LOCAL WINDS*

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INTRODUCTION AND DEFINITION

The terms of the complete equation of atmospheric motion are determined by a number of forces which, through their interplay, cause the movements of the atmosphere. These effective forces are gravity, hydrostatic pressure, friction, and the Coriolis force.

If the Coriolis and friction terms are negligibly small, we deal with *Eulerian wind* equations in which the acceleration can be measured by the pressure gradient. However, if the Coriolis term is large as compared to the acceleration and friction terms, so that the pressure gradient is balanced only by the deflective force of the earth's rotation, we speak of *geostrophic winds*. As a third possibility, the friction terms may be so large that they are of the same order of magnitude as the pressure gradient term, and the Coriolis and acceleration terms may be neglected. In that case the wind blows approximately in the direction of the pressure gradient and it is called an *antitriptic wind* [42].

The antitriptic wind introduces the field of winds restricted to relatively small areas. Their range is of the order of 100 km or less. In this category belong the land and sea breezes, the mountain and valley winds, the jet-effect (*Düseneffekt*) winds, and, at least in so far as their characteristics are determined by the orography of the ground, the foehn and bora winds. These winds of locally restricted influence are classed under the general name *local winds*. They are an elementary yet meteorologically very interesting part of the movement of the atmosphere and a phenomenon which always attracts the attention of the layman.

In the following sections the basic principles of these local winds will be discussed. These winds very closely fulfill almost all conditions of the antitriptic wind since they blow roughly at right angles to the isotherms and isobars and are limited in the vertical to a relatively low altitude.

LAND AND SEA BREEZES

Description of the Phenomenon. In coastal areas, especially on tropical coasts and on the shores of relatively large lakes, we can observe in the course of a day the reversal of onshore and offshore winds, called land and sea breezes. The phenomenon takes the following course:

A few hours after sunrise, depending on location and season, a sea breeze develops, particularly on calm summer days. This sea breeze usually has a noticeable cooling effect, particularly in the tropics; it continues throughout the daylight hours and dies down around sunset. After that the seaward-blowing land breeze appears. This phenomenon, which is restricted to the coastal area proper and which extends only in tropical regions over a 100-km range, must be attributed to the difference in heat response between land and water. On warm, clear, summer days, the horizontal temperature difference between land and water provides the energy that leads to the development of vertical and horizontal air currents of strictly circulatory character. To the ground observer, only the horizontal parts of the circulation are noticeable; the vertical branches can be observed only indirectly, for example, through the formation of clouds over the land. The daytime sea breeze considerably surpasses in intensity the nocturnal land breeze. This is understandable because of the greater daytime arc of the sun during summer and the increased instability and consequently increased vertical austausch in daytime. The nocturnal cooling, on the other hand, produces an immediate stabilization of the air layers near the ground.

It is noteworthy that the direction of the sea breeze does not remain constant in the course of the day, and that this breeze often sets in with considerable gustiness, sometimes in the form of a protrusion similar to a cold front.

It is logical that the gradient wind, as determined by the over-all weather situation, should be superimposed on this local wind system and should at times obscure or even conceal it completely.

As an example of a typical day with land and sea breezes, the wind, temperature, and humidity recordings of the Danzig airfield, 3.35 km inland from the Baltic Sea coast, from June 3 to June 5, 1932, are reproduced in Fig. 1. This diagram clearly shows the onset of the sea breeze from the northeast and north-northeast, respectively, at 1420 on June 3, and at 1430 on June 5, while on June 4 a gradient wind from west-southwest—west-northwest completely conceals the sea breeze. After a calm from 1930 to 2245 on June 3, and from 2100 to 2220 on June 5, the land breeze sets in from southwest and west, respectively, and continues as a mild breeze until 0600 the following day. During the night of June 4—5, a calm from 0030 to 0130 and a shift in wind direction to the southeast are the only indications of the onset of the land breeze, but here also a gradient wind that subsequently appears hinders its development until 0600 in the morning. The temperature and humidity curves of Fig. 1 show characteristic irregularities in their normal trend at the onset and cessation of the sea breeze.

Figure 2 gives an example of the above-mentioned change in velocity of land and sea breezes in the course of the day, as observed at Hoek van Holland on July 31, 1938. We can see from this figure the increase in the speed of the sea breeze up to the daily maximum between 1300 and 1400 and a concurrent steady shift.

* Translated from the original German.
to the right, which will be explained later. The dying down and continued shifting to the right takes place in similar fashion.

As an example of the tropical form of the land and sea breezes the diagram of velocity isopleths by van Bemmelen of the land and sea breezes at Batavia is reproduced in Fig. 3. It is also an example of the vertical velocity distribution during the course of a day.

Explanation of the Land and Sea Breezes. The cause of the land and sea breezes must undoubtedly be sought in the different behavior of land and water under the influence of an equal external heat supply. Water, as compared to soil, has a larger thermal capacity and its specific heat per unit volume reaches a value 40 per cent larger than that of soil. Although water should have a smaller temperature variation for the first reason, the amplitude of these periodic variations would be \(1.414\) that of the land for the second reason. This difference is equalized through the much smaller heat conductivity of water, and measurements reveal that the surfaces of water and sandy or rocky ground have temperature variations of comparable order. The depth to which radiation penetrates can also not be considered responsible for large temperature differences since the infrared radiation is immediately absorbed in the upper water layers.

However, if we direct our attention to the turbulent mixing of the water by wind and waves, which effects a continuous downward transport of surface heat through large masses of water, we recognize that this mixing is the cause of the relatively small temperature variations. It is now clear that the complete absorption of radiant heat in the surface layers of the ground and the weaker influence of this form of energy on the deeper layers result in an entirely different thermal behavior of land and of water. Thus the temperature conditions of the ground are determined almost exclusively by its physical properties, whereas those of
the water are governed also by apparent conduction or turbulent mixing. In contrast with the strong heating of the air over the coastal region, the air over the strip of water offshore is only mildly warmed, and, as a result, a temperature difference between land and water develops. This difference diminishes toward sunset and reverses during the night.

**Temperature Differences, Pressure Differences, and Pressure Distribution during Land and Sea Breezes.**

The maximum temperature differences (ΔT) are given in the literature as follows: Kaiser [46] gives a range of ΔT from 1.6°C to 10.9°C over a distance of 130 km between Wüstrow and Adlergrund Lightship (anchored in the middle of the Baltic Sea, about 100 km out of Swinemünde) as averages of a period of twenty summer days. Grenander [31] found differences in temperature between 3.6°C and 7.6°C at 1400, and between 1.4°C and 3.1°C at 0700 and 2100. These measurements, taken on the Swedish east coast, involved a distance between land and sea stations of 115 km. However, maximum values were as high as 10°C or more; on the other hand, very small temperature differences occurred on some sea-breeze days. Measurements of ΔT at lakes, as for instance at the Lake of Constance, likewise show a large range of temperature differences, namely values between 0.9°C and 4°C at 1400. We may conclude from these few measurements that ΔT covers a wide range of values, and it is certain that a temperature gradient from sea to land exists in the morning hours. During forenoon a reversal takes place, and in the afternoon an increasing land-sea temperature gradient develops which is the driving force in the formation of the sea breeze. Over inland lakes, where opposite shores show this same behavior, the center of the lake must be a neutral zone. In general it may be assumed that the heating over land during daytime can reach a value five times that over water. Such temperature differences force the development of a pressure gradient and circulation system.

At the beginning of the day, the air pressure at higher altitudes over the land rises, while there is only a negligible increase over the water. As a consequence a drainage of the upper air from the land toward the sea takes place, and during forenoon the pressure close to the surface of the sea begins to rise, while it starts to fall over the land. This developing sea-land pressure gradient is accompanied by an air current in the same direction, that is, the sea breeze. A countercurrent, blowing toward land, is established at upper levels above the surface sea breeze. The circulation in daytime is completed by cumulus-forming convection over land and cloud-dissolving subsidence over water. In the evening the land and the overlying air cool faster than the sea and its overlying air, and a reverse nocturnal circulation develops. This circulation must be of smaller intensity and vertical extent because of the lack of instability and convection.

