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3.6 Land/Sea Breeze

The land/sea breeze is a local, thermally direct circulation arising from differential heating between a body of water and the adjacent land. The circulation blows from the body of water (ocean, large lakes) toward land and is caused by hydrostatic pressure gradient forces related to the temperature contrast. Therefore, the land/sea breeze usually is present on relatively calm, sunny, summer days, and alternates with the oppositely directed, usually weaker, nighttime land breeze. As the sea breeze regime progresses, the wind develops a component parallel to the coast, owing to the Coriolis deflection. The leading edge of the sea breeze is called the sea breeze front.



Figure 1: Schematic showing diurnal change in land/sea breeze.

3.6.1 Historical References

The land/sea breeze is indirectly mentioned in the Bible and has been referenced in the scientific literature since Halley (1686), making it one of longest studied meteorological phenomenons.

Aristotle, *Problemata Physica*:

Why do the alternating winds blow? Is it for the same reason as causes the change of current in straits? For both sea and air are carried along until they flow then, when the land-winds encounter opposition and can no longer advance, because the source of their motion and impetus is not strong, they retire in a contrary direction.

Despite this long history, the traditional "convectional" theory of the land/sea breeze remains quite pertinent. This theory, as stated in Buchan (1860), stipulates that the land and sea breeze is caused when:

Land is heated to a much greater degree than the sea during the day, by which air resting on it being also heated, ascends, and the cooler air of the sea breeze flows in to supply its place. But during the night the temperature of the land and the air above it falls below that of the sea, and the air thus becoming heavier and denser flows over the sea as a land breeze.

3.6.2 Life Cycle

The sea breeze normally starts in the morning, a few hours after sunrise, when the solar radiation heats the boundary layer over land. A classical explanation for the development of a sea breeze is the "Upwards" Theory: The differential heating between land and sea leads to the development of a horizontal pressure gradient, which causes a flow from land towards sea. This flow is called a "return current", even though it may develop before the actual sea breeze. The mass divergence and resulting pressure fall over land and the convergence and pressure rise over the sea initiate the Sea-Breeze close to the surface.

The return current aloft carries the excess of air towards the sea. Cloud development frequently occurs in the ascending part of the circulation, while clouds tend to dissipate over the sea, where the air is sinking. When the air is very dry, as often is the case in spring and early summer, the cumulus clouds may not appear at all. In these cases the use of satellite imagery is clearly problematic for the detection of sea breezes, while it may still be detectable using other remote sensing means, such as sensitive weather radars.

In the afternoon, when the boundary layer heating over land is at its maximum, the sea breeze is normally at its most intense, and can penetrate tens of kilometers - in some cases, even over a hundred kilometers - inland. If the large-scale flow is weak, the direction of the sea breeze often veers with time. This is a result of the Coriolis force having an impact on the air current. Another factor influencing the wind direction along the coast is the regular existence of thermal lows over land in the afternoon. Later in the day, as solar radiation decreases, the sea breeze dies out, the thermals weaken and the cumuliform clouds gradually disappear.

3.6.3 Understanding the Land/Sea Breeze Circulation

Circulation is a measure of rotation within a fluid. Specifically, it is a scalar that provides a measure of rotation within a finite region of the fluid. The *Bjerknes' circulation theorem* is obtained by taking the line integral of the inviscid form of the momentum equations:

$$\frac{dC}{dt} = -\oint \frac{dp}{\rho} - 2\Omega \frac{dA_e}{dt} , \qquad (1)$$

where C is circulation, p is pressure, ρ is density, Ω is Earth's angular velocity, A_e is the area of the integral circuit projected onto the equatorial plane. The first term on the right-hand side represents *solenoidal* or *baroclinic* generation of circulation, while the second term on the right-hand side represents the change in circulation due to rotation of the Earth.



Figure 2: An example of Bjerknes circulation theorem to a mesoscale air mass boundary associated with a horizontal temperature gradient. [From Markowski and Richardson]

Consider the example in Fig. 2. Only the vertical segments of the loop contribute the line integral because the quasi-horizontal segments are along constant pressure surfaces. Thus, we can rewrite Eq.(1) as

$$\frac{dC}{dt} = -R_d \ln\left(\frac{p_b}{p_t}\right) \left(\overline{T_w} - \overline{T_c}\right) - 2\Omega \frac{dA_e}{dt}, \qquad (2)$$

where R_d is the dry-air gas constant, p_b and p_t are the pressure at the bottom and top of the integral circuit, respectively, and $\overline{T_w}$ and $\overline{T_c}$ are the mean virtual temperatures along the vertical segments of the integral circuit on the warm and cold sides of the circuit, respectively.

