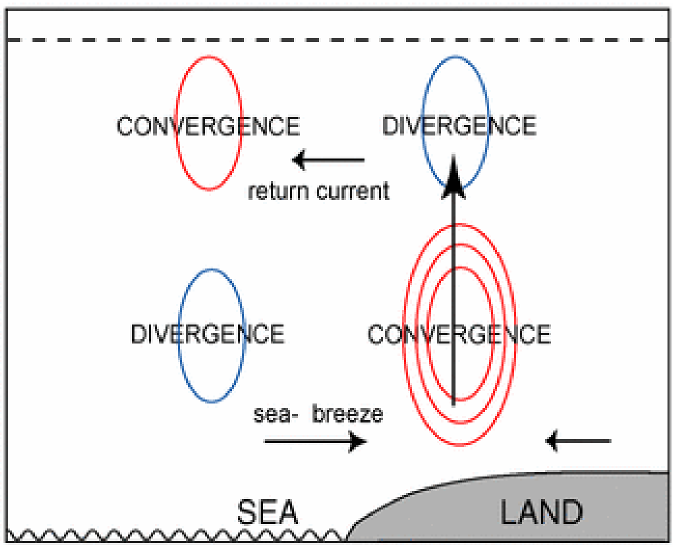
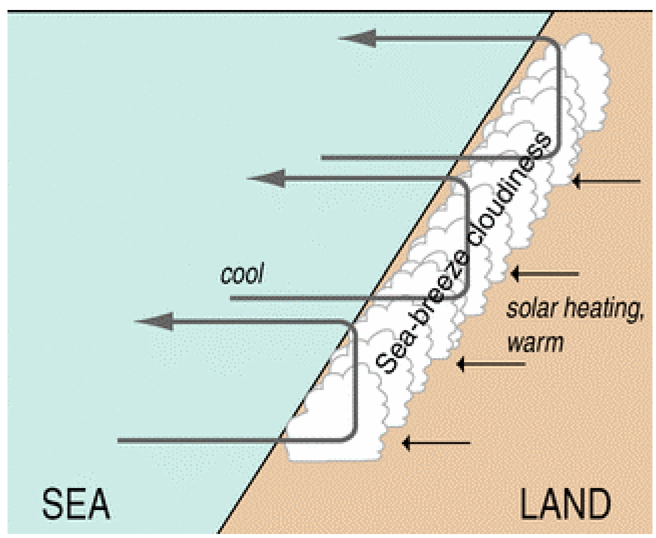
Sea and Land Breezes
METR 4433, Mesoscale Meteorology
Spring 2006
(some of the material in this section came from ZMAG)

Definitions: The 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 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.

The basic structure of the sea breeze is shown below:

