thwart or eliminate the CM cloud formations described above.

14·3. The front models

Fronts are, by definition (see 6·34 and 14·01), the lines of intersection of a surface of discontinuity, separating two air-masses, with another surface—in practice, usually the surface of the earth. The fundamental dynamics of fronts is represented by the formula of Margules (11·21). In this section, descriptive features, especially pertaining to clouds and precipitation, will be added.

In addition to the *absolute and geographical* classification of the fronts given in 13·3, other "relative" classifications are also useful.

According to the *first relative classification*, introduced by T. Bergeron (1934, 1936), a front is called an *anafront* if, along the corresponding surface of separation, the warm air slides upward relatively to the cold air. It is called a *katafront* if the relative vertical motion of the two air-masses has the opposite direction. Diverse types of kinematically possible anafronts and katafronts have been given in 11·23. No type involving a general descent and inherent divergence in the lower layers, however, can exist in nature. The divergence would produce anticyclonic vorticity (cf. 8·82·2), which is incompatible with the existence of the trough of low pressure along the front. Further, since anafronts have a tendency to become sharpened, whereas katafronts are subject to frontolysis (see 11·23), we must expect the anafronts to dominate. In fact, of the four simple front models to be described below, three are anafronts, whereas the fourth has katafront character in the upper layers only.

According to the *second relative classification*, the fronts are called *active* when the warm air ascends along the front surface and *inactive* when it descends (11·23). Since active fronts are associated with pronounced cloudiness and precipitation, and inactive fronts with broken Cirrus and Altostratus, this classification is important in practice.

The *third relative classification* leads to the introduction of *quasi-stationary, warm,* and *cold fronts.* The "upglide cloud system" characteristic of a simple front appears in a typical form at the quasi-stationary front (14·30). If this front begins to move toward the cold side, so that a warm front (14·31) is produced, the cloud system remains essentially the same. This is also the case if the front moves as a cold front slowly toward the warm side, whereas the structure of the cloud system is largely modified if this motion is rapid. Thus, it becomes practical to introduce two different cold-front models (14·32), namely the slow-moving and the fast-running cold fronts.

A cold front often overtakes a warm front, and the combined front system which results is represented by the so-called *occlusion models* (14·33). They differ somewhat, depending on whether the coldest air is found ahead of the warm front or behind the cold front.

The front types presented in the following subsections can refer only to average conditions. The inclination of a frontal surface, the vertical and horizontal extent of the corresponding cloud mass, and other factors will, of course, show considerable variation from one individual case to another. The position of the 0°C isotherm and the ice nuclei level, and the height limits of different hydrometeors, in the vertical sections of figures 14·30·1 to 14·33·2 refer to average conditions during autumn and spring at about 50°N on the west coast of Europe or America.

14·30. The quasi-stationary front model and the upglide cloud system.

The quasi-stationary front and its cloud system have been represented in figure 1 by a vertical cross section and a horizontal "map," covering a narrow strip normal to the front. The map strip shows, in addition to the clouds and precipitation, the frontal wind discontinuity with cyclonic vorticity (see 11·21) and the frontal wind convergence caused by friction in the bottom layer (see 12·30). We have supposed west winds to exist in both air-masses and the front line to run west-east with the colder air to the north of it.

The vertical cross section represents anafront conditions. In the warm air an appreciable upglide motion exists, whereas the vertical motion in the cold air is negligible except just below the front surface, where upglide motion and cloud formation prevail in a "zone of transition." Thus the streamlines in the cold air are

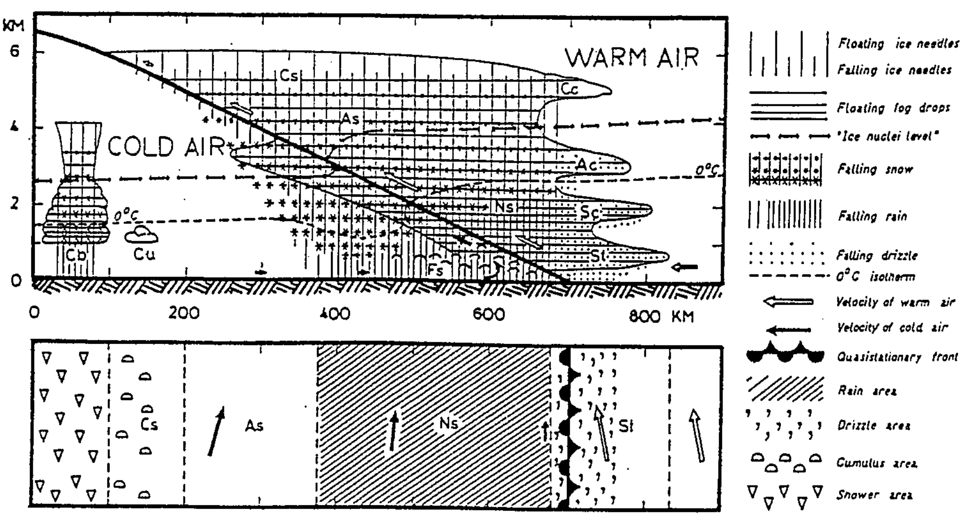

FIG. 14·30·1. The quasi-stationary front represented by a vertical profile and a strip of the surface map.

quasi-horizontal and almost parallel to the front. The ascending motion in the warm air is assumed to reach a level of about 6 km, and the warm air is supposed to be stably stratified for saturated conditions also—an assumption which is not always borne out in nature. If conditional lability exists, Cumulonimbus-like towers will develop above the frontal surface, and the precipitation will be of fluctuating intensity. The multilayered structure of the upglide cloud, which is frequently observed, and which is also indicated in figure 1, must be due to an initial layering of humidity. This stratification in turn would be a product of the vertical mixing which creates a maximum of relative humidity just below each inversion or layer of minimum lapse rate. Where the vertical displacements have become large enough, these separate layers fuse into a compact cloud mass, which extends from the front surface to about 6 km, and within which release or ordinary precipitation is possible.

Up to a level of, say, 4 km the condensation produces supercooled drops of diameter less than 0.05 mm. Above this level, indicated as "ice nuclei level" in our model, and supposed to coincide roughly with the −10°C isotherm, we assume ice nuclei to exist, or to form, so that part of the water passes directly from the gaseous to the solid phase. Above, say, 5 km this latter process dominates, so that the uppermost part of the cloud consists mainly of various types of simple ice crystals (e.g. plates, needles, etc.) and appears as Cirrostratus (*Cs* in the figure). At first the ice crystals are very small and are carried by the ascending current (Cirrostratus nebulosus). Lower down, the largest of them grow by sublimation of water vapor (see 1·08) and acquire real fall velocities, a fact which probably explains the fibrous structure of the Cirrostratus filosus. The fate of the falling ice crystals depends largely on the character of the air-mass through which they fall. Ice particles falling directly through the front surface into the relatively dry cold air very soon evaporate, and do not reach the ground. Crystals falling through the subjacent supercooled water clouds make the latter colloidally unstable (1·08) and grow rapidly by diffusive transport of water vapor from the fluid to the solid phase. Thus, the cloud mass at this level also contains a large amount of falling snowflakes and soft hail, which, sooner or later, fall through the frontal surface. If this surface is traversed at a height greater than, say, 3 km, the precipitation normally evaporates before it reaches the ground. The frontal cloud above 3 km therefore appears as an ordinary Altostratus. Where the height of the frontal surface is less than 3 km, the cloud is, under typical winter conditions when no effective melting of the crystals occurs, an "Altostratus precipitans." During the rest of the year the snow crystals melt in the lower layers and reach the ground as rain, the frontal cloud system in this case being of the Nimbostratus type (*Ns*). In our diagram the latter and more frequent case has been illustrated. We have assumed the melting of the snow to be accomplished at the 2°C isotherm. The assumed lowering of the isotherms in the lower central part of the precipitation area is supposed to be a sign of the cooling effect of the falling snow. Within the whole precipitation area, the air is being moistened, so that eddy diffusion may cause the formation of a low layer of Fractostratus, *Fs*, or even a compact Stratus.

