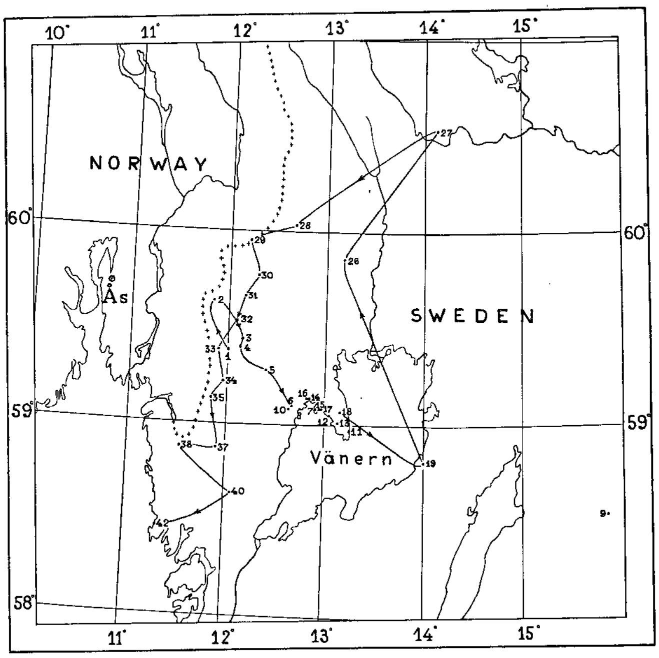


Fig. 23. Distribution of the landing points of the Ås balloons. The numbering of the balloons is the same as that used in the isopleth-diagram.

reaches high up and has a steep profile. Therefore it is more strongly developed at Sealand than in Uccle. We shall soon see how the same lee phenomenon is modified at Ås on the north side of the cyclone track (fig. 25 p. 34).

4. Ås Isopleths.

The map of the chronologically numbered landing points of the Ås balloons is presented in fig. 23. The map in question is of special interest since it shows the resultant direction of upper winds during the various stages of the cyclone passage near to the south of the place of ascent. It will be referred to later.

27 sounding-balloon ascents were at disposal for the construction of the thermo-isopleths (fig. 24). The times of the ascents are indicated by short vertical lines at the ground level and their maximum heights by small crosses. The loss of 6 meteorographs in succession, probably on the field ice of Lake Vänern (see fig. 23) left a serious gap in the observations from 7^h to 16^h on the 16th, but apart from that the density of the soundings is satisfactory. However, a further technical mishap has led to the loss of the bottom part of many of the ascent curves. The isopleths are therefore based on less data in the lower half of the troposphere than higher up. In order to facilitate the interpolation across the gaps, the temperature record of the mountain station Gaustatoppen (1829 m, 120 km W of Ås) has been consulted. Allowance was thereby made for the temperature difference "free atmosphere—mountain" obtained from the complete ascents.

From the beginning of the series air of rather recent arctic origin occupied Scandinavia, and the first Ås ascents show the upper limit of that air mass. This arctic front surface descended nearly to the ground in the evening of the 15th, when maritime polar air masses pushed into the Skagerak, but no change of air mass occurred at ground level at Ås. During the following night the arctic air mass again increased in thickness by advection from the interior of Scandinavia. Its upper limit, however, became more and more diffuse and must therefore be dropped from the isopleth diagram as soon as the more powerful polar front disturbances arrive from the west. The first of these disturbances was an occlusion with a tongue of rather mild maritime polar air behind. The earth front of that occlusion, which was consequently of warm front type, came only to the west coast of Norway. The corresponding warm front surface in the Ås isopleths therefore becomes almost horizontal through the rest of the series and does not reach the ground. For the technical reason explained above, the ascents unfortunately give no information about the exact position and structure of the frontal surface in question after the 16th at 7^h. The same applies to a great part of the frontal surface belonging to the last disturbance, the centre of which went about 250 km south of Ås. The bottom of the warm sector valley has accordingly been assumed (not observed) to reach down to 2.5 dyn km at the moment when the centre of the depression was at its nearest. That moment is not coincident with the moment of lowest pressure in the ground barogram because of the deepening of the travelling depression. The passage of the centre across the meridian of Ås, as determined from the backing of the surface wind, took place at about 20^h, whereas the lowest pressure in the Ås barogram was recorded at 23^h. The rising polar front surface behind the centre is found at 3.24 dyn km in the ascent at 22^h and at gradually higher levels in the following ascents. The maximum altitude of the polar front surface is reached in the forenoon of the 17th, and it descends gently towards the end of the series (as in Sealand).

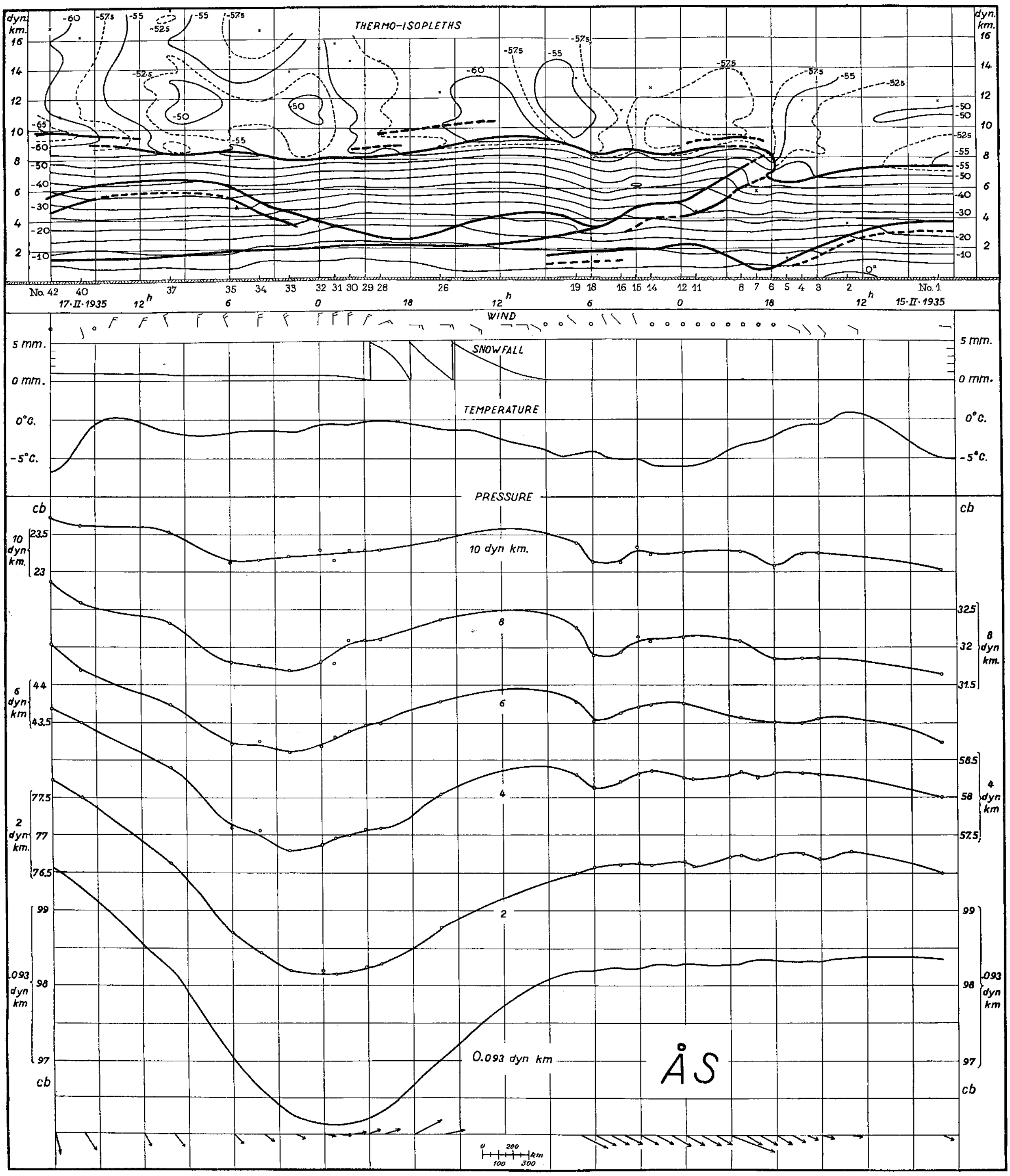


Fig. 24. Ås isopleths.

The tropopause, during most of the period, is without multiple structure and presents no difficult problems of analysis. The tropopause group II (table 7, p. 19) prevails over the tropical air, but in the ascents at 20^h of the 15th and 16^h and 18^h of the 17th group III pre-dominates. The interesting sudden rise of the tropopause in the evening of the 15th, at the first arrival of the tropical air, has already been treated in detail in chapter III (p. 17).

The isopleths during the time interval from the 15th at 18^h to the 16th at 7^h give some information about the structure of the front half of an occlusion (the "rear" of it did not arrive at all because of the vigorous development of the secondary). A well marked warm front surface during this period descends from its junction with the tropopause to the occlusion valley at 3.5 dyn km (16th 6^h). The rise of temperature due to the passage of that warm front surface is, for instance at 6 dyn km, about 10° C. But within the tropical air the rise is immediately followed by a gradual fall of temperature, which goes on until the arrival of the occlusion valley, and by that time the fall has accumulated to 7° C. Just over the occlusion valley the tropical air is thus only slightly warmer than the polar air at the same level before the arrival of the warm front surface. This is different from the picture of a warm front passage with an open warm sector behind, as represented by the Uccle and Sealand isopleths. There a gradual rise of temperature followed the warm front passage, and in Ås a fall. The intermediate case, viz. constant tropical air temperature after the warm front passage, is found in the Lindenberg isopleths.

The relatively low tropical air temperature in the occlusion valley over Ås is probably in the first line due to the lifting of the tropical air that is necessarily involved in the occlusion process (partly also due to the meridional advection as described below for the following young disturbance). The downward slope of isothermal surfaces towards the occlusion valley would, however, be less accentuated in a real synoptic cross-section than in the isopleth diagram. The reason being that, during the 12 hours we are considering, the "age" of the occlusion increases, and thereby also the cooling of its tropical air content progresses.

On the other hand the isopleth diagram is just the right tool for analyzing the connection between the temperature changes and the simultaneous pressure changes at a fixed point. The striking features in this case are mainly the following. The sudden tropopause rise at the arrival of the tropical air has practically no effect on the barograms below. A piece of troposphere has replaced a piece of stratosphere in the column from 6.5 to 9 dyn km, but the mean temperatures of the two pieces differ very little, so that the replacement does not have much pressure effect. Another curious compensation process takes place during the gradual lowering of the warm front surface. Normally, that ought to produce a considerable fall of pressure in all levels underneath, but in this case only a fall of 1 or 2 mb results, because above the warm front surface warm tropical air is being replaced by colder tropical air. This is probably the mechanism normally at work during the approach of a decaying (filling) occlusion system.

During the following 24 hours (morning of the 16th to morning of the 17th) Ås is under the influence of the new vigourously deepening depression, the centre of which passed 250 km S of Ås in the evening of the 16th. The variation of the tropical air temperature during the cyclone passage follows the same phase as that which is known on the south side of the cyclone track. In other words there is a maximum of upper troposphere temperature on the 16th at noon[1] and a minimum of upper troposphere temperature during the night from the 16th to the 17th (incidentally the times of maximum and minimum coincide practically with those at Uccle). The tropopause level oscillates in the same phase, the highest tropopause coinciding in time with the highest troposphere temperature and the lowest tropopause with the lowest troposphere temperature. The stratosphere temperature varies as usual in the opposite phase of that of the upper troposphere. The pressure in the upper troposphere and the lowest stratosphere is finally in phase with the upper troposphere temperature and the tropopause height. In brief, the whole system of Dines-Schedler correlations are in operation. Upper perturbations of

[1] That maximum of upper troposphere temperature coincides with the lacune in the series of soundings mentioned on p. 30. Indirect proofs of its existance can, however, be found in the flight directions on fig. 23 or bottom row of fig. 24. The flight directions are backing from WNW to WSW during the lacune period, which indicates the passage at that time of a wedge of high pressure, that is a maximum in the upper air barograms. That pressure maximum can only exist if it coincides with the assumed maximum of upper troposphere temperature. The distribution of stratophere temperature during the same period has been drawn on the isopleth diagram in analogy with the principles set forth in Geof. Publ. Vol. IX, no.9, fig. 19.

Table 10.

Temperature in the lower stratosphere and at the tropopause during the passage of the cyclone.

Ascent	Temperature at different levels (dyn km)						Tropopause	
	8	9	10	11	12	13	dyn km	temp.
16. 16h	−51.7	−57.3	−59.2	−59.0	−59.0	−59.0	8.70	−57.7
20	−52.8	−56.0	−57.0	−57.0	−57.1	−57.4	8.15	−53.8
21	−53.0	−56.3	−56.7	−57.0	−57.0	−57.0	8.11	−54.0
22	−52.3	−54.7	−55.0	−54.1	−55.7	−55.9	8.00	−52.3
23[1]	−55.1	−54.0	−53.1	−53.8	−54.4	−55.5	8.09	−55.4
24	−51.3	−51.2	−50.5	−49.9	−49.6	−51.4	8.00	−51.3
17. 2	−53.7	−52.1	−51.1	−51.9	−51.2	−53.0	7.97	−53.8
4	−54.2	−53.5	−53.5	−52.2	−52.8	−53.9	8.22	−55.0
6	−53.6	−55.0	−52.7	−51.0	−50.3	−54.5	8.53	−55.8
10	−51.0	−52.2	−52.0	−49.2	−50.4	−50.6	8.38	−54.0
16	−53.2	−59.1	−61.2	−57.3	−57.0	−57.0	9.70	−62.0
18	−51.3	−59.8	−64.5	−60.9	−59.5	−59.9	9.82	−64.8

[1] Probably too cold. Therefore rejected at the construction of the isopleths.

that description would have been satisfactorily explained by the quasi-horizontal meridional advection, arising when a westerly tropical current passes across crests and valleys in the polar front surface (see Phys. Hydr. §§ 182—183). The maximum N elongation produces the conditions observed on the 16th at noon, which normally would be found farther south.[1] The maximum S elongation produces conditions at or a little after midnight, which normally prevail north of the Ås latitude.

At close inspection a new feature however comes in, which is not known from isopleths south of the cyclone track. From the 16th at 20h to the 17th at 10h, the variations in tropopause level are very small, and before the 17th at 2h they are even beyond the limit of accuracy of height determination. In other words the tropopause trough has a flat bottom. The variation of stratosphere temperature nevertheless continues until midnight, and we are presented with the strange phenomenon of a considerable progressive rise in stratosphere temperature over a flat tropopause (see table 10).

The explanation is probably the following. The system of upper perturbations, normally found south of the cyclone track, is connected with the up-gliding over the warm front and down-gliding over the cold front slopes. The necessity of a decrease of the vertical amplitudes of air particles with increasing height leads to the formation and maintenance of the described normal upper perturbation (see Phys. Hydr. § 183).[2] The upper perturbation arranges itself in just the same fashion N of the cyclone track, as long as there is an upper westerly tropical current that assures faster air transport than the cyclone displacement (see fig. 25 I). That condition is always fulfilled in the first stage of the young polar front cyclone, and from that period dates the "normal features" of the Ås upper perturbation. But during the deepening of the cyclone the tropical westerly current decreases in strength north of the cyclone track and is finally reversed to an easterly current, when the aging depression enters the vortex stage (see for instance fig. 40, p. 55). Before the tropical current is reversed from westerly to easterly relative to the earth, it is reversed relative to the system of coordinates attached to the eastward moving cyclone. The reversal will take place earliest at the bottom of the trough of tropical air and later also extend to the upper part of the troposphere. Fig. 25 II and III represent stages of that development, and fig. 25 III is assumed to give the most likely kinematical cross-section of the cyclone about the time when it passed south of Ås. The Ås balloons in that part of the series went eastwards, but the total flights were so short (see fig. 23)

[1] Tropopause surfaces of groups III and IV, otherwise only present S of Ås, are faintly visible at 16h, 20h and 21h.

[2] We want to add to the description in the quoted book, that the damping to zero of the vertical motion is normally reached well above the tropopause, so that the sign of the vertical motion is the same in the lowest stratosphere as at the frontal surfaces underneath.

