The Atmospheric Boundary Layer

• Turbulence (9.1)

• The Surface Energy Balance (9.2)

• Vertical Structure (9.3)

• Evolution (9.4)

• Special Effects (9.5)

• The Boundary Layer in Context (9.6)
Adapted from Dennis L. Hartmann, *Global Physical Climatology*, p. 28 (Copyright 1994), with permission from Elsevier.
Radiative Fluxes

The *net* (downward) *radiation flux* absorbed at the surface is

\[ F^* = F_S^\downarrow - F_S^\uparrow + F_L^\downarrow - F_L^\uparrow \]

where \( F_S^\downarrow \) is the downward shortwave (solar) radiation flux at the surface, \( F_S^\uparrow \) is the upward (reflected) shortwave radiation flux at the surface, \( F_L^\downarrow \) is the downward longwave radiation flux at the surface, and \( F_L^\uparrow \) is the upward (emitted and reflected) longwave radiation flux at the surface.
Radiative fluxes at the surface under clear skies

![Graph showing radiative fluxes over a 24-hour period](image-url)
Surface Energy Balance over Land

The surface energy (flux) balance over land is

\[ F^* = F_{Hs} + F_{Es} + F_{Gs} \]

where \( F_{Hs} \) is the (turbulent) sensible heat flux (positive upward, away from the surface), \( F_{Es} \) is the (turbulent) latent heat flux (positive upward), \( F_{Gs} \) is the (conductive) ground heat flux (positive downward, away from the surface), and the subscript \( s \) denotes surface.
Turbulent and conductive fluxes at the surface under clear skies

Daytime over moist vegetation
Nighttime over moist vegetation
Daytime over a dry desert
Oasis effect during daytime:
hot dry wind blowing over moist vegetation
Soil temperature at various depths under a grass field

(a) Soil Temperature
O’Neill, Nebraska
13-14 August 1953

(b) Soil Temperature vs Depth
O’Neill, Nebraska
13 August 1953
How is the diurnal cycle of the surface skin temperature related to the conductivity of the soil?
Soil temperature at 4 levels below the surface

Adapted from Trans. Amer. Geophys. Union 37, 746 (1956).

Soil temperature at 4 levels below the surface is shown in the graph. The temperature is measured in degrees Celsius (°C) and is plotted against the depth below the surface (meters). The depth levels are 0.75 m, 1.5 m, 3 m, and 6 m, with the 0.75 m level being the closest to the surface and the 6 m level being the deepest. The graph depicts the temperature variations from 1955 to 1956, showing how temperature changes with depth and time.
Soil Temperature

• *Conduction* is down the gradient of temperature.

• The *annual* cycle penetrates to greater depth than the *diurnal* cycle.

• *Amplitude* decreases with increasing depth (for a given forcing frequency).

• *Phase* delay increases with increasing depth (for a given forcing frequency).
Surface Energy Balance of Ocean Surfaces

• Does it make sense to say that $F_{Gs}$ for ocean surfaces is much larger than for land?
• Solar radiation is more readily absorbed by the ocean for two reasons. What are they?
  • Turbulence quickly mixes heat through the ocean mixed layer.
• The specific heat of water is larger than that of soil.
• As a result, the diurnal cycle of ocean surface temperature is almost negligible.
Table 4.1
Properties of Soil Components at 293 K

<table>
<thead>
<tr>
<th></th>
<th>Specific heat ($c_p$) (J kg$^{-1}$ K$^{-1}$)</th>
<th>Density ($\rho$) (kg m$^{-3}$)</th>
<th>$\rho c_p$ (J m$^{-3}$ K$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil inorganic material</td>
<td>733</td>
<td>2600</td>
<td>$1.9 \times 10^6$</td>
</tr>
<tr>
<td>Soil organic material</td>
<td>1921</td>
<td>1300</td>
<td>$2.5 \times 10^6$</td>
</tr>
<tr>
<td>Water</td>
<td>4182</td>
<td>1000</td>
<td>$4.2 \times 10^6$</td>
</tr>
<tr>
<td>Air</td>
<td>1004</td>
<td>1.2</td>
<td>$1.2 \times 10^3$</td>
</tr>
</tbody>
</table>

Bulk Aerodynamic Formulae

• The *bulk aerodynamic method* can be used to estimate the surface sensible and latent heat fluxes, as well as frictional drag on surface winds.

