The atmospheric boundary layer

Turbulent processes in the boundary layer are a key factor in air-pollution meteorology and provide boundary conditions on the dynamics of the atmosphere used in making long-term weather forecasts.

Hendrik Tennekes

The earth's atmosphere is always in motion, and as the winds blow over the surface of this globe, they maintain a "layer of frictional influence" or atmospheric boundary layer that extends from the surface to heights typically of the order of one kilometer. The atmosphere receives much of its heat and virtually all of its water vapor through turbulent transfer processes in the boundary layer. Furthermore, most atmospheric pollutants are released in the boundary layer, and the quality of our daily environment (we all *live* inside the boundary layer!) depends strongly on the mixing capabilities of whatever turbulence happens to be around us.

The turbulent motion in the boundary layer is a very effective agent for the downward transport of momentum, and this momentum flux from the atmosphere into the earth (the surface friction force) has an appreciable influence on the evolution of weather systems. As much as one-half of the atmosphere's loss of kinetic energy occurs in the boundary layer (much of the other half is associated with clear-air turbulence near the tropopause; see figure 1). The loss of kinetic enery of large-scale atmospheric flow patterns is caused by viscous dissipation (conversion of kinetic energy into heat) occurring in the smallest eddies of boundary-layer turbulence; it is characteristic of all turbulent motion that the rate of energy conversion is quite rapid, even if the viscosity—which is a measure for the internal friction in the air—is extremely small.

In this article I will discuss how boundary-layer research is important in air-pollution meteorology and in the general area of long-term weather forecasting. Let us first consider several important boundary-layer parameters, and the general features of turbulence as it affects the circulation and the energy dissipation of the atmosphere.

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Boundary layer

To what height does the boundary layer extend into the atmosphere? The boundary-layer thickness, or height of the mixed layer, may vary from a few tens of meters on clear nights with light winds to about two thousand meters (or more) on sunny summer afternoons. This wide range of heights is caused primarily by the effects of vertical heat transfer in an environment exposed to gravity: upward heat transfer ("thermal convection") enhances the intensity of the turbulence in the boundary layer, and so helps to increase the *entrainment rate* (the rate at which the boundary-layer thickness increases if the mean vertical velocity at the top of the boundary layer is zero). Conversely, downward heat transfer decreases or even suppresses the turbulence in the boundary layer; if the turbulent motion in the upper levels of the boundary layer decays, the boundary layer becomes thinner.

In the relatively rare situations with negligible heat transfer, the boundary-layer thickness depends on the wind speed of the air aloft and the Coriolis parameter (a local measure for the earth's rotation rate). With moderate winds the thickness of such a "neutral" boundary layer is about a kilometer at middle latitudes.

A very striking phenomenon that shows the top of the boundary layer is the occurrence of *cloud streets,* shown in figure 2. If the thickness of the boundary layer is one kilometer, the cloud streets are approximately two kilometers apart.

There are occasions on which the top of the boundary layer can be observed easily. Water vapor is carried upward by boundary-layer turbulence; because the temperature of the air decreases with height, the vapor may reach its condensation level, so that clouds are formed. For example, you may see nice fair-weather cumulus clouds in the upper bulges of the large eddies that distort the interface between the boundary layer and the air aloft. Cumulus clouds always receive their supply of water vapor by turbulent transport through the boundary layer; clearly, the boundary-layer thickness must be at least as large as the height of the cloud base if cumuli are present. Note however that the turbulence inside such clouds is not commonly considered to be part of the boundary layer. This distinction is made because clouds have their own dynamics, powered by the release of latent heat that accompanies condensation of water.

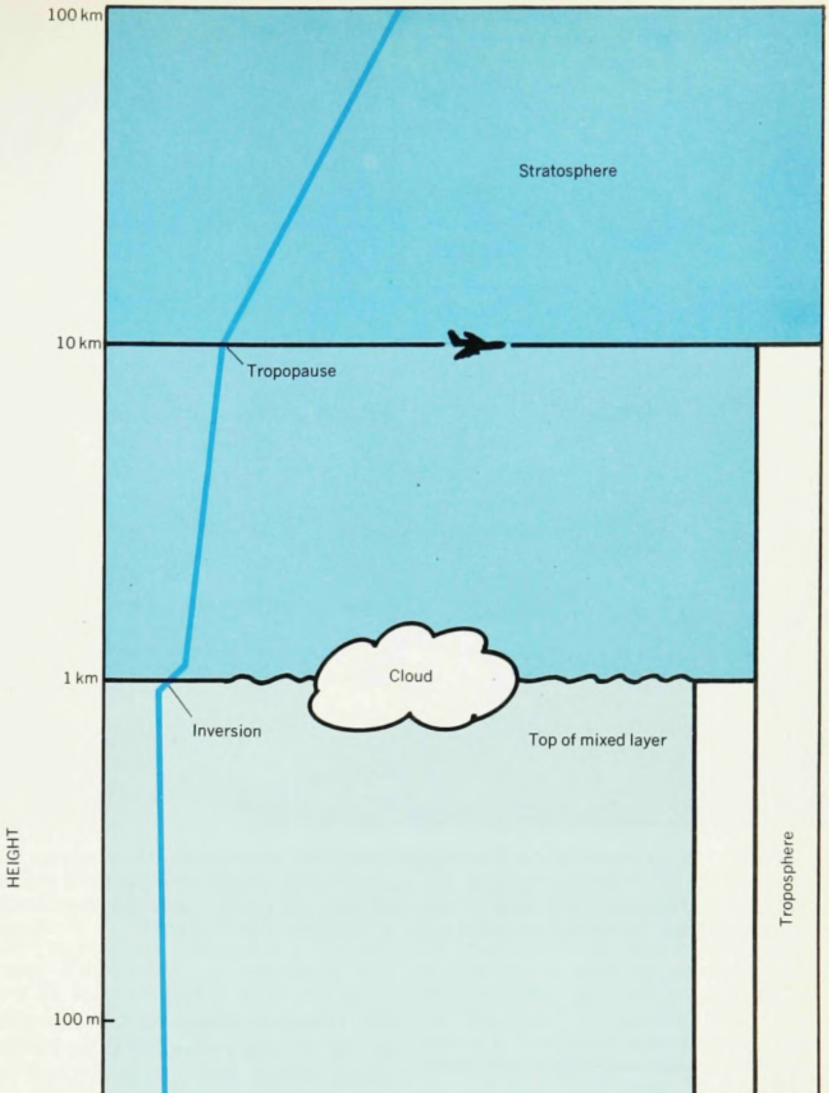
Again, when flying out of any of the world's major airports, you often notice a distinct improvement in visibility as the airliner leaves the polluted boundary layer.

