Static Stability—An Update

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Abstract

Static stability should not be evaluated from the local lapse rate. There is a growing body of observations, such as within portions of mixed layers and forest canopies, showing that the whole sounding should be considered to evaluate stability. Air parcels can move across large vertical distances to create turbulence within regions that would otherwise have been considered stable or neutral according to classical local definitions. A nonlocal determination of static stability is presented that accounts for both the local lapse rate and for convective air parcels moving across finite distances. Such a method is necessary to properly estimate turbulence, dispersion, and vertical fluxes that affect our weather and climate forecasts. Teachers of introductory meteorology courses and textbook authors are encouraged to revise their static stability discussions to follow air parcel displacements from beginning to end.

1. Introduction

Lapse rate, the decrease of temperature with height, has long been used as an indicator of static stability (Petterssen 1940; Haltiner and Martin 1957; Byers 1959; Hess 1959). The classic usage for unsaturated air is given in Table 1, where Θ_v is the virtual potential temperature, and z is height. In dry environments, or in situations where the mixing ratio is relatively constant with height, the potential temperature Θ is usually used instead of Θ_v. For this case, the three stability classifications correspond to: superadiabatic $\partial T/\partial z < \Gamma_d$; adiabatic $\partial T/\partial z = -\Gamma_d$; and subadiabatic $\partial T/\partial z > \Gamma_d$; where T is temperature and Γ_d is the dry adiabatic lapse rate of 9.8°C km^{-1}. A temperature inversion, where $\partial T/\partial z > 0$, is also subadiabatic. The slice method of Bjerknes (1938), which accounts for environmental movement as required by continuity, is a bit more realistic, but can be written (Hess 1959) in a quasi-local form similar to Table 1.

The purpose of this paper is: 1) to discourage use of Table 1 because it is inconsistent with observations, and 2) to suggest a nonlocal approach that more accurately accounts for both the local lapse rate and convective air parcels moving across finite distances. Meteorologists engaged in studies of dispersion, air quality, turbulence, convection, gravity waves, cloud dynamics, severe weather, mountain meteorology, boundary layers, micrometeorology, and agricultural meteorology might find this nonlocal approach useful.

2. Instability and turbulence

In meteorology we are concerned with a variety of measures of stability (static, dynamic, conditional, baroclinic, inertial, slant, etc.), because instability can induce secondary motions that transport heat, moisture, and pollutants, and thereby affect weather and climate forecasts. The secondary circulations can be on the scale of extratropical cyclones responding to baroclinic instability across a hemisphere, thunderstorms responding to conditional instability across the troposphere, thermals responding to static instability across the mixed layer, or small eddies responding to dynamic instability induced by shears near the surface or behind obstacles to the wind. In this paper, we focus on static stability and the associated convective turbulence.

Turbulence is an irregular motion that partially undoes the effects of the instability, analogous to LeChatelier's principle of chemistry (Weast 1968). Instability is usually created within the flow by external forcings that alter one part of the flow relative to the other. For example, radiation might cool cloud top but not cloud base, or friction might retard the wind near the ground but not at higher altitudes. Turbulence then develops to reduce the instability by making the flow state more uniform, for example, by mixing temperature or momentum. When the instability is sufficiently reduced, turbulence decays and leaves the flow in a slightly altered nonturbulent state. In the atmosphere, long-duration forcings such as differential radiative heating or cooling, conduction from the surface, or differential advection can continually destabilize the air, driving turbulence for long periods of time.

TABLE 1. Traditional determination of static stability for unsaturated air.

	Local Lapse Rate		Local Static Stability
Name	$\partial\Theta_v/\partial z$	$\partial T/\partial z$	
superadiabatic	negative	$< -\Gamma_d$	unstable
adiabatic	zero	$= -\Gamma_d$	neutral
subadiabatic	positive	$> -\Gamma_d$	stable

The generic definition of flow stability says that air is unstable if infinitesimal perturbations grow, to paraphrase the *Glossary of Meteorology* (Huschke 1959). The perturbation can be a small-amplitude wave, or a small displacement of an air parcel. If wave amplitudes grow, or if air parcels tend not to return to their starting point, then the flow is unstable and becomes turbulent.

The crux of the problem is that the eventual flow response can grow well beyond the initial infinitesimal perturbation. Most explanations of static stability start by displacing an air parcel an infinitesimal distance from an arbitrary starting height within the ambient environment. This is a logical approach that closely follows the generic definition of stability. However, in unstable conditions the parcel can continue to rise or sink across large distances, forced by the buoyancy of the parcel with respect to the environment. This parcel can buoyantly move through neighboring regions having locally adiabatic or subadiabatic lapse rates, that would have traditionally been classified as neutral or stable. Given the observation of vigorous convective turbulence within these regions, we must conclude that the local lapse rate gives us incomplete information about the response of the flow.