• In kinematic units (K m s$^{-1}$)

\[ F_{Hs} = C_H |V|(T_s - T_{air}) \]

where $C_H$ is a dimensionless *bulk transfer coefficient* for sensible heat, $|V|$ is the wind speed and $T_{air}$ is the air temperature at standard measurement heights (such as 10 m and 2 m), and $T_s$ is the ocean surface temperature.
Bulk Aerodynamic Formulae

• The formula for $F_{HS}$ does not involve a (vertical) turbulent velocity scale. Why not?

• What determines the surface skin temperature $T_s$ over land during a sunny day?

• What happens to $T_s$ at night when the ground surface is cooled by longwave radiation?
Bulk Aerodynamic Formulae

- **Statically neutral conditions over flat land:** $C_H = C_{HN}$ is 0.001 to 0.005 and depends on surface roughness similar to $C_{DN}$ (Table 9.2).

- **Statically unstable conditions over flat land:** $C_H$ is 2 to 3 times larger than $C_{HN}$.

- **Statically stable conditions over flat land:** $C_H$ decreases towards 0.
Table 9.2 The Davenport classification, where \( z_0 \) is aerodynamic roughness length and \( C_{DN} \) is the corresponding drag coefficient for neutral static stability.

<table>
<thead>
<tr>
<th>( z_0 ) (m)</th>
<th>Classification</th>
<th>Landscape</th>
<th>( C_{DN} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0002</td>
<td>Sea</td>
<td>Calm sea, paved areas, snow-covered flat plain, tide flat, smooth desert.</td>
<td>0.0014</td>
</tr>
<tr>
<td>0.005</td>
<td>Smooth</td>
<td>Beaches, pack ice, morass, snow-covered fields.</td>
<td>0.0028</td>
</tr>
<tr>
<td>0.03</td>
<td>Open</td>
<td>Grass prairie or farm fields, tundra, airports, heather.</td>
<td>0.0047</td>
</tr>
<tr>
<td>0.1</td>
<td>Roughly open</td>
<td>Cultivated area with low crops and occasional obstacles such as trees or hedgerows, vineyards.</td>
<td>0.0075</td>
</tr>
<tr>
<td>0.25</td>
<td>Rough</td>
<td>High crops, crops of varied height, scattered obstacles such as trees or hedgerows, vineyards.</td>
<td>0.012</td>
</tr>
<tr>
<td>0.5</td>
<td>Very rough</td>
<td>Mixed farm fields and forest clumps, orchards, scattered buildings.</td>
<td>0.018</td>
</tr>
<tr>
<td>1.0</td>
<td>Closed</td>
<td>Regular coverage with large size obstacles with open spaces roughly equal to obstacle heights, suburban houses, villages, mature forests.</td>
<td>0.030</td>
</tr>
<tr>
<td>( \geq 2 )</td>
<td>Chaotic</td>
<td>Centers of large towns and cities, irregular forests with scattered clearings.</td>
<td>0.062</td>
</tr>
</tbody>
</table>

PART 2
Bulk Aerodynamic Formulae

• *Bulk aerodynamic relationships* exist for water vapor fluxes over water and saturated soil.

• Assume that the surface water vapor mixing ratio = saturation mixing ratio.

• In kinematic units (kg kg\(^{-1}\) m s\(^{-1}\))

\[
F_{water} = C_E |\mathbf{V}| (q_{sat}(T_s, p_s) - q_{air})
\]

where \(C_E\) is a dimensionless *bulk transfer coefficient* for water vapor (\(C_E \approx C_H\)), \(|\mathbf{V}|\) is the wind speed and \(q_{air}\) is the mixing ratio at standard measurement heights (such as 10 m and 2 m), \(T_s\) is the ocean surface temperature, and \(p_s\) is the surface pressure.
Bulk Aerodynamic Formulae

\( F_{\text{water}} \) is related to the latent heat flux \( F_{Es} \) in kinematic units (K m s\(^{-1}\)):

\[
F_{Es} = \frac{L_v}{c_p} F_{\text{water}}
\]

and to the evaporation rate of water (mm/s):

\[
E = \frac{\rho_{\text{liq}}}{\rho_{\text{air}}} F_{\text{water}}
\]
Bulk Aerodynamic Formulae