The mean tangential wind speed acceleration about the integral circuit is $dV_t/dt = (dC/dt)/P$, where $P \approx 2(L+H)$ is the perimeter of the circuit. Here, L and H are the length and mean vertical dimensions of the circuit. Taking realistic values of $p_b = 1000 \text{ mb}$, $p_t = 900 \text{ mb}$, $\overline{T_w} = 300 \text{ K}$, $\overline{T_c} = 290 \text{ K}$, H = 0.9 km, and L = 30 km, the tangential velocity acceleration is $4.89 \times 10^{-3} \text{ ms}^{-2}$. Thus, over the one hour period starting with $V_t = 0$, a tangential wind speed of 17.6 ms^{-1} is generated. In reality, this value will be smaller due to frictional forces.

If we neglect Coriolis $(2\Omega dA_e/dt = 0)$, then the maximum circulation tendency (|dC/dt|) will occur when the inland temperature reaches a minimum or maximum (or in other words when $|\overline{T_w} - \overline{T_c}|$ is maximized). Thus, |C| is maximized when dC/dt = 0, that is, when $\overline{T_w} = \overline{T_c}$. This would occur roughly a few hours after sunrise and sunset. Including Coriolis would of course change this behavior.

The contribution to dC/dt from the Coriolis force acts to oppose the baroclinic generation of circulation. As the land/sea breee develops, it gains a component parallel to the coast. Here the onshore branch would be deflected in the opposite direction as the offshore branch. The penetration of the land/sea breeze therefore increases with decreasing f. In other words, the along-shore wind components grow at the expense of the shore-normal components, which weakens the circulation.

Consider the case of a north–south oriented coastline in the northern hemisphere, with cold air to the west and warm air to the east. This results in a baroclinically-generated counterclockwise (if looking in the positive y direction) circulation about a vertical circuit normal to the coastline. Over time, the lower, offshore branch of the circulation develops a northerly wind component, while the upper, onshore branch develops a southerly wind component. This tips the circuit such that its projection onto the equatorial plane is increased. Thus, $dA_e/dt > 0$, which results in a negative contribution to the circulation tendency.

As a result of the Coriolis and baroclinic generation terms opposing one another, the strength of the land/sea breeze does not lag the diurnal cycle of the peak horizontal temperature gradient. Instead, the maximum land/sea breeze circulations are observed during the pre-sunrise and afternoon hours (*i.e.*, approximately when the maximum and minimum inland temperatures are observed).

3.6.4 Depiction of the Land/Sea Breeze on Satellite

In satellite images, a sea breeze is characterized by a cloud-free area along coastal land areas which protrudes inland. Further inland, away from the influence of sea breeze, cumuliform clouds are often present. The detection of this boundary separating the cloud-free and cumuliform clouds is possible using either visible (cumuliform clouds are seen as white pixel values) or infrared (white to grey pixel values) channels. Water vapor imagery doesn't generally allow the detection of a sea breeze, but in cases where there is deeper convection within the boundary (*e.g.*, sea breeze front), a line of convective cells may be seen. A sea breeze front may appear as an intensifying line of cumulus or cumulonimbus at the leading edge of the sea breeze.



3.7 Organization of Convection

While motion from individual eddies within the boundary layer are statistically random, larger convective scales within the boundary layer may exhibit highly organized behavior.



Figure 3: Vertical velocity perturbations for a shear-free (left panel) and shear-driven (right panel) convective boundary layer. [Adapted from Gibbs and Fedorovich 2014]

3.7.1 Cellular Convection

If wind shear if weak, convection tends to organize itself into hexagonal cellular patterns (see Fig. 3, left panel). This behavior is qualitatively similar to Rayleigh-Bénard (RB) convective cells (a regular pattern of natural convection cells that form in a plane horizontal layer of fluid that is heated from below). This type of convection may also be referred to as *free convection*. There is no formal theory on why convection prefers hexagonal patterns, though it is speculated that it is nature's way of choosing a configuration that minimizes the structure's perimeter-to-area ratio.

