The generation of sea-breezes

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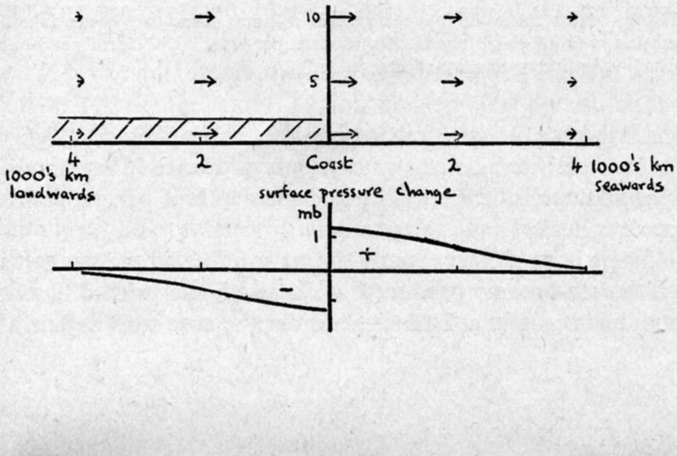
Introduction. The sea-breeze circulation

The reasons for the occurrence of sea-breezes on or near coasts on sunny summer days are well known, at least in broad outline. The land surface, being a poor conductor of heat, responds to the sun's rays during the morning by becoming hotter than the sea. This heat is taken up by the atmosphere, which, as the morning proceeds, becomes heated through a progressively deepening layer by convection currents. Cumulus clouds appearing during the morning over the land often provide visual evidence of this. The heating of the lower layer of the atmosphere over the land, when no such heating occurs over the sea, produces a general drift of air from land to sea, this drift extending through the whole depth of the atmosphere. The same effect would be produced if the land surface were raised about 10 m. This drift is characterised by a low velocity (10–20 cm per sec.) but a large horizontal extent, 3,000–5,000 km on either side of the coastline being affected; the disturbance is propagated from the coastline both landwards and seawards with the speed of sound, and is most easily observed by recording the surface pressure drop which travels inland during the heating period, resulting in the well-known 'heat-low'. These effects are represented schematically in fig. 1.

Fig. 1 – General drift seawards produced by either raising the land surface or heating the surface layers over the land. The accompanying surface pressure change is shown underneath



If the drift were brought about by lifting the land area the surface pressure jump across the coastline would be simply the pressure difference between the two different levels of the land and sea surfaces. When, however, the drift is produced by heating the air over the land, the pressure jump in the lower layers of the atmosphere near the coastline, as soon as it is produced by the general seawards drift, accelerates the surface air in the immediate vicinity of the coast from sea to land, this flow forming part of a general overturning motion of the lower layer (see fig. 2). (It is assumed for the moment that there is no general on-shore or off-shore wind.) This circulation differs in two important respects from the general drift described above; the velocities associated with it are much larger, up to 50 km per hour in magnitude, but its horizontal extent is very much smaller, affecting a distance of, at most, only 100 km from the coastline. Furthermore it is a *circulation* which is set up and not just a general flow in one direction only like the general drift. The whole of this circulation, with which is associated the observed sea-breeze, moves landwards at a velocity of approximately half the maximum surface wind associated with it. It may be regarded as a line-vortex with its axis parallel to the coast propagating landwards along the surface and can be compared in some ways with a thermal, which is also a propagating vortex but has a circle rather than a straight line for its axis and propagates upwards instead of horizontally.

The effects of off-shore and on-shore winds on sea-breeze development

Observers at stations on or just inland from the coast are familiar with the critical effect of the general wind on sea-breeze development, and know that the most pronounced 'frontal' sea-breezes occur on days when there has been a fairly higher wind component blowing off-shore during the morning. Its subsequent replacement by the sea-breeze

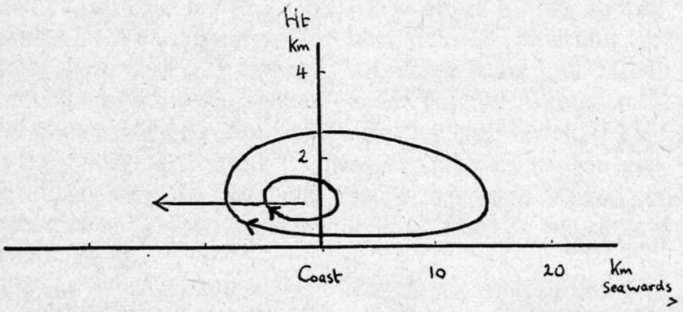


Fig. 2 – Circulation produced by inland heating of lower layers of atmos phere

during the late morning or afternoon is accompanied by a wind direction reversal, temperature drop, humidity rise and often a line of cloud of turbulent appearance. On the other hand, if the off-shore wind is too strong, no sea-breeze is observed. (The *total* wind may be quite strong, and a sea-breeze occur, provided it blows nearly parallel to the coast; it is the off-shore *component* that must not be too large.)