The diagram illustrates the case in which the colder air is of the CM type and the warmer of the WM type, which is most frequent in nature. An observer traveling from the CM to the WM will then see this sequence of clouds: About 600 km from the front he will see the vertical extent of the CM clouds decrease and simultaneously the first part of the upglide cloud system appear at a level of about 6 km. When the Cirrostratus has transformed into Altostratus (500 to 300 km from the front), convection decreases and the Cumuli tend to disappear. Next the Altostratus will be observed gradually to transform into Nimbostratus with rain. Finally nothing but the typical low WM cloud, the Stratus (*St*), is observed, which produces only drizzle or light rain, since the interaction between the solid and the supercooled fluid phase of water is missing.

14·31. The warm-front model. The cloud system given in figure 14·30·1 is, in a way, typical for all anafronts. Some modifications of structure will, however, occur when the fronts are moving. We shall first consider the warm front, whose displacement is directed toward the colder air (fig. 1). In order to facilitate the later applications of the figure, we have assumed the front surface to slope downward from right to left. The distribution of the velocity has been referred to a system of co-ordinates attached to the moving front.

The horizontal map strip across the front is, on the whole, similar to that for the quasi-stationary front (fig. 14·30·1). It shows a cyclonic vorticity of the surface wind at the front corresponding to a trough in the pressure field (11·21), a frontal convergence—which is caused more by deviation from geostrophic flow (see 15·10) than by friction—and a divergence in the cold air, at 400 to 600 km ahead of the front. This divergence will, in nature, usually take place in the region of the high-pressure wedge preceding the front (cf. 15·12). The vertical cross section in figure 1 shows that the warm front has the same anafront character as the quasi-stationary front (fig. 14·30·1). Owing to frictional influence, however, the inclination of the warm-front surface will decrease toward the ground (cf. 10·64·2°). An upglide motion is assumed to take place in the warm air from 0 to 6 km.

If the warm air is again assumed to be a WM, stable also for saturated air, whereas the cold air is a CM, the cloud system will be practically identical with that

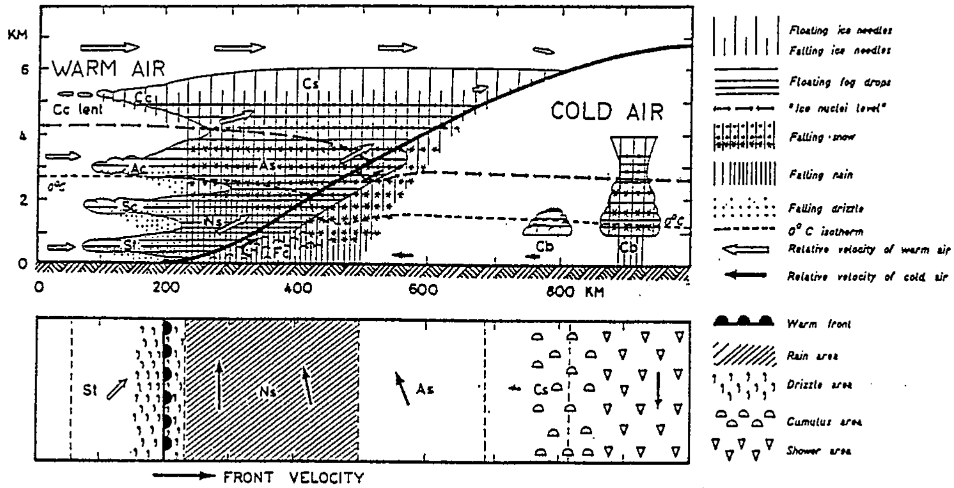

FIG. 14·31·1. The warm front represented by a vertical profile and a strip of the surface map. The wind arrows represent the air motion in a co-ordinate system attached to the front.

discussed in 14·30. A locally fixed observer will see the system pass by as a characteristic upglide cloud system, and on the whole will note the same distribution of the meteorological elements as an observer moving through a quasi-stationary front from the cold toward the warm side. Before any signs of the frontal clouds themselves are visible and while CM clouds may still prevail, fine Cirrus uncinus may be observed well above the warm-front surface (above the top of the diagram in fig. 1) drifting approximately parallel to the front with the cold air to the left. The individual elements of these clouds are constantly dissolving, but they are replenished from the horizon so that the cloud amount continues to increase. About 600 km ahead of the approaching front the CM clouds begin to shrink and the real frontal clouds appear, Cirrostratus, Altostratus, Nimbostratus with Fractocumulus, and near the front Stratus from which drizzle may fall. If the velocity of the front represented in figure 1 has an average value of 50 km h^{-1}, the observer will be under the Cirrostratus for about 2 hours and under the Altostratus and the Nimbostratus for 4 hours each before the front passes. The dimensions given in the models (figs. 14·30·1 and 1) for the cloud system are of course mean values. The inclination of the front, here assumed to be 1/100, is, in fact, somewhat variable. Furthermore, depending on the varying humidity of the prefrontal cold air, the precipitation from the Altostratus may fall for a longer or shorter time through the air of the cold wedge before it evaporates. Consequently values up to 50 per cent larger or smaller than those given in the models may often be observed in nature. Since, moreover, the velocity of the front may have any value from close to 0 km h^{-1} up to 100 km h^{-1}, the local duration times of the cloud regimes may vary within wide limits.

14·32. The cold-front models. A cold front moves, by definition, toward the warm side. As was stated in 10·64·2°, frictional influences will in the lowest 2 or 3 km make the profile of the cold front steeper than that of the stationary front, and will also affect the distribution of the ascending and descending motion of the air around it. The latter influence being somewhat different for slow-moving and fast-running fronts, it has proved useful to introduce two different cold-front models.

The slow-moving cold front, or cold front of the first type, has been represented in figure 1. As in the case of the warm front, the motion has been referred to a system of co-ordinates attached to the front. The slow-moving cold front is on the whole similar to the warm front as far as the cloud system is concerned. The rear edge of the cold-front cloud system is established at the level of transition from anafront to katafront conditions (5 km in the model). The front edge of the cloud system acquires a vertical build-up because in the upper part of the cloud system the air moves forward relatively to the cold front at the ground.

The composition of the cloud as regards ice and water is, in principle, the same as in the previous models. The postfrontal cold air usually has a descending motion like that assumed in figure 1. Therefore, a short period of clear skies is frequently observed just behind the front before the typical CM clouds begin to appear.

An observer at a fixed locality will see the slowly moving cold front advance as a cloud wall, partly of Cumulonimbus type, provided internal WM clouds in the warm air do not hide the view. At the front the wind generally veers rapidly and drops in strength. Rain falls in a zone of maximum width 300 km behind the front. Next the Altostratus becomes thinner and sometimes changes into Cirrostratus, followed by a short spell of fair weather preceding the normal CM showers. The evolution of the weather is thus similar to that found by an observer passing through a stationary front from the warm into the cold air.