The ratio of sensible to latent heat fluxes at the surface is the *Bowen ratio*:

\[ B = \frac{F_{Hs}}{F_{Es}} \]

- Over the oceans \( B \) decreases as \( T_s \) increases from about 1 in the polar regions to 0.1 in the Tropics.
- Over land the evaporation rate depends on soil moisture and *transpiration*.
  - Irrigated crops: \( B \sim 0.2 \)
  - Grassland: \( B \sim 0.5 \)
  - Semi-arid regions: \( B \sim 5 \)
  - Deserts: \( B \sim 10 \)
Water potential gradient

Outside air $\Psi = -10.0$ to $-100.0$ MPa

Leaf $\Psi$ (air spaces) = $-7.0$ MPa

Leaf $\Psi$ (cell walls) = $-1.0$ MPa

Trunk xylem $\Psi$ = $-0.8$ MPa

Root xylem $\Psi$ = $-0.6$ MPa

Soil $\Psi$ = $-0.3$ MPa

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Transpiration

- Xylem sap
- Mesophyll cells
- Stoma
- Water molecule
- Atmosphere
The *bulk aerodynamic method* for momentum gives a *drag law*:

\[ u_*^2 = C_D |V|^2 \]

where \( C_D \) is a dimensionless *drag coefficient*, \( |V| \) is the wind speed at 10 m, and \( u_*^2 \) is the magnitude of the downward momentum flux at the surface.

\( C_D \) varies from \( 10^{-3} \) over smooth surfaces to \( 2 \times 10^{-2} \) over rough ones (Table 9.2).

\( C_D \) also varies with stability like \( C_H \) does.
Bulk Aerodynamic Formulae

• The *bulk aerodynamic method* for momentum gives a *drag law*:

\[ u_*^2 = C_D |V|^2 \]

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\( C_D \) varies from \( 10^{-3} \) over smooth surfaces to \( 2 \times 10^{-2} \) over rough ones (Table 9.2).

From HW on turbulent fluxes:

4. The friction velocity \( u_* = 0.3 \text{ m s}^{-1} \).
   (a) What is the magnitude of the surface stress? Use \( \rho = 1.2 \text{ kg m}^{-3} \).
   (b) If \( h = 500 \text{ m} \), how much would the average ABL wind velocity change over 24 h due to the surface stress alone? Assume that the wind velocity and surface stress vectors are parallel. *Decreases \( \sim 15 \text{ m/s over 24 hours} \).*
   (c) What additional forces act to maintain the ABL wind?
Table 9.2 The Davenport classification, where \( z_o \) is aerodynamic roughness length and \( C_{DN} \) is the corresponding drag coefficient for neutral static stability.\(^a\)

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Bulk Aerodynamic Formulae

• The bulk aerodynamic method for momentum gives a drag law:

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where \( C_D \) is a dimensionless drag coefficient, \( |V| \) is the wind speed at 10 m, and \( u_*^2 \) is the magnitude of the downward momentum flux at the surface.

\( C_D \) varies from \( 10^{-3} \) over smooth surfaces to \( 2 \times 10^{-2} \) over rough ones (Table 9.2).

Example:
Surface wind speed = 10 m/s.
Surface classification: Open.
\( z_0 = 0.03 \) m, \( C_{DN} = 0.0047 \).
\( u_*^2 = C_D |V|^2 = 0.0047 \times 10^2 \) m\(^2\)/s\(^2\) = 0.47 m\(^2\)/s\(^2\).
\( u_* = 0.7 \) m/s.
Bulk Aerodynamic Formulae

• The *bulk aerodynamic method* for momentum gives a *drag law*:

\[ u_*^2 = C_D |V|^2 \]

where \( C_D \) is a dimensionless *drag coefficient*, \( |V| \) is the wind speed at 10 m, and \( u_*^2 \) is the magnitude of the downward momentum flux at the surface.

\( C_D \) varies from \( 10^{-3} \) over smooth surfaces to \( 2 \times 10^{-2} \) over rough ones (Table 9.2).