The main difference between the disturbances which develop with no general wind, and with an off-shore wind is that whereas in the first case the circulation immediately starts to spread inland, in the second case the circulation develops over the sea and, depending on the strength of the off-shore wind component, may or may not penetrate back over the coast against the general wind.

In order to understand the effect of the off-shore wind, consider the temperature distribution in the air over the sea after heating over the land has proceeded for a few hours. Fig. 3 illustrates the difference in heating of air columns over the sea for off-shore winds of 10 km per hour and 30 km per hour. The same diagram serves for the two cases if the horizontal distance scale is adjusted. There is no temperature increase of the air in the column at A which passed over the coastline before heating started. The column at B was warmed slightly in the lowest layers and then passed over the coast after which time no further heat was added. The column at C received rather more heat before passing over the coast, while the columns at D and E, still over the land, have received the full amount of heat available up to the time considered. The large scale drift through the full depth of the atmosphere takes place as in the case of no general wind and is only slightly modified; the distance of a few tens of kilometres through which the heated air has passed out to sea is small compared with the 2,000 km or so affected by the drift. But now the pressure gradient produced in the lowest layers by the large scale drift is distributed over the full distance between the columns A and D rather than over a narrow region near the coastline as with no off-shore wind. The circulation develops as before, but is correspondingly weaker since this pressure gradient is smaller. It will, however, be of much larger horizontal extent.

Remembering that this circulation is *superposed* on the originally undisturbed off-shore wind, it is readily appreciated that the occurrence of a sea-breeze at the coast (and further inland) will depend on whether the landward velocity of the air in the lower part of the circulation becomes greater than the off-shore wind during or soon after the heating period. If the off-shore wind is too strong or the heating too weak the sea-breeze will not occur because the columns become too widely spaced (e.g. as given by scale (b) in fig. 3) and the surface pressure gradient too weak to produce a circulation strong enough to reverse the flow in the lower layers. All that will happen will be that the surface wind will be reduced in strength over a region a few tens of kilometres wide some distance off-shore.

With an initial *on*-shore wind component the heated columns in fig. 3 will be distributed landwards instead of seawards and the situation becomes similar to that shown in fig. 3 if the coastline is transferred to between A and B. The elongated circulation will develop over the land and be indicated by a general strengthening of the surface wind over a strip a few tens of kilometres wide some distance inland.

Effect of atmospheric stability and surface friction on sea-breeze generation

As is well known, the height to which thermals penetrate depends not only on the amount of heating supplied at the land surface but also on the initial temperature distribution in the lower layers of the atmosphere. In fig. 4 there are shown two cases in which the same amount of heat is added,

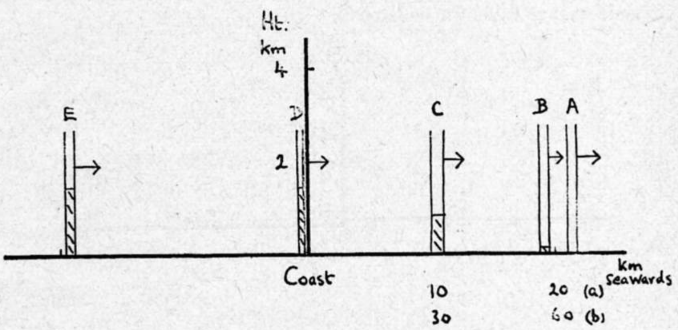


Fig. 3 – Heat taken up by vertical air columns after two hours inland heating with off-shore wind of (a) 10 km per hour, (b) 30 km per hour (shading indicates amount of heat added)

but whereas in the first where the atmosphere is highly stable, the thermals can extend only to about 2 km, in the second there is little stability and the depth of penetration is up to 3 km.