Cellular convection may be classified as either *open* or *closed*. Open cellular convection is characterized by walls of cloud surrounding open, cloudless regions that contain descending motion. Conversely, closed cellular convection is characterized by rings of clouds surrounding cloudy areas that contain ascending motion. Cellular convection is most common over oceans. Open cells tend to form over warm (relative to the overlying air) water, while closed cells tend to form over cold water. Cellular convection over land is almost exclusively open. Closed cells can form over land when the boundary layer is capped by extensive cloudiness. In these cases, radiative cooling at the cloud top is the likely factor. Studies have also shown that open cells tend to form when the large-scale vertical motion is directed downward, while closed cells tend to favor large-scale, upward-directed vertical motion.

Though qualitatively similar to laboratory experiments of RB convective cells, boundary layer cellular structures differ in a few ways. First, the atmospheric boundary layer is not capped by a rigid lid like is the case for RB convection. Second, the aspect (width-to-height) is approximately 3 for RB cells, while values approaching 30 have been observed in the atmosphere. Third, the boundary layer contains dynamical instabilities that can affect cellular organization beyond that expected from thermal instabilities alone. Finally, RB only yields open cells when liquids are used, and closed cells when gases are used. As previously stated, both cellular structures are observed in the atmosphere.

3.7.2 Horizontal Convective Rolls

If wind shear is relatively strong, convection in the boundary layer may organize in the form of counterrotating vortices, called horizontal convective rolls (HCRs; see Fig. 3, right panel). This type of convection may also be referred to as *forced convection*. HCRs span the depth of the boundary layer and have aspect ratios that generally range between 3 and 10, depending on the dominant forcing.



Figure 4: Schematic of HCRs in the boundary layer. [From Markowski and Richardson]

HCRs produce coherent perturbations in momentum, temperature, and moisture. This is due to the coherent updrafts and downdrafts associated with HCRs. The ascending and descending branches provide an efficient means of vertically transporting thermal energy, moisture and momentum across the convective boundary-layer.

These branches can locally destabilize a region of the atmosphere and lead to rapid horizontal variations in temperature and moisture (*i.e.*, CAPE). HCRs also modulate surface fluxes and the boundary layer depth. These rolls can also lead to cloud streets (Fig. 6) if there is enough boundary layer moisture. These clouds generally form in the ascending branch of the HCR.



Figure 5: Schematic of HCRs. [From http://www.zamg.ac.at]



Figure 6: Cloud streets caused by HCRs. [From Markowski and Richardson]

It has been suggested that HCRs form when a critical value of buoyancy flux is reached and are favored over cells when the

$$-\frac{z_i}{L} \lesssim 25,$$

where z_i is the depth of the boundary layer and L is the *Obukhov length*, given by

$$L = \frac{u_*^2}{\kappa \beta \theta_*}.$$

Here, $\kappa = 0.4$ is the von Kármán constant, $\beta = g/\overline{\theta}_v$, and u_* and θ_* are the turbulence velocity and turbulence temperature scales, respectively.

The Obukhov length, of order one to tens of meters, is the characteristic height scale of the dynamic sublayer. In other words, it represents the depth over which mechanical turbulence generated by vertical wind shear dominates buoyant production of turbulence. This parameter varies with stability as: L > 0 under stable conditions, L < 0 under unstable conditions, and $L \rightarrow \infty$ (undefined) under neutral conditions.

Diurnal behavior of $-z_i/L$

- Surface heating from early morning to midday leads to increases in z_i and $|\overline{w'\theta'}|$ (and thus $|\theta_*|$)
- Mixing during this time reduces vertical wind shear (and thus u_*)
- Combined, this leads to $-z_i/L$ increasing from early morning to midday
- At some point, $-z_i/L > 25$, which favors cellular or disorganized convection
- This diurnal trend is consistent with observations that have shown the early formation of HCRs which later transition to cellular/disorganized convection

However, several studies have shown no correlation between roll formation and $-z_i/L$, or other paramters involving mean shear, surface fluxes, or static stability. Possible explanations include non-linear interactions between roll and subroll scales and/or gravity waves excited by boundary layer convection or free-atmosphere processes.