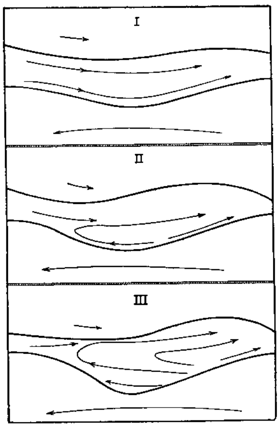


Fig. 25. W—E kinematical cross-section N of the W—E cyclone track. The stream lines represent the assumed air motion relative to the moving cyclone. I: nascent cyclone, II and III later stages of the deepening cyclone. (The streamlines should not be interpreted as approximate trajectories, because of the rapid evolution of the cyclone, and especially because of the great components of air motion perpendicular to the cross-section.)

that a considerable part of the tropical air must have had less eastward speed than that of the cyclone. An important consequence of this is the beginning of up-glide motion from the trough of tropical air in westward direction. This upward motion in fig. 25 III also affects the upper troposphere, so that the tropopause trough gets filled by accumulation of troposphere air underneath. According to the Ås isopleths the stratosphere still displays the high temperature acquired by horizontal advection and descending during the normal regime over the young cyclone, while the underlying part of the troposphere has started ascending. This kinematic system, which is represented schematically in fig. 25 III, supplies a tentative explanation of the thermo-isopleth distribution over Ås during the cyclone passage.

The upper air pressure minimum some hours after the cold front passage, which was found in the Uccle, and still more accentuated in the Sealand, barograms, is to be found also over Ås, but it is much weaker than at the stations south of the cyclone track. At 8 dyn km the entire pressure fall, during the approach of the said pressure minimum, was at Uccle 17 mb, at Sealand 19—20 mb and at Ås about 8 mb only. Interpreting as before the upper air pressure minimum as a "lee effect" in the westerly tropical current on the eastern side of a polar air ridge, we find it natural that the intensity of the phenomenon decreases very rapidly from the south to the north side of the cyclone track, because with such kinematical conditions as indicated on fig. 25 III there is no lee effect any more.

It will be seen later in the Lauttakylä isopleths (that represent the conditions still farther north of the cyclone track and at an older stage of the cyclone), that there is nothing left of an upper pressure minimum behind the pressure minimum at the ground. On the contrary, the lowest pressure occurs there earlier in the upper than in the lower layers. The synoptic picture of these conditions can be consulted on fig. 40 (p. 55), which gives the stream lines of the tropical current just above the polar front surface. A lee eddy has formed in that current above the cold front slope. Its position is such that it must have produced pronounced pressure minima in the Uccle and Sealand upper air barograms, but only slight effects in the corresponding Ås barograms. Lauttakylä, finally, is not affected by that tropical air eddy at all.

5. **Lauttakylä Isopleths** (fig. 26).

8 sounding-balloon ascents are available. The most serious gap in the observations occurred during the night 16th—17th, and the two nearest ascents before that gap did not reach the stratosphere. The construction of the thermo-isopleths in the region in question has been carried farther than strictly allowed on the basis of the Lauttakylä ascents alone. In particular, at the drawing of the upward jump of the tropopause at the change from polar to tropical air underneath (16th at 17^h), the Ås isopleths have been taken as a guide.

The presence of the tropical air in the upper troposphere over Lauttakylä is first revealed by the ascent on the 16th at $21^h 55$. The frontal zone that separates it from the polar air underneath is rather diffuse as compared with what it was at Ås at the corresponding level, but the wet-bulb potential temperature at the front surface has very nearly the same value as was found at Ås (see fig. 11). The dissolution of the frontal structure has apparently been carried further by the continued ascent and inherent cooling of the tropical air. The next ascent on the 17th at $10^h 02$ presents similar conditions and there is up to that time no doubt about the real existence of the frontal surface. The two last ascents, however, give no definite indication about the position of the frontal surface, which is therefore indicated as hypothetical towards the left end of the diagram. It is worth noting that the temperature in the upper troposphere, which is presumably still tropical in origin, does not fall during that last period. It is the polar air underneath that becomes warmer and thereby makes the polar front surface disappear in the temperature field. Probably, this is a subsidence effect connected with the polar air anticyclone over northern Finland.

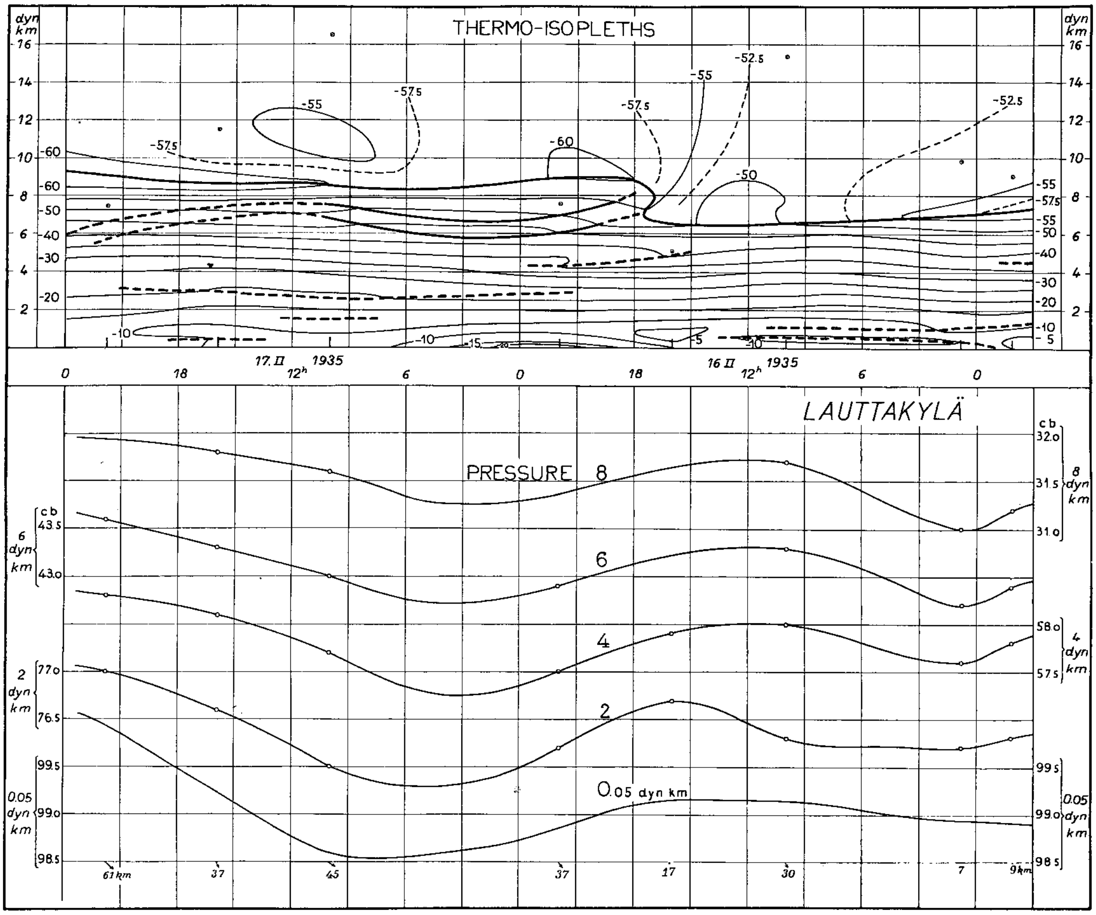


Fig. 26. Lauttakylä isopleths.

The polar air in the Lauttakylä diagram presents several surfaces of subsidence at various levels. Some of them can be connected from ascent to ascent, others not. Near the ground there is most of the time an inversion of temperature, that reaches occasionally as high as 1.5 dyn km. During the early part of the 16th a turbulent layer with normal convective lapse-rate starts forming from the surface upwards, but it does not reach beyond 0.58 dyn km, and is of course annihilated by radiational loss of heat at the ground during the following night. The upper limit of that turbulent layer is drawn with the same sort of line as the surface of (assumed) subsidence higher up, but it is of course another kind of phenomenon.

The barograms corroborate the result from Ås that there is no such steep profile pressure minimum in the upper layers as that found south of the cyclone track. Furthermore it shows (as Ås did not show) an earlier pressure minimum in the upper levels than at the ground. This striking result is however based on two ascents only, and it is desirable to have it verified by more cases of similar nature.

V. Aerological Synoptics.

1. General Remarks.

The most complete sets of upper air observations (including those of mountain observations) are available at or about the synoptic hours of the 15th at 7^h and at 18^h, the 16th at 7^h and at 18^h and the 17th at 7^h, the same at which the synoptic maps (fig. 2—6) are valid. For each of these synoptic hours a set of maps representing the topography of the isobaric surfaces 100, 90, 80 . . . . . 20 and 10 cb has been constructed.[1] To save space we publish only some of these maps, namely the topographies of the 100, 80, 60, 40 and 20 cb isobaric surfaces. The relative topographies between the successive standard isobaric surfaces, which have of course served during the construction of the absolute topographies, have been omitted on the published maps. Instead we have entered on the upper air maps the values of the specific volume

$$s = \frac{R\,T_v}{p}$$

at the standard pressures $p = 100$, 80, 60, 40 and 20 cb respectively. These specific volumes are plotted at each station together with the dynamic height of the isobaric surface in question. The stippled curves represent the synoptic distribution of specific volume at the isobaric surface, the topography of which is shown by the full lines.

The juxtaposition of the two sets of curves may serve two (or more) purposes. Firstly they show the orientation in space of the isobaric-isosteric solenoids, and secondly they provide the base for determining thermal advection. In order to serve this latter purpose the isosteres have been numbered also with their respective values of virtual temperature T_v, and thereby furnish a representation of the temperature distribution along the isobaric surface. At the same time the direction of the geostrophic wind is given by the direction of the contour lines of the isobaric surface (high absolute topography values to the right when looking with the wind), and its strength is given by the formula

$$V = \frac{g}{2\,\omega \sin \varphi} \tan \alpha,$$

where α is the angle of slope of the isobaric surface. Admitting the geostrophic wind as a first approximation to the real wind in the free atmosphere, we can easily deduce from the topography of the isobaric surface and the temperature distribution at the same surface the sign and approximate amount of the thermal advection

$$\frac{\partial T_v}{\partial t} = -\mathbf{V} \cdot \nabla T_v$$

due to the horizontal air motion.

At each one of the synoptic hours we will also compare the flight directions of the sounding balloons with the direction of the geostrophic wind in the various layers. Tables 11—15, which are compiled for that purpose, show that in many cases it is possible to establish with certainty a systematic deviation of the real wind direction from that of the isobars, also above the frictional layer. From such deviations in direction between real and geostrophic wind the acceleration component along the path of the air particle can be found (qualitatively) according to the well known law: If the wind in the free atmosphere, where friction is known to play a negligible part, deviates towards low pressure, the air must have a component of acceleration forwards along its path, and vice versa if it crosses the isobars towards high pressure it will have a backward acceleration component (retardation).

In general the accuracy of the maps of the upper air pressure distribution is not very great. For instance, details like the concentration of isotherms and the bend of isobars at the frontal surfaces easily disappear in the open network of aerological stations. The study of such details will not be possible until the synoptic-aerological data have improved considerably.

2. 15th of February, 7^h (fig. 27).

The map in the upper left corner of the page is put in to recall the arrangement of air masses at the time in question. The warm front of the approaching western disturbance extended from Western Ireland down along the French West coast. The Sealand, Trappes and Uccle ascents went up through the rather diffuse warm front surface and furnish the data upon which the topography of the warm front surface (stippled lines) has been based. The tropopause (contour lines in full), which is high over the approaching warm sector, slopes down to lower levels along the tropopause valley over Central and Northern Europe. The warm front surface and the tropopause intersect along a line North Sea—Western Germany—N. Italy.

The 100 cb map (right bottom corner on page 37) shows the conditions already known from the ordinary synoptic map. The isotherms indicate quite cold conditions over Scandinavia, especially in the interior. The 80 cb isotherms maintain the picture of the cold reservoir over Scandinavia although with less accentuated extremes, since we are now above the surface inversions. The 60 cb and 40 cb isotherms repeat a similar picture with a cold tongue extending from the northern reservoir down to central Europe.

[1] The applied method is described in V. Bjerknes: Dynamische Meteorologie und Hydrographie, Vol. I, Statik, Kap. 7, and need not be reiterated here.

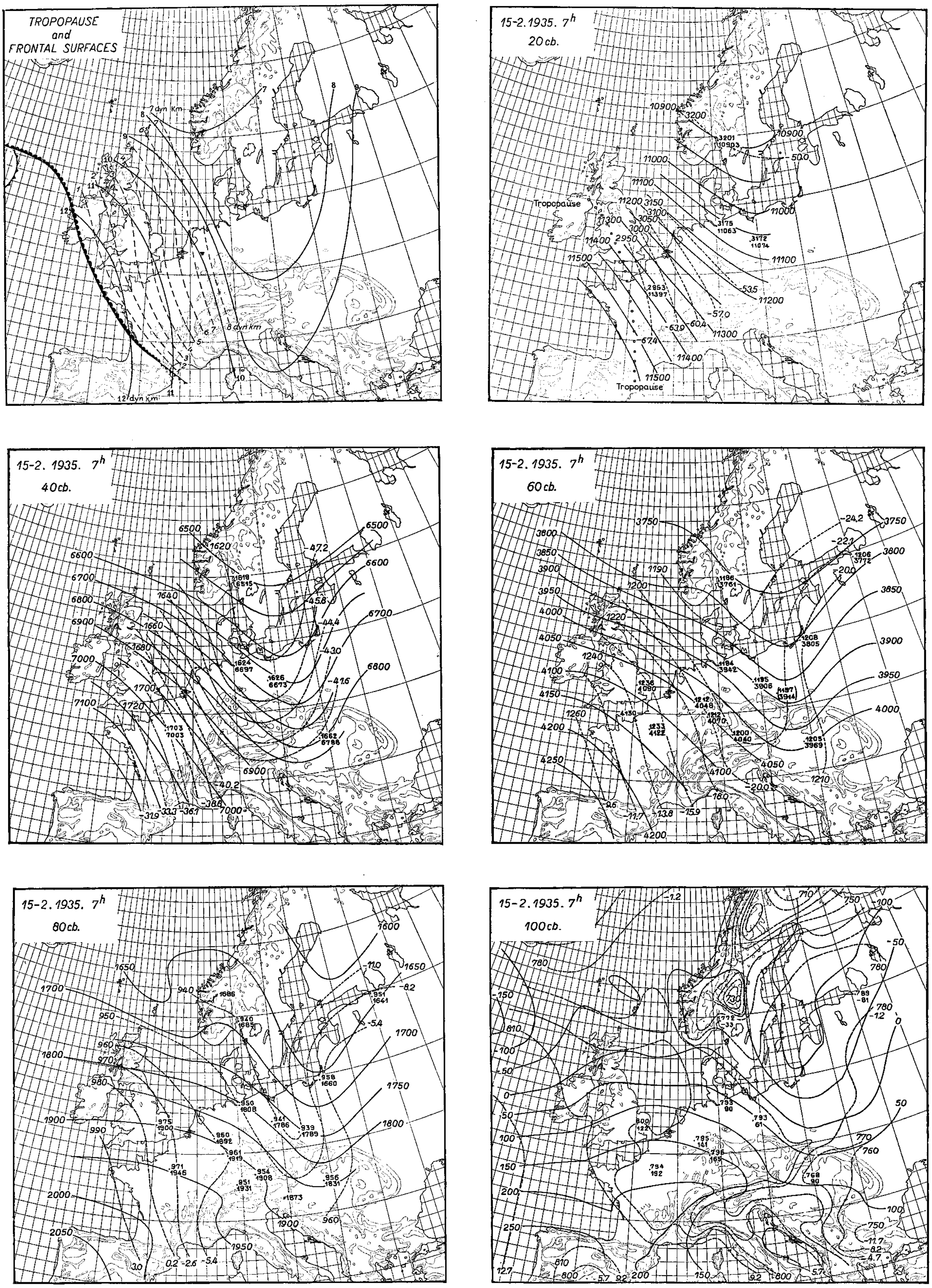


Fig. 27. Upper left: Topography (dyn km) of the tropopause and frontal surfaces.
Other maps: Topography (dyn m) of standard isobaric surfaces and distribution of specific volume (m³/ton) and virtual temperature (°C) at the standard isobaric surfaces.

