If we compare the TKE equation (5.1) with the MKE equation (5.4b):

$$\frac{\partial(\mathrm{TKE/m})}{\partial t} = \ldots - \overline{u_i'u_j'}\,\frac{\partial \overline{U_i}}{\partial x_j}$$

$$\frac{\partial(\mathrm{MKE/m})}{\partial t} = \ldots + \overline{u_i'u_j'}\,\frac{\partial \overline{U_i}}{\partial x_j}$$

we see that they both contain a term describing the interaction between the mean flow and turbulence. The sign of these terms differ. ***Thus, the energy that is mechanically produced as turbulence is lost from the mean flow***, and vice versa.

5.5 Stability Concepts

Unstable flows become or remain turbulent. Stable flows become or remain laminar. There are many factors that can cause laminar flow to become turbulent, and other factors that tend to stabilize flows. If the the net effect of all the destabilizing factors exceeds the net effect of the stabilizing factors, then turbulence will occur. In many cases, these factors can be interpreted as terms in the TKE budget equation.

To simplify the problem, investigators have historically paired one destabilizing factor with one stabilizing factor, and expressed these factors as a dimensionless ratio. Examples of these ratios are the Reynolds number, Richardson number, Rossby number, Froude number, and Rayleigh number. Some other stability parameters such as static stability, however, are not expressed in dimensionless form.

5.5.1 Static Stability and Convection

Static stability is a measure of the capability for buoyant convection. The word "static" means "having no motion"; hence this type of stability does not depend on wind. Air is statically unstable when less-dense air (warmer and/or moister) underlies more-dense air. The flow responds to this instability by supporting convective circulations such as thermals that allow buoyant air to rise to the top of the unstable layer, thereby stabilizing the fluid. Thermals also need some trigger mechanism to get them started. In the real boundary layer, there are so many triggers (hills, buildings, trees, dark fields, or other perturbations to the mean flow) that convection is usually insured, given the static instability.

Local Definitions. The traditional definition taught in basic meteorology classes is local in nature; namely, the static stability is determined by the local lapse rate. The local definition frequently fails in convective MLs, because the rise of thermals from near the surface or their descent from cloud top depends on their excess buoyancy and not on the ambient lapse rate.

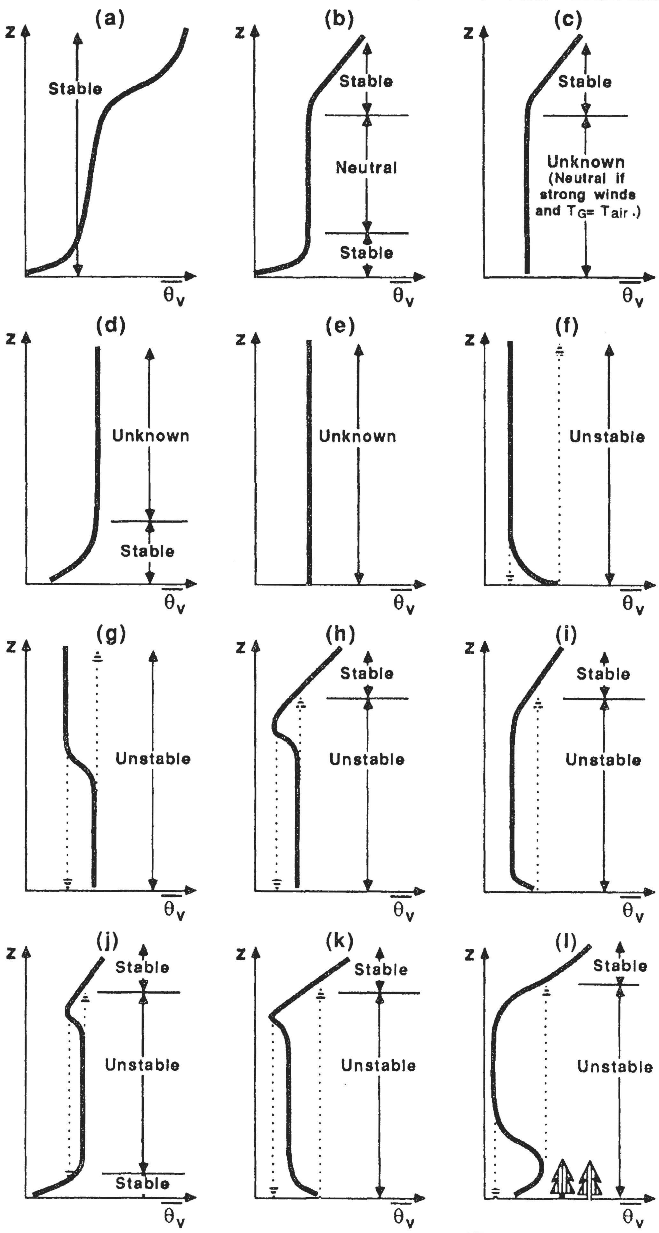

Fig. 5.17 Static stabilitiy as a function of the $\overline{\theta_v}$ profile. Dotted lines denote parcel movement.