The boundary-layer thickness is important not only to air-pollution research, but also to all parts of atmospheric boundary-layer research. The height of the boundary layer is a *scale height* (length scale) for boundary-layer phenomena, a variable that needs to be known in order to nondimensionalize turbulence properties, and—far more importantly—to determine the nondimensional parameters on which the transfer coefficients depend, as an input for the parameterization of cloud dynamics and as a boundary condition for the dynamics of the atmosphere as a whole as it relates to weather forecasting.

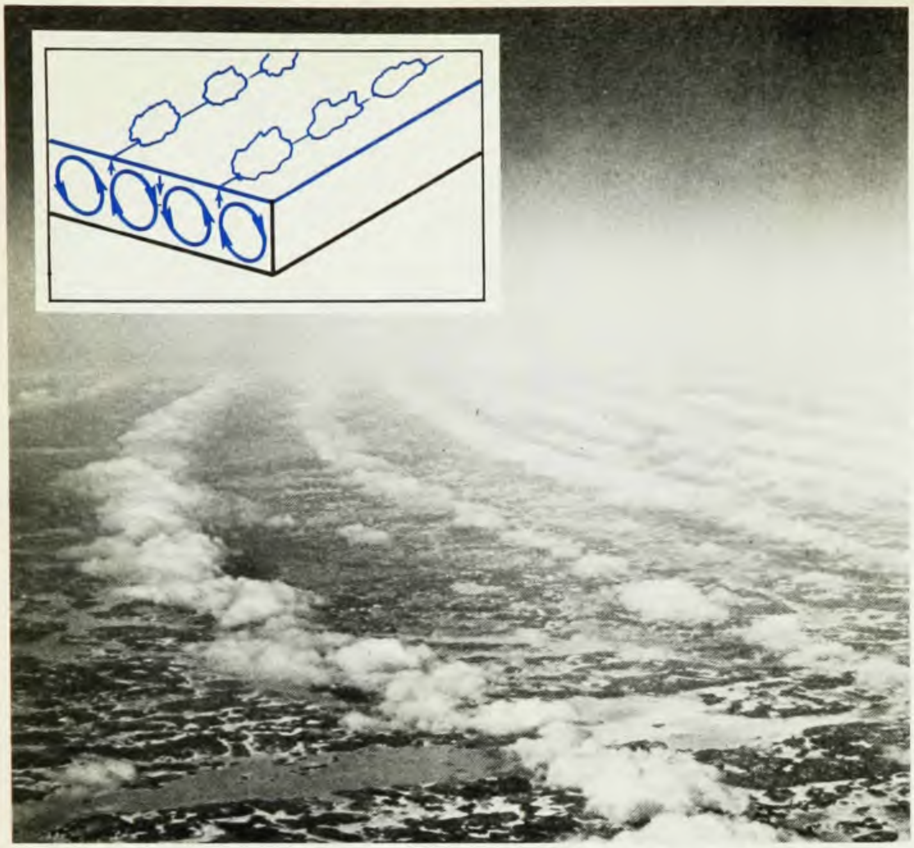
Turbulence

If our definition of turbulence were to include all flow fields that have a chaotic ("random") nature, such that statistical methods are needed to describe their averaged properties, we would have to conclude that all of the atmosphere is in a turbulent state of motion all the time. For example, weather systems (the irregular patterns of low- and high-pressure areas on the weather map) might be viewed as a kind of quasi-two-dimensional turbulence driving the atmosphere's general circulation, which is the wind field obtained by averaging over many seasons.