**Example:**
Surface wind speed = 10 m/s.
Surface classification: Open.
\( C_{DN} = 0.0047 \).
\[ u_*^2 = C_D |V|^2 = 0.0047 \times 10^2 \text{ m}^2/\text{s}^2 = 0.47 \text{ m}^2/\text{s}^2. \]
\[ u_* = 0.7 \text{ m/s}. \]

*How does one determine \( C_D \)?*
c. Aerodynamic methods

From a practical point of view it is more useful if one can relate these fluxes to more standard observations such as mean wind speed and sea-air temperature and humidity differences, i.e.,

\[
\begin{align*}
\tau/\rho &= -uw = CDU^3, \\
H_s/(\rho C_p) &= wT = CTU \Delta T, \\
H_L/L &= E = wq = C_q U \Delta q
\end{align*}
\]

(14)

where \( \Delta T \) is the difference between sea surface temperature and the air temperature at a reference height and \( \Delta q \) the corresponding mean moisture difference. Roll (1965) discusses the derivation of these equations and suggests that \( CD \approx CT \approx C_q \) for conditions not too far from neutral.
R/V Flip (Floating Instrument Platform)

https://www.youtube.com/watch?v=azZlcoPl_CU
c. Aerodynamic methods

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\[
\begin{align*}
\tau / \rho &= -\overline{w w} = C_D U^2 \\
H_z / (\rho C_p) &= \overline{w T} = C_T U \Delta T \\
H_L / L &= E = \overline{w q} = C_q U \Delta q
\end{align*}
\]  

where \( \Delta T \) is the difference between sea surface temperature and the air temperature at a reference height and \( \Delta q \) the corresponding mean moisture difference. Roll (1965) discusses the derivation of these equations and suggests that \( C_D \approx C_T \approx C_q \) for conditions not too far from neutral.

\[
u_*^2 \equiv \frac{\tau}{\rho} = -\overline{w' w'} = C_D U^2
\]

Fig. 5. \( u_*^2 \) vs \( 10^{-3} U^2 \). The straight line corresponds to \( C_D = 1.52 \times 10^{-3} \).
Bulk Aerodynamic Formulae

• Over the oceans, an increase in surface wind speed produces larger waves, which increases the drag.
BEAUFORT FORCE 0

WIND SPEED: LESS THAN 1 KNOT

SEA: SEA LIKE A MIRROR
BEAUFORT FORCE 2
WIND SPEED: 4-6 KNOTS

SEA: WAVE HEIGHT .2-.3M (.5-1FT), SMALL WAVELETS, CRESTS HAVE A GLASSY APPEARANCE AND DO NOT BREAK
BEAUFORT FORCE 4

WIND SPEED: 11-16 KNOTS

SEA: WAVE HEIGHT 1-1.5M (3.5-5FT), SMALL WAVES BECOMING LONGER, FAIRLY FREQUENT WHITE HORSES
BEAUFORT FORCE 6
WIND SPEED: 22-27 KNOTS

SEA: WAVE HEIGHT 3-4M (9.5-13 FT), LARGER WAVES BEGIN TO FORM, SPRAY IS PRESENT, WHITE FOAM CRESTS ARE EVERYWHERE
BEAUFORT FORCE 8
WIND SPEED: 34-40 KNOTS

SEA: WAVE HEIGHT 5.5-7.5M (18-25FT), MODERATELY HIGH WAVES OF GREATER LENGTH, EDGES OF CREST BEGIN TO BREAK INTO THE SPINDRIFT, FOAM BLOWN IN WELL MARKED STREAKS ALONG WIND DIRECTION.
BEAUFORT FORCE 10
WIND SPEED: 48-55 KNOTS

SEA: WAVE HEIGHT 9-12.5M (29-41FT), VERY HIGH WAVES WITH LONG OVERHANGING CRESTS, THE RESULTING FOAM, IN GREAT PATCHES, IS BLOWN IN DENSE WHITE STREAKS ALONG WIND DIRECTION. ON THE WHOLE, SEA
BEAUFORT FORCE 12

WIND SPEED: 64 KNOTS

SEA: SEA COMPLETELY WHITE WITH DRIVING SPRAY, VISIBILITY VERY SERIOUSLY AFFECTED. THE AIR IS FILLED WITH FOAM AND SPRAY
Bulk transfer coefficients over the ocean