The 20 cb surface is over the greatest part of the map situated within the stratosphere, and the warmest air on that map is accordingly found in the north. Over SW Europe the 20 cb surface is under the tropopause, and their probable line of intersection is inserted and marked "tropopause".

In the barocline parts of the troposphere the pressure gradient generally increases with height. Above the same regions the stratosphere is usually barocline too, but with the opposite horizontal temperature gradient, so that the pressure gradient there decreases with height. Furthermore, within the troposphere the region W of the cold tongue of air exhibits a veer of the isobars with height, and to the E of it the isobars back with height.

Table 11. 15. II. 7^h.

Flight direction (°), distance (km) and ceiling (cb), as compared with geostrophic wind direction (°) and speed (m/sec) at standard isobaric surfaces.

Station		Trappes	Hamburg	Lindenberg	Budapest	Ås
Flight	Dir. dist.	318° 100	315° 185	291° 209	210° 85	294° 77
	pmin cb	17.5	4.1	3.3	24.4	17.6
Geostrophic wind	10 cb		290° ?	280° ?		
	20	320° 42	310° 35	285° 27		300° 24
	40	320° 29	315° 40	310° 30	240° 30	305° 25
	60	310° 18	310° 38	315° 35	250° 28	310° 20
	80	290° 12	305° 26	315° 33	270° 24	310° 13
	100	260° 12	305° 20	310° 23	280° 19	315° 9

From the comparison of flight directions and geopstrophic directions (table 11) the following results are obtained. In Trappes the flight direction practically coincides with the geostrophic direction, as measured on the 40 cb and 20 cb maps, although the geostrophic direction below 40 cb deviates systematically to the left of that. The conclusion seems inevitable that the real wind in a considerable part of the column (probably the upper part) has deviated to the right of the geostrophic wind. In other words, it is a case of retardation of the air particles passing the Trappes column. The same is the case over Hamburg, whereas over Lindenberg the deviation from geostrophic flow seems to be small. Budapest and Ås have the opposite deviation of that shown by Trappes and Hamburg. The amount of deviation at Budapest appears unreasonably large, which must however be due to the uncertainty in the construction of the isobaric topography at the edge of the aerological network.

The geostrophic wind measured on the 100 cb map at the warm front near Scilly is 19 m/sec. It strikes the warm front at an angle of 70° and its component normal to the front is about 18 m/sec. The 40 cb level intersects the warm front surface just E of Uccle,[1] where the geostrophic wind is about 32 m/sec from 315°. That wind vector strikes the contour lines of the warm front surface at an angle of about 37°, and its component normal to the contour lines is about 19 m/sec. Thus, the increase of geostrophic wind with height does not lead to any appreciable increase with height of its component normal to the warm front surface, because the geostrophic wind veers with height so as to approach the direction of the contour lines of the warm front surface. In the situation under discussion the real wind direction must have deviated to the right of the geostrophic direction, as was shown for the upper air over both Trappes and Hamburg (Table 11). The up-slope wind component at the warm front surface in the 40 cb level would thus be smaller than 19 m/sec, perhaps even smaller than 16.7 m/sec, which was the average speed of the warm front normal to itself during the spell 7^h to 18^h of the 15th.

This rather rough computation has been presented in order to show why a warm front surface, that

[1] There is no warm front trough on the constructed upper pressure maps, perhaps in part because of the insufficient number of aerological stations.

reaches all the way up to the tropopause, does not produce up-glide cloud in its uppermost part. In that region the tropical air probably advances slower than the entire front phenomenon and glides slightly downwards on the warm front surface. This is usually corroborated by the altostratus shield, that does not keep intact up beyond the 4 or 5 km contour line on the warm front slope. We shall return to the same phenomenon during the description of the upper air conditions 12 hours later. At that time the warm front phenomenon is situated in the middle of the aerological network and can be better analyzed.

The advective change of virtual temperature $\frac{\partial T_v}{\partial t} = -\boldsymbol{V} \cdot \nabla T_v$ is positive all over Western Europe both within the polar and the tropical air, because the wind cuts across the isotherms from warm to cold. It is important to bear in mind that this "warm advection" takes place all the way up to the tropopause, although the wind direction there is from NW. In the stratosphere that same NW-wind produces cold advection, as can be seen from the 20 cb map. These conditions are always found in the foremost part of young depressions.

3. 15th of February 18^h (fig. 28).

The warm front has advanced across England and Western France. The line of junction of the warm front surface and the tropopause runs from Ås to Vienna, about 900 km ahead of the warm front. It is at this time that the foremost edge of the tropospheric tropical air advances over the lowest strata of the receding polar air stratosphere, producing thereby the folding of the tropopause, which was revealed by the series of Ås ascents (p. 17). This folding of the tropopause takes place over a zone, hardly exceeding 100 km in width, and it is therefore quite possible that it also existed 12 hours earlier, but was concealed in the space between ascents.

The general temperature distribution is similar to that of the 15th at 7^h, but a further rise of temperature has taken place in the tropical current over western Europe. At the 40 cb surface it is now even warmer over England than over Spain. The coldest air is still found over Scandinavia. There is warm tropospheric advection over the whole of western and central Europe (not only at the warm front surface but also at both sides of it), and the inherent veer of isobars with height is clearly seen. The tropopause has risen over the same region and now intersects the 20 cb surface farther NE than in the morning. The stratosphere (to the NE of the tropopause line on the 20 cb map) shows cold advection and backing of the isobars with height.

Table 12 shows that the wind component towards high pressure, which was found in the morning at Trappes and Hamburg, is now to be found at all stations. This is immediately seen from the data at Trappes, Uccle, Vienna and Ås where the flight direction deviates to the right of the geostrophic direction in every layer, but also Sealand, Hamburg and Lindenberg show flight directions to the right of the resultant geostrophic vector from the layers traversed (allowance made for the different vertical thickness of the layers). Such a unanimous result from all stations must be trustworthy, and we are consequently entitled to infer from it that a great part of the atmosphere is in a state of retardation during its flow across the investigated area.

Table 12. 15. II. 18^h.

Flight direction (°), distance (km) and ceiling (cb), as compared with geostrophic wind direction (°) and speed (m/sec) at standard isobaric surfaces.

Station		Sealand	Trappes	Uccle	Hamburg	Lindenberg	Vienna	Ås
Flight	Dir. dist.	293° 195	319° 225	315° 252	305° 140	318° 164	342° 123	301° 123
	pmin cb	15.3	9.6	22.9	7.7	5.5	15.0	14.2
Geostrophic wind	10 cb	280° 50	310° 40		290° 30	305° 20	315° 15	295° 40
	20	295° 60	315° 48	310° 50	310° 36	330° 30	340° 25	300° 45
	40	295° 46	310° 38	305° 38	315° 40	325° 31	340° 27	300° 45
	60	290° 40	300° 35	290° 34	270° 20	305° 24	315° 23	285° 28
	80	280° 30	290° 30	270° 30	270° 21	300° 20	305° 20	270° 28
	100	275° 30	280° 29	250° 33	260° 21	275° 18	305° 14	100° 5

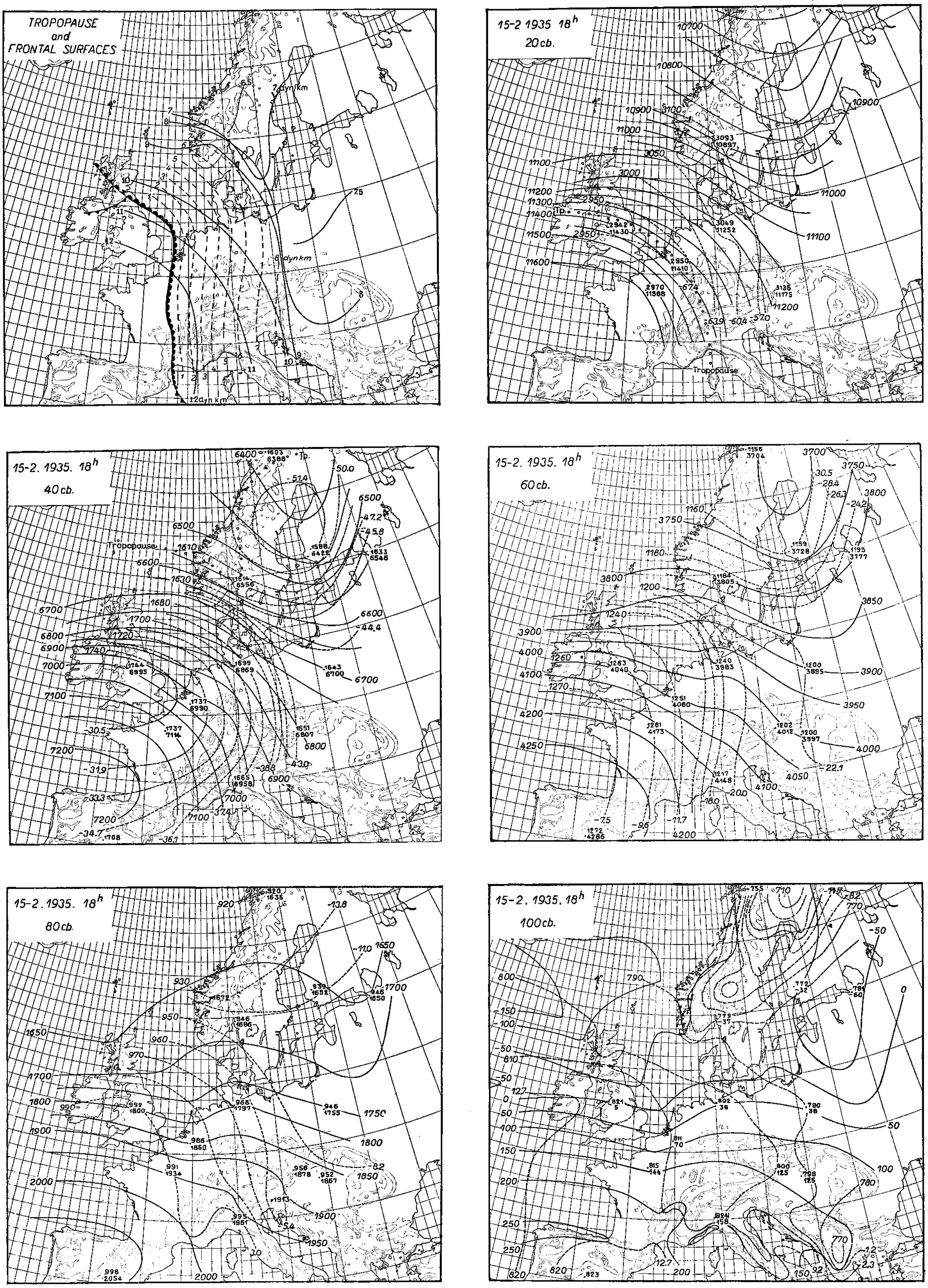


Fig. 28. Upper left: Topography (dyn. km) of the tropopause and frontal surfaces.
Other maps: Topography (dyn m) of standard isobaric surfaces and distribution of specific volume (m³/ton) and virtual temperature (° C) at the standard isobaric surfaces.

The upper air pressure maps lend independent support to that statement. From the shape of the isobars can be seen that a strong current enters from the W over the British Isles (up to 60 m/sec over Sealand) and spreads like a fan over the European continent, where it evidently decreases in speed. Similar conditions also reign 12 hours later, so that the acceleration of individual particles,

$$\frac{d\boldsymbol{v}}{dt} = \frac{\partial \boldsymbol{v}}{\partial t} + \boldsymbol{v} \cdot \nabla \boldsymbol{v},$$

is given with good approximation by the second right hand term alone. That term will generally have a horizontal component perpendicular to the wind direction (centrifugal or centripetal component), with which we are not concerned for the moment. But, in addition, it has a component along the wind direction, and this component is obviously directed backwards in the case of a fan-wise spreading current. In other words, we find from the synoptic upper air maps the same retardation component which was inferred independently from the comparison of flight directions and geostrophic wind. We shall return to this question in connection with the maps of the 16th at 7^h (p. 45).

At this point we have found it useful to supplement the synoptic upper air maps with two synoptic vertical cross-sections, the one from Spain to N Scandinavia (which will be denoted: "meridional cross-section"), the other from the warm sector over England to the cold region over Poland ("zonal cross-section").

The "meridional cross-section" (fig. 29) is based on the data from ascents at Madrid, Trappes, Uccle, Hamburg, Ås, Lauttakylä and Riksgränsen. This cross-section was defined as running along a great circle Trappes—Ås, which passes near enough to Madrid, Uccle and Riksgränsen to make their observations directly usable. The Hamburg and Lauttakylä ascents, which are a little off the line, are also rather well fitted into the diagram. The deviations from simultaneity are greatest in Madrid and Lauttakylä, but at those points local variations with time were insignificant. Therefore, for all practical purposes, the cross-section can be considered as being synoptic and as referring to the 15th of February at 18^h.

On top of the diagram (fig. 29) the ascent curves are traced, and small arrows are entered to point out the discontinuities, which have been taken as representing the polar front, the single or the multiple tropopause respectively.

The polar front surface (marked F) is traversed by the Uccle ascent very close to the ground. The Hamburg ascent shows the usual feeble lapse-rate in the frontal zone, and the front surface proper (marked F) is allocated at the point from where upwards normal lapse-rate is reestablished. At Ås the polar front surface and the tropopause join, as described in connection with the Ås isopleths. The resulting inclination of the polar front surface appears steepest at the upper part of the slope, but that is due to the fact that the meridional cross-section intersects the contour lines rather obliquely in the lowest part of the slope near Uccle.

At this time the tropopause shows good examples of the multiple structure earlier studied in detail on some of the isopleth diagrams. The tropopause groups, originally defined by the isopleths of the short interval ascent series, can now be studied geographically. Well into the domain of the tropical air (Madrid) the tropopause group numbered 5 is the only one represented. The same tropopause can be identified at Trappes, Uccle and Hamburg (arrows numbered 5), where it eventually fades away. Another tropopause, numbered 4, can be found at Trappes and Uccle. Tropopause no. 3, the most conspicuous one of those represented in Hamburg, is not present farther south in Uccle and is just distinguishable farther north in Ås. The tropopause numbered 2 is the lowest one of those found in tropical air regime. It extends from Hamburg, where it only produces a slight irregularity of the ascent curve, to Ås where it is the actual tropopause in the conventional meaning.

When representing the topography of a multiple tropopause system on synoptic maps we would, of course, too much confuse the picture if we tried to draw the contour lines of each one of them. Therefore, in all such maps in this memoir the position in space of a multiple tropopause is represented by an "average tropopause" in the middle of the bundle of tropopauses. When sloping, this average tropopause has a steeper inclination than that of each individual tropopause. For instance, in the meridional section the average inclination of the tropopause bundle from Trappes to Ås (not counting the downward jump at the latter station) is 0.32/100, but the individual tropopauses in the bundle show inclinations not greater than 0.17/100.