There appears to be a paradox between the commonly used local definitions of static stability and the practical techniques to determine the turbulent state of the air. Some meteorologists already utilize nonlocal approaches whenever they estimate the buoyancy of an air parcel by comparing its temperature with that of the surrounding environment. For example, in thunderstorm situations an air parcel might rise from the surface to the midtroposphere, where its continued instability is measured by the parcel–environment temperature difference, not by the local lapse rate of the environment. The midtroposphere region might be highly turbulent at mesoscales due to the thunderstorm convection, even though the local environmental lapse rate is subadiabatic. Such parcel versus environment measures of static stability are also appropriate and, indeed, necessary for unsaturated situations.

Other meteorologists apparently consider lapse rate and stability to be virtually synonymous. The scientific literature is filled with improper statements such as "unstable lapse rate" or "neutral lapse rate," where the stability adjectives "unstable" or "neutral" are inappropriately used to describe the lapse rate. Indeed, this paradigm has reached some textbooks. It is recommended here that the concepts of lapse rate and stability be separated. Meteorologists should take care to use the proper terminology such as "superadiabatic lapse rate" or "adiabatic lapse rate" when describing local temperature gradients, and use the terms "unstable" or "neutral" only when describing stability.

For the present discussion of static stability, it is important to separate the concept of movement of displaced parcels from movement of the ambient mean wind. Although additional destabilizing factors such as mean wind shear (dynamic instability) can compete with or add to static forces to make flow unstable, this paper focuses on only pure static stability. Thus, ambient wind shear is not considered here.

There is a growing list of situations, discussed in section 3, for which the local virtual-temperature lapse-rate measure of stability fails to properly indicate the presence or potential for turbulence. Similar difficulties occur in the oceans, where density would be used in place of virtual potential temperature. A proper definition of static stability, such as is given in section 4, should consider the whole sounding and not just a local segment. With this definition, many apparently paradoxical or unphysical observations, such as countergradient heat fluxes, can be logically explained. It is hoped that these advances in our understanding of stability and turbulence will be useful to the broader atmospheric and oceanic science communities, where there is a real need to know the turbulent state of the flow.

3. Observations of nonlocal effects

Large eddies almost always exist in the turbulent portions of the atmosphere. A turbulent structure or eddy is considered to be large if its size is on the same order as the turbulent domain, or of the same order as the mean flow scales associated with the production of turbulence. Examples include convective thermals of the same 1- to 2-km diameter as the mixed-layer depth, mechanical eddies of the same size as the 100-m-thick shear region of the surface layer, wake turbulence vortices of the same size as the flow obstacle, and convective plumes and microfront curtains of the same thickness as the convective surface-layer depth. Spectral decomposition of turbulent fields shows that the larger-size eddies contain most of the turbulence energy and are often responsible for the greatest transport.

We can see these large coherent structures with our eyes and with special sensors. Smoke from stacks can be seen looping up and down in convective mixed layers. Fair-weather cumulus clouds form within the tops of convective thermals, and make visible the large diameters of the thermals. Shear-driven eddies are visible in the swirling snow. Blowing leaves swirl in wake vortices behind buildings and vehicles. Remote sensors such as lidar, radar, and sodar also detect and photograph aerosol, moisture, and temperature structures associated with these eddies. Specially instrumented towers and aircraft can measure temperature,