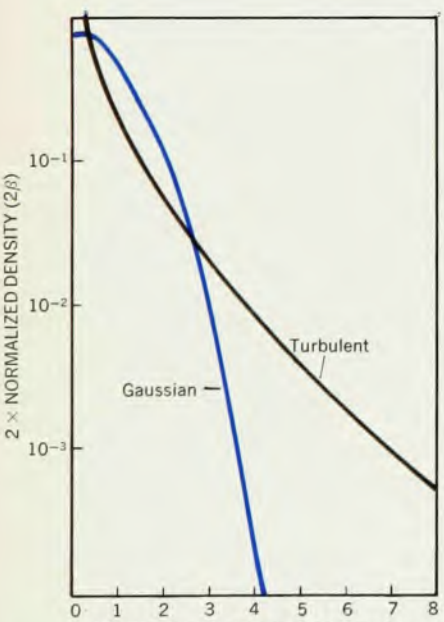
Weather systems are approximately two-dimensional flow fields because



The vertical structure of the atmosphere on a sunny day. In most of the atmospheric boundary layer, the entropy distribution is approximately uniform because of intense turbulent mixing. The mixed layer is capped by an inversion; the air aloft is stable, but the air in the surface layer has an unstable stratification. On clear nights with little wind, the mixed layer is generally much thinner and the temperature distribution is unstable because there is not enough turbulence to prevent large temperature gradients from forming. Figure 1



Cloud streets. Large eddies in the atmospheric boundary layer create cloud streets under favorable conditions. The cloud streets are approximately aligned with the wind direction. If the thickness of the boundary layer is one kilometer, the cloud streets are approximately two kilometers apart. (Photograph taken by H. A. Panofsky over Finland.) Figure 2



Probability density of velocity derivatives in atmospheric turbulence (black) compared with a Gaussian probability density (color). Notice the difference between the "tails" of these distributions. This is a normalized plot; x is the signal of the velocity derivative divided by its standard deviation. The parameter β represents the normalized density. Figure 3

the effective thickness of the atmosphere is much smaller than its horizontal extent (we all intuitively take "the wind" to be a vector in the horizontal plane).

We run into semantic confusion unless we adopt a more restrictive definition. Therefore, we call a chaotic flow field "turbulent" only if it is also intrinsically three-dimensional. In this way, we emphasize one of the key features of "ordinary" turbulence: the amplification of vorticity fluctuations by the mechanism called "vortex stretching." The vorticity vector (defined as the curl of the velocity vector) is a measure for the angular velocity of small fluid volumes; the equations governing its evolution relate to those expressing the conservation of angular momentum—vorticity grows when angular momentum is compressed into a smaller cross section.

Vortex stretching is the mechanism that gives hurricanes and tornadoes their destructive force. It is absent in two-dimensional flow.[1] As vortex stretching proceeds, the diameter of active vortices decreases. In turbulence, this occurs over a wide range of scales, with ever smaller vortices feeding on the strain rates of larger ones. Something akin to the ultraviolet catastrophe would result if the viscosity of the fluid were zero: extremely large values of enstrophy (the squared modulus of the vorticity vector) could be created at ever increasing wave numbers (small wavelengths). In reality, the catastrophe is prevented by the fluid's viscosity, however small, because it limits the wave-number range over which random motion can be maintained. The smallest wavelength in turbulence is called the "Kolmogorov microscale"; in atmospheric turbulence, its value is typically about one millimeter.

Within our definition, we can state that most parts of the atmosphere are only occasionally and intermittently in a turbulent state of motion. Turbulence *does* occur in cumulus clouds, in the jet stream and along the frontal surfaces that separate air masses; but most of the time conditions in the atmosphere prevent the generation and maintenance of turbulence. This state of affairs is true for most of the troposphere, but certainly for the stratosphere (which, as its name implies, has a layered structure).

The atmospheric boundary layer is unlike the rest of the atmosphere in that it is commonly in turbulent motion. Actually, most of the time turbulent mixing of heat and moisture in the boundary layer is so effective that the vertical distributions of entropy and water-vapor content are roughly uniform over the bulk of the boundary-layer thickness. For this reason, most meteorologists refer to the boundary layer as the "mixed layer." Note the usage: not mix-*ing*, but mix-*ed*, because in first approximation, it appears to be fully homogenized, at least often enough to justify the adjective (I must confess to a secret admiration for this kind of semantic precision).

Research in atmospheric turbulence and boundary-layer meteorology is performed for a variety of reasons, ranging from the need to provide data required for air-pollution forecasts to the desire to improve our fundamental understanding of turbulent-flow phenomena. As far as the basic physics of turbulence is concerned, the atmospheric boundary layer happens to be a readily accessible flow with extremely large turbulence Reynolds numbers. The Reynolds number is a nondimensional measure for the importance of inertia terms relative to that of the viscous terms in the equations governing fluid flow; it is defined as $\mathrm{Re} = ul/\nu$ where u is a characteristic velocity of the flow field, l is a characteristic length, and ν is the kinematic viscosity (diffusivity for momentum) of the fluid in which the flow occurs. For air at sea-level temperature and pressure, $\nu = 15 \times 10^{-6}$ m^2 sec^{-1}; for turbulent eddies in the atmospheric boundary layer (with $u = 0.5$ m sec^{-1} and $l = 300$ m as rep-

A dinner-table experiment

A fountain pen and a glass of water are all that is needed for a thought-provoking demonstration of some elements of turbulence dynamics. Very gently lower a drop of ink from the tip of the pen onto the surface of the water. Extreme care is called for, so that the results cannot be blamed on poor initial conditions. Watch what happens (see photo below). A convoluted flow pattern develops soon after the descending drop of ink has acquired some vorticity through viscous friction with the surrounding water. The drop quickly develops into a vortex ring; the ring acquires more vorticity as its diameter increases; it goes unstable suddenly, spinning off several vortices as it breaks up; the new vortices grow and break up, and so on, in a breathtaking cascade of vortex dynamics that is finally stopped when the vortices have become so diluted that viscosity can prevent further instabilities.

