# <u>Mesoscale Meteorology: Sea, Lake, and Land Breeze Circulations</u>

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*Introduction*

Breeze-type circulations result from differential heating (daytime warming, nighttime cooling) of land and water surfaces. They are perhaps the best-known of a larger class of differential heating boundaries. For a given heating, the requisite differential heating primarily results from land both warming and cooling more rapidly than water due to their different specific heat capacities; other contributions arise from more heating going to evaporation and to greater depth over water. The daytime circulation is a *sea breeze* when the water body is an ocean or sea; *lake breeze* when the water body is a lake; or *river breeze* when the water body is a river. Unless stated otherwise, these notes use sea breeze to refer to sea, lake, and river breezes. The nighttime circulation is a *land breeze* independent of the water body type. The name describes *from* where the breeze blows; e.g., a sea breeze blows from water to land. Each are primarily warm-season phenomena, with land warmer than water during the day and cooler than water at night.

The horizontal pressure gradient magnitude across a sea breeze circulation is on the order of 1 hPa per 50 km; for smaller-scale water bodies, this magnitude is correspondingly smaller. Breeze circulations result from this horizontal pressure gradient and accompanying flow acceleration, as is described physically in more detail later in these notes. Sea breezes extend to a depth of 0.5–1 km with $\|\mathbf{v}\|$ 3–10 m s$^{-1}$ maximized above the frictional layer (~100–200 m above the surface). The return flow is deeper (~2–3 km deep) but weaker ($\|\mathbf{v}\|$ ~ 2–5 m s$^{-1}$). Sea breezes and their return flow develop nearly simultaneously. Land breezes are generally weaker than sea breezes. The typical width of the sea or land breeze's leading edge is ~100–250 m. A slope of ~1/10–20 is found along the breeze's leading edge; a short distance rearward, the slope decreases to ~1/100. Horizontal convergence along the breeze's leading edge is $O(10^{-3}$ to $10^{-4}$ s$^{-1})$, supporting upward vertical velocities in excess of 1–2 m s$^{-1}$ along the breeze's leading edge.

*Sea breeze fronts* separate a comparatively warm over-land air mass from an advancing air mass that originates over comparatively cold waters. Sea breeze fronts are similar to density currents: denser, cooler air rushes under and displaces upward warmer, less dense air. The same is true for land breeze fronts, except propagation is from land to water. Both sea and land breeze fronts are akin to synoptic-scale cold fronts, except on smaller spatiotemporal scales. However, it should be noted that land breeze fronts are *not* retreating sea breeze fronts; the two are distinct features. A sea breeze front may propagate a great distance inland during the day, where it dissipates shortly after sunset, with a subsequent but separate land breeze front developing shortly thereafter along the immediate shoreline as the land's temperature cools below that over water. Sea breeze front passage is accompanied by reduced temperature, increased relative humidity ($r_v$ rises and $r_{vs}$ decreases), a wind direction from the upwind water body, and sometimes increased wind speed. Land breeze front passage is similar, except with a wind direction from the upward landmass.

Sea and land breezes are important contributors to the warm-season thunderstorm climatology at tropical to subtropical latitudes, particularly near irregularly-shaped coastlines and peninsulas. At such latitudes, the over-land air mass is typically conditionally unstable with large surface-based

CAPE and minimal CIN, such that locally-enhanced ascent along the sea or land breeze front is sufficient to lift parcels to their LFCs. Thunderstorm formation is less common on higher latitude sea and land breeze fronts due to less supportive ambient thermodynamic air mass properties.

In the absence of appreciable synoptic-scale flow, sea breeze intensity is directly proportional to the cross-breeze temperature difference. This implies stronger sea breeze fronts in spring and early summer when the climatological land-water temperature contrast is largest. Despite this, thunderstorm formation along the sea breeze is most frequent when the ambient environment is most supportive of thunderstorms (e.g., a warm, moist ambient air mass with large surface-based CAPE), which generally occurs during meteorological summer in the United States. Underlying soil characteristics also control circulation intensity; e.g., breeze circulations are climatologically weaker over Louisiana than adjacent states since the southern part of the state is dominated by marshes with a similar heat capacity to water. Breeze circulations are also weaker in proximity to smaller, shallower water bodies that warm and cool more rapidly than their larger counterparts.