In the cold-front model of the second, fast-running type, represented in figure 2, the warm air above 3 km

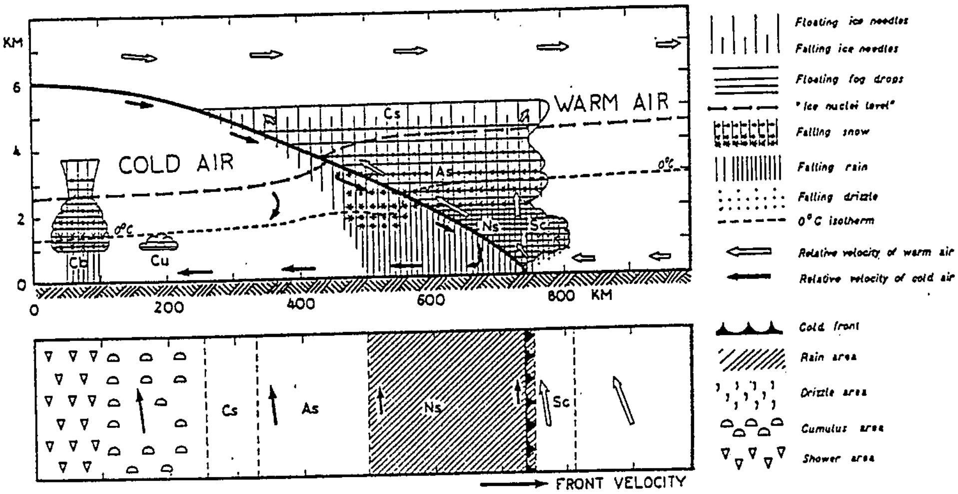

FIG. 14·32·1. The slow-moving cold front represented by a vertical profile and a strip of the surface map. The wind arrows represent the air motion in a co-ordinate system attached to the front.

moves forward relatively to the front itself. Correspondingly the character of the front is changed at the height of 3 km from that of an anafront to that of a katafront. Simultaneously the front at the upper levels is subjected to a frontolysis (indicated in our diagram by a broken front line) caused by the downward motion. The cloud system differs somewhat, depending on whether the warm air is stable, as assumed in figure 2, or unstable for saturated air. In the latter case Cumulonimbus towers develop on top of the frontal cloud system, and the precipitation is of a showery type. The great velocity of the upper warm air causes the top of the cloud system to extend rather far ahead of the front, where it breaks up into gradually thinner Altocumulus layers. Altocumulus lenticularis is sometimes seen to form part of the prefrontal cloud shield. The Altocumulus is too thin to produce precipitation, but where it changes into solid cloud masses extending beyond the ice nuclei level, precipitation is formed and may reach the ground. The area of precipitation has its rear limit fairly near the front and is thus mainly prefrontal. The cold air has a sinking motion which is especially pronounced in the "head" of the cold wedge, where the air so to speak, "rolls" forward. The general descent of the cold air as a rule produces a zone of fair weather behind the fast-running cold front. The approach of a front of the second type toward an observer situated at a fixed locality may at first be indicated by banks of Altocumulus lenticularis; each bank is dissolving individually, but more of them pull up from the horizon and are followed by a thick cover of Altostratus character. The rain begins up to 100 km ahead of the front and ceases at, or shortly after, the front passage. The remaining Altostratus is gradually transformed into Altocumulus opacus, and the sky clears quite suddenly at a short distance behind the front. The fair weather lasts until the ordinary CM convection begins.

The principal difference in precipitation as between the two cold-front types lies in the width of the rain area. As shown by the map strips in figures 1 and 2, the front of the first type has a broad postfrontal rain area, the front of the second type a narrow and mainly prefrontal one.

To make the cold-front models fit low-latitude condi-

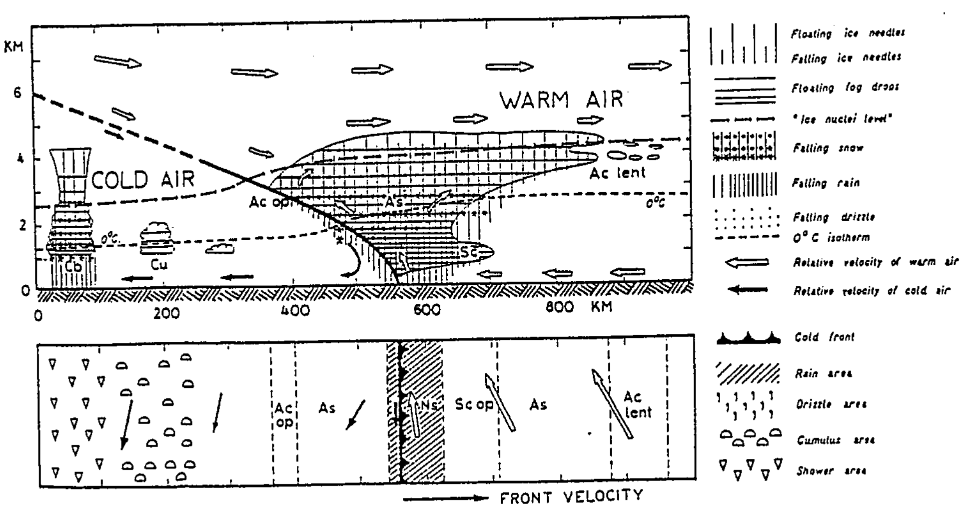

FIG. 14·32·2. The fast-running cold front represented by a vertical profile and a strip of the surface map. The wind arrows represent the air motion in a co-ordinate system attached to the front.

tions, with moist, warm air supplies, Cumulonimbus and showers should be inserted over a rather wide prefrontal zone (see 14·20·6°).

14·33. The occlusion models. The simplest composite fronts are formed when a cold front overtakes a warm front while lifting the warm air, which initially separated them, off the ground. This process is called "occlusion" (for the role of occlusions in the life cycle of frontal cyclones, see 15·00). If the prefrontal cold air at the warm front has the same properties as the postfrontal air at the cold front, no discontinuity will remain at the earth's surface. This case of *neutral occlusion* is, of course, seldom encountered in nature, since the two cold air-masses practically always differ in their physical properties. Therefore, a surface front will generally continue to exist, even after the two original fronts have joined. Depending on whether the coldest air is found ahead of the warm front or behind the cold front, two different cases have to be distinguished.

When the coldest air lies ahead of the original warm front, the system of frontal surfaces is called a *warm-front occlusion;* this system has been represented in figure 1 in co-ordinates attached to the surface front. The horizontal velocity distribution shows cyclonic vorticity concentrated both at the surface front and below the "triple point" of the vertical cross section. The slope of the surface separating the two cold wedges will, according to the formula of Margules (11·21), be steeper than that of the upper warm-front surface. In fact, at the former surface the temperature discontinuity is small but the wind discontinuity large, being the sum of the discontinuities at the original warm and cold fronts.

The cloud systems and rain areas are, at least in the first stages of the occlusion, composed of those described for the warm front and the following cold front, and are mostly of the fast-running and partly frontolyzed type as assumed in figure 14·32·2. As the cold front advances relatively to the warm one, the upglide cloud system at the latter is continually lifted and displaced forward relatively to the surface front. Therefore the rear limit of the corresponding rain area is transferred to a position clearly within the wedge of coldest air. Later a new Stratus system may be formed at the secondary frontal surface which separates the two masses of cold air. In the first stages of the occlusion, however, this cloud system cannot attain any considerable thickness, because the air in the wedge of the upper cold front had been sinking before it started climbing the warm-front slope. The newly formed Stratus system will consist only of minute cloud droplets and will produce drizzle or light rain except in the case when ordinary precipitation falls through it from above. As the occlusion grows colder, the triple point in the vertical section moves upward along the warm-front surface. Therefore the streamlines of the air motion, referred to the surface warm front, must intersect the cold-front surface as indicated in the vertical profile in figure 1.