This experiment has everything a physicist sensitive to visual concepts could hope for: The dramatic nature of flow instabilities, the "random" flow field arising from simple initial conditions, the crucial role of vorticity, and—last but not least—the presence of a cascade mechanism that creates motion at ever smaller scales.

Actually, a detailed description of this experiment and its relevance to turbulence concepts requires a careful analysis of what is happening. For one, the Reynolds numbers involved are rather small, and the "turbulence" here cannot maintain itself for more than a few seconds. Also, the scale of the region in which ink penetrates increases as the descent proceeds. Therefore, the largest scales of motion increase with time, and the "spectral energy cascade" is not nearly as rapid as it might otherwise be. Incidentally, this is an advantage to the observer because the scale-increasing tendencies in this flow prevent the small vortices from becoming too small to be easily observable. Furthermore, kinetic energy is supplied to this flow by the acceleration of gravity as it performs work on the slight density differences involved. This buoyant forcing continues to be effective throughout all but the final (dissipative) stages of the process. In "ordinary" turbulence on the other hand, the energy supply enters mainly at the large-scale end of the spectrum. Finally, the "cascade" in this particular experiment is driven by self-induced straining of vortex rings; at no stage in the process does one get the impression that larger vortices participate in the straining. Nevertheless, the extremely rapid rate at which a vortex ring yields its vorticity and kinetic energy to its offshoots as it breaks up is characteristic for the vigor of the vortex-stretching mechanism.

resentative values) the Reynolds number is as large as 10^7, a value that is extremely hard to obtain in the laboratory.

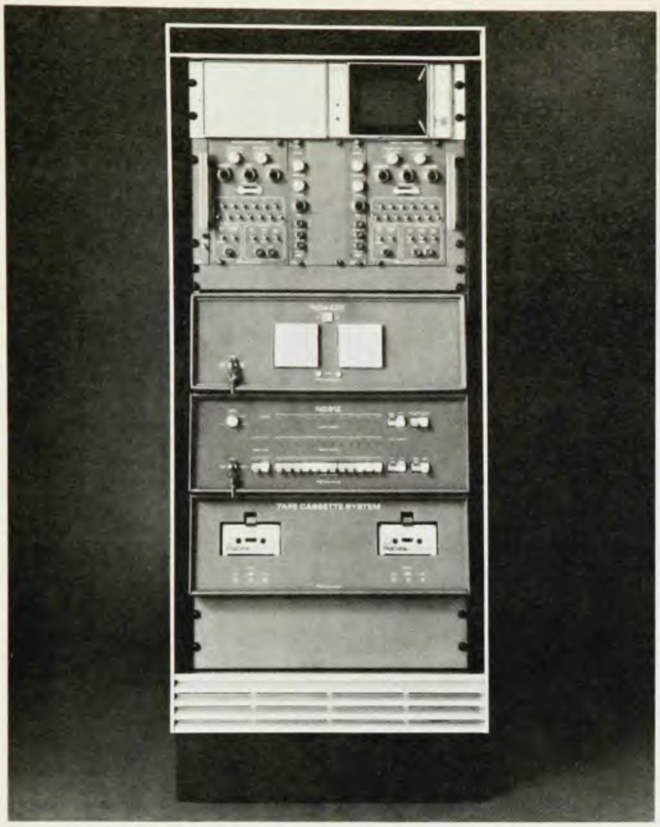
The large dissipation rate of kinetic energy in turbulent flows is made possible by the rapid transfer of energy from larger to smaller scales of motion. This "spectral energy cascade" is driven by random stretching of vortex lines (the vortex-stretching mechanism that was referred to above.) Such a cascade happens when an ink drop is carefully released in a glass of water (see box on this page). The theoretical models of this process are all formulated for the asymptotic case in which the Reynolds number tends to infinity; it is evident that these theories can be tested best by comparison with data from geophysical flows. For example, at the small-scale end of the turbulence spectrum, where viscous dissipation removes kinetic energy, vortex stretching causes probability distributions of vorticity and related variables that are quite different[2] from Gaussian (figure 3). This effect, which is referred to as the "intermittency of the small-scale structure," is most pronounced at the extremely large Reynolds numbers found in geophysical flows. The atmospheric boundary layer is used in this way as a high-Reynolds-number laboratory.

Data on the microstructure of turbulence are needed for many applications, too. Since an appreciable fraction of the kinetic-energy losses of atmospheric motions occurs in the small-scale structure of boundary-layer turbulence, all information on the dissipative eddies is welcome. Also, the small-scale structure of the temperature fluctuations in atmospheric turbulence is responsible for the scattering of electromagnetic waves; it causes twinkling of stars and limits the resolution of telescopes.

Small-scale structure

The small-scale structure of turbulence plays an important role in turbulent diffusion of pollutants. The molecular diffusivity of air is extremely small; yet concentration fluctuations are erased quickly in a turbulent flow. This process is so effective that it is often taken for granted: on stirring cream in a cup of coffee, no one worries about the rate at which the mixture is homogenized. On the other hand, the very presence of paint shakers in hardware stores proves that mixing is not always accomplished easily. What is involved here? The contaminant diffusivity of paint is very small, but its viscosity is quite large compared to that of air. Consequently, the small-scale structure of the velocity field in the can of paint being mixed is poorly developed, with the smallest wavelengths remaining fairly large due to the overwhelming effects of viscosity. Clearly, the rate at which the mixing process proceeds is a function of width of the turbulent energy spectrum.