The intensity of a sea, lake, or river breeze is greatest near the immediate shoreline and weakens with inland extent for two reasons. First, the cross-breeze temperature difference is greatest near the shoreline and diminishes with inland extent as insolation and sensible heat fluxes act to warm the initially-cooler air mass after it departs from its source region. Second, the Coriolis force acts to deflect the wind into a shore-parallel direction with time and, by extension, inland extent. The dependence on the Coriolis force means that, all else being equal, sea, lake, and river breezes at lower latitudes will progress further inland than at higher latitudes. Further, these factors result in sea and land breezes having greatest intensity in the afternoon and predawn hours, with intensity remaining approximately constant thereafter until near sunset or sunrise, respectively.

Sea and land breeze intensity and propagation are controlled by the synoptic-scale flow (assumed in gradient wind balance) and stability. Onshore flow weakens convergence and ascent along the sea breeze front’s leading edge and reduces the land-water temperature difference across the sea breeze front due to the advection of water-cooled air over land. These factors result in a weaker sea breeze front, but one that propagates inland more rapidly. Land breeze fronts are similarly weaker for offshore flow. Sea breeze fronts are stronger, but propagate inland less rapidly, with offshore flow; the same is true with onshore flow for land breeze fronts. Both are stronger when the environmental lapse rate is steeper (i.e., static stability, proportional to the change in potential temperature with height, is low); both are weaker when the environmental lapse rate is smaller. This helps to explain why sea breeze fronts are stronger (and deeper) than land breeze fronts: stability is lower over land during the day, near peak heating, than over water at night.

*Conceptual Overview: Thickness Perspective*

A straightforward interpretation of the sea, lake, land, and other breeze circulations is obtained from consideration of the hypsometric equation, obtained by substitution of the ideal gas law into the hydrostatic equation and subsequent integration between two pressure surfaces $p_{bot}$ and $p_{top}$ (noting that we also substitute a layer-mean virtual temperature to take it out of the integration):

$$z_{top} - z_{bot} = \frac{R_d \overline{T_v}}{g} \ln\left(\frac{p_{bot}}{p_{top}}\right)$$

This equation describes the thickness (i.e., change in height) between two isobaric surfaces $p_{bot}$ and $p_{top}$ as being directly proportional to the layer-mean virtual temperature. Higher layer-mean virtual temperature results in greater thickness, implying lower $z_{bot}$ at $p_{bot}$ and higher $z_{top}$ at $p_{top}$.

Consider the case of a sea, lake, or river breeze. For a layer of limited (e.g., 100 hPa) depth near the surface, layer-mean virtual temperature and thickness are higher over land than over water. This results in horizontal height gradients on isobaric surfaces at the top and bottom of the layer, with lower heights over land at the bottom and over water at the top of the layer. The resulting horizontal pressure gradient force causes an acceleration toward land at the bottom and toward water at the top of the layer. The overturning circulation is completed by ascent along the leading edge of the sea breeze over land and by descent along the back edge of the sea breeze over water.

*Conceptual Overview: Circulation Perspective*

We start with the three-dimensional momentum equation, neglecting friction, applied on constant height surfaces:

$$\frac{d\mathbf{v}}{dt} = -2\vec{\Omega} \times \mathbf{v} - \frac{1}{\rho}\nabla p + \mathbf{g}$$

Here, **v** is the three-dimensional wind vector, **Ω** is the three-dimensional angular velocity vector, and **g** is the gravitational vector. All other terms have their conventional meaning.

Consider some closed area, such as a box or circle, with area *A*. We wish to take the line integral of the momentum equation over this closed area:

$$\oint \frac{d\mathbf{v}}{dt} \cdot d\mathbf{l} = -2\oint \left(\vec{\Omega} \times \mathbf{v}\right) \cdot d\mathbf{l} - \oint \left(\frac{1}{\rho}\nabla p\right) \cdot d\mathbf{l} + \oint \mathbf{g} \cdot d\mathbf{l}$$

Here, *d***l** is a line segment locally tangent to the perimeter of the closed area *A*.

