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ATMOSPHERIC STABILITY AND THE DISPERSION OF PESTICIDES

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ABSTRACT. Applicators have long recognized that temperature inversions dramatically influence the behavior of pesticide material released into the atmosphere during pesticide application. This behavior can be explained in terms of the relationship between vertical atmospheric temperature gradients (stability) and atmospheric mixing (turbulence). This relationship is not intuitively obvious, but once it is understood, it provides the applicator with a tool to use for anticipating pesticide drift potential. The term inversion is synonymous with a stable thermal layer in the atmosphere. In stable thermal layers, mixing is suppressed and both dispersion and translation (or mean movement) of material is slow. This is typical of clear nights. Conversely, under clear, sunny conditions mixing is strong and dispersion is rapid. This situation indicates an unstable thermal layer.

INTRODUCTION

In this paper stability is defined as the variable that describes the vertical temperature structure of the atmosphere (dT/dz, where T is temperature and z is height). It encompasses the 3 cases of stable (inversion), neutral, and unstable (instable) atmospheres. This definition of stability is similar to that of "environmental lapse rate," which is the vertical distribution of temperature in the atmosphere at a given place and time. The reduction of atmospheric temperature into 3 subcategories of stable, neutral, and unstable establishes 3 discrete categories of atmospheric dispersion characteristics.

The pressure (p) increases with depth in the atmosphere and, therefore, so does temperature (T). The term potential temperature (θ) is used by meteorologists to describe the temperature at height with the pressure effect removed. In most pesticide application scenarios, it is not necessary to consider this effect, so T can be used instead of θ. This is because the entire material release and most of the dispersion takes place in a narrow vertical range in the atmospheric surface layer. In some applications, such as higher altitude space spraying for mosquitoes, it is necessary to use θ when considering stability.

The *Glossary of Meteorology* (Huschke 1959) defines instability (unstable), neutral stability, and static stability (stable) as:

> Instability—A property of the steady state of a system such that certain disturbances or perturbations introduced into the steady state will increase in magnitude, ...
>
> Neutral Stability—Under such conditions a parcel of air displaced vertically will experience no buoyant acceleration.
>
> Static Stability—The stability of an atmosphere in hydrostatic equilibrium with respect to vertical displacements, usually considered by the parcel method. The criterion for stability is that the displaced parcel be subjected to a buoyant force opposite to its displacement, e.g., that a parcel displaced upward be colder than its new environment.

The result of unstable and stable temperature profiles on ambient air motion is shown in Figs. 1 and 2. In Fig. 1, the unstable temperature profile causes vertically displaced air to move vertically away from its height of origin. In Fig. 2, the stable temperature profile causes the air to return to its height of origin.

STABILITY AND DISPERSION

To understand the role of atmospheric stability in dispersion, it is critical to understand the interaction among meteorological variables. A point that is often overlooked is that the surface of the earth has a large role in driving the lowest level of the atmosphere. Thermal energy is radiated by the sun, with the wavelength inversely proportional to the temperature of the radiating body, according to Wien's Displacement Law:

$$\Lambda_{max} = C/T_s, \quad (1)$$

where C is a constant and T_s is the temperature at the surface of the sun. This produces short-wavelength (high-frequency) energy that is transmitted through a clear atmosphere and absorbed by the surface of the earth. The surface of the earth heats due to the absorbed thermal energy and reradiates this energy at longer wavelengths (lower frequency) than those produced by the sun. The energy radiated from the surface of the earth is readily absorbed by the atmosphere. Thus, the atmosphere is largely heated from below.

On clear nights the surface loses heat according to Stefan's Law:

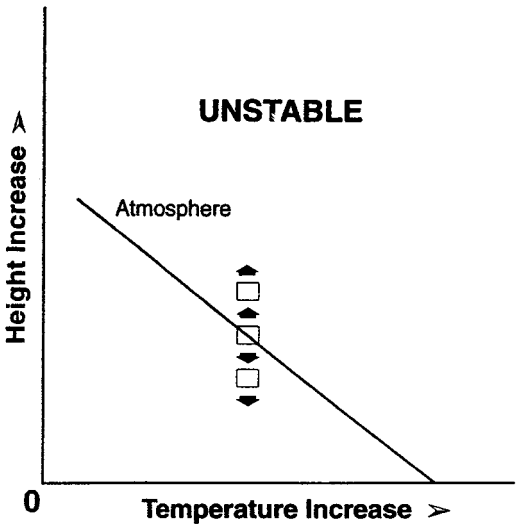

Fig. 1. Depiction of an unstable temperature profile.