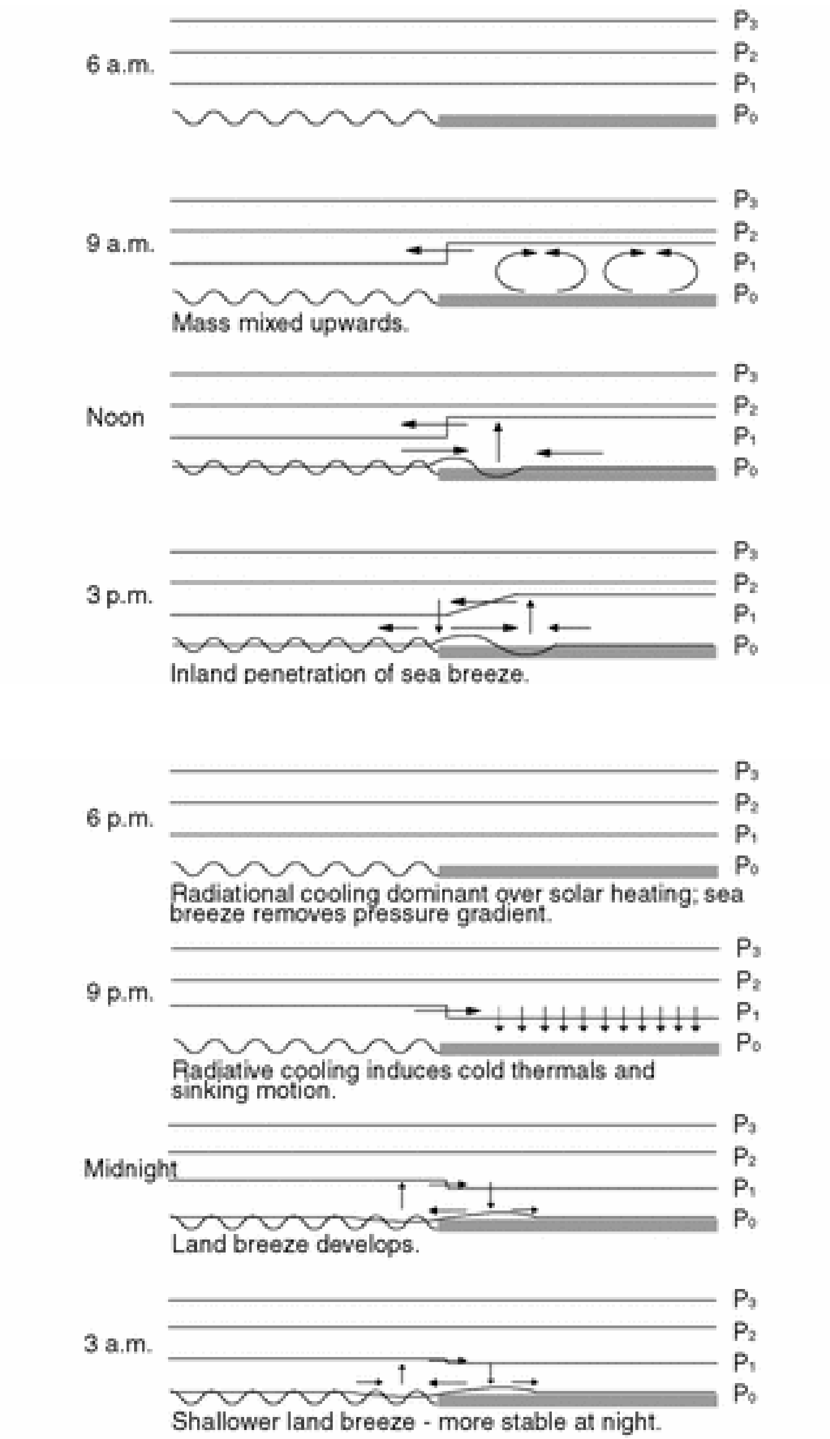




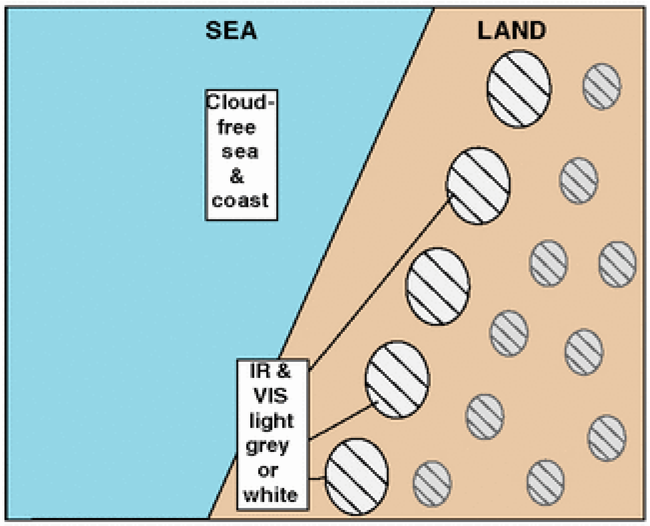
Life Cycle. The sea breeze normally starts in the morning, a few hours after sunrise, when the solar radiation heats the boundary layer over land. A classical explanation for the development of a sea breeze is the "Upwards" Theory: The differential heating between land and sea leads to the development 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