An observer watching a young warm-front occlusion passing by will see a combination of a warm-front and a cold-front cloud system. If the occlusion is not too young, the rain will be of relatively short duration and will cease soon after the "triple point" has passed. Near this point there may be a somewhat showery type of precipitation produced by the upper cold-front system. When the occlusion is growing older—i.e. when the triple point is advancing to great heights and the secondary warm front is becoming feeble—the precipitation from the cloud system may become insignificant or vanish altogether. The passage of the triple point will sometimes, and that of the surface front will always, be indicated by a veering of the surface wind.

When the coldest air occurs in the rear of the initial cold front, the composite system of frontal surfaces is called a *cold-front occlusion*. This occlusion has been represented in figure 2 in a co-ordinate system attached

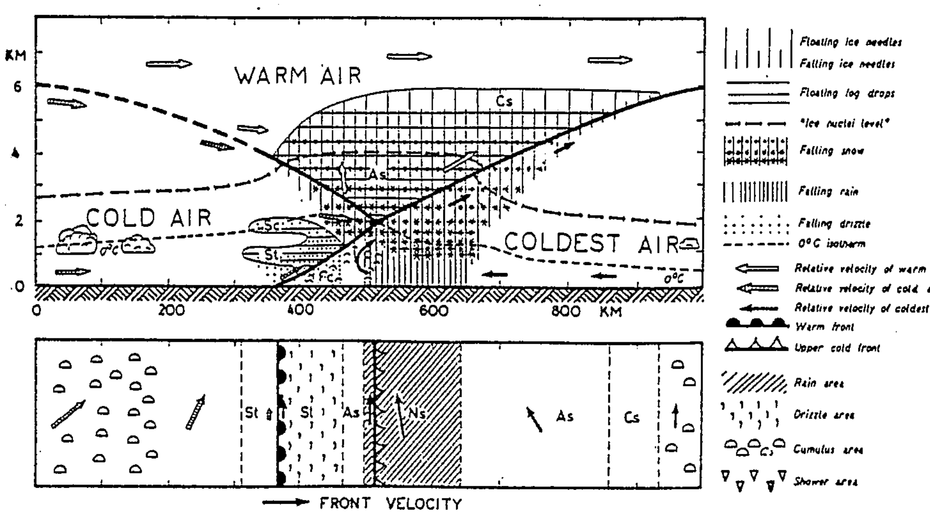

FIG. 14·33·1. The warm-front occlusion represented by a vertical profile and a strip of the surface map. The wind arrows represent the air motion in a co-ordinate system attached to the front.

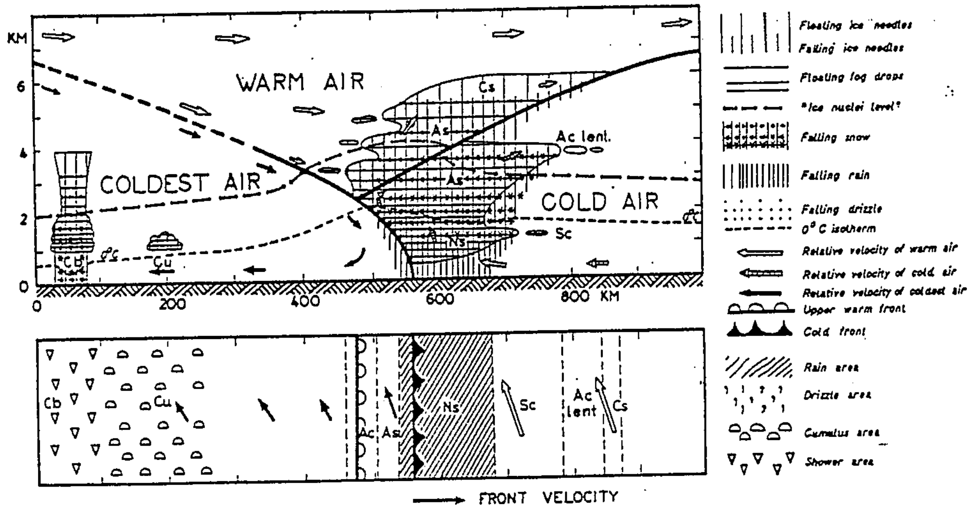

FIG. 14·33·2. The cold-front occlusion represented by a vertical profile and a strip of the surface map. The wind arrows represent the air motion in a co-ordinate system attached to the front.

to the surface front; the cold front has been assumed to be of the fast-running and partly frontolyzed type. The slope of the lower cold-front surface, separating the two cold wedges, will according to Margules' formula be steeper than that of the upper cold-front surface.

The cloud system in the cold-front occlusion is also formed by superposition of two separate frontal cloud systems, but they seem to contract in width during the process. To an observer stationed at the ground the approaching cold-front occlusion appears at first to have the character of a comparatively feeble warm front. The occlusion character is later revealed by the sudden arrival of the cold-front cloud system associated with a marked veering of the wind. The rain reaches its maximum at that stage and then decreases as the cold-front Altostratus becomes higher and thinner. When the cold-front occlusion grows old it is often hard to distinguish from a simple cold front.

14·4. The occluded-cyclone model

In chapter 15 the problem of wave cyclones will be attacked through combined synoptic and dynamic methods. These discussions will show, in conformity with experience, that a frontal wave generally leads to the creation of a moving cyclone, which passes through the various stages illustrated by figure 15·00·1. Since the system of cloud and precipitation in such a cyclone is built up from the elementary systems already described for each type of front, it may be practical to represent it at this stage. Figure 1 illustrates an "ideal" occluded frontal cyclone, moving from west to east, at the time of its maximum of intensity; the coldest air is supposed to be situated ahead of the warm front. Isohypses of the frontal surfaces for 2, 4, and 6 km elevation show the linking together of warm- and cold-front slopes along an upper warm-air trough coinciding with the upper cold front.

The part *a–b* of the surface front is approximately parallel to the isobars and thus of the quasi-stationary type. The part *b–c* of the front is a typical warm front with cyclonic kinks in the isobars and a broad prefrontal rain area. The air on the cold (right) side of the front is assumed to be Polar air. At a sufficient distance from the front this air shows the typical CM showers (▽). Nearer to the front only ordinary Cumuli (⌓) can form, because the instability convection is suppressed by the frontal surface or by surfaces of subsidence within the CM itself (cf. fig. 14·31·1). The air on the warm side (assumed to be maritime Tropical air) is characterized by Stratus layers, accompanied by drizzle near the front. The front section *c–d* is a cold front with sharp cyclonic kinks of the isobars, separating mT and mP. The northern part of the cold front, with narrow and prefrontal rain area, is of the second type, whereas the southern part is assumed to be of the first type, characterized by a broad postfrontal rain area and a relatively gentle slope of the frontal surface (as indicated by the isohypses). Where the postfrontal Polar air becomes sufficiently deep it contains the typical CM showers, whereas near the front only ordinary Cumuli can form. From the point *c* northward to *e* the figure shows an occlusion of the warm-front type, in which the cold front has overtaken the warm front and is to be found as an upper cold front *c–e'*. The occlusion has in its northernmost part developed into a so-called *back-bent occlusion* (cf. 15·00), represented by *e–f*, which has cold-front character. Thus a *false warm sector* is formed, which sometimes seems to cause a regeneration of the cyclone at later stages. Such regeneration must be preceded by a lengthening of the trough of the back-bent occlusion and a strengthening of the temperature contrast across it.