The land-sea pressure gradient near the surface at night and toward morning, as compared to a sea-land gradient during the day, is clearly discernible in Fig. 4.

![Fig. 3.—Velocity isopleths for the land and sea breeze in Batavia. (After van Bemmelen [70].) locally](image)

![Fig. 4.—Average daily period of the air pressure on twenty sea-breeze days at the Baltic Sea. The solid line refers to Swinemünde; the dashed line, to Adlergrund Lightship. (After Kaiser [46].)](image)
demonstrated numerically by superimposing the pressure fields of land and sea breezes and gradient wind [54].

Development of Land and Sea Breezes as a Function of Geographical Location, Season, and Time of Day. On tropical coasts, the land and sea breezes appear with great regularity [9, 11, 53, 70], because the clear sky there causes large variations in temperature over the land in the course of the day. Also the usually weak general air motion does not interfere with the development of local winds. It is only in India that the land breeze is completely obscured from May to September by the strong monsoon, which acts as a steady sea breeze during that season. Partial superimposition of mountain and valley winds on land and sea breezes may also cause peculiar wind conditions, as for instance on the coast of Samos. In higher latitudes these local winds appear almost exclusively, or at least preferably, during the warmer seasons, since only then can sufficiently large temperature or pressure differences develop. In the cooler climates of higher latitudes, as for instance at the shores of the Baltic Sea [46], we can expect land and sea breezes, even in summer, on not more than about 20 per cent of the days. The role of solar radiation becomes apparent in a brief summary (Table I) of the probability of a sea-breeze day for different

TABLE I. RELATIONSHIP BETWEEN CLOUDINESS AND SEA-BREEZE PROBABILITY

<table>
<thead>
<tr>
<th>Cloudiness (per cent)</th>
<th>0-50</th>
<th>60-80</th>
<th>90-100</th>
</tr>
</thead>
<tbody>
<tr>
<td>Probability of a sea-breeze day (per cent)</td>
<td>90</td>
<td>39</td>
<td>27</td>
</tr>
</tbody>
</table>

amounts of cloudiness at the Black Sea. Little cloudiness and strong sunshine are decisive in promoting the occurrence of land and sea breezes. In polar regions the phenomenon disappears almost completely and occurs only once in a while on particularly clear summer days.

In the tropics (Batavia) [70] we find about 40-50 per cent of land-breeze occurrences and 70-80 per cent of sea-breeze occurrences to be fairly reliable values for the dry season. During the rainy season both frequencies increase; the land-breeze probability rises to between 60 and 80 per cent, that of the sea breeze to more than 80 per cent. Ramdas [63] gives frequencies for extratropical land and sea breezes in Karachi, India, (25°N) for every month of the year (Table II). We can see from this table a 100 per cent occurrence during the summer months in contrast with the low percentage in the winter. Similarly, in etesian climates (as for instance the Mediterranean climate) spring has land and sea breezes on 31 per cent of the total number of days, the first half of June on 82 per cent, July on 91 per cent, and autumn on only 35 per cent of the days.

Finally, the phenomenon is in almost all regions a function of the time of day, since its periodic course is a consequence of the diurnal temperature variation. Usually, the sea breeze starts between 1000 and 1100, reaches its maximum velocity around 1300 to 1400, and subsides toward 1400 to 2000, whence it is replaced by the nocturnal land breeze. These approximate times naturally vary with the season and with climatic and local differences.

Intensity, Vertical Extent, and Range of Land and Sea Breezes. The height of the sea-breeze layer varies with climatic and local conditions. Its altitude ranges from 150 m at medium-sized lakes to 200-500 m at large lakes and the seacoast. Extending to 1000 m in moderately warm climates, the sea breeze reaches altitudes of 1300-1400 m in tropical coastal regions, as can be clearly seen in Fig. 3. In India, maximum altitudes of 2 km have been observed. In these areas, the nocturnal land-breeze layer is rather shallow by comparison. In Batavia, for instance, it reaches only to about 200-300 m. The difference in height between land and sea breezes is smaller in the temperate zones.

The intensities of the sea breeze cover the entire Beaufort scale. This breeze is of small force at lake and sea shores in the temperate zone (0 to 3 Beaufort); only at the seacoast do some peak values reach 4 to 5 Beaufort at noon. In the tropics, however, the wind may rise to storm intensity with the onset of the sea breeze. A particularly strong increase occurs on coasts with cold ocean currents offshore. While the horizontal speeds are of the order of meters per second, the vertical components are only of the order of centimeters per second.

The landward range of the sea breeze is estimated by many observers at 15-50 km in the temperate zones. Some values, for instance, are 16-32 km in New England, 20-30 km at the Baltic Sea, 30-40 km in Holland, up to 50 km in Jutland, 15 km on the Flemish coast, 40 km in Albania, more than 50 km on the northern coast of Java, and 40-50 km in Sweden. However, land and sea breezes are often augmented by mountain-valley wind effects which are difficult to separate from them. In tropical countries the sea breeze reaches 50-65 km, sometimes even 124-145 km into the interior. The seaward range of the much weaker land breeze appears to be everywhere much smaller. At the Baltic Sea, for instance, it extends only to 9 km.

Regarding the vertical temperature distribution in the temperate zone where condensation is relatively rare, nearly adiabatic or slightly superadiabatic gradients are to be expected. In the tropics, however, superadiabatic gradients are the rule, but probably reach only to the upper boundary of the sea breeze and decrease rapidly above it.

Conrad's Minor Sea Breeze, or Sea Breeze of the First Kind. The wind designated by Conrad [13] as "minor sea breeze" does not progress from the sea
The Cold Front-Like Sea Breeze, or Sea Breeze of the Second Kind. The character of the sea breeze having a normal, front-free, and steady development is modified if a wind determined by the general weather situation hinders its regular course. This sea breeze, which develops in opposition to the gradient wind, is characterized by its retarded beginning (usually as late as 1500 to 1600), by its front formation at sea and its slow progress toward the coast, and finally by a pronounced break-through of a cold front-like character (distinct gust with wind shift of 180 degrees). The development of this cold-air invasion can be considered to take place in the following way:

In the morning, the offshore gradient wind carries warmed air from the land out to sea and thus displaces seaward and weakens the pressure gradient between land and sea. An air-mass boundary is thus formed at sea against the cool sea air. The warmed land air, which accumulates in a nearly adiabatic layer, is highly turbulent and is able to carry along some of the stably stratified, cold sea air and is forced to rise with it. For a while, a stationary equilibrium between the two air masses may be maintained by an increasingly steepening frontal surface as is depicted in Fig. 5. However, with further heating this equilibrium breaks down, the system becomes unstable, and the sea air now breaks through toward the land in the form of a front. This sufficiently explains the gusty onset and the cold-air character. If we consider that the largest vertical temperature gradients are not reached until 1200 to 1400, the retarded onset of the sea breeze is also understandable.

Theory of Land and Sea Breezes. A complete theory of the land and sea breezes must necessarily consider (in addition to the gradient force caused by land-water temperature difference) (1) the influence of the vertical turbulent heat exchange, (2) the turbulent friction of the air motion, and (3) the influence of the earth’s rotation. In all existing theories these contributory factors receive only partially satisfactory consideration. Usually, the land and sea breezes are treated as simple, antitriptic currents caused by the unequal heating of land and water.