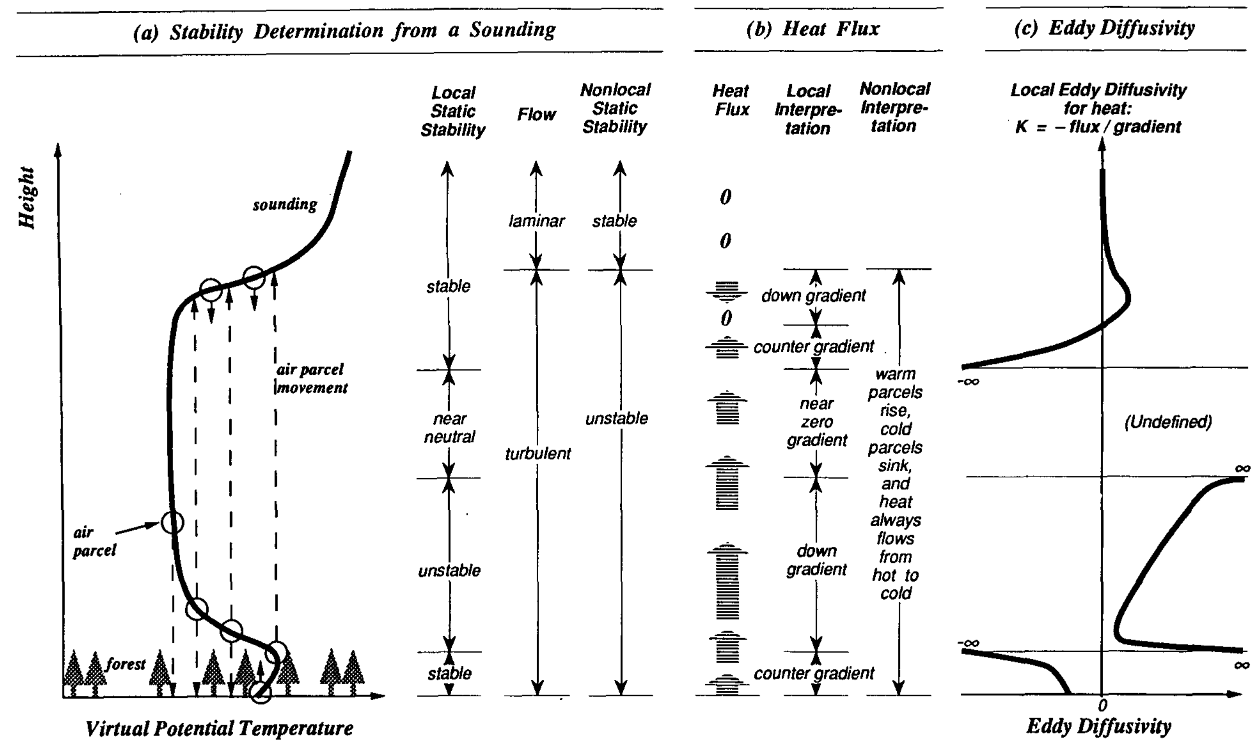


FIG. 1. Local vs. nonlocal interpretation of stability and flow. (a) Typical observed virtual potential temperature sounding (heavy solid line) showing a convective mixed layer over a forest canopy. Small circles represent air parcels, and dashed lines show parcel movement. Local static stability is based on local lapse rate; nonlocal stability is based on parcel buoyancy and movement; and flow state is based on nonlocal stability. (b) Typical observed heat-flux magnitude and sign are shown schematically by length and direction of the wide shaded arrows; local interpretation is made comparing the observed flux to the observed local gradient; and nonlocal interpretation is based on air parcels. (c) Local eddy diffusivity calculated from the observed heat flux and temperature gradient. The local stability, local heat-flux interpretation, and local eddy diffusivity are unrealistic and/or inappropriate for the true flow state.

humidity, and wind deviations within the eddy structures, using conditional sampling techniques.

Local interpretations of static stability can lead to improper expectations about turbulence intensity and pollutant dispersion for flows containing large eddies. For example, the average potential temperature is approximately constant with height near the middle of the convective mixed layer (Fig. 1a). If stability is estimated from local conditions, then the adiabatic lapse rate there would be incorrectly associated with statically neutral conditions, moderate turbulence, and coning smoke plumes. Similar problems occur just above the middle of the mixed layer, where the local increase of potential temperatures with height would incorrectly suggest statically stable conditions, weak turbulence, and fanning smoke plumes. Both classifications are inconsistent with the observations of statically unstable conditions, vigorous convective turbulence, and looping smoke plumes in both regions.

The so-called Zeroth Law of Thermodynamics (heat flows from hot to cold; Perry et al. 1963) is apparently violated when nonlocal processes are ignored. The heat flux is observed to be upward (positive) in both the middle of the mixed layer and just above the middle (Fig. 1b). When only the local lapse rate is considered, this leads to the improper conclusions that heat flows where there is no gradient in the middle mixed layer, and flows from cold to hot in a countergradient process at higher altitudes (Deardorff 1966).

Theories such as eddy-diffusivity theory (K-theory) require unrealistic interpretations of the physics for flows with large eddies. K-theory assumes that fluxes are associated with only small-size eddies, and thus occur down the local gradient in a manner analogous to molecular diffusion. Given observations of temperature and vertical heat-flux profiles, one can solve for the apparent eddy diffusivity: K = –flux/gradient. The resulting values of K are plotted schematically in Fig. 1c for a typical mixed layer. At some heights, K is negative; at others it is infinite. Weil (1990) notes similar problems with the more-sophisticated top-down bottom-up K-theory. Negative or infinite diffusivities are not consistent with the basic diffusion analogy upon which K theory is based. In fact, infinite or discontinuous parameter values in any theory are clues that the theory is not scientifically appropriate for the situation. An eddy-diffusivity closure approximation, such as sketched in Fig. 1c, that utilizes the local temperature gradient of Fig. 1a is clearly not a useful parameterization for finding heat flux (Fig. 1b) for such a convective situation. Although such problems with K-theory have been known for years, K-theory continues to be used in inappropriate situations.

Similar inconsistencies are observed in forests heated by the daytime sun. The horizontal or time-average potential temperature is observed to increase with height from the ground to near the forest canopy top (Fig. 1a). Local interpretations would incorrectly suggest that the air is statically stable (Fig. 1a), smoke would fan, vertical mixing is weak, the heat flux is countergradient (Fig. 1b), and the eddy diffusivity is negative (Fig. 1c). However, observations (Gao et al. 1989) show that heat flux is positive in that region, and is associated with nonlocal convective sweeps of cooler air penetrating the canopy from above. Furthermore, bulk transfer or drag approximations that are a function of local stability (e.g., a function of lapse rate or Richardson number) would yield surface fluxes that are too small for this forest case.