There is an almost equally dramatic effect on the strengths of the sea-breezes produced when airstreams with different stabilities cross a coastline. Consider the case with a general off-shore wind. The circulation produced in the more stable case after two or three hours heating is restricted to a shallower depth of atmosphere, the upper stable layers acting as a strong damping mechanism. On the other hand, when the stability is small, the circulation extends through a much greater depth of atmosphere. The depth of landward flowing air is much greater in the second case as is apparent

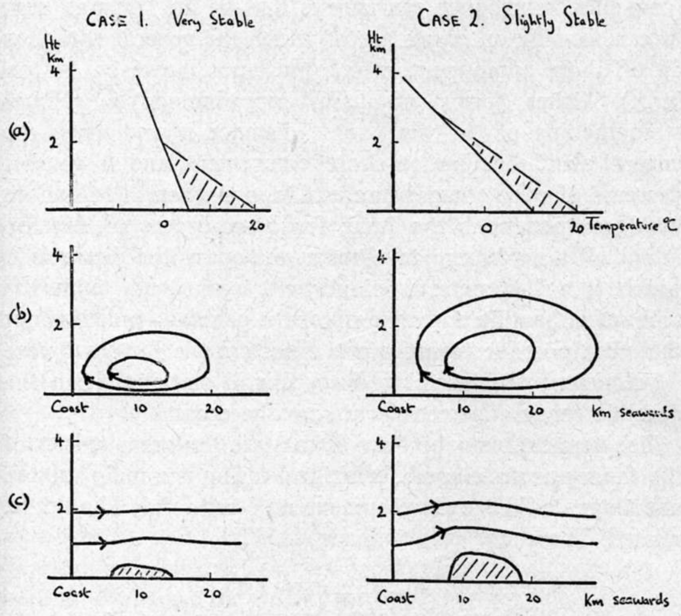


Fig. 4 – Effect of stability on depth of sea breeze.
(a) Temperature distributions inland. The shaded areas, representing total heating, are the same in the two cases.
(b) The sea-breeze circulations.
(c) The sea-breeze circulations superposed on a general off-shore flow to give streamlines of air motion. The air in the shaded region is moving towards the coast.

when the full flow patterns are drawn (fig. 4). Although the surface velocities are different in the two cases, the horizontal pressure gradient produced at the surface by the large-scale drift is the same, being determined almost completely by the total heating and being very little affected by the depth of convection.

Surface friction modifies the sea-breeze in the first 50 to 100 m above the earth's surface as it modifies any air flow moving over a rough surface. The main effect is to slow down the surface layers, and, because of the earth's rotation, also to introduce a wind component along the pressure gradient, i.e. parallel to the coast. This latter effect is treated more fully in the next paragraph. The main point considered here concerns the different degrees of modification of sea-breezes with off-shore and on-shore winds.

With an off-shore wind, surface friction, in retarding the surface layers, reinforces the circulation produced by the pressure gradient and does not start to oppose the on-shore surface flow until this has been established. On the other hand, with an initial off-shore wind, surface friction still further retards the landward flow and opposes the establishment of the sea-breeze circulation from the outset. Furthermore, since the surface drag is approximately proportional to the square of the surface wind speed, this effect increases rapidly as the circulation attempts to develop. This important difference seems to the author largely to account for the strong 'frontal' characteristics of sea-breezes associated with an initially off-shore wind, whereas such characteristics are absent from sea-breezes associated with initial on-shore winds. These differing effects are illustrated diagrammatically in fig. 5.

Effects of the earth's rotation; the wind component parallel to the coast

No mention has so far been made of the effect on the sea-breeze of the earth's rotation. To an observer in space at rest relative to the earth's axis of rotation any line on the earth's surface (not at the equator) moving round with the earth turns about a vertical axis. Because of this, winds recorded as steady by observers on the earth's surface represent air which is *accelerating* relative to the observer in space and this acceleration must be supplied by a horizontal pressure gradient. This is why the wind appears to blow mainly along rather than across the isobars, a steady wind requiring a pressure gradient perpendicular to its direction to maintain it. In the Northern Hemisphere, to an observer with his back to the wind high pressure is on the right. Thus, when a sea-breeze circulation is generated, for example with no initial wind and consequently no pressure gradient *along* the coast, the on-shore wind is not in equilibrium since there is no pressure gradient generated parallel to the coast. The air therefore starts to accelerate *parallel* to the coast towards the side where high pressure would be required to maintain a steady wind of the same strength. This means that in the Northern hemisphere the air starts to accelerate to the right. If initially there is a steady on-shore or off-shore wind and a corresponding pressure gradient along the coast, the same effect occurs in association with the sea-breeze circulation and may be regarded as superposed on the initial wind. Some typical surface wind distributions are shown in fig. 5. As is indicated in fig. 5 (d) all the above arguments hold if there is an initial wind component parallel to as well as normal to the coast.