An older treatment of a stationary circulation is that by Jeffreys [42], who applied it primarily to the sea breeze and to monsoon winds. He considers friction, pressure, and Coriolis force and reaches a solution in which the daily wind change is in phase with the daily temperature oscillation. This result does not conform to reality. The application of this theory to the extended monsoon currents allows the calculation of the height of the monsoon reversal as well as of the amplitude of the surface pressure variations connected with the circulation. The agreement of these calculations with observation proves to be satisfactory.

V. Bjerknes and his collaborators [8] consider the land- and sea-breeze circulation a periodic current around isobaric-isosteric solenoids in which the wind is ninety degrees out of phase with the change in density. Accordingly, the sea breeze would start at the time of the greatest heating and would reach its greatest intensity at the time of the smallest temperature difference between land and sea in the evening. This is not in accordance with observation.

Kobayasi and Sasaki [52] as well as Arakawa and Utsugi [2] base their theories on Lord Rayleigh’s convection theory [64]. They consider vertical and horizontal heat transfer in addition to turbulent friction of the air motion, whereby their basis for calculation becomes more complete than that of Jeffreys. Their solutions of the problem allow a comparison between theory and observation, although in order to reach complete agreement a heat conductivity one hundred times greater than that derived by Taylor from observations must be assumed. A further shortcoming of the theory is the height to which the circulation (including the countercurrent) extends must be known, whereas it actually should result from the boundary conditions of the theory. This theory still neglects friction, which, as had already been pointed out by Godeske [30], is responsible for the phase shift between density and wind-velocity variations.

A simple elementary theory of land and sea breezes is due to F. H. Schmidt [67]. In this theory he is less concerned with giving a complete explanation of the circulatory movement than with clarifying characteristic phenomena of the land and sea breezes, such as the phase shift between wind and temperature or the influence of the earth’s rotation. A definite temperature distribution in accordance with observations is assumed, and the entire system of currents is calculated, taking the compressibility of the air into account. The Coriolis force is also introduced into the calculations, and thus
Schmidt succeeds in explaining quantitatively all facts of this atmospheric phenomenon in a satisfactory manner. Most certainly, his theory has greatly contributed to a better understanding of local winds.

Schmidt's assumption concerning temperature is characterized by the fact that the entire horizontal and vertical temperature distribution is given. He superimposes on the normal vertical temperature decrease the daily radiation temperature wave which is assumed to reach its maximum amplitude near the coast at a distance inland amounting to half the wave length of the temperature oscillation. This amplitude is supposed to decrease exponentially with altitude. However, for reasons of simplicity, he neglects the fact that the maximum amplitude is a function of time which, after all, should be of some importance to the theory. The variation in air density caused by the radiation temperature wave can be calculated; its amplitude also decreases exponentially with altitude. Part of the pressure variation is ascribed to the resulting density variation, and the remainder is caused by divergence and convergence of the compressible air. This influence of divergence is also assumed by Schmidt to decrease exponentially with altitude. If the pressure gradient is known, the horizontal velocity component can be calculated by application of the Guldberg-Mohn friction formula. Finally, the friction constant is assumed to be a quantity that decreases with altitude according to the expression $e^{-r}$, where $r$ is a constant representing the decrease of friction with height and $z$ is the height. This is necessary in order to obtain sufficient variation with altitude of the sea breeze's starting time. We shall not discuss here how far such assumptions are justified; moreover, it would be preferable if the theory itself would yield the variations with altitude of all these quantities. Nevertheless, the theoretical determination of the deviation of the wind direction from the perpendicular to the coast and of the shift of the wind in the course of a day gives a satisfactory result.

The most recent work on the theory and observation of land and sea breezes has been published by Pierson [61]. In this work the theoretical considerations of F. H. Schmidt are considerably improved. Pierson makes assumptions regarding the temperature contrast between land and water that closely approximate reality and he takes into consideration not only the Coriolis force but also that type of friction which is used in the derivation of the Ekman spiral. A solution is given of the Navier-Stokes equations for laminar flow with consideration of the apparent eddy viscosity. Theoretical hodographs illustrate the variations of land and sea breezes under the influence of this kind of friction, the Coriolis force, and the variable pressure-gradient force at all altitudes. The temperature contrast between land and water and its periodic variation during the course of a day as well as its variation with altitude are assumed, with due consideration for eddy diffusion (W. Schmidt, see [61, p. 9]), in a manner similar to that employed by F. H. Schmidt. From this the density and pressure distribution can be computed, and the wind field can be obtained from the equations of motion; the equation of continuity is unnecessary here. A comparison of the theoretical results with observations made at Boston (42°N), Madras (13.4°N), and Batavia (6°S) shows good agreement.

Recently, Haurwitz [37] made an interesting and important contribution to the theory of the land- and sea-breeze circulation. In contrast to previous investigators, he chooses Bjerknes' circulation theorem as his point of departure. With it he proves that the intensity increases, not only as long as the land-water-temperature difference increases, but that it keeps growing until this difference disappears. Thus, the phase shift between temperature difference and wind maximum would be a quarter of the period, that is, six hours. Haurwitz shows that the introduction of frictional influences causes a considerable decrease of this phase shift which leads to a better agreement with observation. The daily shifting of the sea breeze, which was clearly observed in several locations, can be explained without difficulty as an effect of the Coriolis force. Haurwitz' work excels in its logical, all-inclusive consideration of all factors that are of importance for the development of land and sea breezes. However, as in the work by F. H. Schmidt, the theory of the land- and sea-breeze circulation is incomplete. In all these investigations the temperature contrast between land and water at the surface and at all levels above it is assumed. A complete theory should furnish the temperature distribution with altitude as well as the circulation from a given temperature contrast at the surface. The temperature distribution with altitude depends not only on the vertical turbulent heat exchange, but also on the circulation itself. For this reason, with a given boundary condition of temperature at the surface, the equation of heat conduction (as in the theory of the slope winds, see p. 666) must be incorporated in the theory.

If we approach the land and sea breeze as a single circulation cell in the sense of Lord Rayleigh's convection theory [64], we come considerably closer to the problem of these local winds. Furthermore, with this method we can take vertical as well as horizontal heat transfer and turbulent friction into account. The solution must yield not only the entire temporal development of the periodic current system, but also its dimensions as a function of heat supply or of the land-water temperature difference, respectively. I have recently attempted such a solution [18], which actually furnishes the required results in full conformity with observations. For the land-water temperature difference near the surface, which we have assumed as given, I used the simple harmonic function $\vartheta = Me^{at} \sin \omega t$, where $x$ is the normal to the coast, $\vartheta$ the potential temperature, $M$ the amplitude of the temperature variation, $\Omega = 2\pi/(\text{sidereal day}) = 7.292 \times 10^{-3}$ sec$^{-1}$, and the length of the circulation cell is fixed as $L/2 = \pi/L$. This solution permits the utilization of any desired form of the land-water temperature difference, provided it is expressed by a Fourier series; the solution given above then holds for every one of its terms.

This solution, based on the assumption that $\vartheta$ is
where

\[ u = u(x)e^{i\alpha x} + Be^{-i\alpha x}, \]

\[ \vartheta = Ce^{i\alpha z} + Dz e^{-i\alpha z}, \]

and similar expressions for \( u, v, \) and \( \omega. \)

The boundary conditions are

\[ z = 0, \quad w = 0, \quad \text{and} \quad \vartheta = M, \]

and for large values of \( z \) the circulation becomes negligibly small; then, the constants \( A, B, C, \) and \( D \) are fixed whereas \( a \) and \( b \) are determined by the day frequency and the other values given above.

The solutions for \( u \) and \( \vartheta \) are:

\[ u = \frac{rM}{(a^2 - b^2 i)} \left[ a e^{-i\alpha x} + b e^{i\alpha x} e^{i\alpha z} \right] e^{-i\alpha x}, \]

\[ \vartheta = \frac{rM}{(b^2 - a^2)} \left[ e^{-i\alpha x} - e^{-i\alpha z} \right] e^{i\alpha z} \sin \alpha z. \]

The factors \( r \) and \( \alpha \) can also be expressed by the constants given above. It should be noted that all these quantities, including \( a \) and \( b, \) are complex numbers and that with \( e^{i\alpha} \) a separation of the real and imaginary terms would require extensive calculations.