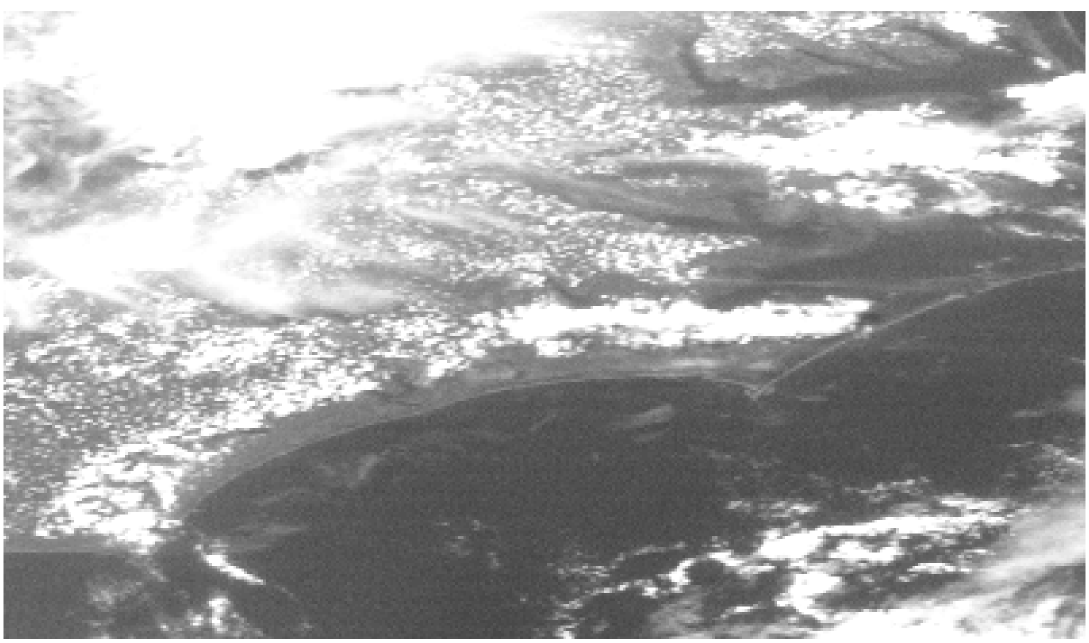
The return current aloft carries the excess of air towards the sea. Cloud development frequently occurs in the ascending part of the circulation, while clouds tend to dissipate over the sea, where the air is sinking. When the air is very dry, as often is the case in spring and early summer, the cumulus 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.



Depiction 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 VIS (cumuliform clouds are seen as white pixel values) or IR (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 can be seen. A sea breeze front (the narrow zone separating the air over the land from the air streaming from the sea) may appear as an intensifying line of Cu or Cbs at the leading edge of the sea breeze.





The sea breeze often behaves like a gravity current, which is a mass of dense air that propagates into the ambient air by virtue of the hydrostatic pressure gradient between them. Thunderstorm cold-air outflows also behave like density currents and the rotor-type circulation at their leading edge is due to pressure-density solenoids (see below) between the dense and less dense air. As will be discussed later, points of intersection between the sea breeze front and horizontal convective rolls represent regions favorable for the initiation of convection.

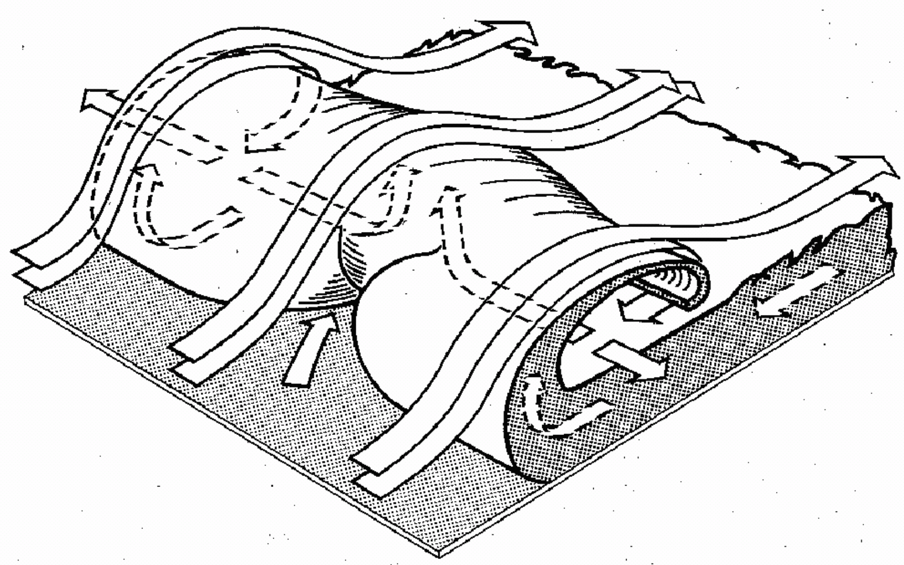


FIG. 64. Schematic representation of the flow at a gravity current head. (After Simpson *et al.*, 1977.)

Dynamics. The general physical mechanisms of the sea breeze have been known for quite some time and can be explained by pressure-density solenoids associated with land-water surface inhomogeneities. Because of the strong heating of the land the mean temperature of the air above the sea is lower than the temperature of the air above the land (see figure below). Thus, if the pressure distribution at the surface is practically uniform, the isobaric surfaces somewhat higher in the atmosphere are inclined in such a way that they are directed downwards towards the sea. However, the surfaces of equal density (called isosteric surfaces), are inclined in the opposite direction, i.e. towards land (because of the higher temperature). This means that the isobaric and isosteric surfaces do not coincide and this results in the generation of circulation or vorticity.