![Graph of Bulk Transfer Coefficients over the Ocean](image)

- **$C_D$**: Drag coefficient
- **$C_H$**: Heat transfer coefficient
- **$C_E$**: Moisture transfer coefficient

**Figure 9.13**: Variation of bulk transfer coefficients for drag ($C_D$), heat ($C_H$), and moisture ($C_E$) with wind speed over the ocean. The coefficients decrease with increasing wind speed, reflecting the decrease in turbulence and skin friction at higher speeds.
**Bulk transfer coefficients over the ocean**

- For speeds of the mean vector wind < 5 m/s, the bulk formulae are not valid.
- Why not?
- How could the formulae be modified so they are valid in such situations?
Exercise 9.2 Consider a column of air initially of vertically uniform \( \theta \) over cold land, capped by a very strong temperature inversion that prevents boundary layer growth. This air column advects with speed \( U \) over a warmer ocean surface with potential temperature \( \theta_s \). (a) How does temperature vary with distance \( x \) from shore? (b) At any fixed distance \( x \) from shore, how does the air temperature vary with wind speed?

[Hint: Use Taylor's hypothesis: \( \frac{\partial \theta}{\partial t} = U \frac{\partial \theta}{\partial x} \).]
(a) What is $\theta(x)$?
(b) What is $(\partial \theta / \partial U)_x$?

\[ \frac{\partial \theta}{\partial t} = \frac{F_{Hz}}{z_i} \]

\[ \frac{\partial \theta}{\partial t} = U \frac{\partial \theta}{\partial x} \]  
(Taylor’s hypothesis)
(a) What is $\theta(x)$?
(b) What is $(\partial\theta/\partial U)_x$?

\[
\frac{\partial \theta}{\partial t} = \frac{F_{Hs}}{z_i} \\
\frac{\partial \theta}{\partial t} = U \frac{\partial \theta}{\partial x} \quad \text{(Taylor’s hypothesis)}
\]

\[
\frac{\partial \theta}{\partial x} = \frac{1}{U} \frac{F_{Hs}}{z_i}
\]
(a) What is $\theta(x)$?
(b) What is $(\partial \theta / \partial U)_x$?

\[
\begin{align*}
\frac{\partial \theta}{\partial t} &= \frac{F_{Hs}}{z_i} \\
\frac{\partial \theta}{\partial x} &= \frac{1}{U} \frac{F_{Hs}}{z_i} \\
\frac{\partial \theta}{\partial t} &= U \frac{\partial \theta}{\partial x} \quad \text{(Taylor's hypothesis)} \\
F_{Hs} &= C_H U (\theta_s - \theta)
\end{align*}
\]
(a) What is $\theta(x)$?

(b) What is $(\partial \theta / \partial U)_x$?

\[
\frac{\partial \theta}{\partial t} = \frac{F_{Hs}}{z_i}
\]

\[
\frac{\partial \theta}{\partial x} = \frac{1}{U} \frac{F_{Hs}}{z_i}
\]

\[
\frac{\partial \theta}{\partial x} = C_H \frac{\theta_s - \theta}{z_i}
\]

\[
F_{Hs} = C_H U (\theta_s - \theta)
\]
(a) What is $\theta(x)$?
(b) What is $(\partial \theta / \partial U)_x$?

\[
\frac{\partial \theta}{\partial t} = \frac{F_{Hs}}{z_i}
\]

\[
\frac{\partial \theta}{\partial x} = \frac{1}{U} \frac{F_{Hs}}{z_i}
\]

\[
\frac{\partial \theta}{\partial x} = C_H \frac{\theta_s - \theta}{z_i}
\]

\[
F_{Hs} = C_H U (\theta_s - \theta)
\]

\[
\theta - \theta_s = (\theta(0) - \theta_s) \exp(-C_H x / z_i)
\]
(a) What is $\theta(x)$?
(b) What is $(\partial \theta / \partial U)_x$?

\[ \frac{\partial \theta}{\partial t} = \frac{F_{Hs}}{z_i} \]
\[ \frac{\partial \theta}{\partial x} = \frac{1}{U} \frac{F_{Hs}}{z_i} \]
\[ \frac{\partial \theta}{\partial x} = C_H \frac{\theta_s - \theta}{z_i} \]

\[ \frac{\partial \theta}{\partial t} = U \frac{\partial \theta}{\partial x} \quad \text{(Taylor's hypothesis)} \]

\[ F_{Hs} = C_H U (\theta_s - \theta) \]

$\theta - \theta_s = (\theta(0) - \theta_s) \exp(-C_H x / z_i)$

Depends only on downstream distance.
The Global Surface Energy Balance

The bulk aerodynamic formulae have been used to estimate the global distribution of the terms in the surface energy balance.

