Mesoscale Meteorology: Lake-Effect Precipitation

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Introduction

As relatively cold air passes over a relatively warm body of water, taken generally here as a lake, sensible and latent heat fluxes are directed from the lake toward the atmosphere. This warms and moistens the air compared to air which does not pass over the lake. The sensible heat flux also triggers planetary boundary layer turbulence, resulting in (quasi-)mixed layer formation near the surface. If surface heat fluxes can sufficiently warm and moisten near-surface air and turbulence can build a sufficiently deep mixed layer, a modest amount of surface-based instability (CAPE) can be generated. The release of this instability brought about by some lifting mechanism such as lower tropospheric meso- to synoptic-scale convergence and/or friction differences between lake and land can result in moist convection formation. This moist convection is known as *lake-effect precipitation*.

Lake-effect precipitation is most common downwind of medium- to large-size mid-latitude water bodies. The size criterion is related to the need for the near-surface air to be sufficiently warmed and moistened; the longer its residence time over a comparatively warm water body, the greater the likelihood that such warming and moistening occurs. The latitude criterion is empirical; lake-effect precipitation can occur at any latitude so long as a sufficiently cold air mass passes over a sufficiently warm water body. This is most common at mid-latitudes, where the water bodies are typically unfrozen during the period when relatively cold air passes over them; as the latent heat of sublimation is much less than that of evaporation, freezing significantly reduces the latent heat flux. In addition, larger (and deeper) water bodies tend to stay unfrozen for a longer time than their smaller counterparts due to the greater heat contained within them to greater depths.

The seasonal cycle exerts a significant control on when lake-effect precipitation typically occurs. Lake-effect precipitation is most common in late fall through mid-winter when lake temperatures are relatively high and air temperatures (with equatorward-directed continental polar air masses) are comparatively cold. In late winter and early spring, lake temperatures are at their annual lows while air temperatures begin to warm; here, the high specific heat capacity of water is important as it ensures lake temperatures both cool and warm much more slowly than the air or dry ground. On shorter time scales, lake-effect precipitation typically occurs in the immediate wake of a cold front, where the requisite antecedent air mass is most likely to be present, and thus in proximity (in time and space) to its companion middle- to upper-tropospheric trough. The times of the year during which lake-effect precipitation is most frequent lead to most such precipitation as snow.

Necessary Ingredients: Thermodynamics

Lake-effect precipitation requires the presence of surface-based instability and some mean(s) by which it can be released. An initially cold, dry air mass located beneath a cold frontal inversion is strongly negatively buoyant. It is only through warming and moistening of this air as it passes over a relatively warm lake that surface-based instability may be generated.

Recall that the surface sensible and latent heat fluxes can be expressed as:

$$Q_h = -\rho c_p c_h \left\| \overline{\mathbf{v}_{10m}} \right\| \left(\overline{T_{2m}} - \overline{T_{sfc}} \right)$$

$$Q_e = -\rho l_v c_e \left\| \overline{\mathbf{v}_{10m}} \right\| \left(\overline{r_{v-2m}} - \overline{r_{v-sfc}} \right)$$

Each are functions of the difference in their respective quantities – temperature for sensible heat, water vapor mixing ratio for latent heat – between the air just above ground-level and the surface itself (here, taken as some body of water). Positive sensible and latent heat fluxes result when the differences are negative; i.e., the water body is warmer or moister than the air above. Fluxes are larger for faster wind speed or greater air-sea disequilibrium, though lake-effect precipitation is more likely when the upstream air mass is comparatively moist (less lake modification necessary to sufficiently moisten the air). Surface sensible and latent heat fluxes warm the near-surface air more than air at higher altitudes.

The immediate post-cold frontal environment is characterized by turbulent vertical mixing to the altitude of the cold frontal inversion, or approximately 1 km behind a typical cold front, resulting from positive surface sensible heat flux from a comparatively warm surface to the comparatively cold air above. In lake-effect precipitation environments, surface sensible heat flux over water is typically larger than that over land. This results in the formation of a superadiabatic layer above the surface while increasing buoyancy-driven turbulence intensity, thereby helping to maintain or even grow mixed layer depth as air passes over the water body.