These inconsistencies can be eliminated and observations explained with nonlocal measures of static stability and flow. The flow sketched in Fig. 1a is indeed turbulent throughout the mixed layer and forest canopy because the static stability is nonlocally unstable, as exhibited by buoyant parcel movements. The dashed lines in Fig. 1a connect the starting points of displaced parcels (sketched by the small circles) to their neutral-buoyancy ending points (at the arrow tips). The subregion of warm rising parcels from the forest top overlaps with the subregion of cold sinking parcels from the mid–mixed layer, leaving the whole layer unstable and turbulent. Within the forest canopy, the cold parcels sinking from the mid–mixed layer (a positive heat-flux contribution) more than offset the cool parcels forced to rise from the forest floor by continuity (a negative heat-flux contribution), yielding the observed net positive heat flux. Warm buoyant parcel rise also causes the positive heat flux just above the mid–mixed layer, regardless of the opposing local temperature gradient.

In light of these observations, it is clear that a different method should be used for determining static stability that considers both the local lapse rate and nonlocal parcel movement, and that nonlocal turbulence parameterizations should be developed for weather and climate forecast models. Concerning the turbulence parameterizations, such proposals have already been presented in the literature, including the spectral diffusivity of Berkowicz and Prahm (1979) and the transient turbulence theory of Stull (1984; Ebert et al. 1989). Concerning a definition of static stability that works regardless of eddy size, such an updated definition is discussed next.

4. A nonlocal definition of static stability

Four preliminary steps are necessary prior to determining static stability:

1) The whole environmental sounding should be known. The vertical bounds of this domain should be at solid surfaces, or at heights where nonlocal parcel vertical motions are known not to be significant. For example, the bounds can be the ground and the stratosphere. For a simplified presentation in a textbook, the bounds could be two hypothetical solid surfaces. Soundings that do not end at such bounds lead to uncertainty in the stability determination, as described later.

2) Conceptually, air parcels should be displaced a small distance initially up and down from all possible starting points within the sounding domain. Practically, we need only test (displace) parcels up from the relative maxima and down from the relative minima of virtual potential temperature in the sounding.

3) Air parcel movement after the initial displacement should be based on parcel buoyancy. This buoyancy is measured by comparing virtual temperatures of the parcel and the environment at the same height. In dry environments, or environments of nearly constant mixing ratio with height, the temperature can be used instead of virtual temperature for simplicity. Buoyant motion is associated with cold parcels that sink and warm parcels that rise.

4) Displaced air parcels that would continue to move buoyantly from their origin should be tracked all the way to their ending point of neutral buoyancy (neglecting overshoot), which is the height where the parcel virtual temperature equals the environmental virtual temperature. By neglecting overshoot, we ignore the possibility that parcel kinetic energy could allow nonbuoyant penetration through thin stable layers toward regions where the parcel is again buoyant. This happens in nature to some extent, and could be utilized in a more sophisticated nonlocal definition of stability (not pursued here).

Once parcel movements have been tracked, the stability of portions of the sounding domain must be determined in the following order:

a) Unstable: those regions within which parcels can enter and transit under their own buoyancy; that is, regions where they move in the same direction as their buoyant force vector. This works even if the parcel originated a finite distance away from its destination. Note that individual parcels need not traverse the whole unstable region. For example, if a subregion is buoyantly traversed by one air parcel, and a partially overlapping subregion is traversed by a different buoyant air parcel, then the whole region formed as the union of those two subregions is identified as "unstable." The special case of contiguous but nonoverlapping unstable subregions can occur in nature and laboratory tanks (e.g., two turbulent mixed layers separated by a strong temperature or density step), and should be classified as adjacent but independent unstable layers.

b) Stable: those regions of subadiabatic lapse rate that are not unstable.

c) Neutral: those regions of adiabatic lapse rate that are not unstable.

d) Unknown: those top or bottom portions of the sounding that are apparently stable or neutral but do not end at a material surface such as the ground. The reason is that above or below the known sounding region might be a region of cold or warm temperature, respectively, that could provide a source for buoyant parcels.

An efficient two-step method for determining the stability of a sounding is to use the classical local-lapse-rate definitions to make a first guess for all layers. If any layers are unstable according to the classical definition, then a parcel from the bottom of the unstable layer should be lifted until it becomes neutrally buoyant. Similarly, a parcel from the top of the first-guess unstable layer should be lowered to its equilibrium level. The domain of such buoyant parcel movement takes precedent over any first-guess stable or neutral layers, and defines the nonlocally unstable regions.

The nonlocal stability for saturated air can also be determined using the above definitions. Namely, parcel temperature change should follow the moist or pseudo-adiabat (as appropriate for the assumptions regarding liquid-water fallout) over that portion of its vertical motion for which it is saturated. As before, the unstable regions are defined as the domains in which parcels move in the same direction as their buoyancy force, for parcels that were initially displaced only an infinitesimal distance.