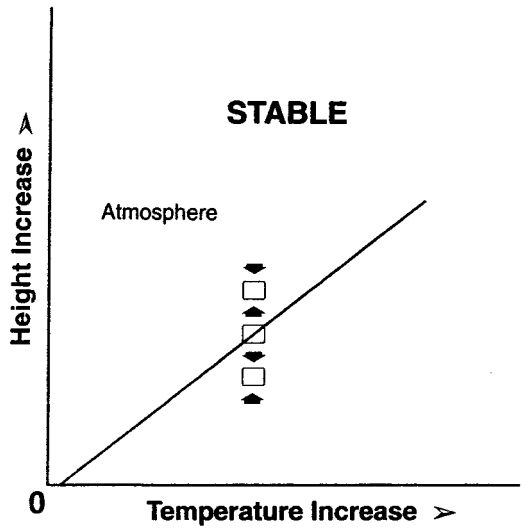

Fig. 2. Depiction of a stable temperature profile.

$$\Delta E = \sigma(\epsilon_1 T_1^4 - \epsilon_2 T_2^4), \quad (2)$$

where ϵ is the emissivity and σ is a constant, so that the surface of the earth (T_1) views outer space (T_2), which has a temperature approaching absolute zero. This causes the surface to rapidly lose heat, cooling the atmosphere from below under clear sky nocturnal conditions. Water vapor and clouds interfere with these relationships because they absorb and reflect radiated energy. Also, transmission, reflection, and absorption/emission of radiant energy vary on a continuous spectrum versus wavelength. This spectrum is a material property of the media encountered by the radiation. See Rosenberg (1974) for a more extensive introductory review of the topics covered in this section.

When the atmosphere is heated from below, the air underneath becomes less dense than the air above. This imparts a buoyant force to the air in the upward direction. Thus, the warmer air rises. As it does, it moves into colder air and rises faster until it is sufficiently mixed with the surrounding air and reaches an equilibrium position. Typically, the air overshoots the equilibrium position and immediately begins to fall. Also, as the warm air lifts off the surface, other air rushes in to replace it. All of this leads to the vigorous mixing that occurs under clear, sunny (unstable) atmospheric conditions.

After sunset, if the sky is clear, the ground cools and the air on the surface becomes colder than the air above. As the night progresses, the surface becomes colder through radiative heat loss and the air on the surface becomes colder through conduction. This colder (stable) layer deepens until sunrise.

For thermal stratification of the atmospheric surface layer to take place *in situ* (either stable or unstable), wind speeds must be low. If a strong (even moderate) wind is blowing due to regional pressure gradients, the thermal stratification will be obliterated and the surface layer will be mixed. Under these conditions, the atmospheric surface layer will tend toward neutral stability.

In the case of a neutrally buoyant gas, dispersion is markedly influenced by stability. In the unstable case, the gas is dispersed rapidly with pronounced vertical spreading, lifting, and injection into the upper air flow. In the neutral case, the gas is dispersed in a more regular (Gaussian-like) plume. In the stable case, dispersion is damped so that material may hang in the air near the source and remain fairly concentrated for long periods. This may be an undesirable situation because this relatively concentrated airborne material is subject to density flows and changing meteorological conditions. In some mosquito control situations, a lingering, suspended droplet cloud is the desired scenario.

DEFINING DRIFT

The preference of applicators regarding optimum release conditions is controlled by the objective of the application, the release scenario (aerial, boom sprayer, air blast, etc.), and the amount of drift desired in the application. Application objectives and release scenarios are discussed in the agricultural engineering, pesticide science, and entomological literature. However, the definition of drift in a given situation

impacts the way the applicator responds to the various atmospheric stability conditions.

In some applications, the concern is to avoid damage to neighboring, nontarget crops. In this case, movement of a large amount of material a short distance off-target will cause problems. This would call for avoiding stable atmospheres, because the slow dispersion rates cause fine spray particles to hang in a coherent cloud that can move off-target and remain at a relatively high concentration. If the target is a volume of atmosphere and the objective is to maintain a high concentration of suspended droplets, then a stable atmosphere is an advantage. In stable atmospheres, drift can cause a relatively high-concentration spray cloud to move off-target. Drift may result from the interaction between a stable atmosphere and terrain effects (density flows) as discussed below.

Some forest applicators would prefer to release spray over a forest canopy under relatively high wind, neutral atmospheric conditions. The energy of the wind causes the material to penetrate the canopy and affect in-canopy pests. In this scenario, a small amount of the fine spray material can drift off-target on the order of tens of kilometers. This preference has been developed in applications where the treated acreage is extensive and human population density is low. In this example, a small amount of material drifts a long distance due to the high wind energy.

Unstable conditions could probably result in the longest distance drift of a tiny proportion of the released material because strong updrafts caused by surface heating can inject material into the upper air flow.

TURBULENCE SPECTRA AND LENGTH SCALES

In describing atmospheric mixing in detail, the frequency (inverse of the wavelength) of the motions that atmospheric turbulence is comprised of need to be examined. Turbulence can be conceptualized as composed of circular eddies that move along with the mean flow. Examining the diameter of these circular motions yields information about distance and effectiveness of dispersal.