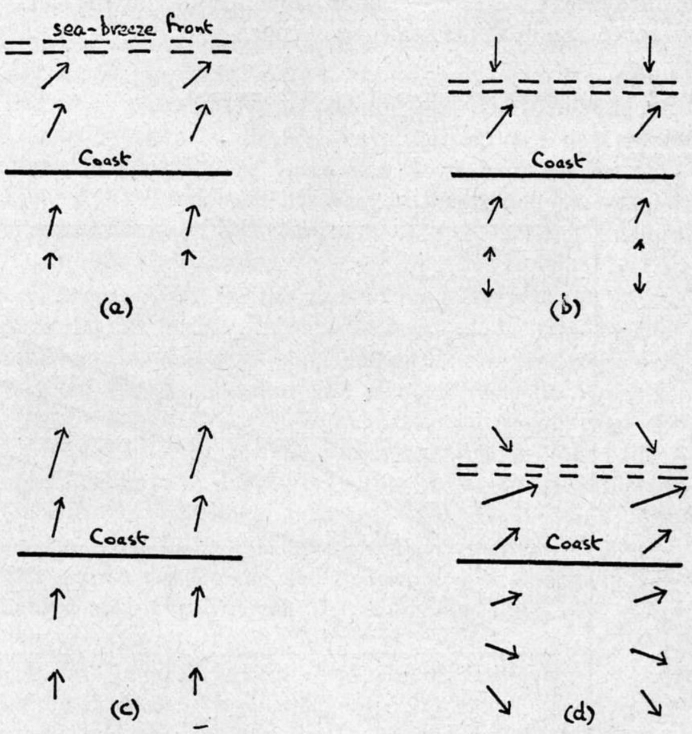


Fig. 5 – Surface wind speed and direction with (a) no initial wind (b) an initial off-shore wind (c) an initial on-shore wind (d) an initial wind with off-shore component and component parallel to coast.

Another important consequence of the earth's rotation is that as the flow becomes more parallel to the coast the spread of the sea-breeze inland becomes much less rapid and eventually ceases altogether. Sea-breezes are rarely observed more than 60 km inland in Europe. This effect also limits the speed towards the coast of the circulation which develops out to sea with a strong off-shore wind.

Surface cooling over the sea. Sea fog

In the discussion in the second paragraph it was assumed that the only heat transfer affecting the lower atmosphere was from the land surface and that this was distributed upwards by thermals. However, when there is an initial off-shore wind, before the sea-breeze commences warm air leaving the land is brought into contact with a cooler sea surface. The small scale eddies responsible for the surface drag in the first few tens of metres also cool the lower part of the atmosphere (the so-called friction layer). If in this cooling process the air temperature drops below its dew point a shallow layer of sea fog is formed, and subsequently speads inland with the sea-breeze. The formation of sea fog is likely to be assisted by a wind component *parallel* to the coast since this generates more eddies in the friction layer. Furthermore, a late sea-breeze is likely to bring in a deeper fog layer than an early one since the air will have spent more time over the sea and the friction layer grown correspondingly thicker.

Even if sea fog is not produced, the thin friction layer will be marked by a higher humidity than air which has remained over land during the heating period, and it is often this feature, rather than a temperature change, which, together with a wind change, is the more definite confirmation of the arrival of the sea-breeze.

The characteristic 'curtains' of cloud as observed by glider pilots (see e.g. Wallington, 1959) are most probably formed when the moist friction layer ascends in the convergence region at the sea-breeze front.