5. Examples

a. Idealized examples

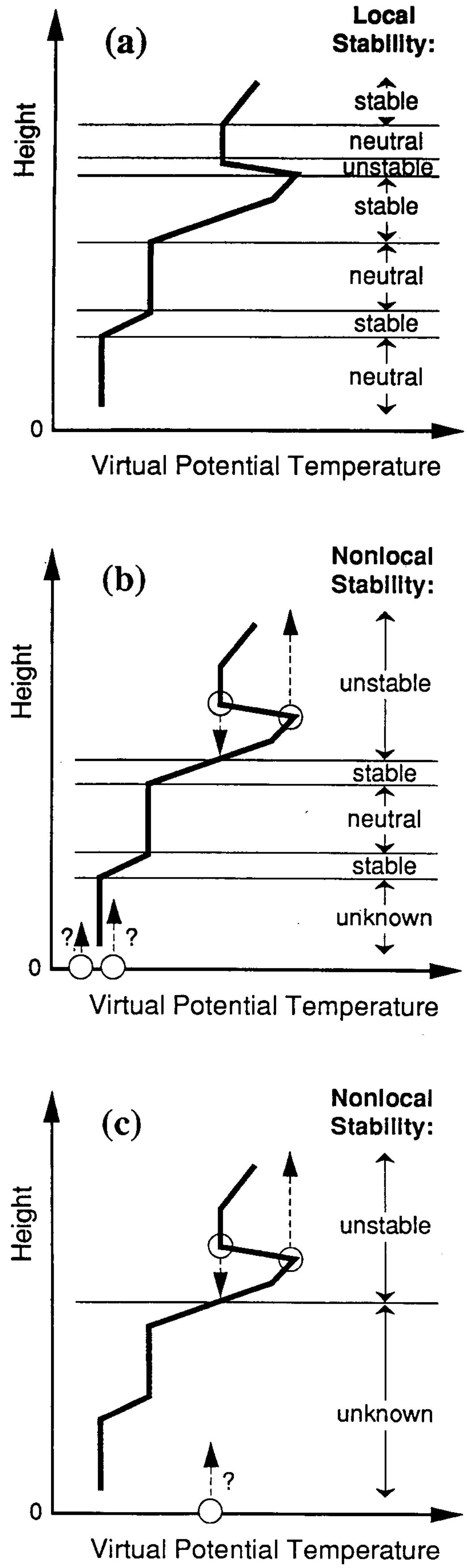
The nonlocal definition of static stability appears somewhat complex, but is clearer when plotted in an example. An idealized sounding is plotted in Fig. 2a, along with the classical "local" static stabilities. Sample buoyant-parcel movements and nonlocal-stability descriptions are shown in Fig. 2b. The top "unstable" region consists of two overlapping subregions of buoyant-parcel movement. Even though the bottom of the "unstable" region is locally stable to parcel displacement, cold parcels from above would tend to sink through it, creating convection and turbulence. This exemplifies how parcels from different origins can give different stability classifications at the same destination, and how the "unstable" classification dominates.

The middle adiabatic layer in Fig. 2b is trapped between two stable layers that insulate it from buoyant parcels that might otherwise penetrate into it. This adiabatic layer thus is classified as "neutral." Such a sandwich configuration is an easy clue to help discriminate between neutral layers and unstable mixed layers. The "bread" of this neutral sandwich can either be stable layers or surfaces such as the ground or sea surface.

The bottom adiabatic layer of Fig. 2b is classified as "unknown," because the end of the sounding does not terminate at the ground. Between the lowest point of the sounding and the ground might be a superadiabatic layer caused by solar heating of the earth, which would be a source region for buoyant thermals, and which would require that the adiabatic layer be classified as "unstable." If there is external evidence that there are no buoyant thermals across that lowest adiabatic region, then the classification could be changed to "neutral." Such evidence could include net radiative heating or cooling, turbulence measurements, observations of smoke plume dispersion, or knowledge that the earth's surface is colder. In particular, ground-surface temperature measured with either radiation sensors or in situ devices can be compared with air temperature to estimate static stability of the air adjacent to the surface. However, in the absence of such additional knowledge, the honest classification is "unknown."

The careful reader might note that if a very strong superadiabatic layer exists at the ground, then the whole sounding could be "unstable." Thus, Fig. 2b is not accurate, and would be better labeled "unstable" in the present unstable part and "unknown" everywhere else (see Fig. 2c). Other readers might protest that we would be neglecting potentially useful information by having such a large "unknown" region. However, that "unknown" region could indeed have vigorous convective turbulence if conditions near the surface were right. This is the reason for desiring in step 1 in section 4 that the sounding be plotted all the way to solid surfaces, because to not do so introduces substantial uncertainty into the static-stability determination of much of the sounding. These principles are demonstrated in additional sounding examples presented by Stull (1988) in his Fig. 5.17, with the correction that his part "f" should be "unstable."