When the solution (7) is worked out for latitude \( \phi = 45^\circ \) and various values of the friction coefficient \( \sigma, \) it yields circulation systems of the land and sea breezes that are in full agreement with observations. Table III lists the basic factors of the circulations for different values of \( \sigma \) with and without consideration of the Coriolis parameter. As is usually the case, the phases are referred to a maximum land-water temperature difference at 1200. It can be seen from Table III that the height of the land or sea breeze lies at roughly 400 m and rises with increasing friction from 320 m to 500 m. It is interesting to note that this altitude is somewhat reduced under the effect of the Coriolis force. Naturally, friction diminishes the velocity of the land and sea breezes to a considerable extent, namely from about 5.5 m sec\(^{-1}\) to 2 m sec\(^{-1}\) for every centigrade degree of temperature difference. For average friction conditions and a maximum land-water temperature difference of 5C over the distance \( L/2, \) we arrive at a maximum sea-breeze intensity of about 10 m sec\(^{-1}\), which agrees quite well with observations. Under these conditions the vertical velocities reach maximum values of about 2 cm sec\(^{-1}\) per centigrade degree of temperature difference, which is also a reasonable value.

Of special interest is the phase of the land and sea breezes. If friction and the Coriolis force are neglected, the phase shift between the maximum temperature difference and the maximum intensity of the sea breeze is 4.7 hr. This shift decreases rapidly to 1.4 hr with increasing friction. In the case of \( \sigma = 2.5 \times 10^{-4} \) sec\(^{-1}\) the maximum intensity of the sea breeze follows the maximum temperature difference closely, as is borne out by most observations. The influence of the Coriolis force on the component perpendicular to the coast is small, as is to be expected. Naturally, a cross velocity \( v \) appears instead, whose phase shift with the velocity perpendicular to the coast is 10.9 hr, or nearly 12 hr. This phase shift stays constant up to the height of the zero layer and then changes sign. Figure 6 is a vector diagram for the case \( \sigma = 2.5 \times 10^{-4} \) sec\(^{-1}\) and \( f = 1.031 \times 10^{-4} \) sec\(^{-1}\) (\( \phi = 45^\circ \)). The figure shows clearly...
the oblique position of the flow ellipse relative to the perpendicular to the coast, in good agreement with observations (see Fig. 2).

Table III. Circulation Characteristics Computed for Various Values of Friction Coefficient and Coriolis Parameter

<table>
<thead>
<tr>
<th>$X \times 10^4$ (sec$^{-1}$)</th>
<th>$Y \times 10^4$ (sec$^{-1}$)</th>
<th>Phase (hr)</th>
<th>Amp. near surface (m sec$^{-1}$)</th>
<th>Phase (hr)</th>
<th>Amp. near surface (m sec$^{-1}$)</th>
<th>Alt. of zero layer (m)</th>
<th>Upper countercurrent</th>
<th>Max. amp. (m sec$^{-1}$)</th>
<th>Alt. of max. wind (m)</th>
<th>Max. amp. aloft (cm sec$^{-1}$)</th>
<th>Alt. of max. wind (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>0</td>
<td>12</td>
<td>$M$</td>
<td>16.7</td>
<td>5.43$M$</td>
<td>320</td>
<td>1.94$M$</td>
<td>650</td>
<td>16.6</td>
<td>4.21$M$</td>
<td>320</td>
</tr>
<tr>
<td>0.5</td>
<td>0</td>
<td>12</td>
<td>$M$</td>
<td>15.1</td>
<td>4.46$M$</td>
<td>340</td>
<td>1.11$M$</td>
<td>670</td>
<td>15.1</td>
<td>3.45$M$</td>
<td>340</td>
</tr>
<tr>
<td>1.0</td>
<td>0</td>
<td>12</td>
<td>$M$</td>
<td>14.1</td>
<td>3.68$M$</td>
<td>305</td>
<td>0.80$M$</td>
<td>700</td>
<td>14.1</td>
<td>2.86$M$</td>
<td>365</td>
</tr>
<tr>
<td>2.5</td>
<td>0</td>
<td>12</td>
<td>$M$</td>
<td>13.4</td>
<td>1.70$M$</td>
<td>500</td>
<td>0.26$M$</td>
<td>920</td>
<td>13.6</td>
<td>1.37$M$</td>
<td>500</td>
</tr>
<tr>
<td>2.5</td>
<td>1.031</td>
<td>12</td>
<td>$M$</td>
<td>13.1</td>
<td>1.84$M$</td>
<td>500*</td>
<td>0.25$M$</td>
<td>920*</td>
<td>13.1</td>
<td>1.76$M$</td>
<td>500*</td>
</tr>
</tbody>
</table>

*Approximate.

Above the zero layer there is a countercurrent whose vertical extent exceeds that of the land- and sea-breeze layer by a factor of 4 to 5. This countercurrent has a weak maximum that barely reaches a quarter of the intensity of the land and sea breezes near the surface. Theoretically, further circulations exist above this circulation cell of land and sea breezes and its countercurrent. However, their intensities are so low as to be negligible for all practical purposes. Thus, in contrast to others, this theory fixes a definite upper boundary to the land- and sea-breeze circulation. This upper boundary is, according to the theory, independent of the intensity of the land-water temperature contrast and a function only of the structure of the atmosphere, the Coriolis force, and the turbulent heat transfer. Friction tends to raise this boundary, while the Coriolis force tends to lower it.

MOUNTAIN AND VALLEY WINDS

The Mountain-Wind Circulation and Its Components. As in the case of coastal areas, local temperature and wind conditions occur in the vicinity of large mountain ranges that are often so strong in their effect that, locally, they modify or even obscure the general weather conditions. Here, thermal differences create a circulation system which, in daytime, consists of a lower current toward the mountains and an upper current in the opposite direction. In the region of the European Alps, Burger and Ekhart [12] have actually shown that this upper compensation current flows away radially from the mountains toward the neighboring plains in daytime. A corresponding flow system was found on the east slope of the Rocky Mountains by Wagner [71] and Ekhart [23]. This upper compensation flow is naturally less pronounced (speeds of the order of 15 cm sec$^{-1}$) than the lower current, since it is more strongly affected by the wind system determined by the general pressure distribution, the influence of which is difficult to separate from the local effect. Figure 7 shows these conditions schematically.

The Thermal Slope Wind. The difference in temperature between the air heated over the inclined mountain slopes and the air at the same altitude over the center of the valley causes the phenomenon of air rising in daytime along the slopes of moun-

![Fig. 6.—Theoretically calculated flow ellipse of the land and sea breeze under the influence of friction ($e = 2.5 \times 10^{-4}$ sec$^{-1}$) and the Coriolis force ($f = 2.2 \sin \phi = 1.03 \times 10^{-4}$ sec$^{-1}$; $\phi = 45^\circ$) when the maximum temperature difference between land and sea is at 1200 LMT. The vector diagram at the left shows the mean winds during the sea-breeze period (0730-1830 EST) at Logan Airport, Boston, Mass. (based on 40 cases). (After Defant [18].)
tains, well known to every mountain climber. These winds start one fourth to three fourths of an hour after sunrise, and blow uphill in daytime. They reach their greatest intensity at the time of maximum insolation and reverse their direction in the evening (about one fourth to three fourths of an hour after sunset). Because of the stronger insolation, they are especially well developed on the southern slopes and are weaker or almost nonexistent on the northern slopes. This wind prefers the ravines and gullies of the usually eroded slopes and is hardly noticeable on the projecting ridges. Numerous pilot-balloon observations in the upwind drafts on the slopes of the mountains north of the Inn valley near Innsbruck [43-45, 65] have clearly demonstrated the existence of such currents.