Consider the combination of a mixed layer of sufficient depth, capped by an inversion at least 1.5 km above ground level, and near-surface air that has been sufficiently warmed and moistened by surface sensible and latent heat fluxes as it passes over the water body. A surface-based parcel in this environment will have a relatively low LCL. As the mixed layer is typically associated with relatively steep lapse rates, an air parcel that ascends beyond its LCL is likely to be positively-buoyant, even if marginally so, over the layer between the LCL and inversion. The release of this positive buoyancy results in lake-effect precipitation. Lake-effect precipitation typically ends as synoptic-scale stability, measured by inversion strength (higher) and altitude (lower), increases.

A general rule of thumb is that lake-effect precipitation requires the 850 hPa temperature to be at least 13°C colder than the water body temperature; for a water body whose surface is at or near sea level, this implies a near-dry adiabatic lapse rate, as often found with a mixed layer, over the ~1.5 km layer between 850 hPa and the water body's surface. For typical Great Lakes wintertime lake temperatures of 0-5°C, this requires T_{850} < -8°C to -13°C, resulting in surface-based CAPE of order 0-200 J kg^{-1} over the vertical layer between the surface and inversion bottom.

Smaller temperature differentials do not negate the ability for air to be warmed and moistened as it passes over a water body, but rather limit the likelihood that surface-based instability will be present to fuel moist convection. In such cases, however, if sufficient mechanical lift exists to enable precipitation formation, *lake-enhanced precipitation* may result. Here, moisture added to the near-surface air increases the amount of precipitation that would otherwise result.

The course text describes a simplified mathematical framework through which the warming and moistening an air parcel experiences as it crosses a hypothetical water body may be estimated.

Neglecting diffusion and horizontal turbulent flux terms, the Reynolds-averaged thermodynamic equation can be written as:

$$\frac{d\bar{\theta}}{dt} = -\frac{\partial\left(\overline{w'\theta'}\right)}{\partial z}$$

If we apply this equation near the surface (e.g., at 2-m) and assume that $\overline{w'\theta'}$ varies linearly with height (e.g., as in the idealized profiles from our planetary boundary layer lecture), this equation can be expressed using a centered finite difference for the right-hand side:

$$\frac{d\bar{\theta}_{2m}}{dt} = -\frac{\left(\overline{w'\theta'}\right)_H - \left(\overline{w'\theta'}\right)_{sfc}}{H}$$

Here, the finite difference is taken over the depth H of the mixed layer and involves the turbulent vertical heat flux terms evaluated at H and the surface. If we approximate $\left(\overline{w'\theta'}\right)_H \approx -0.2\left(\overline{w'\theta'}\right)_{sfc}$, we obtain:

$$\frac{d\bar{\theta}_{2m}}{dt} = \frac{1.2\left(\overline{w'\theta'}\right)_{sfc}}{H}$$

Or, alternatively, using the parameterization for $\left(\overline{w'\theta'}\right)_{sfc}$ given in the course textbook,

$$\frac{d\bar{\theta}_{2m}}{dt} = \frac{1.2c_h\left\|\overline{\mathbf{v}_{10m}}\right\|\left(\theta_{sfc} - \overline{\theta_{2m}}\right)}{H}$$

If we integrate this over some time interval Δt, assuming all terms can be represented by a time-averaged value, we obtain:

$$\Delta\bar{\theta}_{2m} = \frac{1.2c_h\overline{\left\|\overline{\mathbf{v}_{10m}}\right\|\left(\theta_{sfc} - \overline{\theta_{2m}}\right)}}{\overline{H}}\Delta t$$

The time interval Δt can be expressed as the distance the air parcel travels over some water body – i.e., the length L of the water body, for air passing over an inland lake – divided by the velocity of the air parcel – here, $\overline{\left\|\overline{\mathbf{v}_{10m}}\right\|}$. Making this substitution, we obtain:

$$\Delta\bar{\theta}_{2m} = \frac{1.2c_h L\overline{\left(\theta_{sfc} - \overline{\theta_{2m}}\right)}}{\overline{H}}$$

Thus, with knowledge of the heat exchange coefficient, the length of the portion of a water body over which an air parcel passes, the average air-sea potential temperature disequilibrium, and the average mixed layer depth, one can estimate the warming experienced by an air parcel crossing over a given water body. The hypothetical example given in the course text assumes c_h = 0.002, L = 100 km, $\overline{H}$ = 1 km, and $\overline{\left(\theta_{sfc} - \overline{\theta_{2m}}\right)}$ = 10 K, such that $\Delta\bar{\theta}_{2m} \approx 2.4$ K.