In the case of the unstable atmosphere, the longest diameter eddy (λ_{max}) is controlled by the vertical rise of the buoyant air. The air rises off the surface and reaches a height where it begins to drop, or is entrained in an overlying airflow. This height defines λ_{max} in the unstable flow. The λ_{max} tends to increase through the day when conditions are sunny and still. The length can be on the order of thousands of meters. Conversely, in a stable layer, the damping of disturbances leads to a much smaller λ_{max}. Also, the development of these shallow, stable layers requires that the atmosphere be still, therefore turbulence tends to be intermittent and quickly damped out. The final case is in neutral stability; in this case there is often a steady wind mixing the atmosphere. The λ_{max} tends to be related to the surface roughness and may reflect the length of obstacles such as terrain features or buildings. Over flat terrain, friction leads to shearing and the eddies may roll along the surface. The λ_{max} in these flows tends to be intermediate between the stable and unstable cases. The implication to dispersion is that in an unstable flow, a material may be rapidly dispersed by a large eddy. In a stable atmosphere, eddying motions of this magnitude do not exist. In a neutral flow, the regional wind may transport material quickly relative to the eddying motion of the associated turbulence.

Table 1. The Pasquill Stability Classification scheme shows the relationship between wind speed, cloud cover, and solar radiation, where A is strongly unstable and F is strongly stable. This scheme is commonly used in air dispersion modeling (Pasquill 1974).

Wind speed at 10 m (m/sec)	Insolation			Night	
	Strong	Moderate	Slight	≥4/8 Cloud	≤3/8 Cloud
2	A	A–B	B	F	F
2–3	A–B	B	C	E	F
3–5	B	B–C	C	D	E
5–6	C	C–D	D	D	D
6	C	D	D	D	D

CLASSIFICATION

There are many ways to classify atmospheric stability. They range from fundamental indicators such as heat flux and net radiation to general, more intuitive schemes that might use cloud cover and time of day, for example. Three commonly used indicators are the Pasquill Stability Classification, Richardson Number, and Monin–Obukhov Length, which range from simple to complex, respectively.

The Pasquill Stability Classification (Table 1) is a widely used, straightforward approach that can be performed using just wind speed and general categorization of cloud cover or insolation. This classification incorporates many of the principles discussed in this section. It is most

valid over a wide area, because it uses total sky cover as opposed to a small area or point. Stability varies continuously in the atmosphere, which is a 3-dimensional fluid. It is impossible for this simple classification to capture either the horizontal or vertical variability.

The second stability indicator is the Richardson Number (*Ri*, this formulation is known as the dimensionless, gradient Richardson Number):

$$Ri = \left[g\left(\frac{d\theta}{dz}\right)\right]\left[T_a\left(\frac{du}{dz}\right)^2\right]^{-1}, \quad (3)$$

where g is gravity, u is wind speed, z is height, and T_a is an average layer air temperature. As stated in the first section, T may be substituted for θ near the surface. This number indicates that the buoyant force (convective turbulence) is counteracted by the ambient windstream (mechanical turbulence) that breaks up buoyant eddies and entrains air that does not have the same buoyant force. The Richardson Number is negative in unstable conditions and positive in stable conditions (this is based on the convention that $d\theta/dz$ is negative when temperature is decreasing upward).

The third stability indicator is known as the Monin–Obukhov Length (L). This measure has units of length but is arguably difficult to interpret as a physical length. L is calculated as:

$$L = -\frac{u_*^3 \cdot c_p \rho T}{k_a g H(1 + 0.07/B)}, \quad (4)$$

where u_* is the friction velocity, c_p is the specific heat (here, of air), ρ is the density of air, k_a is the von Karman constant, H is the heat flux, and B is the Bowen ratio. L approaches infinity in purely mechanical turbulence; and is on the order of (-10) under very unstable conditions; on the order of (-100) under breezy, unstable conditions; and on the order of (1) under stable conditions. L was introduced as a length scalar in the atmospheric surface layer, which accounts for the interaction of mechanical and convective turbulence. A detailed discussion of the terms that comprise Equation 4 is beyond the scope of this paper. These terms are defined in the basic meteorological references at the end of the paper. Equation 4 shows the dependence of L on both u_*, which can be thought of as the characteristic layer velocity, and H, which is a function of both surface heating and the effectiveness of heat transfer from the surface by the surface layer atmosphere. L is often used to nondimensionalize z in the surface layer $(-z/L)$. This scaling is often employed in the research literature on the atmospheric surface layer. Nondimensional scaling allows sampling programs in the atmosphere to be compared when the relevant length scales are a function of the stability. Refer to Panofsky and Dutton (1984) for a detailed discussion of Monin–Obukhov Length.