Here the thickness of the slope wind layer, as measured perpendicular to the slope, varies periodically with the wind intensity. Maximum values up to 260 m have been measured. However, as a rule, the thickness of the layer lies between 100 and 200 m. The thickness is less for the nocturnal downslope wind. The uphill wind continuously entrains, along its path, air from the space over the valley, so that the thickness of the affected layer is steadily increased in the direction of the uphill flow, and the layer becomes wedge-shaped. Naturally, the intensity of these slope winds varies greatly with the local differences in the slope and its exposure. Also, these winds can seldom be observed in their pure form and are often weakened or strengthened by extraneous wind conditions. On the average, the slope-wind speeds in the direction of the slope amount to about 2-4 m sec\(^{-1}\), according to measurements. Projecting parts of the slope cause a detachment of this current from the slope and thereby an increased vertical movement which can be utilized by soaring pilots to gain altitude. In fact, the entire phenomenon of thermal slope winds is of greatest importance in soaring. The existence of thermal slope winds is often indicated by isolated cumulus clouds over summits or chainlike cumulus formations along ridges. Velocities of 13 m sec\(^{-1}\) have been measured in these updrafts. The nocturnal downslope wind shows a lesser vertical extent and lower velocities.

The highest velocity of these slope winds does not occur close to the slope surface, but at a definite distance from it. Higher up, it rapidly decreases again or is supplanted by other wind conditions. The current is extremely sensitive to changes in the insolation, and a temporary shading of the slopes will cause an immediate response by the wind. Also the difference in the starting time of the current is determined by the varying time at which insolation begins on slopes of different exposure. Thus, the phenomenon is often weakened at the time of the maximum temperature because of shadows cast on the slopes by extensive cloud formations. Little is known about the thermal structure of the affected layer, because temperature measurements normal to the slope would be necessary for its actual determination. Such measurements have been made only up to about 20 m above the slope, since greater heights are difficult to reach. It is known that the layer of air adjacent to the slope shows strong superadiabatic gradients. During the cool downdraft at night, on the other hand, a strong increase of potential temperature (inversion) is often noticeable. The time span during the reversal of the wind is characterized by an isothermal air layer near the ground. At present, we have only a theoretical picture of the structure of the entire slope-wind layer, which, however, is probably very close to reality (see Figs. 11 and 12).

A variant of the thermal slope wind is the glacier wind. This wind, a shallow downflow along the icy surface of the glacier, continues all day regardless of insolation. Its thermal cause is the continuously present temperature difference between the glacier surface and the free air at the same altitude. For this reason, the glacier wind has no diurnal period as does the slope wind. Since the wind is a function of this temperature difference, it reaches its maximum intensity and greatest vertical extent (between 50 m and 400 m) in the early afternoon, according to measurements by Tollner [69] and Ekhart [21]. The glacier wind always appears, even on cloudy days. In daytime, it fades out soon after leaving the glacier because its kinetic energy is dispersed by the ground friction. The glacier wind often collides with the upvalley wind and then slides under it. At night, after leaving the glacier, it blends into the mountain wind which has the same direction. A characteristic of the glacier wind is its strongly turbulent flow which, in a more moderate form, is also a feature of the nocturnal slope wind.

The Mountain and Valley Winds. The phenomenon of the daily wind change along the axis of large valleys is known in all mountainous countries. In daytime, from about 0900 to 1000 until sunset, an upvalley, or so-called valley wind blows. At night an opposite downvalley or so-called mountain wind appears which continues into the early morning hours after sunrise. Numerous investigations of this phenomenon, including soundings of its vertical structure, have been made in many mountainous countries [19]. Mountain and valley winds are best developed in the wide and deep valleys of the Alps. The shape of the valley's cross section and the inclination of the valley bottom are of little influence on these winds. As a matter of fact, in fairly level valleys, such as the valley of the Inn River in Tirol, these winds are particularly well developed. They occur most frequently during persistent high-pressure situations in summer and are thus a typical fair-weather
phenomenon of summer. Nevertheless, mountain and valley winds may also occur on cloudy days and in winter when they become manifest mostly in a modification of the general wind. The vector diagram of the winds in the Inn river valley near Innsbruck in Fig. 8 shows a good example of mountain and valley winds.

![Vector diagram of the mountain and valley winds in the Inn valley (Innsbruck) on June 19, 1929, 100 m above the valley bottom. The dash-dotted lines indicate the axes of the Inn valley; upstream is to the left. Numbers indicate local mean time. (After Ekhart [19].)

This diagram of the wind conditions in the course of a clear day, June 19, 1929, shows that the maximum of the valley wind with speeds of 5-6 m sec\(^{-1}\) occurs shortly after 1500. The maximum of the mountain wind with speeds of 3-4 m sec\(^{-1}\) in the opposite direction occurs at around 0700. In general, the winds blow in the longitudinal direction of the valley's axis.

As an example of the height and intensity of the mountain and valley winds at Innsbruck, an isopleth diagram by Ekhart is reproduced in Fig. 9. The wind velocity distribution shows the characteristic transition from mountain to valley wind. The values are averages of several fair summer days in 1929 and 1931. According to the diagram, the wind maximum is to be found at an altitude of about 200-400 m, while the velocities near the ground are reduced because of the influence of friction. This increase of the valley wind with altitude is characteristic of all valley winds, as is the fact that their average velocities are higher than those of the nocturnal mountain wind. Furthermore, the diagram shows that the valley wind starts almost simultaneously in all layers up to relatively high altitudes. The starting time of the valley wind is closely related to the width of the valley and thus to the size of the mass of air involved. This starting time changes with the season, that is, with the magnitude of the diurnal temperature variations.

The height of the valley wind usually reaches to, or somewhat above, the flanking mountain ridges. The more stably stratified mountain wind, on the other hand, is confined to lower levels. The upper boundary of the mountain and valley wind system usually arches several hundred meters over the mountain crests. There, the transition into the general wind system usually takes the form of an abrupt wind shift or a calm.

The Maloja Wind. The Maloja wind, named after the watershed between Engadine and Bergell, Switzerland, where it appears particularly well developed, is a variation of the mountain and valley wind [4, 6, 7, 10, 35, 36, 39, 50, 58, 59, 72, 74]. It is a mountain wind which blows down valley both day and night. The phenomenon must be attributed to the fact that the valley wind of one valley reaches over a pass into another valley. In the mountain and valley wind system there, it acts as a disturbance, or, more specifically, as an abnormal development of the valley wind. The development of this anomaly is decisively determined by which one of the valleys involved has the larger diurnal temperature amplitude and thus effectively extends its circulation into the other valley across the pass. The question has not been satisfactorily answered whether this phenomenon can be entirely explained by the further increase of the diurnal temperature variation beyond the pass, or whether it is a purely inertial extension of that valley wind which is the more strongly developed. Sometimes, as was shown at the Arlberg Pass in Tirol [22], a strong upslope wind in one valley can, after crossing a pass, augment a valley wind in another valley.

Even in the absence of a pass, the interplay between thermal slope winds and typical mountain and valley winds sometimes produces considerable anomalies of the latter. Veering or backing of the valley wind on respective orographic slopes and cross circulation in narrow valleys, owing to strong insolation on one slope, have been observed. Cyclic wind shifts in the course of a day are caused by the fact that the upslope wind starts before the valley wind, and the downslope wind before the mountain wind.