Consider the left-hand side of this equation. From the chain rule, we can show that:

$$\frac{d}{dt}\left(\mathbf{v} \cdot d\mathbf{l}\right) = \frac{d\mathbf{v}}{dt} \cdot d\mathbf{l} + \mathbf{v} \cdot \frac{d\left(d\mathbf{l}\right)}{dt}$$

The second right-hand term represents a change in the change of position along the perimeter of the closed area following the motion – or a change in velocity. It can be written as:

$$\mathbf{v} \cdot \frac{d\left(d\mathbf{l}\right)}{dt} = \mathbf{v} \cdot d\mathbf{v} = \frac{1}{2} d\left(\mathbf{v}^2\right)$$

Consider the line integral of this around a closed region, where the start and end points along the closed area are identical: the value of $\mathbf{v}^2$ is equal at the start and end of the line integral. Thus, the positive and negative contributions in the line integral must cancel, such that is has a net value of zero. Thus,

$$\oint \frac{d\mathbf{v}}{dt} \cdot d\mathbf{l} = \oint \frac{d}{dt}(\mathbf{v} \cdot d\mathbf{l}) = \frac{dC}{dt}$$

Here $C = \oint \mathbf{v} \cdot d\mathbf{l}$, with $C$ being *circulation*, quantifying flow rotation around the closed area. $C$ equals the average vorticity in the closed area and is positive for counterclockwise rotation.

The first right-hand side term of the line-integrated form of the momentum equation describes the circulation contribution from the Earth's rotation. It can be shown that:

$$-2\oint (\vec{\Omega} \times \mathbf{v}) \cdot d\mathbf{l} = 2\Omega \frac{dA_e}{dt}$$

Note that Ω on the left-hand side is a vector and Ω on the right-hand side is a scalar (= 7.292 x $10^{-5}$ rad $s^{-1}$). $A_e$ is the projection of the closed region's area on a horizontal plane intersecting the Equator; it is equal to $A$ at the poles and equal to zero at the equator.

The second right-hand side term of the line-integrated form of the momentum equation is known as the solenoidal term. It describes circulation changes in a baroclinic atmosphere, where density (and temperature) varies on an isobaric surface. We can expand this term as follows:

$$-\oint \left( \frac{1}{\rho} \nabla p \right) \cdot d\mathbf{l} = -\oint \frac{1}{\rho} \left( \frac{\partial p}{\partial x} \mathbf{i} + \frac{\partial p}{\partial y} \mathbf{j} + \frac{\partial p}{\partial z} \mathbf{k} \right) \cdot (dx\mathbf{i} + dy\mathbf{j} + dz\mathbf{k})$$

Completing the dot product and simplifying the result, we obtain:

$$-\oint \frac{1}{\rho} \left( \frac{\partial p}{\partial x} dx + \frac{\partial p}{\partial y} dy + \frac{\partial p}{\partial z} dz \right) = -\oint \frac{dp}{\rho}$$

Finally, the third right-hand side term of the line-integrated form of the momentum equation is a gravitational term. Expanding this term and using the definition of the geopotential, we obtain:

$$\oint \mathbf{g} \cdot d\mathbf{l} = \oint (g\mathbf{k}) \cdot (dx\mathbf{i} + dy\mathbf{j} + dz\mathbf{k}) = \oint g dz = \oint d\Phi$$

For a closed path, where flow around that path starts and ends at a common location, there is no net change in geopotential; all positive and negative contributions cancel. Thus, this term is zero.

Consequently, the circulation equation is given by:

$$\frac{dC}{dt} = -\oint \frac{dp}{\rho} - 2\Omega \frac{dA_e}{dt}$$

This equation is *Bjerknes' circulation theorem* and describes the *change* in (relative) circulation following the motion: solenoidal term (the absolute circulation) minus the circulation associated with the Earth's rotation. We now wish to consider the contributions from each in isolation.