Conditions for the occurrence of a sea-breeze

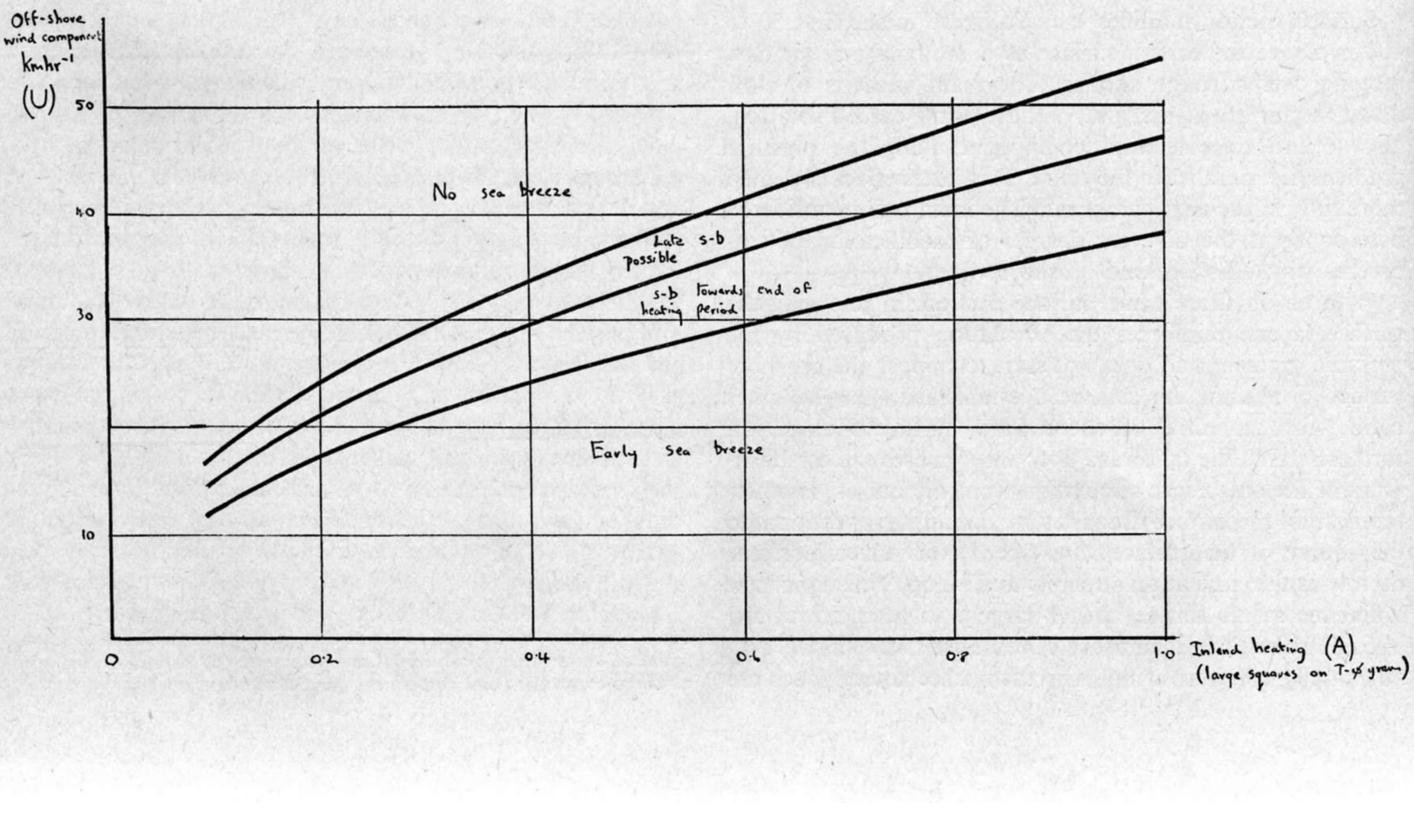
From the above discussion it is clear that whenever heat is supplied to the atmosphere from a land surface then in the neighbourhood of the coast the general flow pattern will be disturbed by this heating process on a horizontal scale of a few tens of kilometres, quite apart from smaller scale thermals distributing the heat vertically. The most spectacular form this disturbance can take is that of the 'frontal' sea-breeze, usually on coasts across which the general wind has an off-shore component (which must not, however, be too large). Other forms of disturbance include a general strengthening of the wind some distance inland when the general wind has an on-shore component and a general decrease of wind some distance off-shore when the off-shore wind component is too large for a sea-breeze to develop inland. If a sea-breeze develops at the coast, the distance to which it will penetrate inland will again vary with the amount of heating and strength of the general wind. It is not therefore possible to lay down exact criteria for sea-breeze development applicable to every inland station within the possible range of penetration from the coast.