The Theory of Mountain and Valley Winds. The development of the theory through many decades [15, 27, 33, 34, 50, 66, 74] finally culminated in the work by Wagner [71-73], who made an intensive study of the daily pressure and temperature variations in the free atmosphere within the valleys of the Alps. These investigations led to the result that at a certain altitude above the valley, usually at about the height of the surrounding ridges, the otherwise different diurnal pressure variations are completely equalized. This level of equalization was designated by Wagner as the "effective ridge altitude." A further important result of these investigations was establishment of the fact that the diurnal temperature variations in the valleys up to the...
height of the ridge were more than twice as large as the variations within a similar layer over the plain. As a consequence, a pressure gradient from the plain to the valley must exist during the day, whereas the reverse gradient must appear at night. The equalization of pressure differences at the effective ridge altitude requires that the pressure differences be largest close to the valley bottom and that they decrease with height so as to disappear completely at the effective ridge altitude.

From these deductions an adequate theoretical explanation of the mountain and valley winds can be derived. In full accord with observations, the unequal rates of decrease in the amplitude of the temperature variations over the plain and the valley bottom create a pressure gradient of the observed magnitude whose direction is toward the valley during the day and away from it at night. The air current thus generated fills the entire valley and decreases with altitude in accord with observations. Thus, there is a thermal cause for the pressure differences, and the air current is theoretically established as a circulation with a reversal of direction at night. This theory also includes the observed circulation between the plains and the mountains and the necessary existence of the upper compensation current. Furthermore, the pressure gradient, according to Wagner, is the result of the combined effects of the inclinations of valley bottom and ridge line. A nonparallel course of these inclinations, such as a downslope of the valley bottom and simultaneous upslope of the ridge line, would cause an extension of the existing gradient. In other words, if the ridge line rises beyond the pass altitude toward the neighboring valley, the gradient existing in the first valley would run in the same direction also in that neighboring valley beyond the pass. In this case the wind would reach over the pass as a Maloja wind. Thus, this abnormal phenomenon of the mountain winds also fits into Wagner's theory.

This theory furthermore requires two wind systems to preserve the stationary condition in the valley. One of them consists of the horizontal inflow of the valley wind, produced by the static conditions; the other one is the thermal slope wind system which takes care of the outflow over the flanking ridges. The combined wind systems have an additional effective source of energy in the heat given off by the heated slopes. The slope winds have a further important function in that they continuously add heated slope air to the air over the valley through that branch of the slope-wind circulation which descends over the valley center. Thus, the temperature in the center of the valley cross section is continuously increased during the day and decreased at night, when the heating conditions are reversed.

The mechanism interlocking the upslope and downslope winds with the mountain and valley winds in the course of a day, described in great detail by Wagner [73], is shown in Fig. 10. The great importance of the thermal slope winds as integral members of the larger circulation between plains and mountains is obvious. On the basis of a theoretical treatment of the stationary slope wind by Prandtl [62], the present author recently extended these considerations to the nonstationary case [17].

Prandtl's theory has particular significance, because a deeper insight into the mechanism of the thermal slope-wind current was gained through the introduction of turbulent heat conduction and turbulent friction. A disturbance of the temperature field, as well as static instability, always appears in the air layer above a heated surface and thus tends to establish turbulence.

Let us assume the spatial distribution of the potential temperature $\vartheta$, together with the temperature disturbance which stems from the heat transfer along the heated slope, in the form:

$$\vartheta = A + Bz + \vartheta'(n),$$  \hspace{1cm} (8)
where \( A \) is a constant at \( z = 0 \), \( B \) is the lapse rate of potential temperature, \( z \) is the vertical, \( n \) is the normal to the slope, and \( \theta' \) is the disturbance in potential temperature caused by heat conduction in the direction \( n \). We can then expect the velocity \( w \) of the stationary current along the slope to be a function of \( n \) only.

The interplay of heat transfer and heat conduction furnishes the differential equation,

\[
g \beta \theta' \sin \epsilon + \frac{\nu_k}{B \sin \epsilon} \frac{\partial^4 \theta'}{\partial n^4} = 0, \tag{9}
\]

where \( \beta \) is the coefficient of expansion, \( g \beta \theta' \sin \epsilon \) is the acceleration in the upslope direction, \( \epsilon \) is the angle of the slope, \( \nu \) is the coefficient of turbulent friction, and \( k \) is the coefficient of turbulent heat conduction.

The usual solutions of equation (9) are

\[
\theta' = C e^{-n/t} \cos \frac{n}{l},
\]

and

\[
w = C \sqrt{g \beta_k e^{-n/t} \sin \frac{n}{l}}, \tag{10}
\]

where

\[
l = \sqrt{\frac{4 \nu_k}{g \beta B \sin \epsilon}}.
\]

and \( C \) is the temperature disturbance at the surface of the slope. The actual form of the solution is shown schematically in Fig. 11.

![Schematic profiles (normal to the slope) of the wind speed \( w \) and temperature \( \theta' \). (After Prandtl [62].)](image)

I have checked the correctness of these theoretical results by a detailed comparison with actually measured average values of \( \theta' \) and \( w \) and have found a splendid agreement, except for the disturbing gradient influences in the upper layers. Furthermore, it is noteworthy that the resulting values of the austausch due to impulse transport (virtual friction) and of that due to transport of the heat content (virtual heat conduction) are of a plausible magnitude. Figure 12 presents the theoretical distribution of the potential temperature during upslope and downslope winds, respectively, for which no observations are available. The characteristics of the stratification become very apparent.

![Theoretical distribution of the potential temperature over a mountain slope during (a) upslope wind and (b) downslope wind. (After F. Defant [17].)](image)

The theory can be extended to include the oscillatory nature of the slope wind by simply multiplying, as a first approximation, the solutions (10) of the stationary case by the factor \( \cos \Omega t \). Because of the fourth root in the expression for \( l \), variations in the values of \( \nu \) and \( k \) [17, pp. 441-444] are of little influence on the solution. The average air transport by the slope winds can be roughly computed from the calculated and the observed average profiles of the slope wind. We can then estimate how long it would take until the rising heated slope air has replaced the air masses over the valley bottom. The interest of this question becomes apparent if we consider that this process of warming the air in the valley center has an effect on the pressure gradient and thus on the generation of the mountain and valley winds. Calculations show that a slab of air 1 m thick, 1500 m long (the width of the valley), and 1750 m...
Alps, especially in foehn, conspicuous local winds can be observed which blow from the mountain ranges down to the coast. In North America the foehn, known as the chinook, has a climatic significance over a more or less wide belt in the lee of the mountains. These local winds have become known under various names; however, today the general term foehn is used.

Although particularly well developed in the European Alps, especially in Switzerland and Tirol, such foehn winds blow also in Greenland from the inland ice over the mountain ranges down to the coast. In North America the foehn, known as the chinook, has a climatic influence on a very wide belt east of the Rocky Mountains. In Argentina, this wind is known as the zonda and blows down from the Andes. Local winds with foehn characteristics are also well known in Japan, New Zealand, and in eastern and central Asia. There is hardly a mountain range where the foehn is completely lacking. After all, the foehn will appear wherever prevailing winds must pass over a mountain barrier. Such barriers thereby become great divides of weather and climate.

The foehn is also noted for characteristic weather elements other than high temperature and low humidity. There is always extraordinarily good visibility; the mountains appear unnaturally close and clear and assume steel-blue to purplish hues. The clouds as forming and dissolving, because they remain stationary in spite of strong winds.

The foehn is a gusty wind, and temperature and humidity curves therefore always show irregularities to a varying degree during a foehn. In the Alps the foehn is always strongest where north-south valleys open into the plains or into large east-west cross valleys. This latter form is characteristic, for instance, at Innsbruck. It was there that the investigation of this phenomenon was most intensively pursued (see, for example, the work by Trubert, v. Ficker, A. Defant, Ekhart, and Pernter [14, 20, 24, 26, 60]).