Figure 1 provides us with additional idealized examples for the determination of static stability. Air parcels shown in the middle of the superadiabatic region just above tree top are unstable if lifted or lowered. If lifted an infinitesimal amount from their starting point, they are warmer than the surrounding



environment and continue to rise a finite distance to the top of the mixed layer. When lowered, those parcels sink all the way to the surface because they are colder than the environment (regardless of the local lapse rate within the forest canopy). Parcels close to tree top, however, are statically unstable if lifted, but stable if lowered. By conceptually tracking parcels from all possible starting heights of the sounding for this example, we find that some parcels move finite distances under their own buoyancy, while others tend to resist displacement. The extent of the nonlocally statically unstable region for this example is the region between the surface and the inversion capping the mixed layer, because that is the contiguous domain covered by the superposition of all parcel displacements moving under their own buoyancy.

In some parts of the boundary layer, turbulent motions appear anisotropic in even one dimension (Ebert et al. 1989). For example, the air parcels shown at the top of the turbulent mixed layer in Fig. 1 resist displacement over small distances. However, continuity of air requires that some of these parcels move downward in response to the warmer rising parcels that take their place. This is just the entrainment process that is known to occur near the mixed-layer top. The net result is that upward-moving parcels come a great distance from near the surface or tree top, while downward-moving parcels move a shorter distance. Analogous but inverted asymmetries occur near tree top, or near the surface if there are no trees. These asymmetries are closely tied with the finite distances traveled by the air parcels.

b. Real case studies

Turning to examples based on real data, Fig. 3 shows a virtual potential-temperature sounding measured on 15 June 1983 by the National Center for Atmospheric Research (NCAR) Queen Air aircraft (N306D) during the 1983 Boundary Layer Experiment (BLX83, Stull and Eloranta 1984). This was a 12-min slant descent flight roughly 40 km long over relatively flat farm and pasture land near Chickasha, Oklahoma. If we were to examine only the aircraft sounding (solid line with error bars) and utilize the classical stability definition based on local lapse rate, we would incorrectly guess that the bottom 500 m are nearly neutral, the region between 500–1700 m is slightly stable, and the remainder is more strongly stable.

However, there is abundant additional information that leads us to the correct conclusion that the bottom 2000 m are nonlocally unstable. First, synoptic conditions were favorable for strong solar heating of the ground, with an anticyclone of 1018-mb sea level pressure centered over the region. Skies

FIG. 2. Idealized examples of static-stability determination. Small circles represent air parcels, and dashed lines show buoyant-parcel movement. (a) Traditional determination of stability using the local lapse rate. (b) Likely nonlocal stabilities if surface heating or cooling is small. (c) Honest stability classification if the magnitude of surface heating is truly unknown.

were mostly clear during the period of the sounding. A pyranometer attached to the NCAR PAM II Portable Automated Mesonet surface weather station (at 35°00'27.2"N, 97°52'22.4"W) recorded 1000 (±20) W m^{-2} insolation. A nearby net radiometer operated by Argonne National Laboratory (ANL) measured 700 (±10) W m^{-2} net radiation near the ground. Any of these factors by themselves would provide evidence that solar heating of the ground could trigger warm buoyant thermals to rise.

More direct evidence shows that the surface and the near-surface air did become hot. A downward-looking infrared thermometer (PRT-5 radiometer) on the aircraft measured a radiometric ground-surface skin potential temperature of 37 (±4)°C, averaged over a 20-km-long flight leg flown 50 m above the surface. This is plotted as a large black square data point in Fig. 3. The PAM station below the aircraft recorded a 30-min average virtual potential temperature in the air, at 2 m above ground, of 31.6 (±0.6)°C, also plotted as a large black square data point. These warm temperatures were indeed associated with a large positive (warm air upward) turbulent sensible heat flux: 100 (±20) W m^{-2} measured by the aircraft at roughly 50 m above ground, and 96 (±5) W m^{-2} measured by ANL on a mast 10 m above a uniform pasture, both using eddy-correlation methods.

The most direct evidence of unstable conditions and nonlocal convection was from lidar observations of aerosol-laden thermals rising from near the surface to peak heights of about 2000 m above ground, with an average mixed-layer depth of about 1650 m (Nelson et al. 1989). Aircraft measurements during the same period confirmed the presence of a convective mixed layer as indicated by the large turbulence kinetic energy, high humidity, and a linear decrease of kinematic heat flux with height. The observation of 5% to 10% cumulus clouds was an easy visual clue of convection.

For evidence of nonlocal countergradient heat fluxes across forest and crop canopies, and the associated turbulent coherent structures causing sweeps, ejections, and temperature-ramp structures, the reader is referred to papers by Leclerc et al. (1990), Gao et al. (1989), Shaw et al. (1989), Dlugi et al. (1989), Raupach et al. (1986), Finnigan (1985), and Li et al. (1985).