Figure 1 further presents several vertical cross sections cutting through the cyclone in zonal orientation; the tropopause appears in these cross sections and has been drawn according to the results of 15·30. A profile $\alpha\alpha$, north of the cyclone center, shows only Polar air near the earth's surface, but the Polar front surface

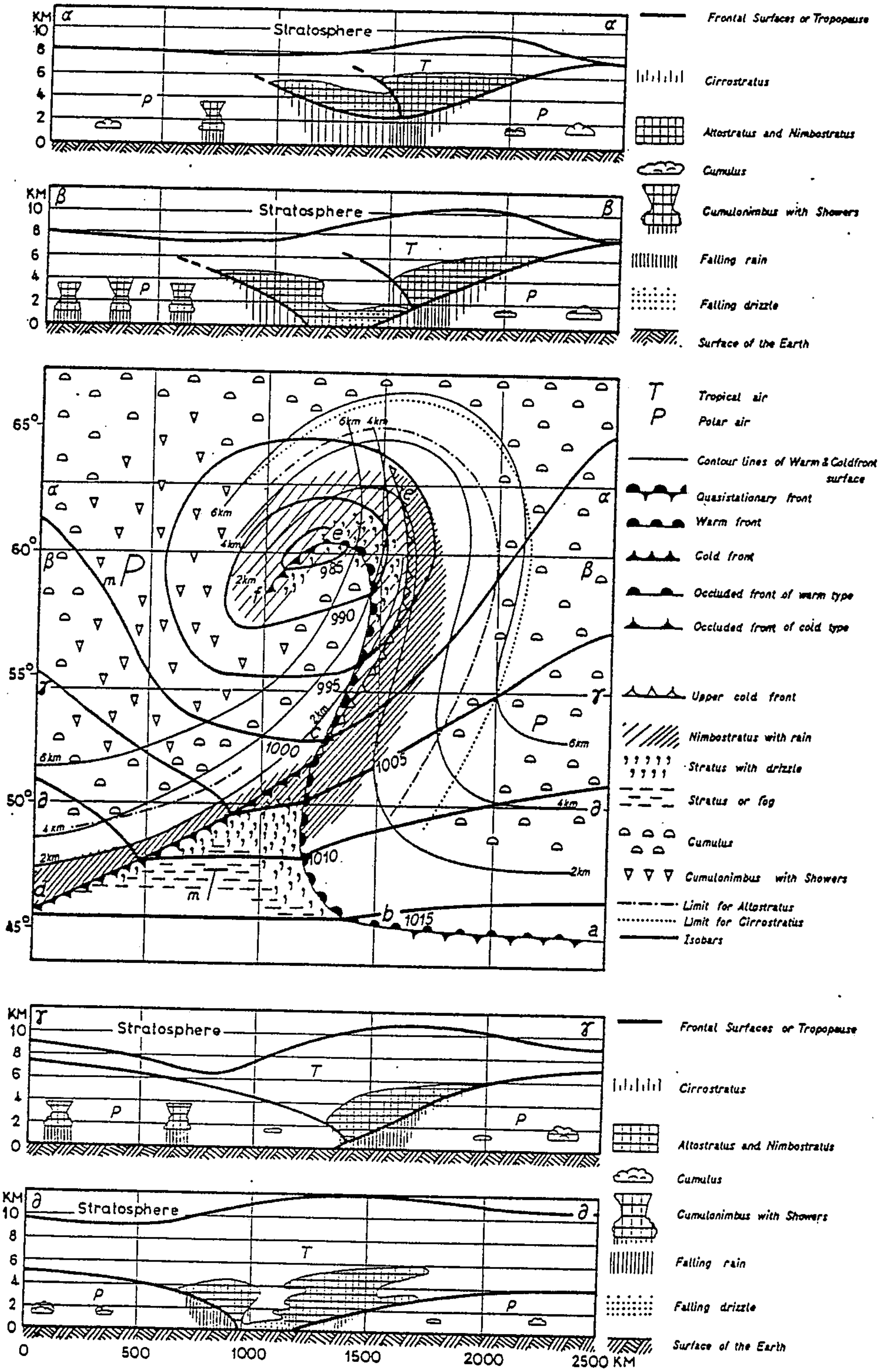

FIG. 14·4·1. A model of a strong occluded cyclone represented by a surface map (middle) and by vertical profiles (above and below).

lowers temporarily while the center passes. The eastern part of the cloud system shown in $\alpha\alpha$ belongs to the tongue of Tropical air which climbs the slope of prefrontal cold air. The western part of the cloud system lies behind the upper cold front and is therefore of polar origin. Also this air is saturated with moisture after a considerable lifting, and produces precipitation. Toward the western end of the profile, where the upper wind veers from southwest to northwest, the upglide motion ceases, and the frontal surfaces vanish because of frontolysis. The next, more southerly, cross section, $\beta\beta$, passes through the occluded front c–e and also through the back-bent occlusion e–f. The warm-front occlusion c–e has a prefrontal rain area, whereas the back-bent occlusion e–f is of cold-front type with mainly postfrontal rain. The air at the surface of the earth is everywhere Polar air, but the center part of it is of a relatively warm variety. The profile $\gamma\gamma$, which does not pass through the back-bent occlusion, shows a simple warm-front occlusion, represented in more detail in figure 14·33·1. Finally, the profile $\delta\delta$ passes through the warm sector somewhat to the south of the triple point c. The Tropical air, which here reaches the ground, is bounded to the east by a warm-front surface represented in detail by figure 14·31·1, and to the west by a cold-front surface of the second type (see fig. 14·32·2).

14·5. The weather type zones of the earth

In section 13·3 the mean geographical positions of the fronts were briefly discussed by means of surface pressure maps. We shall now supplement the maps by a vertical cross section along a selected meridian, showing the frontal clouds and forms of precipitation. As has been seen from the maps in 13·30–13·33, such a meridional profile must differ with longitude and season. Thus, a winter profile in the western Atlantic would show no Arctic front, whereas a more easterly profile would do so. Furthermore, in some profiles the Intertropical front would be situated north of, in others south of, the equator, some of the profiles would show fronts between maritime and continental air, and so on. We present as an example a cross section along 10°W referring to winter conditions (fig. 1). It contains in addition to the Polar front an Arctic front as well as an Intertropical front and also a Trade front, which is, however, ill defined at sea level. By omitting or modifying some of these fronts, it is easy for the reader to construct similar profiles for other longitudes. The vertical cross section has been supplemented by a horizontal "map strip" along the meridian, showing the wind distribution and the forms of precipitation observed at the ground.