Using the same assumptions and approximating $\left(\overline{w'r_v'}\right)_H \approx 0.6\left(\overline{w'r_v'}\right)_{sfc}$, an analogous expression can be obtained for the moistening than an air parcel experiences upon water body passage:

$$\Delta\overline{r}_{v\ 2m} = \frac{0.4c_e L\left(\overline{r_{v\,sfc} - \overline{r_{v\,2m}}}\right)}{\overline{H}}$$

Here, r_{v_sfc} is the mixing ratio defined by the water body temperature – i.e., the saturation mixing ratio at the water body surface. The hypothetical example given in the course text assumes c_e = 0.002, L = 100 km, $\overline{H}$ = 1 km, and $\left(\overline{r_{v\,sfc} - \overline{r_{v\,2m}}}\right)$ = 4 g kg^{-1}, such that $\Delta\overline{r}_{v\ 2m} \approx 0.3$ g kg^{-1}, which for typical wintertime values of surface pressure (~980-1010 hPa) and temperature (~260-273 K) corresponds to an increase of 2-m dew point temperature of approximately 2 K.

As air parcels move downwind of a water body over which they previously crossed, entrainment of cooler, drier air from its surroundings can no longer be overcome by warming and moistening from below. This reduces surface-based instability with increasing distance from the shoreline.

Necessary Ingredients: Lifting

For surface-based instability to be realized or released, there must be sufficient lift to bring an air parcel to its LFC. There are several features and/or processes that may provide this lift.

Due to air mass modification by surface sensible heat fluxes, the near-surface air over the water body is warmer than that over adjacent land regions. Thus, the thickness of the near-surface layer is greater over the water body than over land, such that near-surface pressure is lower over water than over land, establishing a solenoidal circulation with weak mesoscale convergence directed toward the water body's center directly proportional in magnitude to the land-water temperature contrast. This mechanism results in the initial generation of lake-effect precipitation bands. This suggests that lake-effect precipitation bands are most intense over water near the downwind shoreline, where the air has gained close to the maximum heat and moisture that it will from the lake, and will weaken rapidly inland away from the downwind shore as convergence weakens.

Friction is an important contributor to planetary boundary layer lifting in two ways. First, friction acts as a drag on the near-surface wind more over land than over water. Thus, air parcels passing from a water body to land will slow as it does so. This is known as *frictional convergence* and is an important contributor to enhanced lifting along the immediately shoreline. Second, frictional drag acts to reduce the magnitude of the Coriolis force, such that near-surface wind blows across isobars toward areas of low pressure, again more so over land than over water. This will enhance convergence (divergence) along the shoreline closer to the anticyclone (cyclone); for example, a westerly geostrophic wind blowing over an east-west-oriented Northern Hemisphere water body is associated with enhanced convergence on the south (north) side of the water body. These, too, imply rapid weakening of lake-effect precipitation bands inland away from the downwind shore.

Friction-induced convergence into regions of near-surface cyclonic vertical vorticity implies that ascent within the planetary boundary layer is directly proportional to cyclonic vertical vorticity; i.e., near-surface flow that is cyclonically-curved enhances lift compared to that with uncurved or anticyclonically-curved near-surface flow. There is no clear relationship between other synoptic-scale lifting mechanisms and lake-effect precipitation potential, however. Consider, for example, a cold front passage. Ahead of the cold front, warm air advection reduces the magnitude of the 850 hPa to water body surface temperature difference, reducing lake effect potential. Behind the cold front, this difference may again grow large, but lake effect potential will to some extent be modulated by the strength and height of the frontal inversion in its wake. Cyclonic winds about the cold front may shift the wind direction into a more favorable or less favorable direction for a long fetch across the water body. Further, from thermal wind, the horizontal layer-mean temperature gradient with the cold front must be associated with relatively large vertical wind shear, potentially hindering lake-effect precipitation organization. We discuss fetch and vertical wind shear later in these notes.