To estimate the acceleration due to the intersection of the isobaric and isosteric surfaces, we have to recall the definition of circulation (see Chapter 4 in Holton):

$$C \equiv \oint \mathbf{U} \cdot d\mathbf{l}$$

where l is the position vector and the integral is taken over a closed contour C. We obtain the circulation theorem by taking this same line integral over the vector equation of motion in an absolute coordinate system:

$$\oint \frac{D_a \mathbf{U}_a}{Dt} \bullet d\mathbf{l} = -\oint \frac{\nabla p \bullet d\mathbf{l}}{\rho} - \oint \nabla \Phi \bullet d\mathbf{l}$$

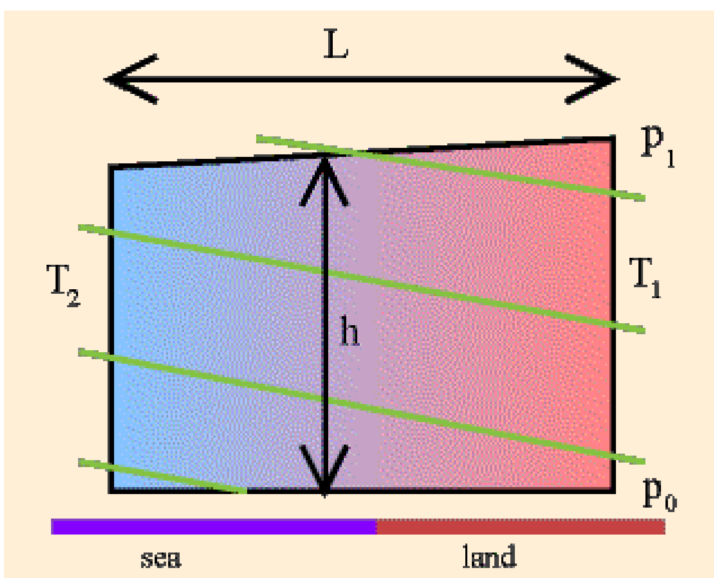
Owing to the definition of the geopotential, the last term vanishes. Also, using a vector identity for the LHS, we end up with

$$\frac{DC_a}{Dt} = \frac{D}{Dt}\oint \mathbf{U}_a \bullet d\mathbf{l} = -\oint \rho^{-1} dp$$

where the term on the RHS is the so-called solenoidal term (zero for a barotropic fluid). Stokes' theorem can be used to rewrite the RHS in terms of the gradients of specific volume and pressure, which represent solenoids (see p. 102 in Holton):

$$\iint_A -\left(\vec{\nabla}\alpha \times \vec{\nabla} p\right)\cdot \vec{k}\, dA = \oint_\Gamma \alpha\, dp$$

Thus, whenever the gradients of density (green lines in figure below) and pressure (solid lines) do not align, as in the vertical cross section of a sea breeze shown below, a horizontal circulation can develop.



To determine the magnitude of this circulation, we rewrite the solenoidal term using the ideal gas law

$$\frac{DC_a}{Dt} = -\oint RT d \ln p$$

Evaluation of this integral along the closed contour (bounding box in the figure shown above) indicates that there is only a contribution of the vertical segments of the contour. This is because the horizontal segments are taken in isobaric planes and the inclination of the isosteric surfaces relative to the isobaric surfaces is neglected. Thus the resulting increase in circulation is:

$$\frac{DC_a}{Dt} = R \ln\left(\frac{p_0}{p_1}\right)(\overline{T}_2 - \overline{T}_1) > 0$$

If <v> is the mean tangential velocity along the contour, then it follows:

$$\frac{D\langle v\rangle}{Dt} = \frac{R \ln(p_0/p_1)}{2(h+L)}(\overline{T}_2 - \overline{T}_1)$$

Taking the following realistic values: P_0=1000 hPa, P_1=900 hPa, T_2-T_1=10 K, L=100 km, h=1 km, results in a rate of change of velocity of about 1.5 x 10^{-3} m s^{-2}. Starting with V=0, this means that after about one hour the velocity will have increased to about 10 m/s. In reality, friction will also increase strongly with increasing wind speed and a more or less stationary situation will develop.