<u>The virtual potential temperature</u> **<u>θv</u>** is an important concept in atmospheric boundary meteorology. It can serve as a stability criterion for an atmosphere with a moisture gradient. When θv is constant, the atmosphere is statically neutral. When it decreases with elevation, the atmosphere is statically unstable. When it increases with elevation, the atmosphere is statically stable.

<u>The virtual temperature (Tv)</u> is the temperature at which dry air would have the same density as the moist air, at a given pressure.
In other words, two air samples with the same virtual temperature have the same density, regardless of their actual temperature or relative humidity.
the virtual temperature is always greater than the absolute air temperature.

<u>The potential temperature</u> is the temperature which would result if the air were brought adiabatically to a standard pressure level P_O=1000 kPa. Combining the first law of thermodynamics with the equatio of state of air gives

<u>Buoyancy</u> is one of the driving forces for turbulence in the BL. Thermals of warm air rise because they are less dense than the surrounding air, and hence positively buoyant.

where r_{sat} is the water-vapor saturation mixing ratio of the air parcel, and r_L is the liquid-water mixing ratio. In (1.5.1a) the potential temperatures are in units of K, and the mixing ratios are in units of g/g. For unsaturated air with mixing ratio r, the virtual potential temperature is:

$$\theta_v = \theta \cdot (1 + 0.61 \cdot r) \qquad (1.5.1b)$$

A derivation of the virtual temperature is given in Appendix D.

As usual, the potential temperature, θ, is defined as

$$\theta = T \left(\frac{P_o}{P} \right)^{0.286} \qquad (1.5.1c)$$

where P is air pressure and P_o is a reference pressure. Usually, P_o is set to 100 kPa quantity Cp·ɵ is called the **dry static energy.**

1.5.2 Example

Problem. Given a temperature of 25°C and a mixing ratio of 20 g/kg measured at a pressure of 90 kPa (900 mb), find the virtual potential temperature.

Solution. First, we must find the potential temperature:

$$\theta = T\,(P_o/P)^{0.286} = 298.16\cdot(100/90)^{0.286} = 307.28 \text{ K}$$

The air is unsaturated. allowing us to find the virtual potential temperature from:

$$\theta_v = \theta\cdot(1 + 0.61\cdot r) = 307.28\cdot[\,1 + 0.61\cdot(0.020)\,] = 311.03 \text{ K}$$

1.6.4 Virtual Potential Temperature Evolution

Given the virtual potential temperature profiles from the previous subsections, it is useful to integrate these profiles into our concept of how the boundary layer evolves. If rawinsonde soundings were made at the times indicated by flags S1 through S6 in Fig 1.7, then Fig 1.12 shows the resulting virtual potential temperature profile evolution.

We see from these soundings that knowledge of the virtual potential temperature profile is usually sufficient to identify the parts of the boundary layer. The structure of the BL is clearly evident.