Aircraft and other observations have been used to describe the general characteristics of HCRs

- A minimum wind speed of about 5.5 m/s is needed throughout the depth of the PBL for rolls to exist
- Rolls tend to be aligned with the ambient wind
- Directional shear is not required for rolls to exist
- Low-level shear is most important in comparison to shear throughout the depth of the PBL
- Rolls are the preferred convective mode (as opposed to more unorganized convection) when a critical surface heat flux has been achieved
- Roll axis can be up to 30 degrees different from geostrophic wind or shear vector
- The average ratio between roll wavelength and PBL depth is between 2.5 and 5

3.7.3 Interaction Between HCRs and the Sea Breeze



Figure 7: Schematic showing the interaction between HCRs and the sea breeze.

Because HCRs and the sea breeze both represent regions of locally enhanced lifting and convergence, their interaction can lead to convective initiation.

It is clear from the figure above that enhanced sea breeze front cloudiness can occur where the HCR updraft intersects the sea breeze front for at least two reasons: (1) the extra lifting provided by the HCR updraft itself; and (2) the enhanced convective instability in the air within the roll updrafts. Recall that the roll circulations cause variations in CAPE across the rolls, principally because the roll updrafts lift moisture from the lowest-levels. This makes air traveling along the roll updrafts more unstable and prone to convective instability, leading to enhanced frontal cumuli at the roll updraft intersection locales.

Note the absence of cloudiness at the downdraft intersection points. The spatial variability of lifting and cloudiness along the sea breeze front appears to be as much due to the influence of roll downdraft suppression as roll updraft enhancement. Once they develop, frontal cumuli establish subsidence on their flanks, and this coincides with the roll downdraft intersection points. Thus, there seems to be a positive convective feedback owing to cloud-scale circulations that maintains the along-frontal variability once that variability has been established. This occurs as long as the rolls are strong enough to create that variability.



Figure 8: Schematic diagram showing the interaction between the sea-breeze front and horizontal convective rolls The sea-breeze front is delineated by the heavy blue barbed line. The circulation along the leading portion of the sea-breeze front is shaded purple. The horizontal vorticity vectors associated with the counter-rotating roll circulations are also shown, as are clouds along the horizontal convective rolls and at the intersection points along the front. The shear vector and low-level winds are indicated with black and green arrows, respectively. [From Markowski and Richardson]

Detailed stages of interaction and convective initiation

Early Stage

Figure 9 spans the early stage from a numerical simulation. It depicts fields that have been averaged along the coastline (which actually resides at x = 45 km, well beyond the left edge of the sub-domain displayed.) Vertical velocity is shown (dashed contours represent descent), along with cloud water contours (in red) and the cool marine air boundary (in blue). The leading edge of the latter is the sea breeze front.



Figure 9: Early stage of HCR and sea breeze interaction.

Intermediate Stage

Figure 10 depicts the intermediate stage which encompasses the most dramatic SBF-HCR encounter. Note the roll located above x = 120 continues to intensify, spawning a deep and strong convective cloud above it. Note also that as the SBF approaches this strengthening roll, its own cloud is suppressed. The circulation associated with the roll updraft, its cloud, and the downdraft that had appeared in between, combined to push less favorable dry mid-tropospheric air into airstream that was flowing above the sea breeze front, and this choked off the sea breeze front cloud. By the last time shown, there is virtually no cloud remaining right above the sea breeze front.



Figure 10: Intermediate stage of HCR and sea breeze interaction.

Intermediate Stage

The final stage is shown in Fig. 11. As the sea breeze front continues to close in on the ever growing roll, several interesting things happen. First, the sea breeze front propagation speed increases. With the absence of condensation warming above the front, the pressure difference across the front increases, leading to faster movement. Second, the frontal updraft strengthens, a logical consequence of the faster propagation. This updraft intensification leads to the regeneration of convection directly above the front, and this convection is now deep and strong rather than shallow. By the time of the bottom panel there are two strong updrafts, separately spawned by the sea breeze front and HCR, hidden within a single cloud shield. The regenerated convection is so strong, releasing so much condensation warming, that the sea breeze front propagation speed slows dramatically around this time. Indeed, the frontal boundary in contact with the surface actually retrogrades during the last few time periods depicted in the figure.



Figure 11: Final stage of HCR and sea breeze interaction.