The relation of the individual tropopauses to the isentropes can be studied in the bottom part of

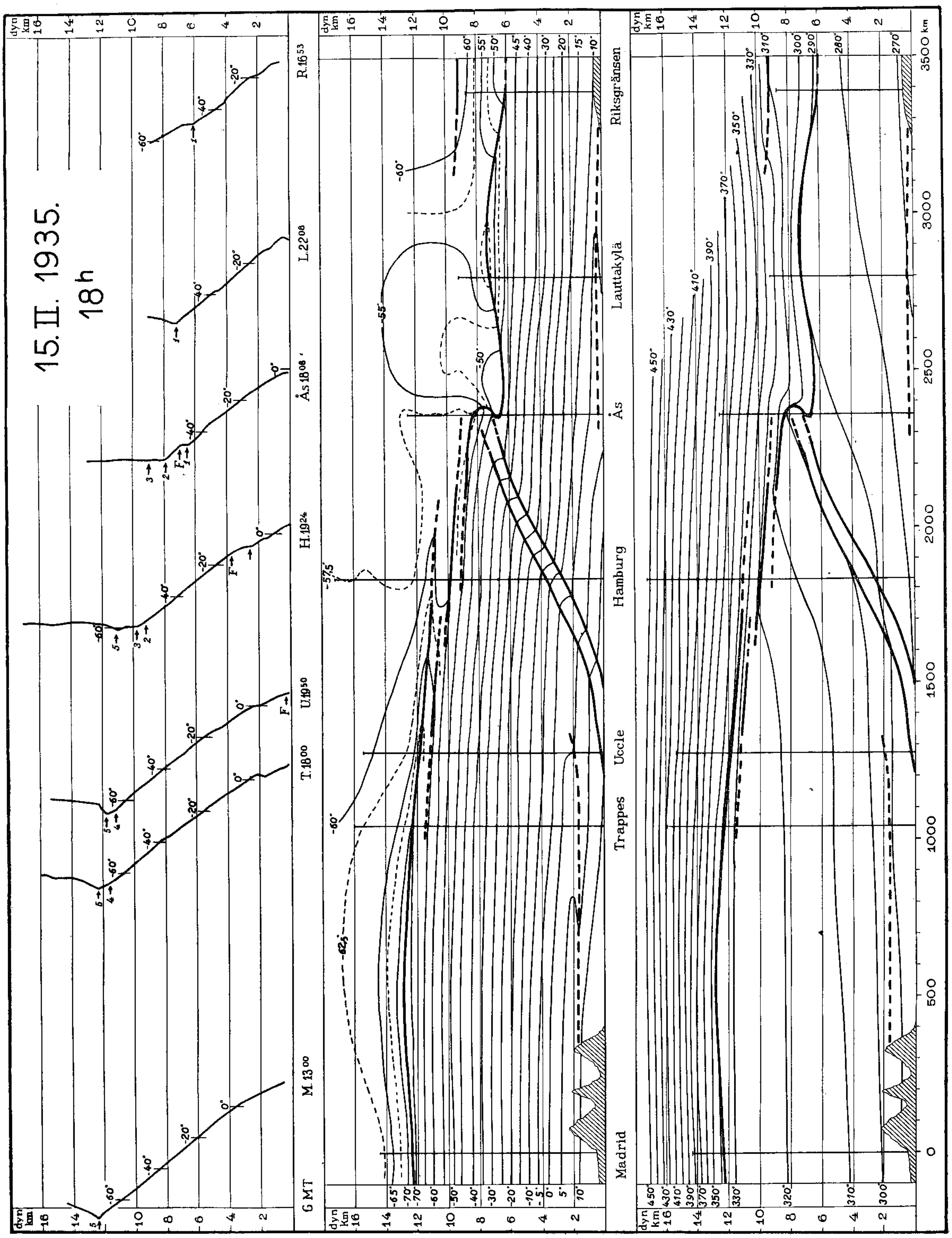


Fig. 29. The "meridional cross-section", evening of the 15th of February, 1935. Top: The soundings used. Middle: Isotherms. Bottom: Isentropes.

the diagram (fig. 29). By definition these tropopauses are quasi-isentropic, but actually they show a systematic slight decrease of entropy northwards.

The polar air tropopause (numbered 1) is non-multiple in Lauttakylä, just as it was at Ås too in the corresponding situation before the arrival of the tropical air. Farther north, at Riksgränsen, there is one discontinuity (numbered 1) at 6.1 dyn km, which perhaps connects with the tropopause at Lauttakylä and Ås in the way shown in the diagram. Above that level the lapse-rate is again considerable, so that another tropopause-like surface, which has been tentatively entered in the diagram, must exist beyond the reach of the ascent. This kind of temperature distribution is in the arctic winter frequent enough to influence the statistically determined average field of temperature.[1]

The tropical air well S of the polar front is quasi-barotrope (already shown in fig. 8), but the same air mass is barocline, with a meridional temperature gradient of about 1° C/100 km, where it overruns the wedge of polar air. The bottom part of the diagram describes the same state of affairs by an upward trend of the isentropes from S to N above the polar front surface. The line of potential temperature of say 300° thus remains between 2 and 3 dyn km in the tropical air S of the polar front, but reaches 8 dyn km at the northermost edge of the tropical air over Ås.

The "zonal cross-section" (fig. 30) is based in the western portion on the Sealand-(Kew)-isopleths for the afternoon of the 15th and in the eastern portion on German and Polish ascents from the evening of the same day. In spite of the lack of absolute simultaneity of the ascents, the diagram ought to represent a fairly reliable approximation to a synoptic cross-section referring to the evening of the 15th.

The general temperature distribution in the "zonal cross-section" is rather similar to that of the "meridional cross-section" because both of them lead from complete tropical air regime with warm troposphere, cold stratosphere and high tropopause to complete polar air regime with cold troposphere, warm stratosphere and low tropopause.

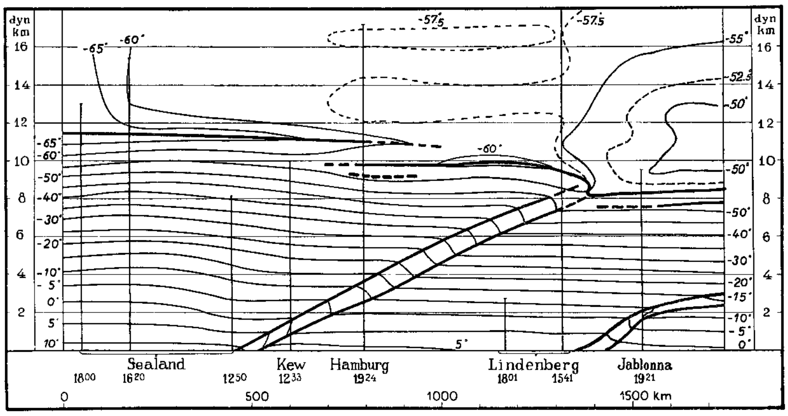


Fig. 30. The "zonal cross-section", evening of the 15th of February, 1935.

The Hamburg ascent, which is the same as that in the meridional cross-section, goes through the "tropopause bundle". The uppermost of the tropopauses (group 5) can be identified in the Sealand ascents, and the middle one (group 3) in Lindenberg. The existence of a well defined warm front surface all the way up to the tropopause is shown by the Lindenberg ascent, and a sudden rise in tropopause level seems likely just E of the ascent under consideration.

It is to be noted that the temperature distribution in this zonal cross-section, especially in the tropopause region, is not in every way typical for a vertical cross-section through a simple warm front surface, because of the influence of the first occluded disturbance to the north (see also p. 23).

An attempt will now be made to study the motion of the warm sector air relative to the moving warm front. In fig. 31 two auxiliary lines have been drawn, which are supposed to be contained in the sloping warm front surface. The one runs from Ås to the nearest point of the warm front over the North Sea and the other E—W along 50° N, in other words not far from the described meridional and zonal, cross-sections respectively. At the points where the auxiliary lines go through the 100, 80, 60 and 40 cb surfaces we have measured the geostrophic wind[2], given by the maps on page 40, and introduced it at corresponding points in the new map (fig. 31). (This can be seen at all 4 points on the line towards Ås, but on the other auxiliary line it has been necessary to omit the vectors at the 2 lowest points in order not to complicate the picture). These absolute wind

[1] E. Palmén, Met. Zeitschr. 1934, p. 17.

[2] This is, of course, a simplification, which introduces some inaccuracy. According to table 12 the real wind deviates to the right of the geostrophic.

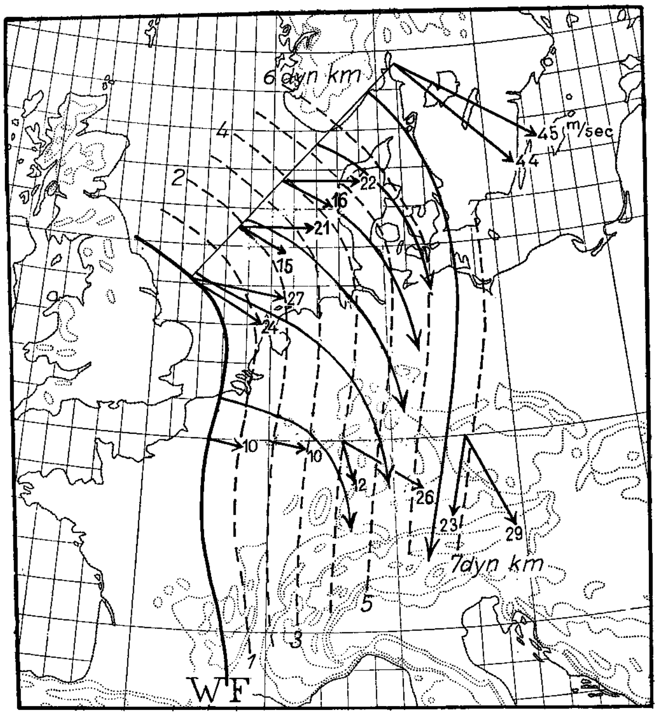


Fig. 31. Stream lines for the geostrophic motion of the tropical air relative to the moving warm front, 15th of February 1935, 18^h.

vectors all have an up-slope component. We now subtract the component of motion of the warm front phenomenon itself (10 m/sec respectively 18 m/sec along the auxiliary lines), and arrive at the relative wind vectors shown in fig. 31.

At the 40 cb points the relative wind is practically parallel to the warm front contour line, which means horizontal gliding along the front surface at that level (about 6.5 dyn km). At all the lower points the relative wind vector has a certain up-slope component, very slight over the North Sea, only a few m/sec, but up to 10 m/sec over France and Western Germany. Stream lines based on the relative wind vectors have been entered and show the up-slope flow bending anticyclonically and approaching asymptotically a contour line of about 6.5 dyn km height. The relative stream lines also serve as an approximate picture of the warm air trajectories relative to the warm front, provided that the kinematical structure of the warm front phenomenon does not change too rapidly.

Fig. 32 gives a schematical vertical cross-section along 50° N of the kinematics of the up-slope current. The relative geostrophic W-components at the warm front surface in this vertical cross-section are the same as those inserted along 50° N in fig. 31. The corresponding vertical velocity components near the warm front surface follow from the kinematic boundary condition. The maximum upward velocity component is found on the lowest part of the slope, where the geostrophic up-slope component is about 10 m/sec and where the inclination of the warm front surface amounts to about 1.1/100, hence the vertical component of the up-gliding air amounting to 0.11 m/sec. The upper stream lines have been given a smaller inclination than that of the warm front surface, thus assuming $\frac{\partial v_z}{\partial z} < 0$ and $\frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} > 0$ in the ascending wedge of tropical air. This assumption is justified by the fact that all ascending motion forced upon a stable medium, like in our case the tropical air, will be damped with height. $\frac{\partial v_x}{\partial x}$ measured eastward along the vertical cross-section is evidently negative, since v_x decreases to zero at the eastern end, and therefore $\frac{\partial v_y}{\partial y}$ must be definitely positive. In our case that must imply an increase of the southward component of the tropical air the farther south we proceed.

It is also important to remember another kinematic property of the ascending wedge of tropical air, namely the veer of the wind with height. The geostrophic directions over Uccle in table 12 show that convincingly, in that the veer with height in the tropical air column, from the 80 cb to the 20 cb level, amounts to as much as 40°. A similar veer with height also takes place in the rest of the ascending wedge of tropical air, because it follows as an inevitable consequence of the slope of isotherms downwards towards the east (which is seen in fig. 32).

In every vertical air column, inside the ascending wedge of tropical air, there is thus an increasing northerly advection with increasing altitude. This would tend to set up an unstable vertical lapse-rate of temperature, but the vertical contraction $\left(\frac{\partial v_z}{\partial z} < 0\right)$ of the originally stable air column has the opposite effect, and the net result is in our case a vertical lapse-rate smaller than, but close to, the saturated adiabatic.

The foremost edge of the warm front altostratus might be expected to coincide with the trajectory along the 6.5 dyn km contour line in fig. 31, which intersects the plane of fig. 32 in the point C. However, if we had been able to base the preceding construction on the real wind, which deviates to the right of the geostrophic, we would have found a somewhat lower position for the horizontal relative trajectory. This restricts the space inside which the altostratus can exist. Furthermore the foremost thin edge of the altostratus probably breaks up by mixing with the air above and below. Actually, the impervious cover of altostratus had its eastern limit approximately along the 4.5 km contour line of the warm front surface.

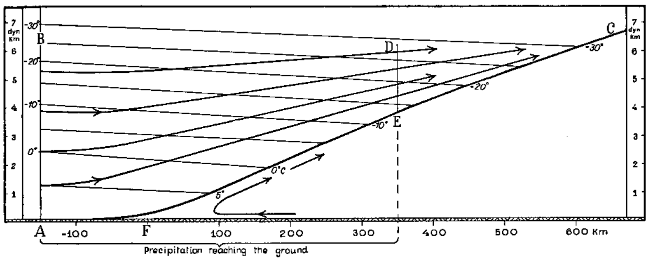


Fig. 32. Schematical vertical cross-section illustrating the stream-lines of the geostrophic air motion relative to the moving warm front.

4. 16th of February 7^h (fig. 33).

The primary depression, now between the Färoes and Shetland, is occluded. The post-frontal polar air is as usual warmer than the pre-frontal polar air and rises above the latter. The occluded front between the two kinds of polar air runs at the ground along the Norwegian and Danish W coast to the region of Hamburg, and the upper cold front, that shows how far up-slope the western polar air has advanced, must be oriented from Hamburg northwards past Oslo. The secondary depression system follows closely behind over Scotland and profits by the continued warm air supply for its further development.

The great scale temperature distribution resembles that of the evening before with warmth over the tropical air anticyclone in the SW and extreme cold in the NE corner of the map. The isobaric contour lines veer slightly with height and the gradient for W and NW winds increases definitely with height, except for the tropical air south of France, in which a reversal of the normal meridional temperature gradient (and also a reversal of the meridional tropopause slope) has taken place.

The tropopause intersects the 20 cb surface along the elongated minimum of temperature from N England over Holland towards the E Alps. To the NE of that line (in the stratosphere) the thermal advection is negative and the geostrophic wind is backing with height.

Table 13. 16. II. 7^h.

Flight direction (°), distance (km) and ceiling (cb), as compared with geostrophic wind direction (°) and speed (m/sec) at standard isobaric surfaces.

Station		Sealand	Trappes	Uccle	Lindenberg	Vienna	Budapest	Ås	Lauttakylä
Flight	Dir. dist.	274° 157	308° 145	308° 255	320° 171	331° 215	315° 120	296° 203	332° 30
	pmin cb	9.3	7.5	15.1	11.6	6.9	9.3	14.0	10.0
Geostrophic wind	10 cb	270° 58	275° 35	280° 50	305° 43	300° 38	270° 32		
	20	275° 65	290° 40	290° 60	305° 45	305° 48	280° 35	290° 29	290° ?
	40	275° 52	285° 28	290° 40	315° 40	320° 33	270° 29	295° 34	280° 16
	60	270° 43	285° 20	285° 33	315° 22	310° 18	280° 19	290° 16	260° 7
	80	270° 35	280° 24	285° 30	285° 15	280° 19	260° 19	180° 8	- 0
	100	270° 35	275° 24	285° 30	200° 5	285° 20	245° 25	115° 5	45° 5

All the flight directions in table 13 deviate from the geostrophic directions in the same sense as was found in table 12. At Budapest and Lauttakylä the deviations are larger than what seems reasonable, which must be due to some uncertainty in the construction of the upper pressure field in the outskirts of the map. At Sealand the deviation is now quite small, but all the other stations show considerable deviations, which must be real. The phenomenon is of course still connected with the retarded flow of particles through the fan-wise spreading current.

In fig. 34 the approximate path of a sample particle of that spreading current has been tentatively constructed, stationary conditions being assumed. The path is supposed to remain at the 60 cb level and is chosen so as to end over Vienna on the 16th at 7^h. The method of construction is one of approximation, starting with the construction of the geostrophic trajectory and then taking into account the deviation of the real trajectory from the geostrophic one.