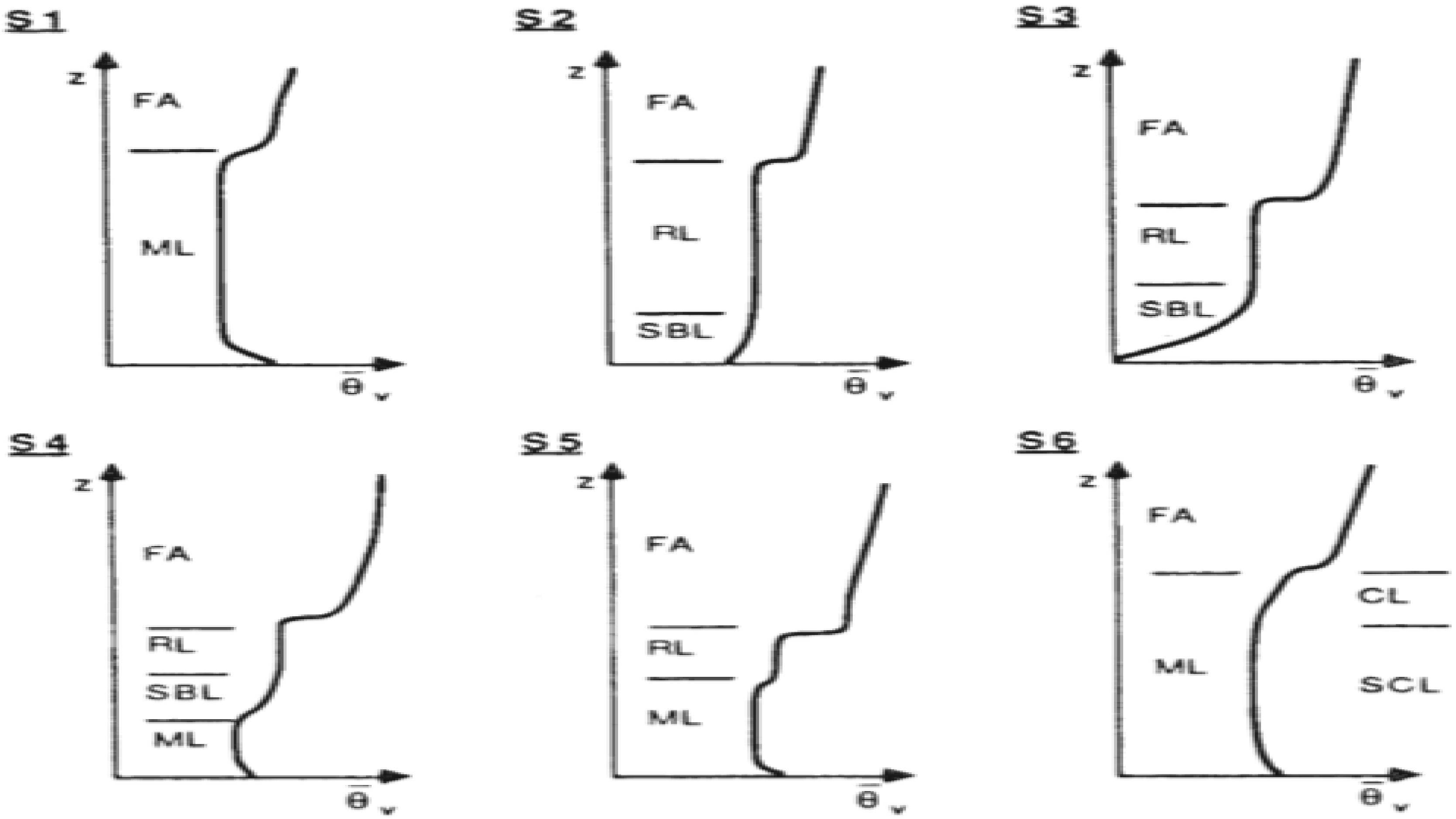


Fig. 1.12 Profiles of mean virtual potential temperature, $\bar{\theta}_v$, showing the boundary-layer evolution during a diurnal cycle starting at about 1600 local time. S1-S6 identify each sounding with an associated launch time indicated in Fig. 1.7.

Stated another way, knowledge of the virtual potential temperature lapse rate is usually sufficient for determining the static stability. An exception to this rule is evident by comparing the lapse rate in the middle of the RL with that in the middle of the ML. Both are adiabatic; yet, the ML corresponds to statically unstable air while the RL contains statically neutral air.