As an example, in the middle 50% of the convective ML the lapse rate is nearly adiabatic, causing an incorrect classification of neutral stability if the traditional local definition is used. We must make a clear distinction between the phrases "adiabatic lapse rate" and "neutral stability". An ***adiabatic lapse rate*** (in the virtual potential temperature sense) may be statically stable, neutral, or unstable, depending on convection and the buoyancy flux. ***Neutral stability*** implies a very specific situation: adiabatic lapse rate AND no convection. The two phrases should NOT be used interchangeably, and the phrase "neutral lapse rate" should be avoided altogether.

We conclude that ***measurement of the local lapse rate alone is INSUFFICIENT to determine the static stability.*** Either knowledge of the whole $\overline{\theta_v}$ profile is needed (described next), or measurement of the turbulent buoyancy flux must be made.

Nonlocal Definitions. It is better to examine the stability of the whole layer, and make a layer determination of stability such as was done in section 1.6.4. For example, if $\overline{w'\theta_v'}$ at the earth's surface is positive, or if displaced air parcels will rise from the ground or sink from cloud top as thermals traveling across a BL, then the whole BL is said to be ***unstable*** or ***convective***. If $\overline{w'\theta_v'}$ is negative at the surface, or if displaced air parcels return to their starting point, then the BL is said to be ***stable***.

If, when integrated over the depth of the boundary layer, the mechanical production term in the TKE equation (5.1) is much larger than the buoyancy term, or if the buoyancy term is near zero, then the boundary layer is said to be ***neutral***. In some of the older literature, the boundary layer of this latter case is also sometimes referred to as an ***Ekman boundary layer***. During fair weather conditions over land, the BL touching the ground is rarely neutral. Neutral conditions are frequently found in the RL aloft. In overcast conditions with strong winds but little temperature difference between the air and the surface, the BL is often close to neutral stability.

In the absence of knowledge of convection or measurements of buoyancy flux, an alternate determination of static stability is possible if the $\overline{\theta_v}$ profile over the whole BL is known, as sketched in Fig 5.17. As is indicated in the figure, if only portions of the profile are known, then the stability might be indeterminate. Also, it is clear that there are many situations where the traditional local definition fails.

5.5.2 Example

Problem. Given the sounding at right, identify the static stability of the air at z = 600 m.

z (m)	$\overline{\theta_v}$(K)
1000	298
800	299
600	299
400	299
200	298
0	295

Solution. Using a local definition in the absence of heat fluxes, if we look downward from 600 m until a diabatic layer is encountered, we find a stable layer with cooler temperatures at 200 m. Before we reach any hasty conclusions, however, we must look up from 600 m. Doing so we find cooler unstable air at 1000 m. Thus, the static stability is unstable at 600 m.

Discussion. The whole adiabatic layer is unstable, considering the nonlocal approach of a cool parcel sinking from above. This sounding is characteristic of stratocumulus.

5.5.3 Dynamic Stability and Kelvin-Helmholtz Waves

The word "dynamic" refers to motion; hence, dynamic stability depends in part on the winds. Even if the air is statically stable, wind shears may be able to generate turbulence dynamically.

Some laboratory experiments have been performed (Thorpe, 1969, 1973; Woods 1969) using denser fluids underlying less-dense fluids with a velocity shear between the layers to simulate the stable stratification and shears of the atmosphere. Fig 5.18 is a sketch of the resulting flow behavior. The typical sequence of events is:

(1) A shear exists across a density interface. Initially, the flow is laminar.

(2) If a critical value of shear is reached (see section 5.6), then the flow becomes dynamically unstable, and gentle waves begin to form on the interface. The crests of these waves are normal to the shear direction

(3) These waves continue to grow in amplitude, eventually reaching a point where each wave begins to "roll up" or "break". This "breaking" wave is called a ***Kelvin-Helmholtz (KH) wave***, and is based on different physics than surface waves that "break" on an ocean beach.

(4) Within each wave, there exists some lighter fluid that has been rolled under denser fluid, resulting in patches of static instability. On radar, these features appear as braided ropelike patterns, "cat's eye" patterns or breaking wave patterns.

(5) The static instability, combined with the continued dynamic instability, causes each wave to become turbulent.

(6) The turbulence then spreads throughout the layer, causing a diffusion or mixing of the different fluids. During this diffusion process, some momentum is transferred between the fluids, reducing the shear between the layers. What was formerly a sharp, well-defined, interface becomes a broader, more diffuse shear layer with weaker shear and static stability.