The geostrophic trajectory leads back to the point marked 60 cb, just north of Sealand, on the 15th

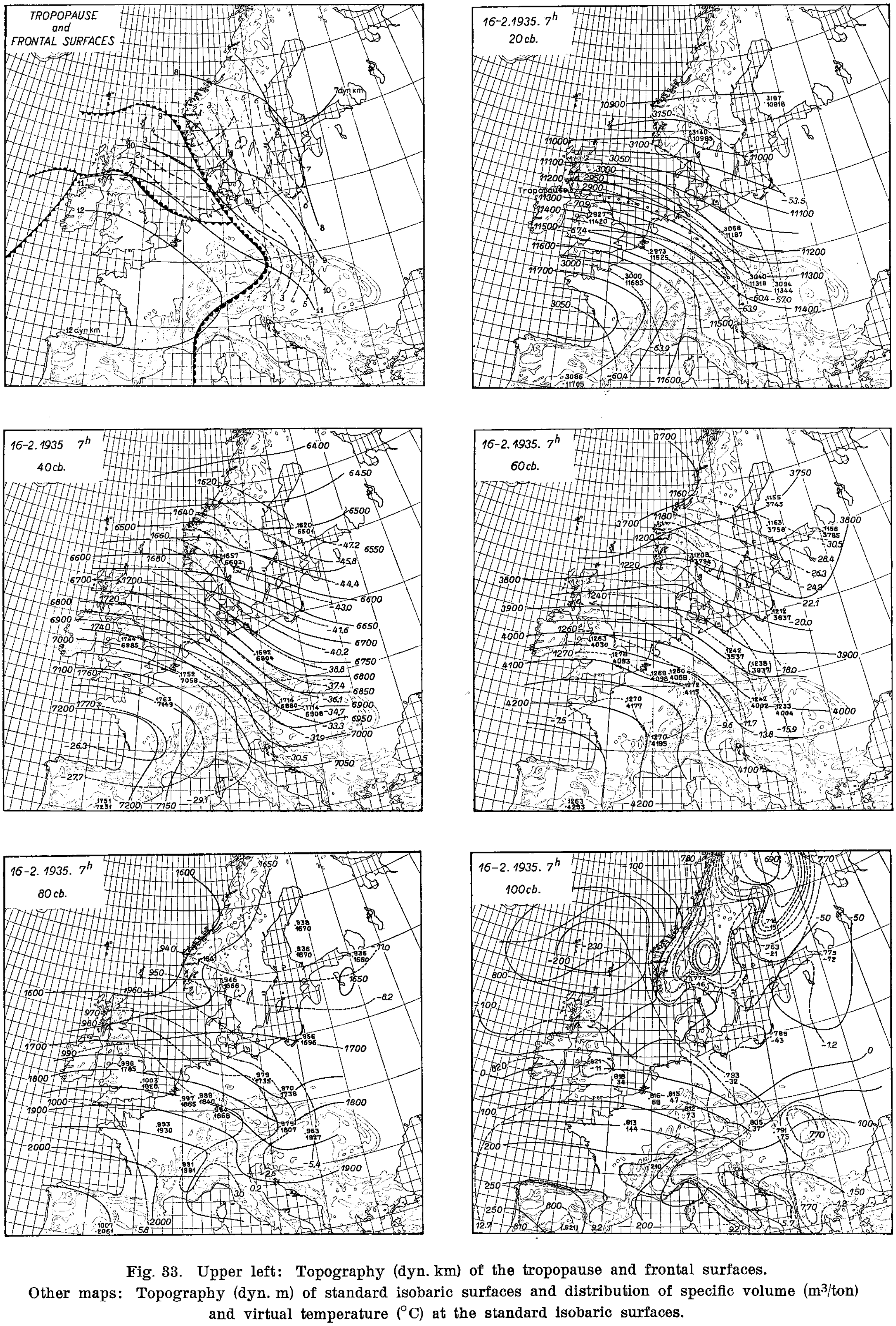


Fig. 33. Upper left: Topography (dyn. km) of the tropopause and frontal surfaces.
Other maps: Topography (dyn. m) of standard isobaric surfaces and distribution of specific volume (m³/ton) and virtual temperature (°C) at the standard isobaric surfaces.

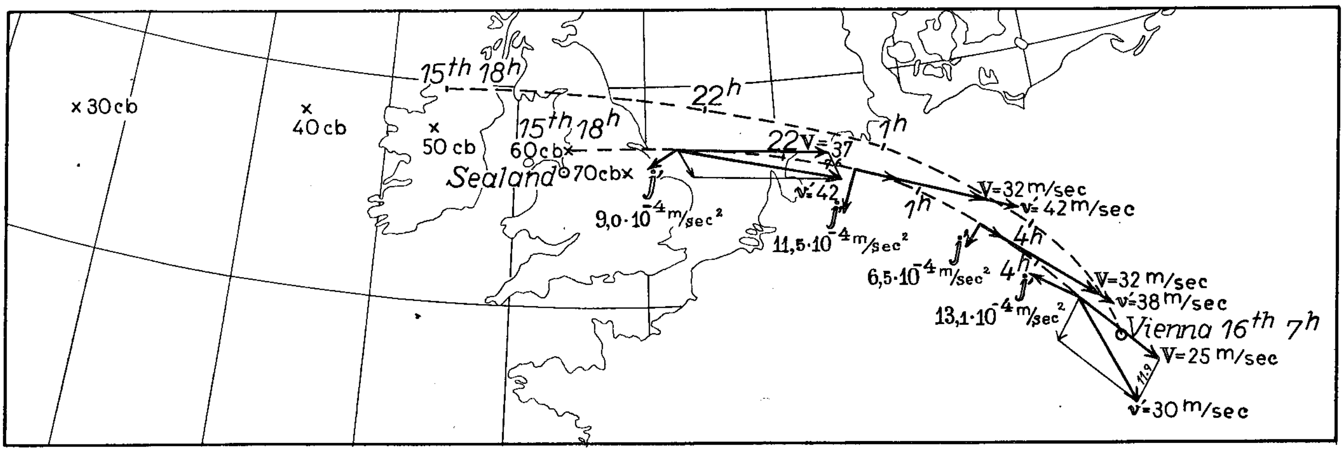


Fig. 34. Approximate path of an air particle in the fan-wise spreading current under stationary conditions.

at 18^h. At that point the geostrophic wind was 270° 43 m/sec, and at Vienna, the end of the trajectory, the geostrophic wind was 310° 18 m/sec. Geostrophic wind values are also available at intermediate points on the trajectory, so that 1st approximation acceleration vectors can be constructed at any selected point. In figure 34 they have been constructed over the time intervals 18^h—22^h, 22^h—1^h, 1^h—4^h and 4^h—7^h and inserted at the middle of those periods (marked **j**'). They all point to the concave side of the trajectory. Furthermore, the westernmost and the easternmost of the **j**' vectors have a component backward relative to the motion of the air particles, which is there retarded.

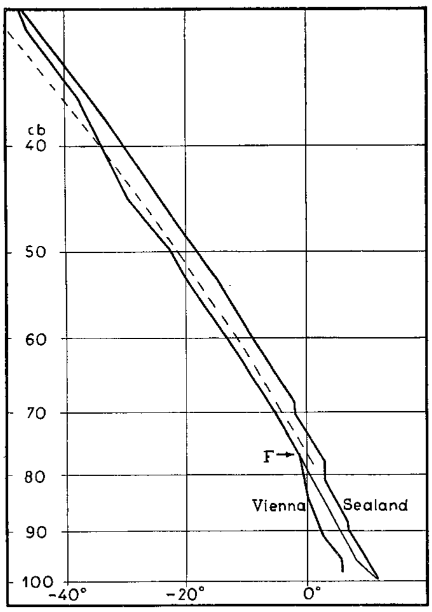


Fig. 35. A comparison of the tropical air at Sealand 15th of February 18^h and Vienna 16th of February 7^h.

The mere existance of these 1st approximation acceleration vectors is of course an indication that the particles would not keep exactly to the geostrophic trajectory. If we introduce the geostrophic wind

$$V_x = \frac{-s\dfrac{\partial p}{\partial y}}{2\,\omega \sin \varphi}, \quad V_y = \frac{s\dfrac{\partial p}{\partial x}}{2\,\omega \sin \varphi}$$

in the equations of frictionless motion

$$\frac{d\,v_x}{d\,t} = -s\frac{\partial p}{\partial x} + 2\,\omega \sin \varphi\, v_y$$

$$\frac{d\,v_y}{d\,t} = -s\frac{\partial p}{\partial y} - 2\,\omega \sin \varphi\, v_x\,,$$

we obtain the relationships

$$\frac{1}{2\,\omega \sin \varphi}\frac{d\,v_x}{d\,t} = v_y - V_y$$

$$-\frac{1}{2\,\omega \sin \varphi}\frac{d\,v_y}{d\,t} = v_x - V_x\,,$$

which tell that the vector difference between the real wind $(v_x,\ v_y)$ and the geostrophic wind $(V_x,\ V_y)$ is directed perpendicularly to the left of the acceleration vector and has a strength $\frac{1}{2\,\omega \sin \varphi}$ times the acceleration vector. In figure 34 the 1st approximation values of this vector difference $(v_x - V_x)$, $(v_y - V_y)$ have been calculated and inserted as obtained from the 1st approximation accelerations. Adding them to the geostrophic winds (marked **V**) we obtain the 1st approximation values of the real wind (marked **v**′). Since the trajectory is anticyclonic throughout, **v**′ is everywhere greater than **V**. Furthermore, **v**′ deviates to the right relative to **V** wherever the acceleration has a backward component. The greatest deviation (24°) is found near Vienna where the air retardation is particularly pronounced. This is just what table 13 shows. In a great part of the air column over Vienna the winds must have been almost 30° to the right of the geostrophic direction in order to produce the observed flight direction. Over Germany, where the 1st approximation acceleration has no component along the trajectory, no deviation in direction between **v**′ and **V** is found, but that may be due to small inaccuracies in the pressure field from which the geostrophic trajectory was derived. Over the North Sea there is again the same deviation of the wind to the right of the geostrophic direction as that found from the flight directions in Western Europe.

The trajectory of the air over Vienna, as resulting from the 1st approximation calculation of the real wind (fig. 34), would evidently lead back to points north of the geostrophic trajectory and actually pass about 200 km north of Sealand. With **v**′ all the time greater than **V** the 1st approximation trajectory is also longer than the geostrophic one and comes from Northern Ireland. One might attempt a further refinement of the result by a 2nd approximation calculation starting with a calculation of 2nd approximation accelerations from the data of the new trajectory, but the data under consideration is hardly worth it, and the result would undergo no essential change.

A further indirect sign that the trajectory of the Vienna air at 60 cb went well north of Sealand can be found from a comparison of the Vienna ascent of the 16th at 7^h and the Sealand ascent of the 15th at 18^h (fig. 35). The Vienna ascent reaches the tropical air at the point F, while the Sealand ascent goes through tropical air throughout. Air particles from near the ground at Sealand would arrive at F after a dry-adiabatic ascent of 350 dyn m and then a saturated adiabatic ascent to 2200 dyn m (as indicated by the thin line). The saturated adiabatic ascent of Sealand air particles higher up in the column would however not lead to as low temperatures as those of Vienna unless upward displacements of up to 4 dyn km are admitted. This is of course unreasonable when the air nearest to the cold wedge is lifted 2.2 dyn km only (compare fig. 32). The probable conclusion must therefore be that the Vienna air also owes its relatively low temperature to a more northern origin than that of the Sealand air, as shown above by the construction of the approximate trajectory.

At the 30 cb level the temperature difference between Sealand and Vienna is almost nil and above 30 cb the Vienna ascent is warmer than that of Sealand. The comparison between the two ascents in those high layers is however of less interest since the 13 hour trajectory of the Vienna air at 30 cb comes from far west of Sealand, where the temperature is unknown. In figure 34 the point in question (marked 30 cb) can be found out in the Atlantic, about 2800 km west of Vienna. Corresponding points have been inserted to show the length of the 13 hour trajectory at other levels. The lowest trajectory constructed (at 70 cb) is about half the length of that at 30 cb. This may serve as a rather obvious warning against the methods based on considering vertical air columns as travelling entities that remain vertical.[1] In our case the column of tropical air, that was vertical over Vienna on the 16th at 7^h, was inclined at about 0.5/100 to the horizontal the evening before.

Another warning should not be forgotten, when dealing with air mass history in a current the speed of which is rapidly increasing with height. If the air trajectory in such a current is not horizontal, the horizontal component of acceleration of the travelling particle will be:

$$\frac{d v_x}{d t} = v_s \frac{\partial v_x}{\partial s} + v_z \frac{\partial v_x}{\partial z}$$

$$\frac{d v_y}{d t} = v_s \frac{\partial v_y}{\partial s} + v_z \frac{\partial v_y}{\partial z}.$$

The first term to the right represents the horizontal acceleration component, computed under the assumption of a horizontal trajectory (v_s = horizontal component of velocity), and the second term to the right represents the additional horizontal acceleration component arising from the vertical displacement of the particle to layers with other horizontal speed than the layer of origin. Only the first term was considered at the construction of the trajectory in fig. 34, since the trajectory was supposed to be horizontal. If it were a rising trajectory, which under the given circumstances seems natural, the decrease of wind speed along the trajectory would have been

[1] See for example E. W. Hewson, Quart. Journ. Roy. Met. Soc. Vol. 63, No. 268, p. 8.

smaller than that computed in fig. 34. In order to get no decrease of horizontal wind velocity along the trajectory we would, however, have had to assume an ascent from the 60 cb nearly to the 20 cb surface on the way from Sealand to Vienna (see table 12), and that is evidently impossible. The accelerations (and the corresponding deviations of the real wind from the geostrophic) constructed in fig. 34 therefore certainly go in the right sense, although they may be a little too large.

5. 16th of February 18h (fig. 36).

The primary occluded depression slows down off the Norwegian coast while the secondary system, still in possession of a large warm sector, moves rapidly eastwards from the North Sea to Southern Scandinavia. The situation is complicated through the existence of two competing secondaries, the oldest one near E Scotland and the newest over Skagerak. Both secondaries are deepening rapidly and soon after amalgamate into one big storm centre. Unfortunately only rather scanty aerological data is available about this important phase of the development of the storm centre, but the large scale features of the upper air maps can nevertheless be considered as reliable.

A new important feature has come into the temperature distribution through the arrival of maritime polar air masses over the British Isles. The isotherms of the tropical air, which were oriented NW—SE behind the warm front, now turn to a SW—NE orientation ahead of the cold front that approaches from NW. The isobaric contour lines accordingly back with height over the western part and veer with height over the eastern part of the warm sector. The anticyclonic curvature of the warm sector current, already visible at the ground, thereby becomes more and more accentuated with height. Inside and up above the cold wedge over the British Isles the current is cyclonic.[1]

The strong concentration of the temperature and pressure gradients on the upper air maps over England is arrived at from a consideration of the shape of the Sealand isopleths (p. 27). The geostrophic wind over Sealand measured on the 40 cb map reaches the high value of about 130 m/sec. Since the trajectory of the air particles in the same region is definitely cyclonic (the air is overtaking the cyclonic trough), the gradient wind is much less, certainly well under 100 m/sec, but no exact value can be deduced with the insufficient synoptic data at hand. However, the maximum strength of upper air wind must be undoubtedly found over England, and east of that region of thronged isobars the current must spread and lose speed. The same conditions of retarded flow of air particles, as illustrated in fig. 34, will prevail east of Sealand, and the wind ought to have a slight component towards high pressure. Table 14 verifies that. Trappes, Uccle, Hamburg and Lindenberg have flight directions deviating to the right of the average geostrophic wind directions. (The deviation of the Hamburg flight is particularly great, but that is also rather much due to the long distance, 620 km, flown in a current of anticyclonic curvature). Ås, which is probably not in the region of diverging isobars, has

[1] The topography of the 40 cb surface shows a remarkable resemblance to the stream-lines of the cirrus motion outlined in Meteorologische Zeitschrift 1931, Heft 8, page 8.

Table 14. 16. II. 18h.

Flight direction (°), distance (km) and ceiling (cb), as compared with geostrophic wind direction (°) and speed (m/sec) at standard isobaric surfaces.