The net upward transfer of energy through the Earth’s surface is

\[ F_{\text{net}}^\uparrow = -F^* + F_{Hs} + F_{Es} \]

where \( F^* \) is the net downward radiative flux, and \( F_{\text{net}}^\uparrow = -F_G \).
2.1 Components of the Earth System

The departures from the zonal-mean, shown in Fig. 2.11 (bottom). The coolness of the eastern oceans relative to the western oceans at subtropical latitudes derives from circulation around the subtropical anticyclones (Fig. 1.16). The equatorward flow of cool air around the eastern flanks of the anticyclones extracts a considerable quantity of heat from the ocean surface, as explained in Section 9.3.4, and drives cool, southward ocean currents (Fig. 2.4). In contrast, the warm, humid poleward flow around their western flanks extracts much less heat and drives warm western boundary currents such as the Gulf Stream. At higher latitudes the winds circulating around the subpolar cyclones have the opposite effect, cooling the western sides of the oceans and warming the eastern sides. The relative warmth of the eastern Atlantic at these higher latitudes is especially striking.

Wind-driven upwelling is responsible for the relative coolness of the equatorial eastern Pacific and Atlantic, where the southeasterly trade winds protrude northward across the equator (Fig. 1.18). Wind-driven upwelling along the coasts of Chile, California, and continents that occupy analogous positions with respect to the subtropical anticyclones, although not well resolved in Fig. 2.11, also contributes to the coolness of the subtropical eastern oceans, as do the highly reflective cloud layers that tend to develop at the top of the atmospheric boundary layer over these regions (Section 9.4.4).

The atmospheric circulation feels the influence of the underlying sea surface temperature pattern, particularly in the tropics. For example, from a comparison of Figs. 1.25 and 2.11 it is evident that the intertropical convergence zones in the Atlantic and Pacific sectors are located over bands of relatively warm sea surface temperature and that the dry zones lie over the equatorial cold tongues on the eastern sides of these ocean basins.
Annual mean surface net radiation
Annual mean surface latent heat flux
Annual mean surface sensible heat flux
Annual mean net downward heat flux into ocean
9.3 Vertical Structure

The surface is anomalously warm or cold relative to the mean temperature at that latitude (Fig. 2.11). The net flux is upward over the warm waters of the Gulf Stream and the Kuroshio current and it is downward over the regions of coastal and equatorial upwelling where cold water is being brought to the surface.

This section considers the interplay between turbulence and the vertical profiles of wind, temperature and moisture within the boundary layer, drawing heavily on the diurnal cycle over land as an example.

9.3.1 Temperature

Depending on the vertical temperature structure within the boundary layer, turbulent mixing can be suppressed or enhanced at different heights via buoyant consumption or production of $T$. In fact, it is ultimately the temperature profile that determines the boundary-layer depth.

Recall that the troposphere is statically stable on average, with a potential temperature gradient of $3.3^\circ C/20862$ km (Fig. 9.15). Solar heating of the ground causes thermals to rise from the surface, generating turbulence. Also, drag at the ground causes near-surface winds to be slower than winds aloft, creating wind shear that generates mechanical turbulence. Turbulence generated by processes near the ground mixes surface air of relatively low values of potential temperature, with higher potential temperature air from higher altitudes. The resulting mixture has an intermediate potential temperature that is relatively uniform with height (i.e., homogenized within the boundary layer). More importantly, this low altitude mixing has created a temperature jump between the boundary-layer air and the warmer air aloft. This temperature jump corresponds to the capping inversion.

![Annual mean net upward energy flux](image.png)

Annual mean net upward energy flux at the Earth's surface as estimated from Eq. (9.21) based on a reanalysis of 1958–2001 data by the European Centre for Medium Range Weather Forecasting. [Courtesy of Todd P. Mitchell.]