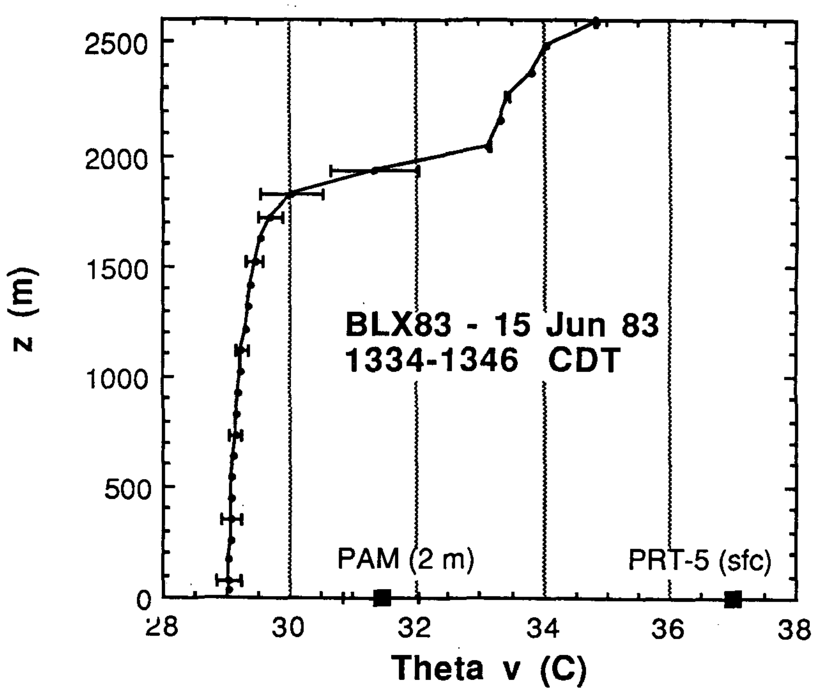
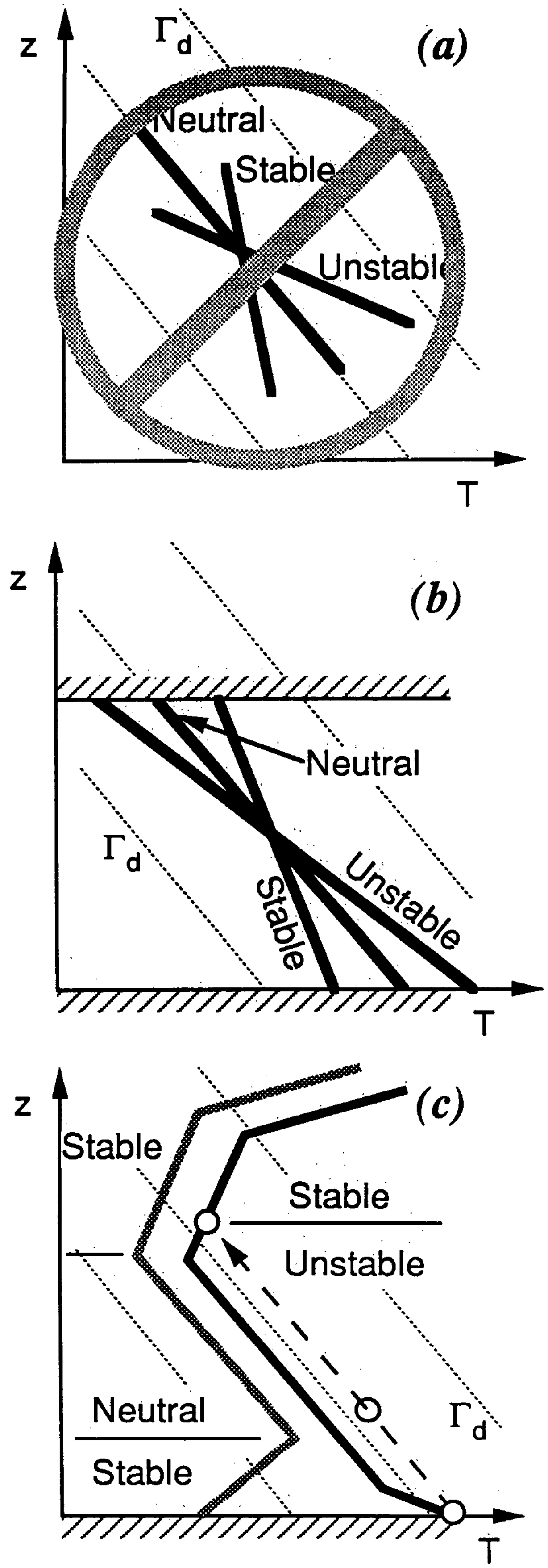


FIG. 3. Virtual potential-temperature sounding (solid line) observed by aircraft during the BLX83 field experiment in Oklahoma. Error bars indicate one standard deviation of data scatter. Large square data points are estimates of surface radiometric skin temperature (PRT-5) and surface air temperature measured 2 m above ground (PAM).