Station		Sealand	Trappes	Uccle	Hamburg	Lindenberg	Ås	Lauttakylä
Flight	Dir. dist.	262° 160	273° 61	270° 270	314° 620	310° 210	256° 121	243° 17
	pmin cb	8.6	47.2	14.9	?	?	13.6	49.3
Geostrophic wind	10 cb	240° 85		275° 28	275° 35	295° 32	270° 37	
	20	240° 120		270° 36	275° 50	310° 40	270° 50	
	40	235° 130	260° 40	265° 43	270° 43	315° 37	265° 48	275° 20
	60	255° 65	255° 36	260° 38	270° 39	310° 33	225° 20	270° 10
	80	270° 40	265° 34	265° 34	270° 35	295° 28	160° 18	- 0
	100	270° 32	265° 29	265° 33	275° 30	285° 24	135° 20	120° 5

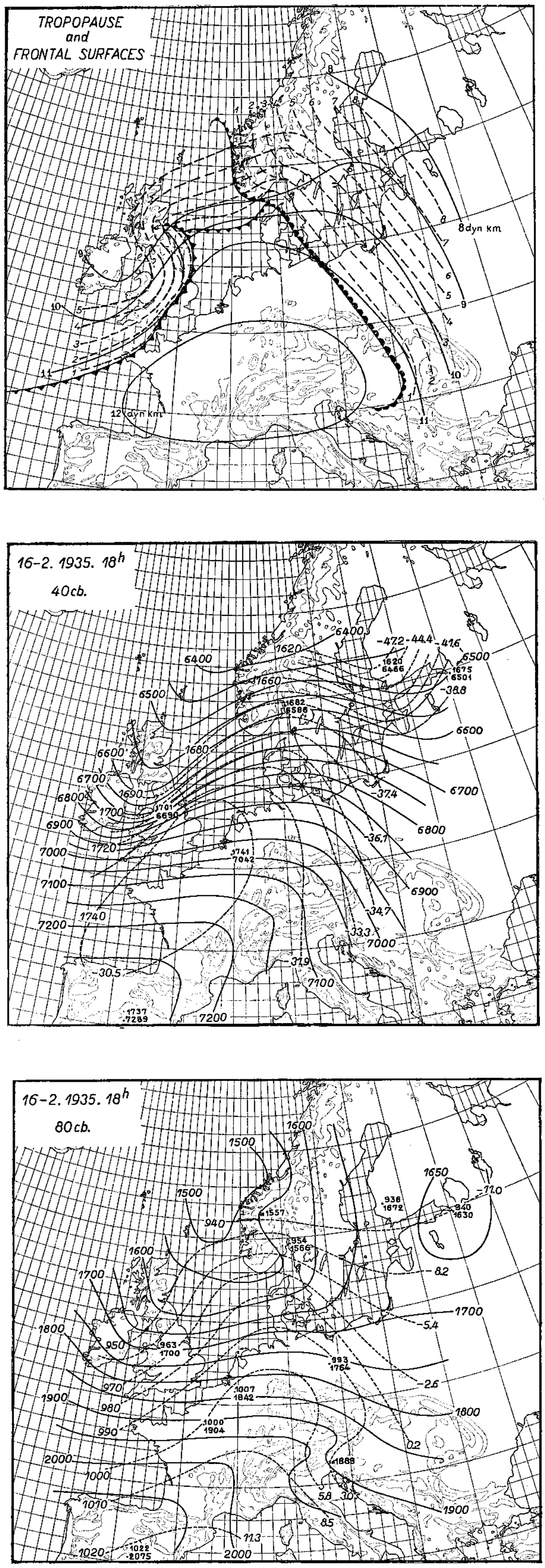


Fig. 36. Upper left: Topography (dyn. km) of the tropopause and frontal surfaces.
Other maps: Topography (dyn. m) of standard isobaric surfaces and distribution of specific volume (m³/ton) and virtual temperature (°C) at the standard isobaric surfaces.

a flight direction indicating a close approach of the upper wind to geostrophic conditions.

Figure 37 shows the approximate flow of the tropical air near the boundaries within which it is enclosed, that is the earth's surface in the warm sector, the warm front surface and the cold front surface. The stream-lines representing the motion of the warm sector air actually must emerge from the cold front surface and end against the warm front surface at finite angles. In fig. 37 the vector lines leading down the cold front slope and up the warm front slope are therefore not the stream-lines of the warm air wind itself but those of its component tangentially to the frontal surface under consideration. That component is, however, just as the wind itself, in the first approximation equal to the geostrophic wind measured at the level in question. This measurement has been carried out on the various maps of isobaric topography, by always selecting points just on the warm side of the frontal surfaces. The resulting streamlines of the tangential component illustrate the following facts:

From the north end of the warm sector (at the Skagerak secondary) the warm air climbs NE-wards along the steepest possible route up the warm front slope. Abundant precipitation results in that region (Ås 16 mm). Farther south the warm air climbs in an E or SE-ward direction and produces less rain intensity. If the warm air motion were constructed relatively to the moving warm front system, as in fig. 31, the stream-lines would have had less upslope component all over and practically none beyond an approximate 6 dyn km altitude.

The motion along the frontal slope, that extends from Southern Norway to Scotland, and in particular the conditions at the frontal bend over the secondary depression E of Scotland, remain unknown. Over England, however, where the frontal and isobaric topographies are approximately known, the described construction of the warm air component tangential to the frontal surface has been carried out. The resulting "tangential stream-lines" point obliquely downwards along the cold front slope. If the motion of the cold front system is subtracted (as will be later shown in fig. 41), the resulting relative flow still goes downwards on the upper part of the cold front surface, but upwards on the lower part of it. The air that rises just at the cold front passage, and produces the short lasting intensive cold front rain (see the Sealand isopleths), must come from the

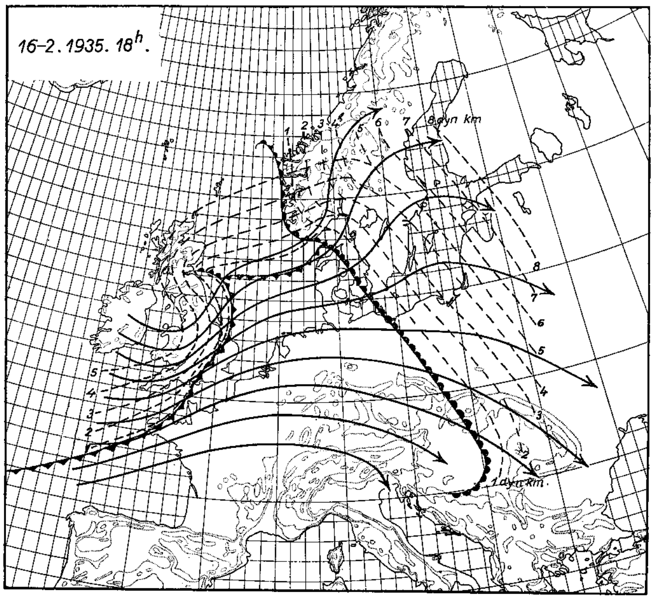


Fig. 37. Approximate flow of the tropical air in the warm sector above the frictional layer and at the slopes of the polar front surface.

lowest part of the warm sector current, that has a smaller eastward component of motion than that of the cold front (partly due to friction, but probably also due to under-normal wind above the frictional layer).

The surprisingly low tropical air temperature above the polar air tongue over the British Isles was already discussed in connection with the Sealand isopleths. The explanation of that temperature of course depends on the life history of the tropical air sample under consideration. The tracing of that life history is not yet possible because the air in question has only just arrived at the western border of the map. We will therefore return to the question when discussing the situation of the morning of the 17th.

6. 17th of February 7^h (fig. 38).

The three principal air masses are distributed as follows: 1) The coldest polar air, originally of arctic, more recently of northern continental life history, occupies the NE part of the map. 2) The maritime polar air occupies the NW part and is rapidly gaining ground over the region previously held by the tropical air. Where the two kinds of polar air meet, along the occluded front from the primary depression off the Norwegian coast, the maritime polar air is overrunning the continental polar air. 3) The tropical air is still present at the

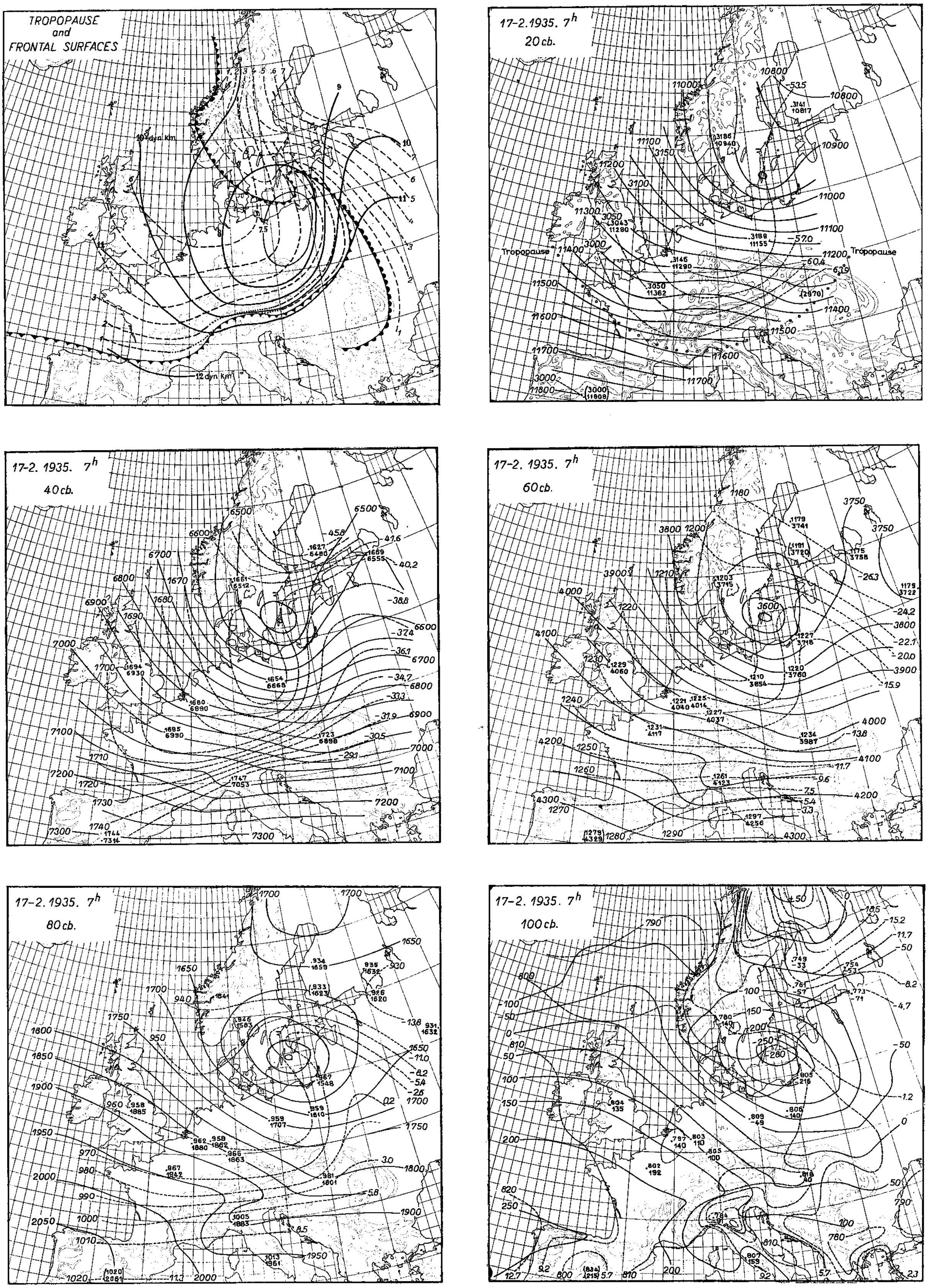


Fig. 38. Upper left. Topography (dyn. km) of the tropopause and frontal surfaces.
Other maps: Topography (dyn. m) of standard isobaric surfaces and distribution of specific volume (m³/ton) and virtual temperature (° C) at the standard isobaric surfaces.

ground over the Mediterranean States, Hungary and part of Poland (although concealed by surface inversions). Furthermore, it is to be found over the wedge of continental polar air as far north as Lauttakylä and Sloutzk, and all the way above the tongue of maritime polar air.

In the cross-section from Madrid to Moscow in figure 39 (constructed according to the methods set forth on p. 41) the arctic-continental type of polar air is seen to the right. It is surmounted by the tropical air, which is reached by the Sloutzk ascent above a well defined warm front surface at 6.8 dyn km. The maritime polar air, in the middle of the cross-section, is bordered by less distinct frontal surfaces for two obvious reasons: the maritime polar air is less cold than the continental type of polar air, and the adjacent tropical air is in that region colder than normal because of more northern "life history" (see p. 57). The front surface above the maritime polar air (marked F on the ascent curves) has been located at the place where the uniform tropical air lapse-rate commences. That transition is unmistakable at Trappes and Uccle (and likewise at Sealand which is not represented in the cross-section) but only just perceptible at the other ascents farther east (see fig. 12). The chosen position of the frontal surface receives some corroboration by the fact that the points marked F on the ascents arrange themselves along a "characteristic saturated adiabatic" as described on p. 14.

The various tropopauses in the multiple tropopause system have been identified by the methods explained on p. 15—22. The numbered arrows on the ascents indicate the number of the tropopause group in classification table 9, p. 21. All the groups from 1 to 6 are represented, since the cross-section passes from regions of extremely low to rather high tropopause positions. All the same, the tropospheric air mass underneath the tropopause is, according to life history, a kind of tropical air, even at the place of lowest tropopause. We must postpone our attempt of explaining this surprising state of affairs till we have also considered the synoptic maps for the upper levels.

The upper pressure maps (fig. 38), which are in this case based on a fairly good network of stations, show the usual strengthening of the westerlies with height but this time rather little change of their direction. The gradient for easterlies is reversed with height over Finland (which is also corroborated by the flight direction of the Lauttakylä balloon, which landed 45 km to the E by S), but nearer to the Baltic depression a narrow zone of weak easterlies seems to persist even up to the 20 cb level. Consequently, a closed circulation must also exist in the tropical air that occupies the upper troposphere above the occluding Baltic depression. That closed circulation in the current of tropical air must have developed from the marked cyclonic bend of the same current over England the evening before.

Figure 40, which has been constructed analogously to figure 37, depicts the stream-lines of the tropical air just above the cold air wedges. The tropical air curving round the N side of the centre climbs up the valley between the two wedges of polar air. Although in itself a weak current its velocity, relative to the NW current of maritime polar air advancing aloft over S Sweden, is certainly of the order 10 m/sec, and considerable snowfall is formed over that region. Probably, this is the general explanation of the extension of the precipitation area to the W and NW of occluding centres.

The tropical air that has been caught into the cyclonic whirl will probably continue to rise till it reaches the level of equilibrium with the surrounding polar air. It will then have got the same temperature as the polar air in the corresponding level, and will henceforth act as part of the polar air mass. This transformation of tropical air into polar air must take place in the upper central part of every large occluding depression, and the process in question is the necessary counterpart to the transformation of polar air into tropical air, so often observed in the low levels of subtropical highs. Just as synoptic meteorology at the ground presents the polar air in various states of degeneration and final transformation into tropical air, synoptic aerology will also present the tropical air in various transitional states on the way to its final transformation into polar air. Unfortunately, no ascent is available from the central part of the Baltic depression, that could have shown how far the transformation from tropical into polar air has proceeded.