The Arctic air, which in figure 1 is assumed to extend from 90° to 70° latitude, is mainly cloudless in its source region (as long as it does not arrive over open sea, where it will appear as a CM). Stratus layers may, however, occur and produce slight precipitation in the form of ice needles, as indicated in the "map strip." Along parts of the Arctic front an upglide cloud system exists, which is assumed to give precipitation in the form of snow north of the front. Just south of it the northward-flowing maritime Polar air generally appears as a stable WM, characterized by a Stratus system from which drizzle falls. Farther south the mP flows toward the warmer ocean, so that the typical CM clouds appear: Cumulonimbus with comparatively narrow ascending towers separated by areas of slowly descending unsaturated air. The showers extend southward to the region where the subsidence under the inclined surface of the Polar front prevents convective clouds from reaching the ice level. A narrow zone of fair weather may exist between the mP showers and the upglide cloud system. The Polar front system is analogous to that of the Arctic front, but rain and not snow will be the characteristic precipitation. South of the Polar front, where the maritime Tropical air appears as a WM, a drizzle region exists, and next a region with CM clouds, which in the 10°W cross section are prevented from giving effective precipitation by the subsidence under the Trade front surface. This front and also the Intertropical front are, at the longitude under consideration, generally without precipitation, the continental Tropical air over Africa being too dry. The cT between 25° and 10°N presents no typical CM or WM characteristics, a fact which, together with the great dryness, accounts for the almost cloudless skies of the desert. The Intertropical front surface (cf. 13·30) slopes gently upward toward the equator, but the temperature contrast along it will be insignificant (more temperature contrast and more depth of the cold wedge will be found in the opposite season, when the southern hemisphere has winter). In the equatorial region, where usually no inversions impede the convection, showers occur. This showery convection tends to efface the Intertropical frontal surface in the temperature field, but the wind shift, from westerly com-

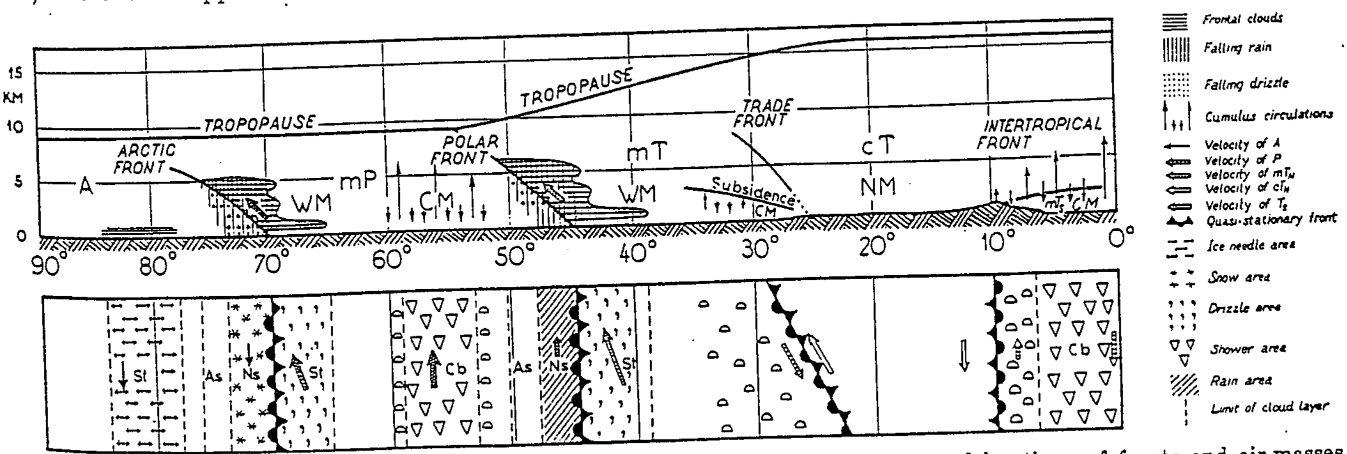

FIG. 14·5·1. Meridional profile and meridional map strip of a model atmosphere with normal locations of fronts and air-masses in winter at 10°W.

ponents in the cold wedge to easterly components above it, usually persists.

The positions of the various fronts may, of course, differ considerably from the mean positions given in figure 1, and temporarily some of the fronts may also be missing. The most mobile front is the Polar front, which sweeps all over the temperate zone and gives that region its highly variable and rather wet climate. The arctic climate has less variability than that of the temperate zone, owing to lesser frequency of frontal disturbances.

The subtropics, with their large anticyclones, have a climate that is seldom troubled by frontal passages. The continental subtropics have dry climates except in the seasons when they are reached by maritime currents (monsoons). The oceanic subtropics also have very little precipitation in the eastern and central parts of the highs. But the southern and western fringes of the highs are frequently under the influence of tongues of showery weather extending from the doldrum zone (13·24).

The equatorial climate as represented symbolically in figure 1 results from the meeting of a cool, moist (mT) and a warm, dry (cT) current. Showers then occur only when the moist current has reached a certain minimum depth. Other meridional profiles, intersecting the tropics in mid-ocean, would show moist currents on either side of the Intertropical front. A zone of showers is then likely to develop at the front, whereas on either side of it the trade-wind inversions keep the CM clouds flat and devoid of precipitation. A wider spread of showers and rain away from the Intertropical front occurs with the "easterly waves" (15·43) and with the tropical cyclones (15·44), both migrating from east to west. Furthermore, remnants of extratropical cold fronts which have descended to low latitudes may also stir up shower activity. In the equatorial ocean areas, not instantaneously affected by any of the types of disturbance mentioned, the atmosphere is stable with moderate Cumulus, over some cold-water areas even stable enough for Stratus. Over equatorial land areas with adequate moisture supply, the afternoon shower activity is a regular feature, and the intensity of showers is greatly enhanced by the above-mentioned migrating disturbances.

LITERATURE CITED IN CHAPTER 14

Bergeron, T.
1928 Über die dreidimensional verknüpfende Wetteranalyse. I. Geofys. Publ., vol. 5, no. 6. 111 pp.
1934 Die dreidimensional verknüpfende Wetteranalyse. II. Dynamik und Thermodynamik der Fronten und Frontalstörungen. Ausg. Zentralverw. hydro-meteor. Einheitsdienstes (Moscow).
1936 Physik der troposphärischen Fronten und ihrer Störungen. Wetter, vol. 53, pp. 381–395.

Byers, H. R.
1937 Synoptic and aeronautical meteorology. New York: McGraw-Hill Book Co. 279 pp.

Chromov, S. P., and N. KončEk (trans. G. Swoboda)
1940 Einführung in die synoptische Wetteranalyse. Wien: J. Springer. 532 pp.

Goldie, A. H. R.
1923 Circumstances determining the distribution of temperature in the upper air under conditions of high and low barometric pressure. Quart. Jour. R. Meteor. Soc., vol. 49, no. 205, pp. 6–16.

Hjorth, H.
1941 Contribution to the air mass analysis in southeastern Norway. Arch. Math. og Naturv., vol. 45, no. 1.

Jaw, J. J.
1935 A preliminary analysis of the air masses over eastern China. Mem. Nat. Res. Inst. Meteor. (Nanking), no. 6. 24 pp.

Neiburger, M., C. Beer, and L. Leopold
1945 The California stratus investigation of 1944. Washington, D. C.: U. S. Weather Bureau. 83 pp.

Petterssen, S.
1940 Weather analysis and forecasting. New York: McGraw-Hill Book Co. 502 pp.

Poljakova, A. N.
1936 Die Verteilung der äquivalentpotentiellen Temperaturen mit der Höhe in verschiedenen Luftmassentypen in Moskau. Jour. Geophys. and Meteor., vol. 6, pp. 512–525.

Schinze, G.
1932 Die Erkennung der troposphärischen Luftmassen aus ihren Einzelfeldern. Meteor. Ztschr., vol. 49, pp. 169–179.
1938 Untersuchungen zur aerologischen Synoptik mit Tabellen. 3d ed. Breslau.

Showalter, A. K.
1939 Further studies of American air mass properties. Mon. Wea. Rev., vol. 67, no. 7, pp. 204–218.