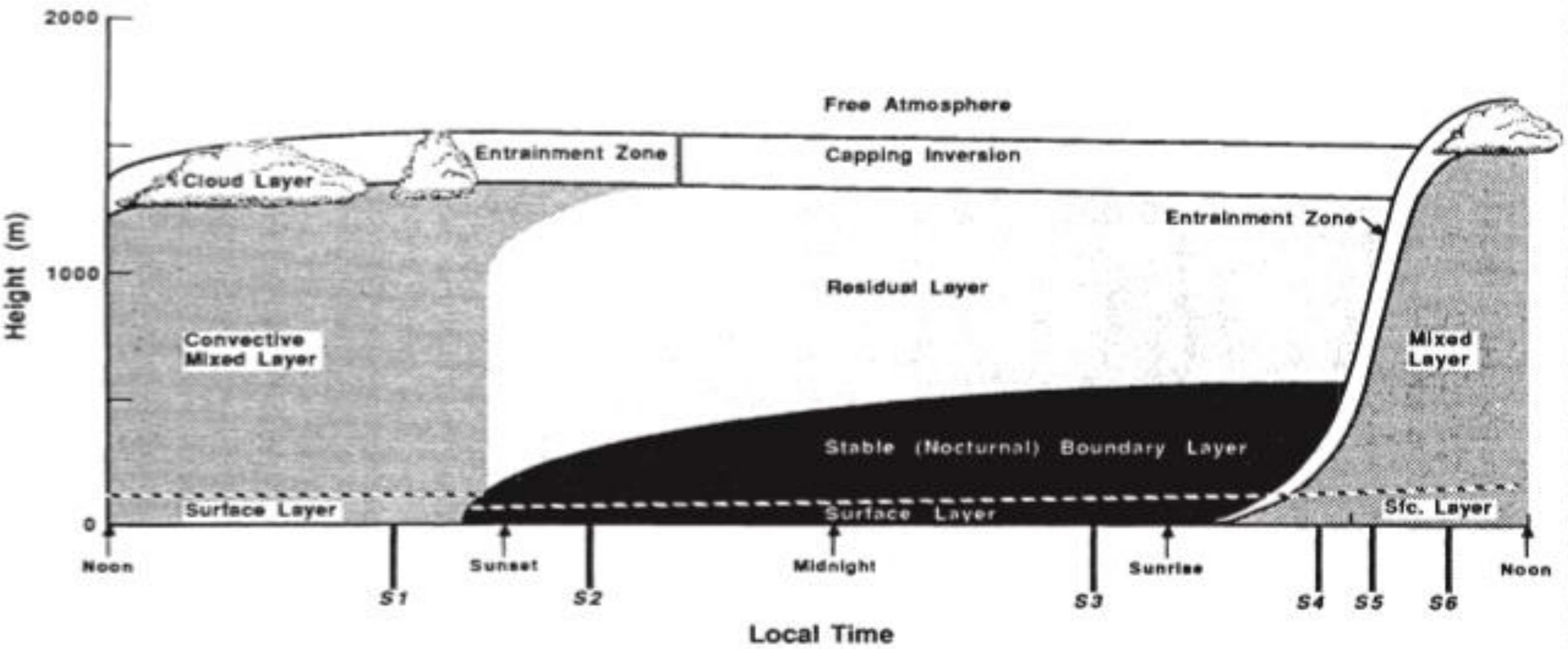


Fig. 1.7 The boundary layer in high pressure regions over land consists of three major parts: a very turbulent mixed layer; a less-turbulent residual layer containing former mixed-layer air; and a nocturnal stable boundary layer of sporadic turbulence. The mixed layer can be subdivided into a cloud layer and a subcloud layer. Time markers indicated by S1-S6 will be used in Fig. 1.12.

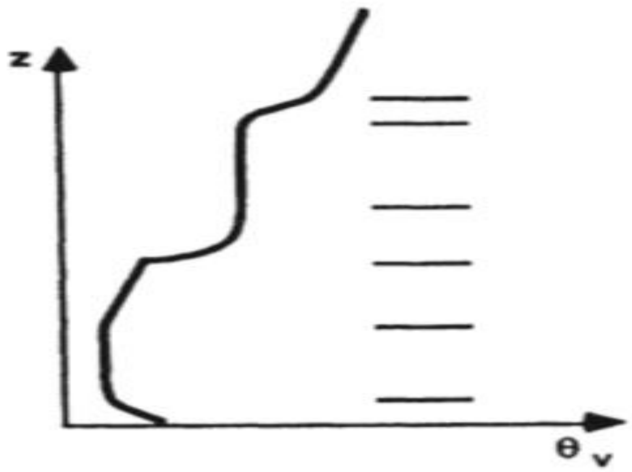
Table 1-1. Comparison of boundary layer and free atmosphere characteristics.