A glance at the photograph in the



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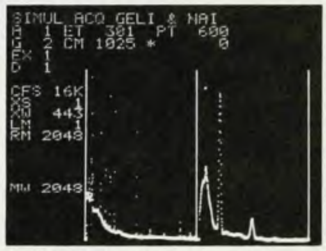
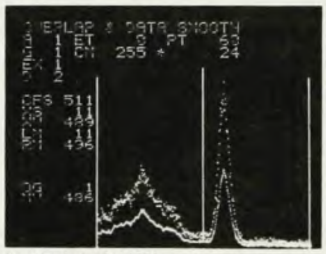


1. 4420 display showing spectra acquired simultaneously from GeLi detector in 4096 channels and NaI detector in 1024 channels.

2. 4420 isometric display of 35 x 40 data point region of interest for detailed spectrum analysis in dual parameter study. Resolution—4K x 4K data points.



3. Display of 4420's spectrum overlap and data smoothing functions.

4a.

NO.	ECL	CENTROID	FWHM	BKGLHGT	AREA	LEFT	RIGHT
1		187.99	1.96	100461.63	59220.50	138	148
		188.13	2.01	32012.00	54938.15		
2		243.44	2.77	62212.51	51464.88	181	192
		243.71	2.68	23576.57	52523.50		
3		260.33	2.17	35917.00	2592.54	196	203
		260.57	2.22	1585.40	2919.48		
4		276.07	2.37	29015.00	2387.00	209	215
		276.23	2.34	1024.29	1992.74		
5		296.40	2.04	47277.76	100277.49	222	234
		296.57	2.09	52700.62	91729.37		
6		352.85	2.58	39487.51	150378.91	266	278
		353.02	2.40	71139.23	140739.26		
7		511.25	3.60	11884.50	1806.13	392	400
		511.09	3.47	542.23	1563.64		
8		608.93	2.40	14480.30	86282.63	467	479
		609.23	2.44	40875.34	82949.45		
9		664.10	2.23	12860.64	3609.72	510	524
		665.20	2.62	1067.02	2319.95		
	10	661.58	2.62	486.63	1056.04		
10		741.75	3.40	7270.00	1189.63	573	581
		741.76	3.21	381.79	1018.87		

4b.

4a. and **4b.** 4420's automatic Peak Search and Calibration for doublet resolution by Gaussian fit:*

a. Display of 12 peak portion of Ra^{226} spectrum selected for peak search showing peak energy computed and displayed automatically.

b. Printout of peak parameters (centroid and FWHM in KeV) showing resolution of doublet in peak 9 after 10 iterations.

*Based on work reported in "Automatic Spectrum Analysis on Minicomputers: Doublet Resolution by Gaussian Fit" by E. D. Von Meerwall, Department of Physics, The University of Akron, Akron, Ohio and M. D. Gawlick, Nuclear Data, Inc., to be published in Computer Physics Communications.

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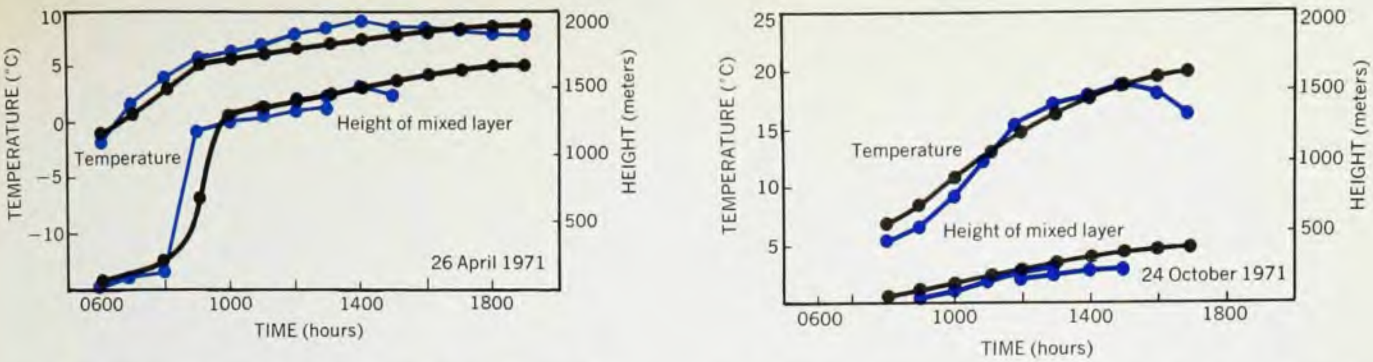