The best that can be done is to state the main results of the theory as developed so far, indicating the main factors affecting sea-breeze development and how they should be related. These are

(1) Amount of inland heating (A).
(2) Off-shore wind component (U).
(3) Stability (indicated by vertical extent of thermals).

Regarding the stability as fixed, the character of the disturbance of the general flow which develops is found to depend on the ratio $\frac{gA}{U^2}$ where g is the acceleration due to gravity. This means that if, say, for a given value of A and U (and stability) a sea-breeze crosses the coast at midday and travels inland a distance of 20 km, then if A is increased to $1\frac{1}{2}$ times its former value, the same kind of sea-breeze will occur if U has approximately $1\frac{1}{4}$ times its former value.

Fig. 6 – Criterion for occurrence of sea-breeze on coast.



Since, however, all velocities are increased by 25 per cent, it will travel rather further inland, about 25 km. Hence a station 24 km inland will observe a sea-breeze on the second occasion, but not on the first.

The curves exhibited in fig. 6 are based on an approximate numerical computation and show under what conditions of heating and off-shore wind or sea-breeze can be expected to cross the coast. An average stability has been assumed with no highly stable layers below about 2 km, but not conditions for vigorous convection. As has already been mentioned, with a less stable atmosphere sea-breezes become more likely, but with an inversion in the first 1 or 2 km the sea-breeze is less likely to occur.

It is not possible to give similar curves based on numerical computations for stations inland, since the approximate method used does not enable the inland penetration to be calculated. It can be anticipated, however, that such curves must differ a little from those in fig. 6 since with light off-shore winds the circulation will not penetrate far inland with small heating rates and for a given station inland at, say, 20 km from the coast there will be a lower cut-off heating rate below which no sea-breeze will penetrate to the station. On the other hand, for larger values of U and quite high values of A the circulation which develops will be controlled considerably in its later stages by the effect of the earth's rotation and with subsequent weakening cannot penetrate far inland against the off-shore wind. The resultant critical curve for a station 20 km inland is thus likely to appear more flattened for merger values of A.

Conclusions

The purpose of this paper is to provide a general account of the physical mechanism of sea-breeze production and to emphasise that the sea-breeze is one of three or four possible forms taken by the disturbance of an airstream near the coast on a sunny day.

The numerical results quoted above are based largely on mathematical theory and are intended mainly as a guide to observers who would like to keep records for their particular locality and would like to know which are the important factors to record. Ideally these are the wind at ½ km (geostrophic wind) well inland, the amount of daytime heating from when the surface temperature reaches the sea temperature and the temperature distribution up to about 3 km.

These data is normally available at meteorological stations and can usually be obtained by arrangement with the meteorological officer. There is ample scope for local investigations of this kind.

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Zusammenfassung

Die Seebrise wird physikalisch erklärt durch das Zusammenwirken von Wind und Sonneneinstrahlung. Der Vorgang läuft folgendermassen ab:

Am Morgen, wenn sich das Land schnell durch die Sonne erhitzt (während das Meer sich kaum erwärmt), dehnt sich die erwärmte Atmosphäre vom Land über die See aus. Obgleich sich diese Störung mit Schallgeschwindigkeit mehrere tausend Kilometer weit ausbreitet, bewegt sich die Luft selbst nur mit etwa 20 bis 30 cm/sec. (Der Effekt ist der gleiche, als ob das Land um etwa 10 m gehoben würde.) Der Abfluss der Luft führt zu dem bekannten «Wärmetief» über Land. Die Folge davon ist eine Druckdifferenz über der Küste am Boden, die einen Druckgradienten von der See zum Land hervorruft (Fig. 1).

Die Luftmassen über dem Meer setzen sich sofort unter dem Einfluss dieses Druckgradienten in Richtung Land in Bewegung und formen eine Zirkulation über der Küste (Fig. 2). Im Gegensatz zu der kleinen Geschwindigkeit der anfänglichen Störung sind die Geschwindigkeiten dieser Seebrisenzirkulation hoch (bis zu 50 km/h), aber dafür ist die horizontale Ausdehnung viel kleiner, höchstens 100 km von der Küste. Der Wirbel bewegt sich etwa mit der halben Maximalgeschwindigkeit landeinwärts, wobei seine Horizontalachse parallel zur Küste bleibt. Dieses Strömungsbild gilt aber nur, wenn keine allgemeine Windströmung existiert.