6. Discussion of appropriateness

There are many mechanisms that can destabilize the flow in a way that allows nonlocal convection. We have already discussed heating of the bottom of the boundary layer resulting in rising warm thermals, and heating at tree-top level allowing cold "thermals" from above to penetrate downward into the forest. Another scenario is associated with altocumulus castellanus clouds caused by warm-air advection under a cooler layer of air. A stratus or altostratus deck that experiences radiative cooling at cloud top and heating at cloud base can become convective. Cold-air advection over a mountain ridge can overrun warmer air to the lee of the mountains, creating instability that can cause deep convection and thunderstorms. All of these unstable regions would probably become more uniformly mixed with time under the action of convective turbulence; however, instability and turbulence could be maintained if there were continued destabilizing forcings.

The question of appropriateness also allows us to address averaging time limitations based on intermittency and patchiness of the turbulence or secondary circulations. For example, what is the static stability of a mesoscale region possessing several thunderstorms that buoyantly rise through an otherwise quiescent and locally statistically stable environ-



ment? Is the whole mesoscale region statically unstable and convective? The answer is probably "yes" if one averages over several thunderstorm lifetimes. A similar conclusion is reached for the boundary layer when one averages over several typical thermal lifetimes. However, when averaged over significantly shorter periods than the lifetimes of the relevant convective element, the environment for both examples might be considered neutral or stable. Thus, we must evoke a limitation that the nonlocal static-stability classification is appropriate for the mean flow only when averaged over time periods greater than convective element lifetimes of the dominant eddies.

Nonlocal measures of static stability reduce to traditional local measures in those real flows for which complete buoyant-parcel movements from origin to destination are small. In particular, neutral and stable classifications utilize local lapse rates for both the nonlocal and the traditional methods. Thus, the nonlocal definition is a general approach with wide applicability.

7. Conclusions and recommendations

A proper definition of static stability should consider the whole sounding, not just a local segment. Given a sounding, air parcels starting from the relative maxima and minima of the virtual potential-temperature profile should be conceptually displaced up and down, and their subsequent buoyant motion tracked until they stop. Buoyancy is determined by comparing the virtual temperature (or density in the case of ocean soundings) of the parcel to that of the environment at the same height. The union of domains traversed by all such parcels moving under the action of their own buoyancy away from their source to their destination defines the "unstable" regions. These are regions where parcel movement is in the same direction as the buoyant force. The remaining regions can be classified as "stable" or "neutral" using traditional local-lapse-rate methods. If parcels from different source locations yield different stability classifications at a destination, the "unstable" classification dominates.

The sounding should be bounded by solid surfaces, such as the ground, or by regions such as the stratosphere where convection is not anticipated. The stability

FIG. 4. Explanations of static stability that are improper (a) can be replaced by idealized figures having bounded domains (b), or by example soundings that extend from the ground to the stratosphere (c), where circles and dashed arrows represent air parcel movement, and the horizontal lines separate regions of different stability. Thin sloping dotted lines are dry adiabats. Thick lines represent portions of atmospheric soundings. Two separate soundings are plotted in (c).

of the bottom or top portion of an incomplete sounding that does not intersect such a surface introduces uncertainty into the determination of the true stability, suggesting a classification of "unknown." Meteorologists with data at a limited set of heights, such as from an instrumented tower, are advised to incorporate other observations (e.g., turbulence, dispersion) or known boundary conditions (e.g., surface heating or cooling, or even time of day and season) to try to eliminate these unknown regions.

The unstable region exhibits turbulence or other secondary circulations that can alter the vertical fluxes of variables and thereby impact weather and climate forecasts. The designation of a region as nonlocally unstable is appropriate if the averaging time defining the mean flow in that region is greater then the lifetimes of individual convective elements.

Examples of the failure of classical local estimates of static stability were given in section 3. It was also shown that air parcel movement is properly tracked in some disciplines such as thunderstorm meteorology, but is not incorporated into the general definition of static stability. It would seem appropriate to make our definition of static stability consistent with our present understanding and practical usage.

Textbook authors are encouraged to incorporate this revised view of static stability into their new editions. They are recommended to leave out the usual local-lapse-rate diagrams and any discussion of the stability of superadiabatic, adiabatic, or subadiabatic lapse rates. Existing figures can easily be modified by extending sounding segments upward and downward until the domain is bounded by solid surfaces or strong inversions (Fig. 4). Authors and teachers are welcome to copy and use Figs. 2 and 4.

Theoretical meteorologists studying statically stable or neutral conditions should be careful before they apply their theories to the real atmosphere. For example, a nonlocal determination of static stability across a subadiabatic region should be made before internal gravity-wave theories are applied there. If there is convective turbulence across the subadiabatic region, then gravity waves might not exist, or might undergo strong nonlinear interactions with the turbulence.

Acknowledgments. The National Science Foundation (NSF) supported this research under grants ATM-8211842 and ATM-8822214. Thanks also go to Pal Arya, John Sigmon, and Qing Zhang for reviewing the manuscript. Data for the case study was provided by Ed Eloranta, Tim Crum, Nick Wilde, Craig Kunitani, and Eric Nelson. The NCAR field observing facility (sponsored by NSF) is gratefully acknowledged for its aircraft, PAM, and raob support during BLX83.

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