During the winter, the great increase in temperature causes a rapid melting of the snow. However, floods seldom occur because the melting is very localized and decreases rapidly with altitude. Also, during a foehn, evaporation is very rapid because of the low relative humidity, and precipitation is confined to the passes, the peaks, and the windward side of the mountains. This distribution of precipitation is the chief characteristic of the foehn and has decisive climatic consequences for the areas on both sides of the range.

The thermodynamic explanation of the foehn is essentially due to Hann [32]. Since the theory belongs to the basic principles of theoretical meteorology, the reader is referred to pertinent textbooks. In general, it can be said that observations are in good agreement with Hann’s theory: On the windward side, cloud formation (particularly the stationary foehn wall) and precipitation must occur at a certain altitude as a result of the condensation. This removal of moisture together with the compression of the air during its descent on the lee must cause dissolution of clouds, warmth, and dryness there.

When air flows across a mountain range it is subjected to the above-mentioned foehn processes, as explained by Hann, and exhibits foehn characteristics on the lee side of the mountain. Such an air flow normal to the mountain ridge can be maintained only by a pressure gradient that is parallel to the ridge. Accordingly, such an air flow exists only when the general weather situation has a very definite pressure pattern, as shown schematically in Fig. 13. The sinusoidal deformation of the isobars, characteristically produced by the thermal pressure effect, forms a bulge of the isobars toward the high pressure on the lee side of the mountains and toward the low pressure on the windward side (often called the “foehn nose”).

Depending on the orientation of the mountains, the air currents, after passing through the thermodynamic process, show the foehn phenomena to a greater or lesser extent. In the case of mountain ranges oriented from west to east (e.g., the European Alps or the Pyrenees) a south wind will blow across the mountains if the low pressure is to the west and the high pressure to the east (see Fig. 13a). Because of its original warmth, the air from the south is well suited to the development of very marked foehn phenomena. For this reason, the south foehn is the most striking type of foehn from the viewpoint of the meteorologist as well as of the layman (see v. Ficker [28, pp. 25–37]). If low pressure lies to the east and high pressure to the west, the air from the north is transported across a mountain range oriented west-east (see Fig. 13b). The cold air piles up to the top of the mountains and then descends on the lee side while undergoing the thermodynamic process. This air, however, exhibits foehn properties only when the foehn process changes its cold-air character to such an extent that, upon arriving on the lee side of the mountains, it
is markedly warmer and drier than the air that was there before. This type of foehn, called north foehn [25; 40; 29, pp. 48-53], is relatively rare.

In the case of a north-south orientation of a mountain range, the development of a foehn requires low pressure to the north and high pressure to the south, or vice versa. Then, a west-east current passes across the mountains (e.g., the Rocky Mountains, the Andes, or those of Greenland, New Zealand, and Scandinavia) (see Fig. 13c). On the lee side, considerable temperature differences in the meridional direction occur; for example, in western Canada the broad warm foehn current, after passing the Rocky Mountains, meets the extraordinarily cold Canadian polar air. The result is increased cyclogenesis or rapid deepening of existing cyclones, which must be considered a consequence of the foehn process. This phenomenon, which is to a much lesser extent associated with the south or north foehn, has perhaps an analogy in the formation of the Genoa cyclone south of the Alps and the Apennines.

The thermal pressure effects during foehn cause in all cases a great increase of the horizontal pressure gradient in the direction of the mountains (as much as 9 mb per 100 km at sea level or at the level of the valleys). The mountain barrier prevents the transformation of such gradients into air motion; or, in reverse, the establishment of such gradients is made possible only by the mountain barrier. Although this gradient is less strong at the level of the mountain ridges, it cannot be explained by the thermal contrast alone.

The factors determining the various types of foehn are the general synoptic situation, the orientation of the mountain ranges, and the type of air masses passing over the mountains. In turn, the thermodynamic foehn process causes changes in the pressure field and in the properties of the air masses involved, which affect the general weather situation.

The warming and drying of air, observed in anticyclones, bear great similarity to the foehn and are caused by the dynamic heating during the descent of air from aloft. As with foehn, clouds dissolve and fair weather without precipitation sets in. In these cases, warm air is always present aloft as is evident from observations at mountain stations or from aerological soundings. This warm air rests on the cold air that lies in the valleys and in most cases shows up as a sharp inversion and decrease in humidity. The descending motion is very slow and must be ascribed to a divergent flow at the surface, directed away from the mountains. This phenomenon is called high foehn or free foehn in the literature [28, pp. 53-58]. Since an anticyclonic situation usually precedes the south foehn, the free foehn over the mountains often forebodes the subsequent development of a foehn current.

The warm foehn current on the lee of the mountains descends into the lowlands rather than ascends as one would expect. The cause of this descent into the valleys is a meteorological problem. Following older concepts, v. Ficker [28, pp. 34-37] has answered this question by resolving the foehn development into several phases (preliminary phase, anticyclonic phase, stationary foehn phase). A period of foehn is generally preceded by an anticyclonic weather situation as the preliminary phase, with very stable vertical temperature distribution. Cold air lies in the valleys, with dry warm air above it and separated from it by an anticyclonic subsidence inversion. Before the onset of the foehn, the cold air moves out of the valleys and away from the mountains under the influence of the pressure distribution. The upper boundary of the cold air is thereby lowered, and warm air from aloft supplants it. When this warm air reaches a station, the anticyclonic phase sets in for this station. The temperature rise and humidity drop connected with very little air motion indicate the anticyclonic foehn. Thus, the warm air does not actually break through the cold air to the valley floor, but follows the cold air which must first flow out of the valley into the plains while its upper boundary subsides. Only when the foehn wall forms on the windward side of the mountains and the horizontal pressure gradient is increased in the direction of the mountains, does the stationary foehn phase set in. In this phase, the foehn proceeds as explained by Hann.

The foehn does not always reach the valley floor as a warm current. If a remnant of shallow cold air remains in the lowlands, the foehn current moves over it. Occasionally, some cold air remains in certain parts of the valley, and the warm air reaches the valley floor only at certain foehn islands which are of great climatic significance. Also at night, when the intensity of the

![Fig. 13.—Windward and leeward effects of mountains on the isobaric pattern at sea level (schematic): (a) south foehn in the European Alps, (b) north foehn in the European Alps, and (c) southeast current over the Scandinavian mountains.](image-url)
foehn current diminishes and the cold air remnants in the valleys are augmented, these remnants may again combine into a shallow cold-air layer covering the entire valley floor. The foehn current then lifts from the valley floor, a fact that shows up in meteorological recordings as marked temperature and relative humidity jumps called *foehn pauses.* Figure 14 shows a particularly good example of the temperature and humidity records during the foehn period from February 2–5, 1904, at four alpine (Inn valley) stations at different altitudes. In this figure all features of the individual foehn phases, foehn pauses, and the characteristic response of the stations to the south foehn are evident.

![Figure 14](image)

**Fig. 14.** Typical south foehn registrations from the Innsbruck foehn area from February 2 to 5, 1904. Thin solid line—Innsbruck (573 m); dashed line—Igls (876 m); dotted line—Heiligwasser (1240 m); thick solid line—Patscherkofel (1970 m). (After v. Ficker [24].)

According to A. Defant, remnants of cold air in valleys form closed systems and may be excited into oscillations (similar to waves on lakes) by the passage of a foehn current over them. These oscillations show up as temperature fluctuations that are not to be confused with the short, periodic pressure fluctuations that occur during foehn.