The great enigma in the thermal structure of the cyclone comes in through the appearance of rather "cold tropical air" over the cold front surface. The phenomenon has been discussed at some length in connection with the isopleths of Sealand and Uccle, so that we shall here only add some words about the

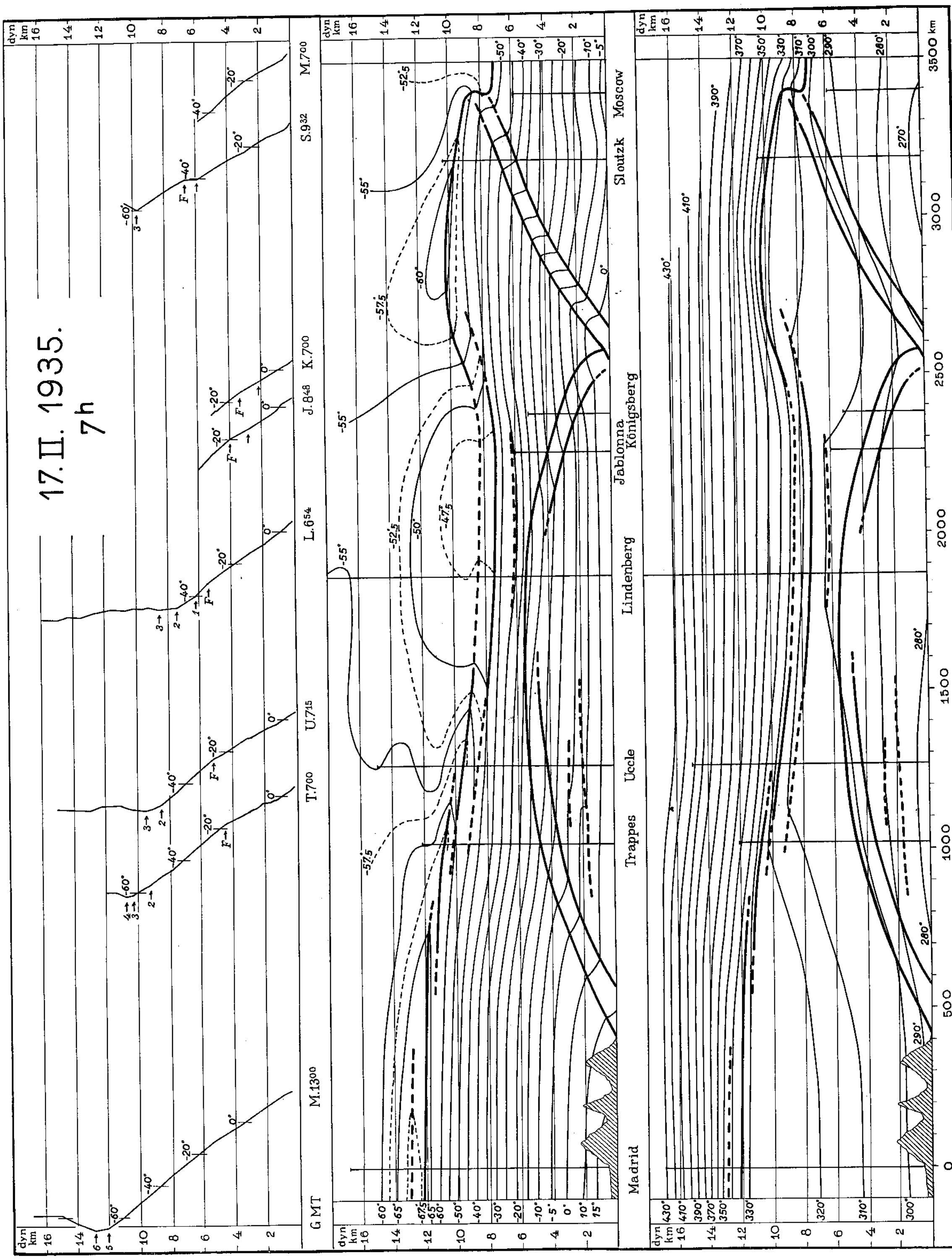


Fig. 39. Vertical cross-section Madrid—Moscow, morning of the 17th of February 1935. Top: The soundings used. Middle: Isotherms. Bottom: Isentropes.

probable life history of the mysterious "cold tropical air" in the light of the synoptic maps we are discussing. The stream-lines of the air above the tongue of maritime polar air, according to figure 40, originate from the NW corner of the map, and if trajectories are constructed they give a similar result. A particle of the "cold tropical air" at 40 cb over Lindenberg would, for instance, have been N of Scotland 12 hours earlier. Still earlier the particle would have belonged to the warm air supply of the family of depressions extending from Iceland to New Foundland. The air particle would be relatively warm for the latitudes just S of Iceland but, after climbing over the tongue of maritime polar air and arriving in lower latitude over Central Europe, it must be colder than the air particles arriving in Europe direct from the Azores region during the 15th and the 16th of February. To put it in brief, the "cold tropical air" seems to be derived from the warm sectors of the next family of depressions travelling farther north than the depression we are examining.[1] The most sudden drop of temperature from the genuine to the cold tropical air occurred at Sealand, where we were compelled to admit a drop of temperature amounting to about 10—15° C in 3 hours in the upper troposphere, that cannot have been reached by the wedge of polar air. The transition between the two kinds of tropical air is probably somewhat less abrupt in Uccle (10° C in 4 hours) and may eventually become quite diffuse.

In the morning of the 17th the 40 cb map (fig. 38) gives the best synoptic picture of the distribution of cold and warm tropical air. The 40 cb surface is namely above the polar front surface, except at Lauttakylä where it is a little below. Apart from this northern edge of the map the isotherms in the 40 cb level therefore depict the synoptic distribution of temperature in the tropical air. The coldest part of the tropical air over Germany (—42° at Lindenberg) is almost as cold as the polar air over Finland (—46° at Lauttakylä) and very much colder than the tropical air over southern Europe (—29° at Pavia) and the relatively warm tropical air E of the cyclone.

The movement of the various parts of the polar and tropical air mass, relative to the moving cyclone, can be studied on the maps fig. 41. We have used

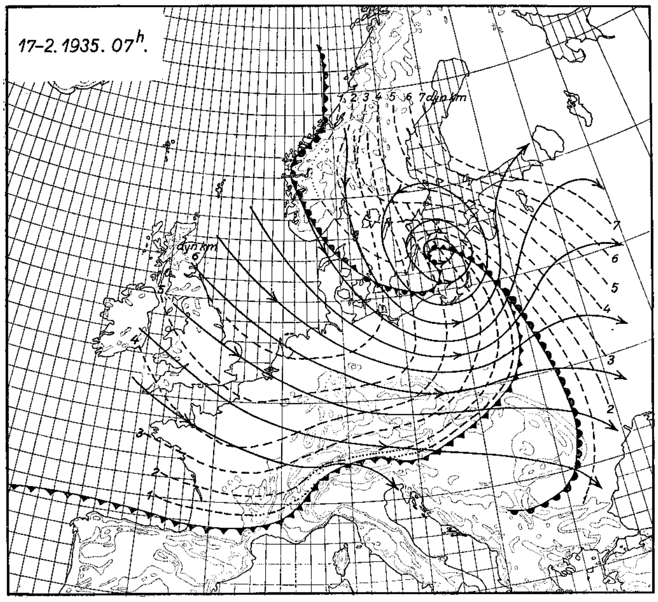


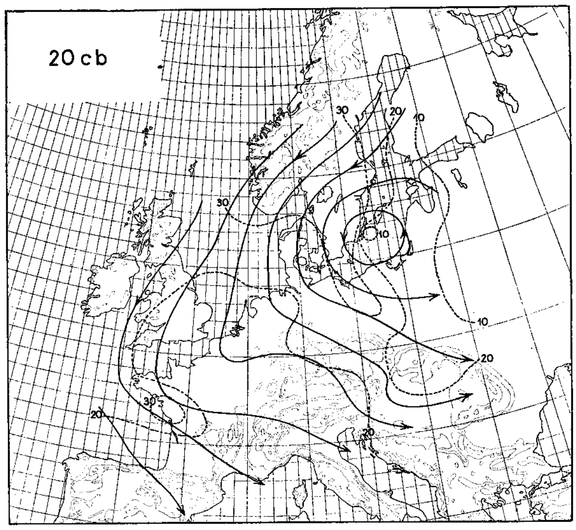
Fig. 40. Approximate flow of the tropical air in the warm sector above the frictional layer and at the slopes of the polar front surface.

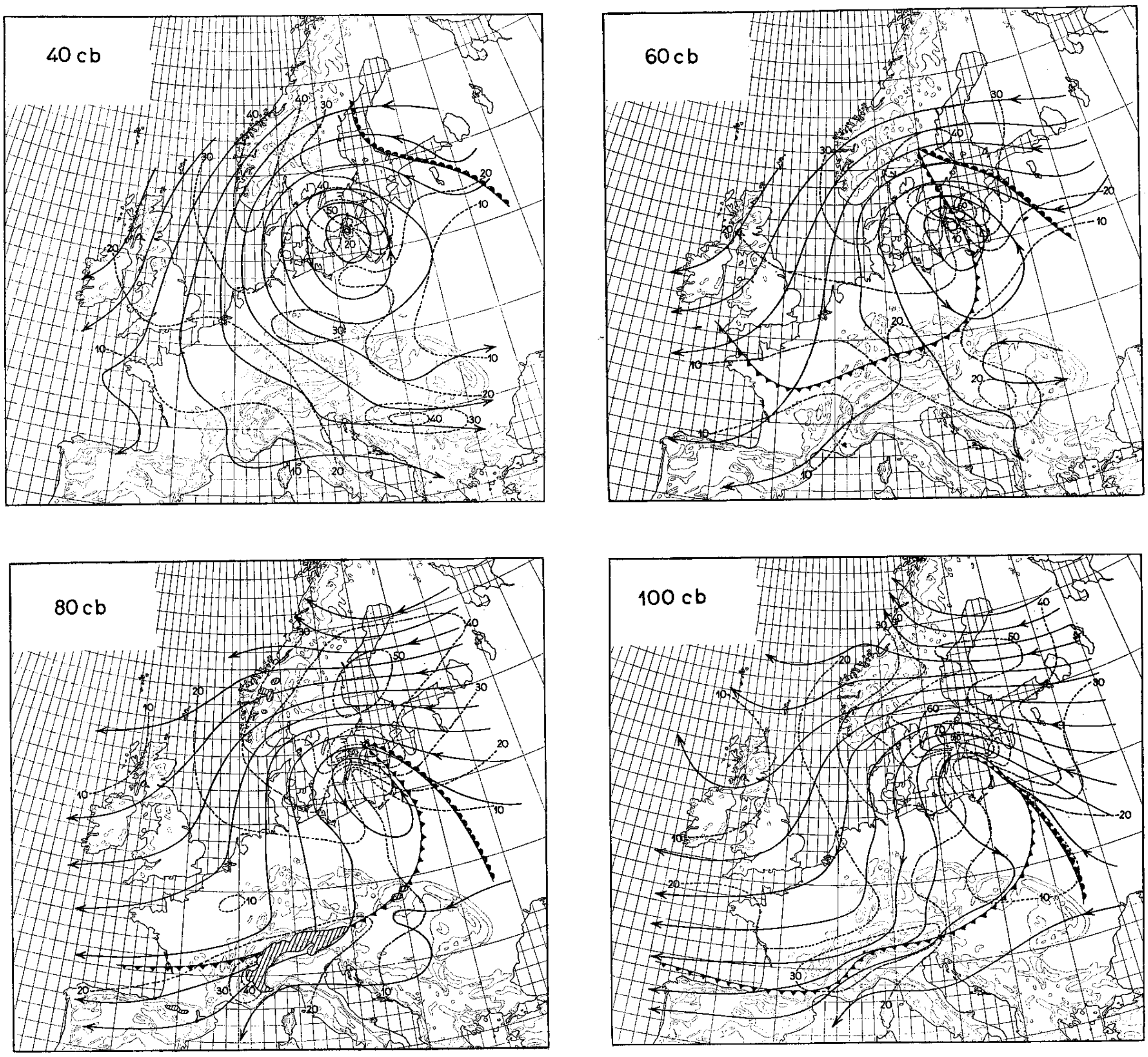
the field of the geostrophic wind as the substitute of real wind measurements, and have subtracted the speed of the perturbation from that geostrophic field of motion. As the speed of the perturbation we have taken the average speed of the cold front on its way from England to Poland, 120 km/hour, and reduced it to 100 km/hour (27.8 m/sec) since the cold front was definitely slowing down towards the end.

We consider first the map of relative stream-lines at the 100 cb level (which is actually close to or even below the ground, but nevertheless may serve as a good enough substitute for a map just above the layer of surface friction). The relative flow goes mostly westward and is of course particularly strong just north of the centre of the cyclone. Over a small region south of the Baltic cyclone the relative flow is directed from the W to NW but its velocity is moderate. The relative flow has a component near to zero normal to the cold front from the Baltic to the Carpathian Mountains, which is, of course, quite natural since we refer the relative flow just to that part of the cold front. The cold front part farther away from the cyclone drags behind, and the

[1] A doubtful question of terminology arises in this and other similar cases. The air behind the warm front, which is on the 17th at 7^h off the Hebrides, had acquired its warmth over the Atlantic in temperate and not in subtropical latitudes. Quite gradually, without any further front passage, the air current under consideration became during the 17th more genuinely tropical with a source region W of the Azores. We have preferred to use the term "cold tropical" for the relatively cold phase of the air current in question.

Fig. 41. Stream-lines of the geostrophic air motion relative to the moving cyclone in the morning of 17th of February, 1935. The intensity of the relative motion is inversely proportional to the mutual distances of stream-lines and is also represented by the stippled curves of equal intensity (m/sec).





constructed relative flow has therefore a component across it.

On all the higher maps the fronts are inserted along the lines where the frontal surfaces intersect the isobaric surface in question.

The 60 cb map is the first one to show a definite component forward across the cold front also in its rapidly moving part. Both the polar and the tropical air must consequently have a down-slope component of motion next to their mutual boundary, the cold front surface. This down-gliding will also take place well below the 60 cb level, say roughly down to 3 dyn km. That means that all cloud products along the cold front, that tower up beyond 3 dyn km, are carried forward, ahead of the front, as a shield of Alto-cloud.

At 40 cb the region of closed circulation is much larger. The air south of the centre, as far away as our map reaches, is overtaking the perturbation. Over Western Europe, however, the air which moved from NW relative to the earth moves from NE relative to the moving cyclone, and is therefore, at any rate for the moment, being left behind. The 40 cb surface is situated above the polar front surface, apart from a small region in the NE corner of the map, and it therefore presents the relative flow of the tropical air round the moving cyclone in the upper troposphere. The dominant feature of the relative field of flow is the advection of tropical air from high to low latitudes behind the centre, which is also the reason for the appearance of cold tropical air where there previously was warm tropical air. Since there is a closed relative circulation around the S side of the cyclone the cold tropical air will continue the encircling of the centre from the SW-quadrant to the SE- and eventually to the NE-quadrant. The final result is then the cold core, which is always found in old cyclones. From this investigation it would appear that the cold core of old cyclones does not altogether consist of polar air, but also, in the higher troposphere, of the described cold tropical air.

At 20 cb the W wind is weaker than at 40 cb, but still a large area south of the depression retains a component from the W relative to the moving perturbation. The southward advection behind the cyclone stands out just as clearly as on the 40 cb map. This southward advection, in addition to bringing cold tropical air, also brings a low tropopause and a high stratosphere temperature to Central Europe. The tropopause topography (fig. 38) and the 20 cb isotherms (fig. 38) verify that.

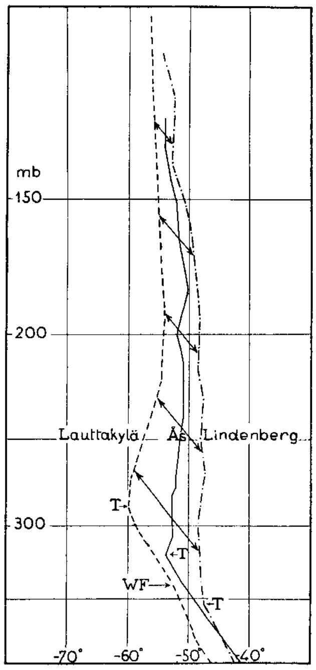


Fig. 42. Hypothetical vertical displacements of the stratosphere. 17th of February 1935.

However, the southward advection cannot alone explain why the tropopause came even lower down at Lindenberg (7.5 dyn km) than it ever did in the very complete series of ascents from Ås north of the cyclone track (lowest tropopause 8.0 dyn km). The closed tropopause depression over Lindenberg (fig. 38) must therefore be partly due to previous descending motion in the stratosphere and the upper troposphere. The tropopause has a remarkably steep slope, of about 1/100, on the SE side of the said tropopause depression, that is, over the lowest part of the cold front slope. Over the remaining warm sector the tropopause is still over 11 dyn km, which is practically as high as it used to be in the previous tropical air regime over Western Europe.

In fig. 42 an attempt is made to estimate the amount and extent of the downward motion, which led to the conditions observed over Lindenberg in the tropopause depression. The "initial state" chosen in fig. 42 is that of the Ås ascent on the 17th at 2^h, at the time of lowest tropopause, and the "final state" is that of the Lindenberg ascent during the passage of the tropopause depression at that place. There is of course no identity of trajectory of these two columns, because the Lindenberg air in the tropopause level did not come from Ås but from the region NW of Scotland (see fig. 41). However, as we have no ascent from that region, we have taken the Ås ascent as a substitute representing roughly what the tropical air near a cyclone at 60° N may look like. The descent needed to transform the Ås stratosphere into that of Lindenberg is nowhere greater than 0.5 dyn km and tends towards vanishing with increasing height. The Ås ascent took place outside the region of closed contour lines round the tropopause depression so that there is no need to assume that downward motion has also taken place there. It is, however, likely that the Ås stratosphere has also descended a little, and that the total descent of a stratosphere like that of Lindenberg may be 1 dyn km or even slightly more, counting from its preceding "normal state".