Property	Boundary Layer	Free Atmosphere
Turbulence	• Almost continuously turbulent over its whole depth.	• Turbulence in convective clouds, and sporadic CAT in thin layers of large horizontal extent.
Friction	• Strong drag against the earth's surface. Large energy dissipation.	• Small viscous dissipation.
Dispersion	• Rapid turbulent mixing in the vertical and horizontal.	• Small molecular diffusion. Often rapid horizontal transport by mean wind.
Winds	• Near logarithmic wind speed profile in the surface layer. Subgeostrophic, cross-isobaric flow common.	• Winds nearly geostrophic.
Vertical Transport	• Turbulence dominates.	• Mean wind and cumulus-scale dominate
Thickness	• Varies between 100 m to 3 km in time and space. Diurnal oscillations over land.	• Less variable. 8-18 km. Slow time variations.

10) a) Given air at a pressure height of 90 kPa (900 mb) with a temperature of 30 °C and a mixing ratio of 20 g/kg, find the virtual potential temperature.

b) Given saturated air at 85 kPa (850 mb) with a temperature of 20 °C and a total water mixing ratio (i.e., sum of vapor and liquid mixing ratios) of 20 g/kg find the virtual potential temperature. (Hint, you might want to employ a thermodynamic diagram.)

11) Given the following virtual potential temperature sounding, identify each layer (example, ML, RL, SBL, FA). Estimate what time of day that sounding was made, and then sketch the virtual potential temperature profile that you might expect four hours later.



- ***Momentum*** is mass times velocity (kg· m/s); thus, ***a Momentum flux*** is (kg· m/s)/(m2.s).

- ***Flux*** is the transfer of a quantity per unit area per unit time. In BL meteorology, we are often concerned with mass, heat, moisture, momentum and pollutant fluxes.

Eddy Flux

The fluid motion can transport quantities, resulting in fluxes.. Thus , the turbulence transports quantities too.

As a conceptual tool, suppose we examine a small idealized eddy near the ground on a hot summer day (see Fig 2.12a). The average potential temperature profile is usually superadiabatic in such surface layers. If the eddy is a swirling motion, then some of the air from position 1 will be mixed downward (i.e., w' is negative) , while some air from position 2 will mix up (i.e., w' is positive) to take its place. The average motion caused by turbulence is $\overline{w'} = 0$, as expected (from section 2.4.3).

The downward moving air parcel (negative w') ends up being cooler than its surroundings (negative θ', assuming that θ' was conserved during its travel), resulting in an instantaneous product $w'\theta'$ that is positive. The upward moving air (positive w') is warmer than its surroundings (positive θ'), also resulting in a positive instantaneous product $w'\theta'$. Both the upward and downward moving air contribute positively to the flux, $w'\theta'$; thus, the average kinematic eddy heat flux $\overline{w'\theta'}$ is positive for this small-eddy mixing process.

This important result shows that turbulence can cause a net transport of a quantity such as heat ($\overline{w'\theta'} \neq 0$), even though there is no net transport of mass ($\overline{w'} = 0$). Turbulent eddies transport heat upward in this case, tending to make the lapse rate more adiabatic.