Solot, S. B.
1943 The meteorology of central Africa. Air Wea. Serv. Tech. Repts., ser. 105, no. 50.
1950 General circulation over the Anglo-Egyptian Sudan and adjacent regions. Bull. Amer. Meteor. Soc., vol. 31, no. 3, pp. 85–94.

Sverdrup, H. U.
1945 Oceanography. *In* Handbook of meteorology (F. A. Berry *et al.*, ed.), pp. 1029–1056. New York: McGraw-Hill Book Co. 1068 pp.

Tschierske, H.
1932 Abhängigkeit der Sicht von der Luftmasse. Wetter, vol. 49, pp. 307–312.

Tu, Chang-Wang
1939 Chinese air mass properties. Quart. Jour. R. Meteor. Soc., vol. 65, no. 278, pp. 33–51.

U. S. Army Air Force
1945 Non-frontal cumulus clouds on the North Atlantic routes. AAF Wea. Sta., Massachusetts Inst. Tech. Repts., no. 8. 16 pp.

Willett, H. C.
1933 American air mass properties. Papers Phys. Oceanogr. and Meteor., Massachusetts Inst. Tech., vol. 2, no. 2. 116 pp.

Chapter 15

DYNAMICAL ANALYSIS OF CYCLONES AND ANTICYCLONES

As the next step toward the study of the weather map, we now proceed to a description and dynamical analysis of entire atmospheric disturbances. In 15·0 the empirical facts about the life cycle of extratropical cyclones are presented in the language of the synoptic models of air-masses and fronts as set forth in chapters 13 and 14. Then follow various attempts at explaining the observed developments through dynamical theories partly developed in earlier chapters. The theories invoked concern, first, the irreversible energy transformation involved in the transformation of a frontal wave to a cyclonic vortex, and, secondly, the trigger mechanism necessary to start the incipient frontal wave. The first of these problems was attacked in 12·41 and the second in 9·73. The observed coupling of the frontal wave with accompanying upper waves adds new problems of wave dynamics to be solved within the scope of general cyclone theory, for which some groundwork was laid in chapter 11.

In 15·1 further studies on the kinematics of extra-[tro]pical cyclones follow, which mainly concentrate on [determin]ing the field of vertical motion. The facts from this kind of study have not yet been properly incorporated into some of the theories of quasi-horizontal wave motion used in 15·0. Moreover, knowledge of the field of vertical motion is indispensable for the proper application of thermodynamical equations to the cyclone problem and, indirectly, for the study of the cloudiness and precipitation of cyclones.

In 15·2 the theory of pressure changes presented in chapter 11 is applied to cyclones and anticyclones, and gives a useful check of the wave theories in 15·0 against the thermodynamical analysis in 15·1.

In 15·3 a new synoptic phenomenon, the slow-moving long wave in the westerlies, is considered with regard to its influence on the motion and development of the faster-moving short-wave phenomena of individual cyclones and anticyclones. The reciprocity of interaction of short- and long-wave phenomena calls for further future concentration of study on the all-embracing atmospheric problem of the general circulation.

Tropical disturbances have only recently been analyzed and classified synoptically, and they present novel and difficult problems of dynamical interpretation. In 15·4 the disturbances encountered on synoptic maps in the tropics have been represented by tentative [dyn]amic models, but many questions as to their life cycle have had to be left open.

Chapter 15, in which theory is for the first time confronted with the complexities of synoptic meteorology, will often show the reader the shortcomings of our present basic theories. One of these shortcomings is due to the fact that the theories in the preceding chapters operate, for the sake of simplicity, with models having sharp discontinuities of temperature and wind, whereas the corresponding models in the real atmosphere display frontolysis in various degrees (cf. 14·00). The great need for improved dynamical insight into the problems of real atmospheric disturbances will become increasingly clear as we come to the problems of daily forecasting (chapter 18).

15·0. Origin and life cycle of extratropical cyclones

The typical development of the extratropical cyclone on a pre-existing front is presented in 15·00, mainly following the ideas of the Polar front theory of 1920. According to that theory the individual cyclone runs through successive frontal wave stages and ends up as a cold vortex, because that development involves an irreversible energy transformation from potential to kinetic. The possible creation of the kinetic energy of the growing cyclone from the potential energy of an initial system of cold and warm air separated by a front is evaluated in 15·01.

The trigger mechanism which breaks the Margules equilibrium of the adjacent cold and warm air-masses is the subject of sections 15·02–15·04. In 15·02 Helmholtz' ideas on the dynamic stability of baroclinic zonal currents are adapted to test the dynamic possibilities for nongeostrophic isentropic upgliding or downgliding across the direction of the fundamental currents. In 15·03 a model of the fundamental currents parallel to a front in "straight westerlies" is presented, on which the criteria for up- and downgliding from 15·02 are tested. In 15·04 the upgliding is shown to be possible as part of the frontogenesis in a confluent current system, and the downgliding on the cold side of the front is shown to be the probable later consequence of the continued sharpening of the front. The first downgliding to reach the ground leads to a cold-front push which starts off the frontal wave.

In 15·05 is shown how a frontal wave necessarily creates an internal wave in the whole baroclinic current above it, and keeps that wave in rather fixed relative phase. The tilt of the troughs and crests of this upper wave is shown to be associated with the process of intensification of the cyclone. In 15·06 the triggering of the dynamic instability observed in the region of the anticyclonic bends of the upper westerlies is shown to lead to unstable waves and "meandering" of the upper current. This may also lead to cyclone formations which are not started in the ordinary way as waves on a surface front. Subsection 15·07 states the empirical evidence for the formation of strong cyclonic storms when unstable shear occurs in the upper troposphere, and underlines the importance of that shear for the intensification of all storms of whatever origin.

In 15·08 the theories of V. H. Ryd and R. Scherhag concerning cyclone intensification of mature cyclones under an upper delta-shaped wind field are cited. The corresponding dynamic model of waves superimposed on a fundamental current of delta pattern has not yet been analyzed.

15·00. The Polar front theory of cyclones. Experience in use of synoptic weather maps in temperate and arctic latitudes shows that most cyclones form as waves on the Polar front (or the Arctic front) and run through a typical life cycle, which is illustrated schematically by maps and zonal profiles in figure 1. The front itself originates through the bringing together of warm and cold air in one of the major cols of the general circulation (see 13·3), and extends initially along the axis of dilatation of the hyperbolic streamline fields of such a col. The most frequent initial condition for cyclone wave formation in the zone of westerlies is the presence of two slightly converging currents (see fig. 1*a*), one cold and one warm, coming from the col area. They are separated by a sloping Polar front surface whose intersection with the ground is quasi-straight and parallel to the wind. The difference in wind speed on the two sides of the front is always such as to produce cyclonic shear in the frontal zone.

For simplicity, in the map pattern of figure 1*a* the quasi-stationary front has been made to run exactly west to east. This front orientation may vary from case to case and is more often west-southwest to east-northeast, or even southwest to northeast, than for instance west-northwest to east-southeast. The growing frontal waves usually start just east of the col. Figures 1*b*–*e* show their normal development into cyclonic vortices. The time scale is, roughly, *a*–*c* 24 hours and *c*–*e* another 24 hours. The map area covered in each of figures 1*b*–*e* is smaller than that of figure 1*a*.