Of special interest is the wave form of the air flow in the free atmosphere on the lee side of hill chains, mountain ridges, or other extended elevations of the ground. When the wave crests of this leeward current reach above the condensation level, the stationary wave becomes visible as a fixed "Moazagotl cloud" on the lee side of the mountains. The Moazagotl cloud consists of one or more cloud banks parallel to the mountain barrier; it forms continuously on the windward side of the wave and dissipates on the lee side. Sailplanes may reach great altitudes in the Moazagotl updraft. Prandtl, Küttner, and recently Lyra [55] have conducted theoretical investigations of these Moazagotl or foehn waves in the lee of mountains. In the development of the theory an increasing number of factors have been taken into consideration; Lyra's recent extension of the theory to include any polytropic stratification brought the theory of sinusoidal air currents on the lee side of barriers to a preliminary conclusion. Figure 15a shows the theoretically computed streamline pattern over a mountain barrier. Such foehn waves with their attendant cloud systems have been observed over numerous mountains. The waves caused by the Rocky Mountains in the northwestern United States have been investigated in detail by Hess and Wagner [38]. A particularly interesting case, in which the waves on the lee side reach an altitude of 40,000 ft, is shown in Fig. 15b.

![Figure 15](image)

**Fig. 15.** Stationary waves in the free atmosphere on the lee side of mountains. (a) Streamlines across a mountain range with Moazagotl clouds on the lee side. (After Lyra [55].) (b) Cross section from Tatoosh Island, Wash., (PTA) to Minneapolis, Minn., (MP) for 0900Z, January 14, 1945; isolines of potential temperature in degrees absolute. Vertical scale: units of pressure-altitude (ft) of U. S. standard atmosphere. (After Hess and Wagner [38]).

During well-developed foehn currents, marked physiological disturbances in men and animals occur on the lee side of mountains. These disturbances are ascribed to the short-period pressure fluctuations during foehn [28, pp. 65–105], but their causes have not been completely explained as yet.

**The Bora.** Hamm's theory, which includes all fallwind phenomena in the mountains, must also explain the cold fall-wind phenomenon. If a very cold air mass passes over a mountain barrier, or if it blows from the interior of a cold land mass over a plateau, the dynamic temperature rise during its descent on the lee side may not suffice to turn this fall wind into a warm foehn. Rather, this wind will be cold upon its arrival in the plain and is called *bora,* after the best-known of these local winds, which occurs on the coast of Dalmatia [48, 49, 57]. There, the cold, continental outbreaks of winter air from Russia find their way through Hungary and over the mountains of Dalmatia to drop over a
low, but relatively steep, slope to the otherwise warm coast of the Adriatic Sea.

The outstanding characteristic of the bora is its extraordinary violence, which often causes heavy damage. The cause of these stormlike fall winds is the liberation of potential energy when the cold air begins to drop to the sea. There are violent gusts of 50–60 m sec\(^{-1}\) called *refjuli*, and barometric fluctuations of 4 mm Hg are not rare in these cases. The bora does not extend far over the sea. However, it induces a heavy sea and atomizes the wave crests to such an extent that a cloud of mist (*fumarea*) is formed over the ocean. The bora of the Adriatic Sea has a pronounced diurnal period of intensity and frequency of occurrence. The maximum intensity occurs between 0700 and 0800, the minimum around 2400. It occurs most frequently about 0600 to 0700, most rarely about 1400.

We may distinguish between cyclonic and anticyclonic bora, according to the origin. The cyclonic bora is dependent on the existence of a low-pressure area over southern Adria, with an especially warm sirocco blowing from the south over the front portion of this low and aloft over the bora. Consequently, the sky is usually cloudy, and there is precipitation. The bora then blows more steadily and covers the entire Adriatic Sea. For an anticyclonic bora to develop, a strong high-pressure and aloft over the bora. For an anticyclonic bora to develop, a strong high-pressure area must exist over central Europe, with an extension of high pressure over Dalmatia which need not be opposed by a cyclone with closed isobars in the south. In that case the bora exhibits a violent character, but does not extend far out to sea (about 10 nautical miles). The sky remains clear, with the exception of a foehn wall over the mountains.

Under favorable conditions, well-developed local winds of bora character occur in many areas in the world. These conditions are (1) a short topographic drop, and (2) a sharp climatic division between a cold plateau and a warm plain so that, under suitable pressure conditions, cascading of the deeply chilled continental air can take place. Especially well known are the bora winds of Novorossiisk on the northern Caucasian shore of the Black Sea \([1, 3]\) and those of Novaya Zemlya \([75]\).

A. Defant \([16]\) has dealt theoretically with the question of the drainage of cold air along a slope. His method of attack, in which he bases a somewhat schematic model on the equations of motion and continuity in a doubly stratified atmosphere, allows a very good estimate of the influences of gravity and pressure gradient on this drainage of cold air masses. It is interesting to note that these two influences are of almost the same order; in other words, not only gravity, but also the pressure gradient, has an influence on the drainage of the cold air. If the steady current is disturbed, it may remain stable up to a critical slope angle, in which case small initial disturbances propagate as waves with a definite period and eventually fade out under the influence of friction. If the critical angle is exceeded, small initial disturbances grow exponentially with time, and the current assumes an unstable and turbulent character. A ground slope of 1:100 seems to be the critical value, fully confirmed by observation. An example of a stable current is the orderly, nocturnal outflow of the mountain wind through slightly inclined valleys; one of unstable flow is the extremely turbulent bora in Dalmatia.

**The Mistral and Jet-Effect Winds.** The bora-like local wind of Provence and the French coast of the Mediterranean up to Perpignan \([5, 29, 51]\) is called the *mistral*. It is a combination of a bora and a wind that is increased through the so-called jet effect. The mistral is due to the drainage of cold air initiated by a high-pressure ridge which frequently extends from the Azores to France, and by a permanent low-pressure area over the warm Gulf of the Lion. Its intensity is increased and its duration prolonged through the constriction of the geographic gate between the Pyrenees and the western Alps. If fresh polar or arctic air enters the Mediterranean basin in the rear of a cyclone that moves off to the east, the bora and the jet effect build up the mistral to a destructive intensity. It should be noted that it is not the shallow cold surface air of the plateau of central France, but the deep break-through of polar air into the western Mediterranean basin that favors the development of the mistral. Similar fall winds and jet-effect winds are caused on the Pacific coast of North America by the break-through of polar continental air; they are known as *northers* \([41]\).

The jet effect, that is, a purely local increase in wind intensity because of certain orographic configurations, is observed in many localities. The convergence of streamlines in a constricted path necessitates a substantial increase in wind speed. The pressure gradients responsible for the jet-effect wind extend only over very short horizontal distances and can be detected only by special investigations. A very detailed study of such conditions in the Vienna basin was made by Margules \([56]\). In that area, the air masses that stream through the gate between the Kahlenberg and the Bisamberg during a west wind display all the phenomena of the jet effect, and the corresponding pressure disturbances are clearly revealed by barograms.

**SURVEY OF DESIRABLE STUDIES OF LOCAL WINDS**

**Land and Sea Breezes.** Further intensive investigation of the vertical structure of land and sea breezes on especially suitable coasts by means of continuous aerological measurements over land and water at various distances from the shore would be most desirable. As regards the theoretical aspects, the problem appears adequately solved.

**Mountain and Valley Winds.** Accurate aerological cross sections along favorably situated slopes up to the ridge are still needed. Likewise, systematic upper-air soundings should be carried out in a particularly favorable valley location from its deepest recesses out into the plains. Special attention should be given here to the interrelation between the slope wind and the valley
The cell circulation for an inclined slope and the connection of the slope-wind circulation with the system of the mountain and valley winds still constitute a theoretical problem.

**Foehn Winds.** As an extension of the work of W. Schmidt [68], a comprehensive monograph on the occurrence of the foehn in all areas of the world, with special emphasis on its climatic importance, is needed. As far as the foehn theory is concerned, its thermodynamic side of the problem, which is a purely theoretical problem, should be paid special attention. As far as the foehn theory is concerned, its thermodynamic aspects appear to be completely explained. The dynamic side of the problem, which is a purely hydrodynamic problem, can be advanced only by special aerological investigations in the different mountain ranges of the world. Special attention should be paid to the problem of wave formation in the lee of the range.

**REFERENCES**

I. General references on the subject of local winds.


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