Forecasts of temperature and boundary-layer thickness for two different days at De Bildt, the Netherlands. Operational forecasts (those that are issued by the weather service to the press and broadcasting media) of maximum temperature and of pollution index are based on equations that predict the endpoints of the curves in this figure. The black lines are the forecasts based on equations given in the article and a related equation for temperature change in the boundary layer. The colored lines show the temperature observed at 1.5-meter height and the boundary-layer thickness estimated from observed temperature profiles. 26 April 1971 was a nice day in De Bildt; the mixed layer grew rapidly between 8 a.m. and 10 a.m. because of weak stability of the air aloft, and after this time the increase was slow because of a pre-existing inversion at 1200 meters. On 24 October 1971 the air above the mixed layer was very stable and surface heat flux was small (this is late fall); because the mixed layer did not grow thicker than about 400 meters, the potential for severe air pollution was high. (Royal Netherlands Met. Inst.) Figure 4

box on page 55 shows how turbulent motion deforms a drop of contaminant into a complicated array of rings, sheets, strings and ribbons. This process continues until the concentration gradients become so large that the molecular diffusivity can take over. In turbulence with a well-developed microstructure, the small-scale eddies are extremely effective at straining the contaminant field into ever thinner ribbons, so that the subsequent molecular exchange process (the final stage of diffusion) can be quite rapid. This is called "strain-accelerated" diffusion;[1] it is evident that this process would not be possible without a spectral energy cascade that maintains the straining motion of very small eddies. In a paint shaker, viscosity calls an early halt to the cascade; the straining thus is much less vigorous, and the time needed to obtain homogenization becomes quite apprecable.

In this context, air-pollution chemistry should be mentioned briefly. Chemical reactions between different species require contact on a molecular scale; the small-scale structure of turbulence generates a highly contorted interface between different chemical species, thus providing an extremely large contact surface and allowing chemical reactions to proceed rapidly. The combustion rate in the turbulent flames of an incinerator is several orders of magnitude larger than that in the smooth, laminar flame of a candle.

Cloud streets

One of the interesting developments in research on turbulent fluxes is a keen appreciation for the role of the large turbulent eddies in the atmospheric boundary layer. It appears that the Coriolis effects caused by the earth's rotation help to maintain the large eddies as quasi-permanent secondary flows shaped like helical rolls or vortices (figure 2). Their axes are roughly aligned with the wind, and their diameters are comparable to the boundary-layer thickness. In conditions with suitable temperature and humidity distributions, the presence of these rolls is manifested by the cloud streets near the top of the boundary layer as I mentioned before: In the updraft region between two counterrotating vortices, water vapor is carried to the condensation level, so that clouds result. The regular pattern of some of these cloud streets suggests that the large eddies in the planetary boundary layer have lifetimes that far exceed those of their laboratory counterparts. Early BOMEX (Barbados Oceanographic and Meteorological Experiment 1969) results suggest that these eddies play a crucial role in the feedback mechanism[3] that regulates the fluxes of heat and moisture; this is in line with the ideas of A. A. Townsend, who proposed a simple feedback loop for turbulent shear flows in the laboratory. The BOMEX results further suggest that the buoyancy associated with the water vapor in the maritime boundary layer can be more important than the buoyancy generated by temperature differences. This would add another item to the list of phenomena for which the water in the atmosphere is responsible.

Air-pollution meteorology

Fluid dynamics, of course, cannot *cure* any air-pollution problems; it can only provide information needed for rational control measures and site planning. Estimates of air-pollution dispersal in the atmospheric boundary layer require knowledge of wind speed, wind direction, turbulence intensity and the thickness of the boundary layer.

In pollution problems that involve urban regions with horizontal scales of more than a few kilometers, the details of the structure of boundary-layer turbulence tend to be less important because turbulence quickly mixes the pollutants uniformly over the entire height of the mixed layer. Also, horizontal turbulent transport is much less effective than transport by mean wind. Therefore, the principal variables in air-pollution dispersal over moderately long distances are the wind-speed vector and the height of the mixed layer. The product of these two quantities is called the "ventilation volume," but one is tempted to call it the "flushing factor"; after all, we are dealing with a kind of sewage. It is obvious that air-pollution hazards are greatest when the flushing factor is small. It is not hard to estimate the average wind speed in the boundary layer if the winds aloft and the surface topography are known; a major research issue these days is the boundary-layer thickness and its variation from hour to hour, from day to day, and from season to season.

What are some of the conditions that control the height of the mixed layer and its temporal and spatial variations? I mentioned earlier the heat flux at the surface: When it is upward, turbulent entrainment is rapid; when it is downward, there may not be enough turbulence to maintain the boundary layer, much less to increase its thickness. Now I should also mention the effects of slowly rising or slowly sinking motions associated with weather systems. It is the sinking air (called "subsidence") that is an obvious potential cause for trouble. Subsidence occurs mainly in high-pressure regions, and when a large high-pressure center sits quietly over a large indus-


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trial city like Pittsburgh for a few days, the quality of its atmospheric environment quickly deteriorates. The density stratification of the air above the boundary layer and horizontal transport (advective changes of boundary-layer thickness, as when a thicker boundary layer flows in from elsewhere) also play significant roles in this problem.

Simple theoretical models that incorporate some of these effects have been formulated in the last few years; the results of these studies may be illustrated by the following equations:[5,4]

$$\frac{dh}{dt} = w_e - w_h$$

$$w_e = 1.4H_s/\rho c_p \gamma h$$

Here, h is the height of the mixed layer, w_e is the entrainment velocity, w_h is the large-scale vertical velocity at the top of the boundary layer (negative in the case of subsidence), H_s is the surface heat flux, ρ is the air density, c_p is the specific heat at constant pressure, and γ is the vertical gradient of potential temperature in the air aloft (a measure for the stability of the air above the boundary layer). These equations are valid for daytime conditions in which most of the turbulent energy in the boundary layer is maintained by the upward heat flux. Two examples of forecasts based on equations like this are given in figure 4.