The nascent cyclone. Figure 1*b* shows the nascent cyclone as a newly formed wave, which moves along the front from west to east with a velocity nearly equal to that of the warm air near the ground (see 10·63, where the same rule is derived for a wave between two barotropic air masses). The corresponding perturbation of the initial pressure field is at this stage only slight, but the typical kinks of the isobars where they intersect the front appear clearly (cf. fig. 11·22·1). The *warm front* ahead of the so-called warm sector separates a concave receding part of the cold air from a convex advancing part of the warm air. Analogously the *cold front* separates a concave receding part of the warm air from a convex advancing bulge of the cold air. The cold bulge just west of the warm sector contains air which has slid down along the frontal slope, and is therefore less cold than the rest of the air-mass.

The wave cyclone. The frontal wave in figure 1*c* shows a further development of the asymmetry which started to appear in the nascent cyclone. Along with that development goes a process of frontolysis (11·23) of a part of the cold front next to the center of the cyclone, because the cold air in that part is sinking. The cold air ahead of the warm front is usually not sinking, and the frontal kink in the isobars there remains sharp, although the frontal temperature contrast at the ground level may be masked by the thin layer of cold air left behind by the receding cold wedge.

In the wave stage of the cyclone the upper perturbation, shown in the tropopause profile, begins to appear. The wave crest of the tropopause (as well as that of the upper isobars and the upper tropospheric isotherms) is situated over the warm-front surface, and the tropopause wave trough over the cold-front surface. This phase lag of the upper perturbation relative to the lower one never fails to appear, and is rather constant from case to case. It is suggested in 15·04 that the anticyclonic vorticity on the crest of the upper perturbation must always form where the lifted warm air of the frontal wave is exposed to horizontal divergence, and the cyclonic vorticity of the upper trough must form through horizontal convergence above the region where downslope motion takes place on the cold-front surface. If this is the fundamental action from below on the upper layers, the constancy of the phase lag is easily understood. It is also possible that the upper layers can develop perturbations independently, as outlined in

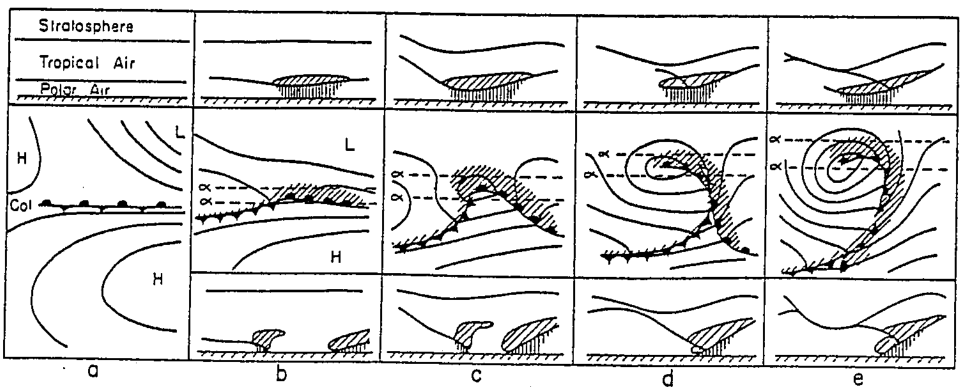

FIG. 15·00·1. Life cycle of cyclones from wave to vortex.

15·06. At any rate, for vigorous cyclone formation a strong development of the upper as well as the lower perturbation is essential. Such co-ordinated growth of the upper and the lower disturbance is assumed to take place in the model development under consideration.

The occluded cyclone From the very beginning the part of the cold front near the center of the cyclone is moving more rapidly than any other part of the front system, and soon catches up with the warm front, as shown by figure 1*d* and *e*. This process, which involves the lifting of the warm-sector air between two wedges of colder air, has (by T. Bergeron) been named the *process of occlusion* (cf. also 10·64·5°). The front that remains where the two cold wedges meet is called an occluded front. The southern profile in figure 1*e* shows the schematic arrangement of air-masses at the occluded front, specialized for the case where the cold wedge ahead of the front is colder than the cold wedge behind it (cf. also the more detailed representation of a warm-front occlusion given in fig. 14·33·1). The cold front has in this case climbed up on the receding cold wedge and is from now on an *upper cold front*. Its projection on the ground lies ahead of the occluded front proper, which separates the two kinds of Polar air at the ground. Figure 1*e* is intended to represent the rather frequent case of a change from a warm-front occlusion to a cold-front occlusion with increasing distance from the center. Where the change takes place there is zero temperature difference between the prefrontal and the postfrontal cold wedge. It is natural to mark that spot by a little gap in the occluded front on the surface map.

The initial asymmetry of the pressure field of the young Polar front wave gradually develops into the formation of a pressure trough pointing backward from the center of the cyclone. The same trough in most cases begins to rotate round the cyclone, and in figure 1*e* it points southwestward from the center. Near the center the *back-bent occlusion* coincides with the trough. Farther out from the center the trough contains no front, and will be referred to henceforth as the *nonfrontal trough*. The nonfrontal trough lies under the trough of the upper perturbation, and the two are of course connected hydrostatically. A vigorous development of the nonfrontal trough is a sure sign of a corresponding active development of the upper trough.

As soon as the cyclone occludes, its speed usually drops. The type of occluded cyclone shown in figure 1*e* results when the decrease in speed is moderate. If the speed drops more decisively, while the cyclonic winds are still strong, the occluded front will wind around the center in spiral shape.

The Polar front theory of the life cycle of cyclones seeks to explain the increase of kinetic energy by a transfer from potential energy. Thus, the lifting of the warm-sector air and its replacement at the ground by the adjacent wedges of cold air in the occlusion process is interpreted as involving a transfer of potential into kinetic energy. This process seems to dominate during the life-cycle stages *a* to *e* and to provide an explanation for the rapid appearance of new kinetic storm energy. All synoptic experience shows the evolution from a frontal wave to a cold cyclonic vortex to be irreversible, which would indicate that the cyclone owes its kinetic energy to the triggering of the energy stored during the preceding period of frontogenesis.

Sometime after the completion of the occlusion process, the continual dissipation of kinetic energy by friction, and also the lifting of cold air in warmer environment, begin to determine the sign of energy transfers. The storm energy thus finally reverts into heat and probably also into potential energy (see also 12·41 and 15·01).

A temporary rejuvenation of old cyclones may sometimes occur if the back-bent occlusion, of the cold-front type, is longer than normal and separates two Polar air-masses of different temperatures. The Polar air trapped between the occluded front and the back-bent occlusion is then warmer than the rest of the Polar air, and may act as a new warm sector and give the cyclone a second brief period of deepening.

15·01. The energy transformations in the frontal cyclone. Attempts at realistic quantitative estimates of the energy transformations in the growing cyclone cannot get past the uncertainties introduced by "neighboring systems"; and if the theoretical treatment is restricted to a piece of atmosphere enclosed within walls, the interpretation of the results becomes speculative. Even so we believe that the "tube experiment" described in 12·41·4° has some bearing on our practical problems, and we shall discuss it here.

The main thing we can learn from the tube experiment is that the rising of the warm air and the simultaneous sinking of adjacent cold air on the opposite side of a frontal surface represent a source of energy of the right order of magnitude for the making of storms. But we are still far from any accurate application of the energy experiment. The tube eliminates all motion parallel to the front, and it is difficult to assess what difference it would make to the energy transformation if the motion parallel to the front were reintroduced. Probably it would enable the warm-air elements to rise higher, during the life cycle of the individual cyclone, than could be inferred from the experiment in the tube; because, through the motion eastward parallel to the front, the rising warm air is continually contacting new cold air, across the front, which has not yet warmed up by sinking. The tube experiment may thus lead to underestimation of the possible production of kinetic energy. Another point to have in mind is the possibility of ascent over a narrow area and compensating descent over a wide area similar to that investigated for the