Die Bildung der sogenannten «Seebrisenfront», die von englischen Segelfliegern häufig benutzt wird, hängt von der vorherrschenden Windrichtung ab und wird durch eine ziemlich starke Gegenströmung (vom Land zur See) begünstigt. Die Stärke dieser Gegenströmung und die Erhitzung des Landes bestimmen, ob die Seebrisenfront auf dem Meer liegen bleibt oder ins Land vordringt. Im Gegensatz zu Figur 2 entwickelt sich das Zirkulationssystem jetzt weiter draussen über dem Meer und erstreckt sich über eine grössere horizontale Ausdehnung.

Ebenso wichtig für die Entwicklung der Seebrise ist die Stabilität der Luftmassen über dem Land. Je nach der Stabilität wird sich die Erhitzung über eine kleinere oder grössere vertikale Strecke verteilen und demnach eine flachere oder höhere Seebrisenfront bewirken (Fig. 4). Wenn man die Vertikalzirkulation und die allgemeine Windströmung überlagert, erhält man das Bild der Fig. 4c, wo die schraffierten Gebiete die Seebrisenkomponente (See, Land, Wind) darstellen. Je nach der anfänglichen allgemeinen Windrichtung und -stärke werden die Seebrisenfront und Bodenwindverteilung nach Fig. 5 verschieden sein. (Das Land liegt oberhalb der «Coast»-Linie.)

In Fig. 5a herrscht kein Wind, und die Seebrisenfront liegt tief im Land. Bei Gegenströmung (5b) liegt sie näher an der Küste, und bei «Rückenwind» (5c) dringt die kühle Seeluft ohne Front tief ins Land vor. Eine Parallelkomponente des Windes zur Küste (5d) ändert das Bild wenig. Die Rechts-

drehung des Windes in allen vier Fällen der Fig. 5 (wenn man von der See nach dem Land fortschreitet) wird auf die Erdrotation zurückgeführt. Die zunehmende Rechtsdrehung des Windes führt schliesslich zu küstenparallelen Winden und verhindert das weitere Vordringen der Seebrisenfront, die selten tiefer als 60 km ins Land eindringt.

In jedem Falle muss die Seebrisenkomponente senkrecht zur Küste stärker sein als die entsprechende Gegenwindkomponente des vorherrschenden Windes, um der Seebrise das Vordringen ins Land zu ermöglichen. Hierbei spielt auch die Oberflächenreibung und ihre Änderung beim Übergang von See zu Land eine Rolle. Im allgemeinen verzögert die Reibung das Vordringen der Seebrise, verstärkt aber ihren frontalen Charakter.

Häufig ist mit der Seebrise Seenebel verbunden. Er bildet sich in der ursprünglich auf das kalte Meer vordringenden Landluft. Später, wenn diese Luft zurückkehrt, verdampft der Nebel über Land, aber die Feuchtigkeit bleibt in der Konvergenzlinie der Seebrisenfront sichtbar, wo sie die typischen «Vorhänge» formt, an denen der Segelflieger die Seebrise erkennen kann.

Die vorgehende Analyse zeigt, dass eine Vorhersage der Seebrise von drei Faktoren abhängt:

1. Über Land erzeugte Wärmemenge (A).
2. Vorherrschende Windkomponente vom Land zur See (U).
3. Stabilität (Höhe der Thermik).

Wenn die Stabilität gegeben ist, wird die Stärke der Seebrise durch die Grösse gA/U^2 bestimmt (g = Gravität). Wenn man A gegen U aufträgt (Fig. 6), zeigt sich dann, dass die Entwicklung der Seebrise durch eine Kurvenschar vorhergesagt werden kann, die die Fälle «keine Seebrise» (oben), «mögliche Seebrise», «späte Seebrise» und «frühe Seebrise» (unten) voneinander trennt. (Die Grösse A in Fig. 6 ist von einem Adiabaten-Diagramm genommen.) Es empfiehlt sich, den Wind in 500 m Höhe und die Erwärmung unter 3000 m Höhe zu benutzen. Die genaue Lage der Seebrisenfront lässt sich auf diese Weise noch nicht vorhersagen.

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