Consider first the solenoidal term; i.e.,

$$\frac{dC}{dt} \approx -\oint \frac{dp}{\rho}$$

Substituting with the ideal gas law, where $p = \rho R_d T_v$, we obtain:

$$\frac{dC}{dt} \approx -\oint R_d T_v \frac{dp}{p} = -\oint R_d T_v d(\ln p)$$

Let us consider a vertical circulation with top and bottom surfaces found along isobaric surfaces at the top and bottom of the sea breeze circulation. The line integral along these segments is zero by definition; $d(\ln p)$ is zero for constant $p$. The vertical segments of this circulation are found on opposite sides of the sea breeze circulation equidistant from the sea breeze front's leading edge; we will assume ascending (descending) motion along the warm (cold) side of the front.

On the warm side of the front, the line integral is equivalent to a traditional integral from $p_{bot}$ to $p_{top}$, i.e.,

$$\frac{dC_{up}}{dt} = -\int_{p_{bot}}^{p_{top}} R_d T_v d(\ln p)$$

Letting $T_v$ be approximated by a layer-mean virtual temperature, this equation becomes:

$$\frac{dC_{up}}{dt} = R_d \overline{T}_{v\,warm} \ln\left(\frac{p_{bot}}{p_{top}}\right)$$

Similarly, on the cold side of the front, the line integral is equivalent to a traditional integral from $p_{top}$ to $p_{bot}$ (since we start at the top of the layer and integrate downward along the line segment), i.e.,

$$\frac{dC_{down}}{dt} = -\int_{p_{top}}^{p_{bot}} R_d T_v d(\ln p) = -R_d \overline{T}_{v\,cold} \ln\left(\frac{p_{bot}}{p_{top}}\right)$$

The total circulation change is the sum of these two components, i.e.,

$$\frac{dC}{dt} = R_d \ln\left(\frac{p_{bot}}{p_{top}}\right)\left(\overline{T}_{v\,warm} - \overline{T}_{v\,cold}\right)$$

Because $\overline{T}_{v\,warm} > \overline{T}_{v\,cold}$, the net circulation change is positive, resulting in stronger ascent in the warm air and stronger descent in the cold air. For fixed $p_{bot}$ and $p_{top}$, given observations of the layer-mean virtual temperature on either side of the sea breeze front one can compute the change in circulation due to the solenoidal term. Because circulation is the line integral of velocity about the perimeter of the closed region, the resulting change in velocity around the closed region can be computed by dividing the circulation change by the closed region's perimeter. Note, however, that both circulation and velocity changes computed in this manner neglect Coriolis and friction.

Consider next the term related to the Earth's rotation, i.e.,

$$\frac{dC}{dt} \approx -2\Omega \frac{dA_e}{dt}$$

The only contribution to this term results from the change following the motion in the projection of the closed region's area onto a horizontal plane intersecting the Equator. Initially, $A_e$ is ~0; a vertical plane of limited depth has negligible projection onto such a horizontal plane. However, as time progresses, the Coriolis force results in the wind being deflected to the right at both the top and bottom of the sea breeze circulation (particularly at higher latitudes). Consider the case the Lake Michigan lake breeze. Start at the Milwaukee shoreline. The initial vertical circulation is entirely in the *x*-*z* plane. As the easterly near-surface flow proceeds inland, it is deflected to the right; conversely, as the westerly near-surface return flow aloft evolves, it is also deflected to the right. This results in a circulation with a component in the horizontal (*x*-*y*) plane; i.e., Milwaukee to Menomonee Falls near the surface, Menomonee Falls back to Milwaukee in the return branch.

Over time, this increases the projection of the closed region's area onto the horizontal plane that intersects the Equator. Because of the leading negative, this counteracts the solenoidal term and results in a smaller increase in circulation following the flow. Consequently, sea and land breeze front intensity is largest in the afternoon and predawn hours; later, the expected circulation increase from continued warmer layer-mean virtual temperature over land for the sea breeze is partially offset by the Coriolis force. Because of the Earth's curvature, this reduction is smaller at low latitudes; the distortion of the closed region into the *x*-*y* plane is smaller, and the projection of this area to the horizontal plane intersecting the Equator is also smaller.