There is also an active research interest in the dynamics of the turbulence at the interface between the boundary layer and the overlying air. This interface is commonly marked by an "inversion"—the top region of the boundary layer has less entropy than the air aloft. Last fall, the Air Force Cambridge Research Laboratories and the British Meteorological Office carried out an extensive field experiment in Minnesota to obtain data on the turbulence dynamics at the inversion base. Some of the instruments they used are shown in figure 5.

Forecasting

Short-term forecasts apparently are rather insensitive to thermal forcing (non-adiabatic heat inputs) and boundary-layer fluxes (in fact, frictionless, adiabatic models are being used for that purpose), but the relatively crude modeling of the boundary layer contributes to relatively poor *long-term* forecasts. Improvements in those forecasts require among other things that boundary-layer dynamics (including friction, kinetic-energy dissipation and all fluxes) be accounted for in better ways.

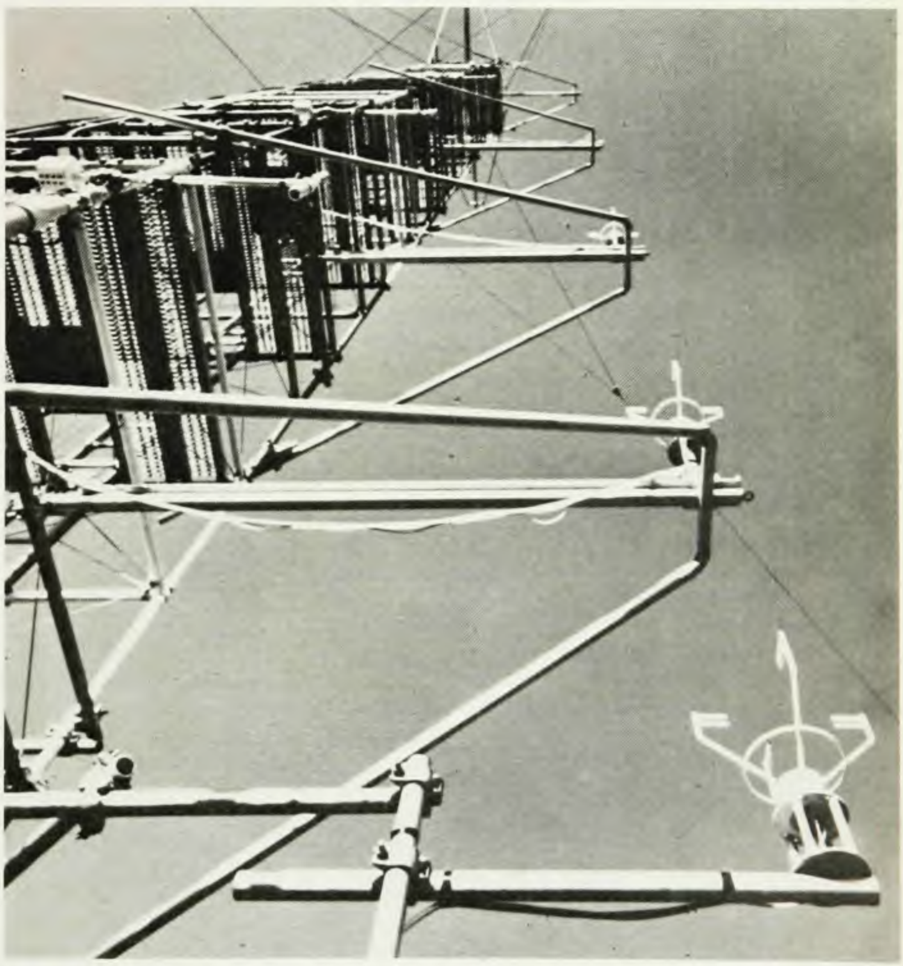
One of the major goals of applied meteorology is the improvement of long-term weather forecasts (those that make predictions for about a week or more ahead). Subjective forecasts based on the sypnotic method (the one that uses weather maps with cold and warm fronts) become quite unreliable after two or three days, and the accuracy of the currently operational numerical prediction methods leaves much to be desired if periods of five days or more are involved. This is so because the time scale of energy exchanges in the troposphere is about one week. The reaction time of the atmosphere to the turbulent fluxes of heat and moisture is about three days, that to the viscous dissipation of kinetic energy is about one day, that to the latent-heat is also one day, and the reaction time to radiative fluxes is about one week.[5] Some of these nonadiabatic effects thus exert a significant influence on the development of the weather after two days or so, and all of them are important to long-term forecasting.

Significant improvements in long-range forecasting are expected from more sophisticated boundary-layer studies, such as the work being done by J. W. Deardorff.[6] He solves the equations of motion for an ensemble of realizations on the large computer of the National Center for Atmospheric Research. These computer simulations—which are so expensive that most university researchers could not dream of doing it—have resulted in data on eddy structures, turbulence intensities, turbulent fluxes and many other properties of the atmospheric boundary layer, both for neutral and unstable conditions (see figure 6). Deardorff first used a fixed inversion "lid" at the top of his boundary layer, but his most recent computer simulations of turbulence in the atmospheric boundary layer include the following features:

- a staggered grid with 64 000 grid points, covering an area of 5 × 5 km in the horizontal plane and extending to a height of 2 km

finite-difference equivalents of six basic equations and 15 auxiliary equations, the latter governing the evolution of subgridscale variances and fluxes

- an exact inversion of the finite-difference divergence of the equation of motion, needed to compute the pressure fluctuations
- surface boundary conditions supplied