However uncertain these attempts of a quantitative estimate of the stratospheric descent, during the formation of the tropopause depression, may be, there can be no doubt about the very existence of that downward motion. The phenomenon is certainly connected with the spreading and sinking of the tongue of maritime polar air underneath and with the descending of the tropical air to the leeward of that polar air tongue.

The tropopause depression is not centrally situated, with respect to the cyclone centre, but keeps to the S or SW of it.[1] The reason must be that ascending motion is still going on in the region where the tropical air is lifted between the two wedges of polar air, in other words over the area of continuous precipitation to the N and E of the centre. No ascent is available just from that region, but also farther N the Lauttakylä ascent reveals signs of ascending motion in the tropopause region. The trajectory in that level goes back towards Ås, so that a comparison with the Ås ascent in fig. 42, as "initial state", is legitimate. As can be seen, the Lauttakylä tropopause (T) is on the same dry-adiabatic as that of Ås (and Lindenberg), but it has risen about 0.6 dyn km. In the stratosphere the upward displacement is damped out with increasing height.

[1] In the cyclone investigated by J. van Mieghem, Mém. Vol. III, Inst. Roy. Météor. de Belgique, 1937, the tropopause depression went down to 6 dyn km at stations more than 1000 km to the SW of the cyclone centre.

The ascending motion of the remaining upper warm sector must eventually come to an end. Then, the tropopause depression may perhaps place itself centrally over the cyclone centre, thus realizing, also in the high layers, the symmetrical circular depression, that is believed to be the last stage in the life history of cyclones.

The comparison of flight directions and geostrophic wind (table 15) does not give so clear results as the previous days, but the question is still worthy of some comment.

Considering first the flights south of the cyclone track, we find that the Sealand balloon deviated to the left of the geostrophic directions, the Trappes, Uccle and probably also the Budapest one deviated to the right, whereas the Lindenberg one had no appreciable deviation. The flight deviations are thus more diversified than before, when they were all oriented to the right of the isobars. This simple unanimity of deviations was due to the acceleration term $\boldsymbol{v}\cdot\nabla\boldsymbol{v}$ in the fan-wise spreading current, which covered the whole map and remained semi-permanent. On the 17th of February the kinematic conditions over Western and Central Europe are, of course, far from being permanent and the accelerations have their most general form

$$\frac{d\boldsymbol{v}}{dt} = \frac{\partial \boldsymbol{v}}{\partial t} + \boldsymbol{v}\cdot\nabla\boldsymbol{v}\,.$$

The term $\boldsymbol{v}\cdot\nabla\boldsymbol{v}$ has the opposite effect of what it had before, for the current is now converging and not spreading. If the kinematic conditions were still semi-permanent, flight directions over Western Europe would deviate to the left of the isobars. Only

Table 15. 17. II. 7^h.

Flight direction (°), distance (km) and ceiling (cb), as compared with geostrophic wind direction (°) and speed (m/sec) at standard isobaric surfaces.

Station		Sealand	Trappes	Uccle	Lindenberg	Budapest	Ås	Lauttakylä
Flight	Dir. dist.	290° 125	298° 120	303° 174	291° 209	285° 198	314° 84	292° 45
	pmin cb	11.7	17.7	11.1	3.3	23.3	10.8	8.0
Geostrophic wind	10 cb	305° 28	265° 25	265° 26	280° 52	-	310° 19	305° 20
	20	310° 28	280° 33	270° 32	285° 52	-	350° 25	315° 16
	40	320° 30	290° 33	290° 39	290° 52	280° 37	360° 27	225° 5?
	60	315° 24	285° 28	295° 33	295° 40	285° 35	10° 22	135° 10?
	80	300° 19	280° 19	295° 30	300° 36	275° 25	35° 23	115° 23
	100	290° 24	295° 14	300° 25	305° 36	270° 20	45° 27	115° 25

the Sealand flight does that, whereas the Trappes and Uccle flights definitely deviate to the right of the isobars. At the latter two stations the term $\boldsymbol{v} \cdot \nabla \boldsymbol{v}$ must evidently have been overcompensated by the term $\frac{\partial \boldsymbol{v}}{\partial t}$. This was also to be expected from the maps of geostrophic flow relative to the moving perturbation (fig. 41), which show that the air over Western Europe does not enter the region of maximum W-wind ahead of it, but is left behind and comes into a region of weaker W-wind.

The zero deviation of the Lindenberg flight presents no difficult problems. Lindenberg is in the middle of the region of maximum wind force, where $\boldsymbol{v} \cdot \nabla \boldsymbol{v}$ is about zero and $\frac{\partial \boldsymbol{v}}{\partial t}$ is also likely to be small.

North of the cyclone track, Ås presents a case of a considerable deviation to the left of isobars, in other words a case of a strong acceleration component parallel to the wind. According to the upper air maps Ås is in a field of converging isobars, which is evidently part of the explanation, but $\frac{\partial \boldsymbol{v}}{\partial t}$ probably also acts in the same sense to produce a strong acceleration component along with the wind.

The wind conditions over Lauttakylä are so complicated that nothing can be concluded about the acceleration.

It will be understood that the dynamical analysis, which has been attempted here with data from table 15, ought to have been based on a more complete system of observations. For a reliable determination of the acceleration in non-permanent situations upper air maps should be available at intervals of 6 hours, preferably even shorter.

7. Isallobaric maps.

The set of maps, fig. 43 and 44, summarize the variations of pressure in space and time during the investigated period. Each map represents the 12-hour change of level of an isobaric surface, obtained graphically as being the difference between the topographic maps of that isobaric surface at the beginning and the end of the 12-hour period. The maps will be referred to as "isallobaric maps", although the change of pressure is not expressed in pressure units but in change of geopotential of selected isobaric surfaces.

During the first 12-hour period (morning to evening of the 15th) the 100 cb surface has a maximum of sinking over the E-coast of England, obviously connected with the advance of the warm front surface. The zone of greatest negative values extends NW—SE, parallel to the warm front and the adjoining occluded front. The same zone of negative values extends from E-Iceland towards the Balkans during the following 12-hour period, but it has then developed two separate isallobaric minima at the Shetlands and over Bohemia respectively. The latter one can be identified with the isallobaric minimum over E-England in the previous interval.

The 20 cb surface, during the same two 12-hour intervals, shows a strong isallobaric high travelling from France to the Adriatic Sea, with the same speed as that of the warm front. The warm front over W- and Central Europe is thus connected with an isallobaric high in the upper layers, situated farther S than the isallobaric low in the lower layers. Over S. France and Italy the influence of this upper isallobaric high is strong enough to make the usual pre-warm-front fall of pressure disappear. Farther N the upper isallobaric high is weaker, and the normal pre-frontal fall of pressure can assert itself underneath.

The 60 cb surface, as a matter of course, must show intermediate conditions between those described for the 100 cb and the 20 cb surfaces, which means in this case a rather "flat distribution" of pressure changes. Particularly striking is the almost unaltered height of the 60 cb surface during the second 12-hour period, while simultaneously the 100 cb surface is sinking and the 20 cb surface is rising over most of the area of the map.

The secondary depression, that formed W of Scotland and which eventually became the deepest depression of the map, dominates the features of the isallobaric maps for the third and fourth 12-hour periods (fig. 44). The centre of its isallobaric low, as determined at the 100 cb surface, is in the third period over the Skagerak, and in the fourth period moves from there to the Baltic. The following isallobaric high arrives over the British Isles in the fourth period. The same isallobaric systems can be easily identified higher up, but their positions and shapes change. For instance in the fourth period, which shows the most clear cut conditions, the isallobaric low extends SW-wards with increasing height, while the isallobaric high is displaced N-wards with increasing height. The zero-line between the two, at the 100 cb surface, runs approximately N—S from Scandinavia to the Mediterranian, but NE—SW from Scandinavia to SW England at the

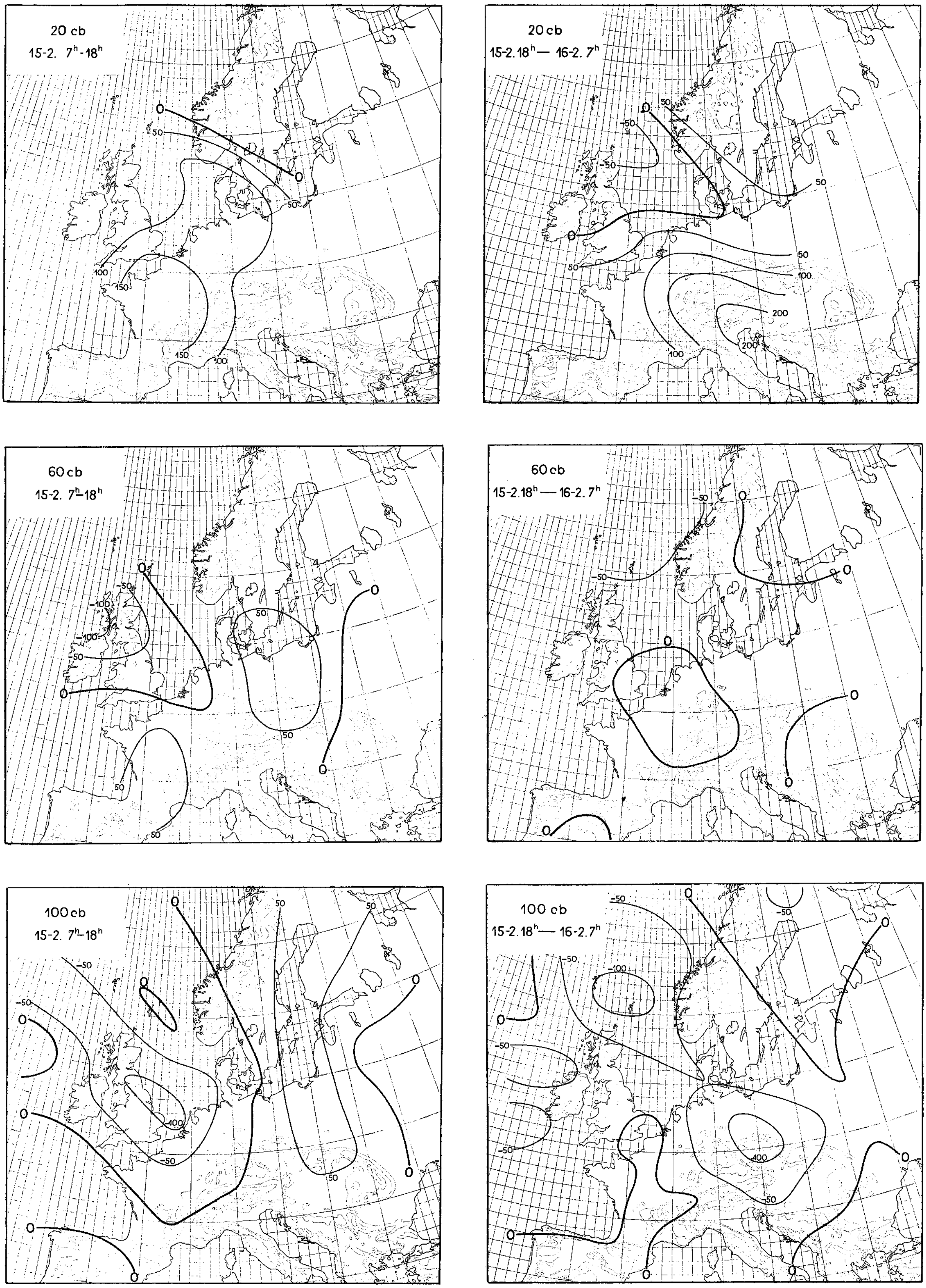


Fig. 43. 12-hourly "isallobars" represented by the 12-hourly change of level (dyn m) of given isobaric surfaces.

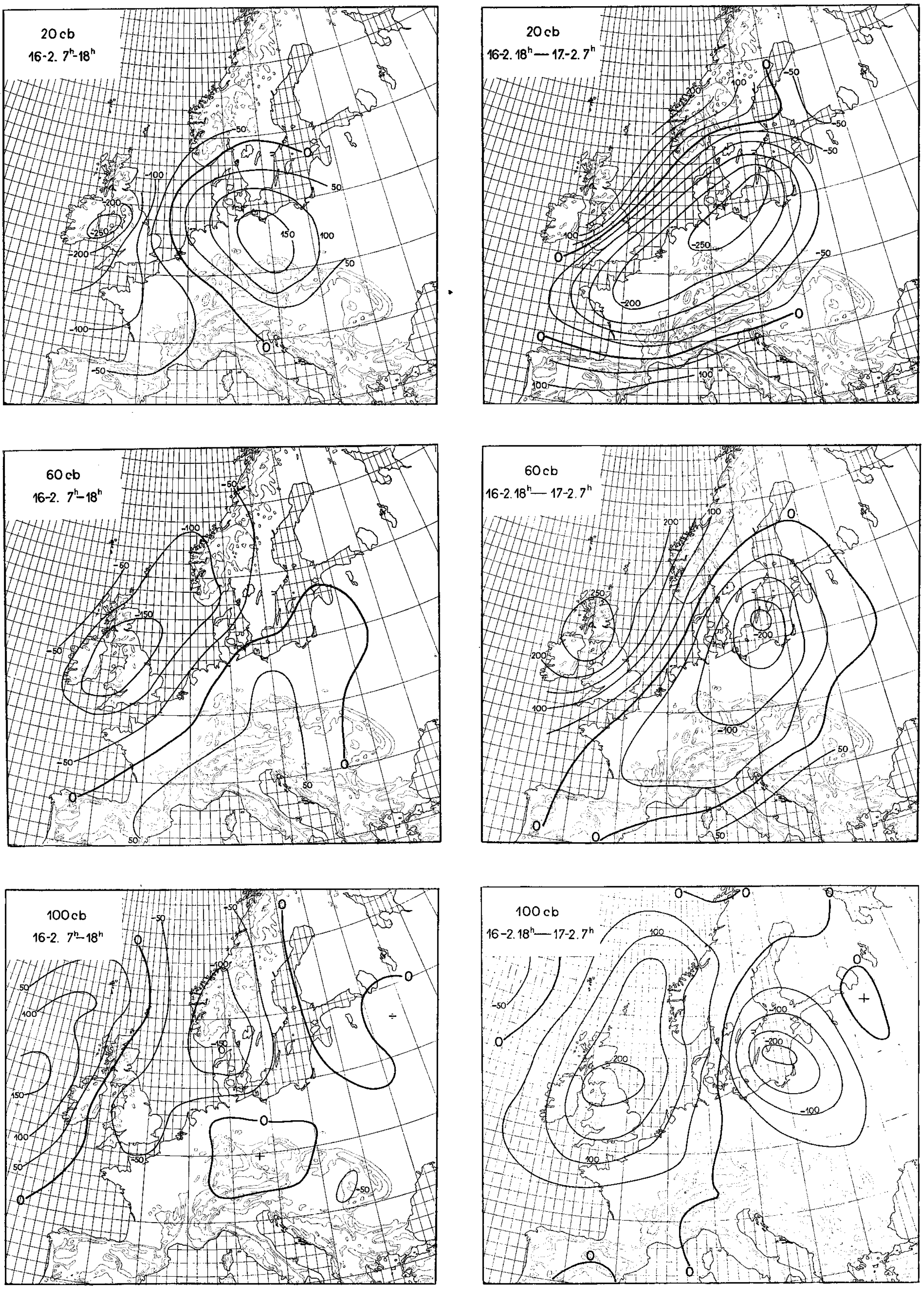


Fig. 44. 12-hourly "isallobars" represented by the 12-hourly change of level (dyn m) of given isobaric surfaces.

20 cb surface. This is but another demonstration of the fact, repeatedly mentioned, that in the barograms to the S of the cyclone track there is an upper air pressure minimum over the cold front slope (the lee-whirl in the tropical current), whereas to the N of the track that upper air pressure minimum approaches to coincidence with the pressure minimum in lower layers and rapidly decreases in intensity.

We purposely abstain from a theoretical discussion of the three-dimensional field of pressure variations, represented in the maps of fig. 43 and 44. It seems advisable to wait for the results of some further synoptic-aerological investigations before attempting to give theoretical interpretations.[1]

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[1] Some guiding principles in that line can be found in J. Bjerknes: "The Upper Perturbations", Mém. de l'Ass. de Météorologie de l'U. G. G. I. Edimbourg 1936, and in: "Theorie der aussertropischen Zyklonenbildung", Met. Zeitschrift, Dec. 1937.

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