Orography beyond the downwind shoreline can also modulate lifting in lake-effect precipitation environments. While the immediate shoreline is often at slightly higher elevation than the water body surface, in some locations terrain height may increase rapidly a short distance inland. In the mid-latitudes, particularly noteworthy examples are the Tughill Plateau in New York downwind of Lake Ontario and, to lesser extent, the mountain ranges (Hurons, Porcupines, etc.) of the upper peninsula of Michigan downwind of Lake Superior.

Other Considerations

The equations for the warming and moistening of an air parcel upon water body crossing indicate a relationship of each with the length L of the water body. This L is a measure of the *fetch* of the water body, or the horizontal distance traveled by the wind as it traverses the water body. Longer fetches increase the residence time for air parcels over the water body, increasing the warming and moistening that they experience via sensible and latent heating over water, in turn increasing the strength of the solenoidal circulation described above. Further, the added moistening fostered by a longer fetch increases the amount of precipitable water vapor if the air becomes saturated by lifting. In general, long fetches (e.g., along the long axis of an elongated lake) promote intense single precipitation bands with embedded cellular convection. Short fetches (e.g., along the short axis of an elongated late) promote multiple narrow, less intense precipitation bands that act much like horizontal convective rolls. Typically, both single and multiple lake effect bands are roughly aligned with the near-surface wind and vertical wind shear vectors.

Lake-effect precipitation is also modulated by wind speed. Single precipitation bands are favored for fast boundary layer wind speeds (~15-30 kt) and long fetches. Fast winds result in large latent and sensible heat fluxes due to their wind speed dependence; long fetches ensure that air parcels encounter such elevated fluxes longer than for shorter fetches. Empirically, faster winds also are associated with more focused solenoidal-induced convergence over water. Entrainment, sensible, and latent heat fluxes over land act to weaken lake-effect bands with progressively further inland extent; however, faster winds enable these bands to propagate further inland before these factors have had sufficient time to act upon the band.

In cases with moderate-strong winds but small fetch, lake-effect precipitation typically organizes into a multiple-banded structure. Inland band propagation is reduced for moderate winds relative to strong winds. Lake-effect precipitation, particularly inland from the water body, is uncommon for light winds. In such cases, the solenoidal-induced convergence described earlier results in the formation of a *mesolow* rather than an organized, inland-propagating convective band. From the continuity equation, near-surface convergence must result in ascent, and this ascent amplifies any existing vertical (relative plus planetary) vorticity within the over-lake environment. This can be shown by the vertical vorticity tendency equation. Neglecting friction and baroclinic generation, the vertical vorticity tendency following the motion is given by:

$$\frac{d\zeta}{dt} = (\zeta + f)\frac{\partial w}{\partial z} + \left[-\frac{\partial v}{\partial z}\frac{\partial w}{\partial x} + \frac{\partial u}{\partial z}\frac{\partial w}{\partial y} \right]$$

The right-hand side forcing terms represent stretching of existing vertical vorticity and tilting of horizontal vorticity into the vertical, respectively. Vertical vorticity increases following the flow for ascent increasing with height and positive ambient vertical vorticity for the stretching term. Through its modulation of air propagation (direction and speed) as air passes from land to water, shoreline curvature may also promote mesolow development.

Finally, vertical wind shear also modulates lake-effect precipitation band structure and intensity. As vertical wind shear (particularly its directional component) increases, band organization tends to decrease. Single (multiple) bands are preferred for directional shear of 0-30° (30-60°). Larger directional shear tends to not be found in lake-effect environments. Vertical wind shear controls the extent to which cooler, drier environmental air is entrained into the lake-effect band. Minimal directional shear allows the comparatively warm, moist air along the lake-effect band to remain relatively isolated from its surroundings, whereas larger directional shear promotes cooler, drier air to infiltrate the lake-effect band along its sides.