These 3 approaches to characterizing stability provide a range from simple, applied approaches to research approaches that require sophisticated data gathering before they can be calculated. The appropriate categorization is specific to the problem being addressed, the level of temporal and spatial resolution required, and the resources available to solve it.

TERRAIN EFFECTS

Movement of a relatively high-concentration, coherent spray cloud for long distances in the atmospheric surface layer can be the result of the combination of a stable layer and sloping ground (terrain effects). This combination can result in what is known as a density flow. As described above, stable stratification typically occurs under low wind speed conditions. The lack of mixing promotes formation of the surface-based cold air layer, which then suppresses mixing in a positive feedback. Even though dispersion is damped in the stable layer, material will not drift because there is no strong mean flow to move the cloud. These considerations change over sloping ground.

Because the cold air on the surfaces is denser than the air above it, it sits on the surface. If this denser air is on a slope, it begins to move downhill in a coherent flow reminiscent of liquid moving downhill. The air in the surface-based stable layer will begin moving on very slight slopes and will be influenced by surface roughness. In mountain valleys, very strong density flows develop downslope as cold air pours off valley sides and flows downhill along the axis of the valley floor.

Density flows provide a mechanism for relatively concentrated spray material to move in the atmosphere. The stable air that is moving downhill retains its slow dispersion characteristics (low dispersion coefficients) so the material remains relatively concentrated. In this type of flow, the air is often moving downslope in a valley or gully. The spray cloud is physically constrained from dispersing laterally by the valley walls. In some cases, the movement of the air generates enough mechanical turbulence to break up the stable layer. In other cases, the density flow can meander for miles or form a deep stable pool of air in a valley bottom. The possibility of a relatively concentrated spray cloud moving far off-site in a density flow provides a worst-case scenario for many applications, especially ground-based applications.

A classic scenario would involve a farmer raising grain on the high plains where dry, clear weather is common. On a very clear, dry morning before sunrise, an applicator sprays the crop. A recently burned-off ditch that trends slightly downhill provides a smooth conduit. As cold air begins to move down the ditch, it is replaced by air from the crop field and spray material moves into the ditch and begins to flow downhill. The ditch outlet, though a substantial distance away, is near a sensitive crop. Some portion of the spray material ends up on the sensitive crop instead of the target crop. A characteristic of this type of event is that it might not happen for a long period of time, but then happens when the weather and other factors (often human activity) interact in just the right combination. The fact that this type of long range drift is nonintuitive and occurs intermittently makes it difficult to study. The nature of this type of drift event also makes it difficult to educate applicators. If this situation could be anticipated, it could be avoided. These events have been recreated experimentally and this type of scenario has led to some of the most costly and confrontational cases of off-target drift of pesticides.

EFFECTS OF HUMIDITY AND OPEN WATER

Water plays an integral role in this discussion of stability and dispersion through its influence on the atmospheric energy balance. Clouds effectively reflect incoming shortwave radiation and absorb outgoing longwave radiation. The net effect is that neither strong stable nor unstable layers will develop under cloud cover. An overcast day generally indicates a neutral atmosphere, although there are exceptions. High atmospheric humidity also reduces the amount of shortwave solar energy that reaches the surface and reduces the longwave energy lost to space. Meteorologists define atmospheric transmissivity and opacity as measures of the radiative properties of the atmosphere itself.

The effect of open water on dispersion can be very pronounced. The higher c_p of water means that it lags the ground surface in heating during the day and in cooling during the night. This lag in heating is augmented by evaporation, which accounts for 90% of available thermal energy (stored in the phase change) at a free water surface. This means that only 10% of the available energy is available to raise the temperature of the air.

The manifestation of the differences in the thermal properties between a water surface and a soil surface is a horizontal temperature gradient. During clear, sunny days, cold air moves off the water and replaces warm air rising off the land, causing a sea breeze. During clear nights the reverse occurs. The sea breeze consists of stably stratified air.

A well-documented phenomenon related to the sea breeze cell is known as fumigation. This is caused by the stably stratified air over water moving inland over a developing unstable air layer known as a thermal internal boundary layer (TIBL). The TIBL is composed of eddies with λ_{max} approximately the depth of the TIBL. An airborne material released at height into the stable layer can move over land, slowly dispersing in the stable air, and suddenly be brought to earth by a large eddy when the top of the TIBL is encountered. This can lead to high surface concentrations of airborne material far from the point of release. Refer to Randerson (1984) for a detailed discussion of this phenomenon.

CONCLUSIONS

Atmospheric stability is one of the controlling factors in the dispersion of neutrally or near neutrally buoyant airborne materials. The optimum stability for the release of spray material in the atmosphere depends on the objective of the application, the source scenario, and the nature of the concerns about drift. Stable conditions can lead to high concentrations off-target. Unstable conditions can lead to a small amount of material being transported great distances.

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