Vertical array of acoustic anemometers on a 32-meter tower at the Air Force Cambridge Research Laboratories field site at Donaldson, Minnesota. Acoustic anemometers determine wind-speed components by measuring Doppler shift in the speed of sound along opposing paths. (Photo: Victor Neumeier.) Figure 5

Time sequence of the vertical velocity field in an extremely large computer simulation of boundary-layer turbulence by J. W. Deardorff (reference 8). Black lines are contours of updrafts and the colored lines are contours of downdrafts. Note the three convection cells starting at the arrows in the first frame; they grow as time proceeds and merge with other updrafts in later frames. (This figure reproduced with the permission of the American Meteorological Society.) Figure 6

by surface-layer similarity laws

- initial conditions taken from early-morning observations at the 1967 Wangara expedition in Australia
- explicit treatment of the turbulent entrainment into the air aloft
- extremely large execution times. Each time step (six seconds) requires one minute of CDC-7600 computer time, and the simulation of one 24-hour cycle recently required 350 hours of NCAR's CDC-7600.

Another approach in modeling does not rely on computer experiments, but postulates models for certain terms in the equations governing the momentum flux, the heat flux, the kinetic energy of the turbulence, the temperature variance and other variables that are either variances or covariances of the turbulent fluctuations involved. The transport equation for the vorticity variance (the so-called "dissipation-rate equation") plays an important role in these schemes, either implicitly or explicitly, because it controls the evolution of the time and length scales in turbulent flows.

The kind of problem that one encounters in this approach is common to all branches of nonlinear statistical physics: Upon averaging the equations governing the evolution of the product of two fluctuating variables, terms are generated that contain triple products of fluctuations and/or other products for which no equation is available (other than in terms of yet higher moments). Research on these "second-order" (also "second-generation") models of turbulent transport is quite intensive at present and promises to continue that way for some years into the future.

Modeling the atmosphere

At this point, it appears worthwhile to have a brief look at some of the more general issues involved in modeling atmospheric flows. The numerical calculations for weather forecasting are generally performed on a grid with a mesh spacing on the order of 400 kilometers horizontally and about a kilometer vertically. Wind, temperature and humidity are computed at each grid point. The detailed properties of the boundary layer cannot be included in such a scheme; the boundary layer has to be represented by parameterized boundary conditions. Now, on a grid that spans a significant fraction of the globe, all of the Netherlands for example (admittedly not a very large country!) stays well within a single grid square. Numerical predictions of atmospheric flow on a *regional* scale thus require a much finer mesh, with a grid points spaced some twenty kilometers apart, say. Finally, numerical calculations of boundary-layer development may require a mesh size of about one kilometer in the horizontal plane. These different kinds of calculations, done by different people (or agencies) for different purposes, somehow have to be matched and made compatible with each other. This is called "telescoping"; it has become a major issue in atmospheric modeling, and many researchers are working on this problem around the world.

For example, a long-term numerical forecast for the entire globe (or for a single hemisphere) requires *some* boundary-layer inputs, no matter how crudely parameterized. These calculations must be performed before the smaller-scale models can be put on the computer because the large-scale forecast has to provide boundary conditions on the edges of the smaller-scale grids. This is why meteorologists are going the route of parameterization of boundary-layer fluxes. However, that doesn't settle the issue. If it turns out that you need to compute boundary-layer fluxes (such as the turbulent transport of momentum and heat) in order to predict boundary-layer development in order to predict the generation, or dissipation of low clouds in order to predict the input of latent heat and changes in the radiation budget in order to predict changes in surface temperature in order to estimate surface fluxes of moisture and heat, where do you cut that circle? Nevertheless, sophisticated boundary-layer calculations will some day find operational use in regional forecasts (such as the prediction of snow fall in the Buffalo–Erie area), in air-pollution forecasts, and in calculations of hurricane development and decay.

New challenges

In closing, I wish to go beyond the scope of this survey and return to a research area that was mentioned briefly at the beginning of this article. The motion in weather systems is very much constrained in the vertical; it is

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kind of quasi-two-dimensional turbulence. Charney[9] calls it "geostrophic turbulence." This is an exciting subject for the turbulence researcher—or, for that matter, any fluid dynamicist—who is looking for new challenges. For example, how does two-dimensional turbulence carry mean momentum from one place to the next, and how is it that the Reynolds stress of a low-pressure system is such that more mean momentum is fed into the westerlies at middle latitudes, where that momentum was greatest to begin with? This illustrates the reverse energy cascade in two-dimensional turbulence: Energy tends to flow toward larger scales, not to smaller scales as in ordinary turbulence. Earlier, I used a figure of speech: weather systems are driving the general circulation. Are they? Do the middle-latitude westerlies ride on cyclones and anticyclones, or do weather systems ride on the atmosphere's belt of eastward zonal winds? Exactly what are the roles of advection, spectral transfer and thermal forcing?

In this context, the subject of atmospheric predictability has to be mentioned. Numerical forecasts are forward integrations of the equations of motion and energy; they use inadequate boundary conditions and initial conditions obtained from a spotty global network of observation points. How fast do errors grow in these calculations? Is the weather two weeks from now predictable in principle, is it predictable in principle but not in practice, or is it altogether hopeless? Several fluid dynamicists are working on problems in this area; I should add that theoretical studies and computer simulations of two-dimensional turbulence are of great significance in the continuing efforts to find answers to the many questions that remain.

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