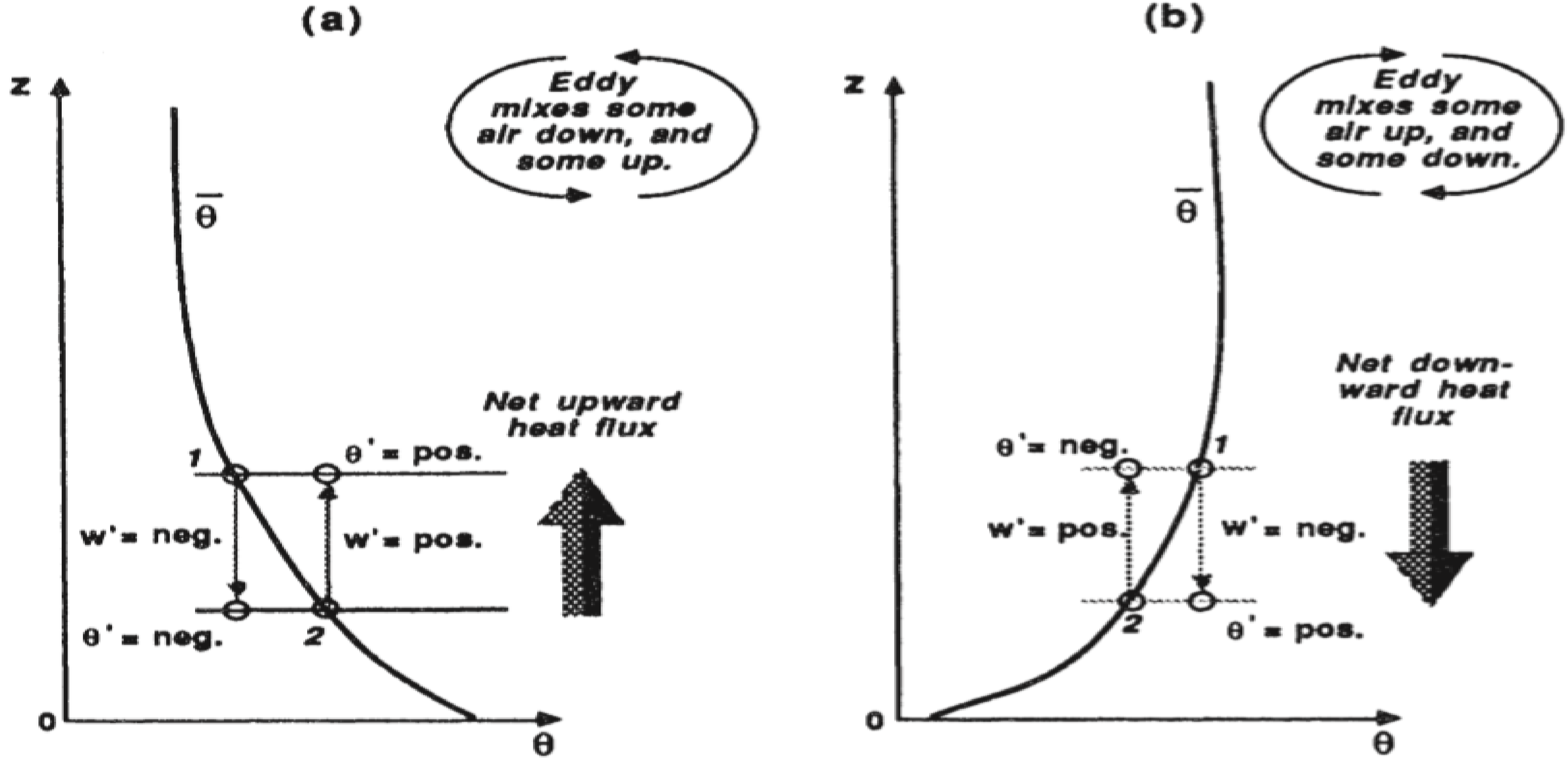


Fig. 2.12 Idealization of the small eddy mixing process, showing (a) net upward turbulent heat flux in a statically unstable environment, and (b) net downward turbulent heat flux in a stable environment.

Next, let's examine what happens on a night where a statically stable lapse rate is present (Fig 2.12b). Again, picture a small eddy moving some air up and some back down. An upward moving parcel ends up cooler than its surrounding (negative $w'\theta'$), while a downward moving parcel is warmer (negative $w'\theta'$). The net effect of the small eddy is to cause a negative $\overline{w'\theta'}$, meaning a downward transport of heat.

$$\text{Vertical kinematic eddy heat flux} = \overline{w'\theta'} \tag{2.7.1a}$$

$$\text{Vertical kinematic eddy moisture flux} = \overline{w'q'} \tag{2.7.1b}$$

$$\text{x-direction kinematic eddy heat flux} = \overline{u'\theta'} \tag{2.7.1c}$$

$$\text{Vertical kinematic eddy flux of U-momentum} = \overline{u'w'} \tag{2.7.1d}$$