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The Effects of Middle Atmosphere Turbulence

on Coupling between Atmospheric Regions

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Neutral atmosphere turbulence exists almost ubiquitously throughout the atmosphere up to heights of about 100 km, and then completely disappears above about 120 km altitude. It's importance in providing coupling in the lower regions of the atmosphere (troposphere and stratosphere) is well known, but its importance for coupling between the middle atmosphere and ionosphere is not so well understood. In this review, we will concentrate on the role that turbulence plays in the coupling between the middle atmosphere and the low levels of the ionosphere. The discussion will include a review of some of the important principles of turbulence, as well as some of the unique features about middle atmospheric turbulence. A detailed collection of measurements of turbulent eddy diffusion coefficients will be presented,
along with some warnings about how these data should be interpreted. The need to understand turbulence at a very fundamental level in order to apply measurements of
such "diffusion coefficients" is stressed.

1. Introduction

One end product of many dynamical motions in the atmosphere is turbulence, and indeed for small scale dynamical features turbulence is in fact the major end-product. This is a highly non-linear, predominantly small-scale process which causes diffusion and, ultimately, atmospheric heating. However, due to the exponential increase of atmospheric kinematic viscosity with increasing height, an altitude is eventually reached above which neutral atmospheric turbulence can no longer exist-any tendency for turbulence to form is quickly damped by the large viscosity. This "transition" altitude shows a variation as a function of time and season, but is typically in the range 95 to 115 km.
Therefore for much of the upper ionosphere, neutral atmospheric turbulence is not directly important. However, studies of neutral turbulence in the middle atmosphere are still relevant to ionospheric work for at least two main reasons. Firstly, larger scale wave motions like tides and internal gravity waves frequently propagate up through the middle atmosphere into the ionosphere, and the existence of turbulence modifies the propagation behaviour of these waves. In other words, the middle atmosphere acts as a "filter" for these waves, and the existence and morphology of turbulence plays some role in defining the filter characteristics. Secondly, although turbulence does begin to damp out above about 95-100 km altitude, it nevertheless does exist in a somewhat modified form up to around 120 km, and the mixing which takes place in the boundary between the middle atmosphere and the lower ionosphere is an important coupling mechanism.

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In this brief review, we will begin by investigating some of the differentways in which neutral atmospheric turbulence is visualized. We will begin with a rather "coarse" view, looking at turbulence simply as a diffusive "damping" process for larger scale waves. Some of the inadequacies of this simple picture will be highlighted, and the need for understanding
at a finer resolution will be emphasized. Throughout the article we will discuss recent
measurements of atmospheric turbulence in the middle atmosphere, including a special
section concentrating on the "turbopause" region, and we will discuss how these observations
relate to middle atmosphere-ionosphere coupling.

2. Visualization of Turbulence

Although turbulence is in general associated with some sort of chaotic behavior, its visualization differs from field to field. To some workers, the details of the fine scale structure are completely unimportant, and turbulence is represented by a simple "K' term in the fluid dynamical equations. For example, the momentum equations can be written fairly generally as

(1)

where p is used to represent density, p is pressure, u is the total velocity vector, v is the
kinematic viscosity, t is time, V is the gradient differential operator, D/Dt represents
differentiation following the motion (=•Ý/•Ýt+u•E•¤), V2 is the Laplacian operator, g is the
acceleration due to gravity, ƒ¶ is the angular rotation rate of the earth, and F represents other
forcing terms (per unit mass) like ion drag, etc. But in many applications, u is written equal
to a mean and a fluctuating component, and if this is substituted into (1) and then averaged in
time or space, an equation results which looks very much like (1), but with u replaced by the
mean velocity alone. However, there are also some remaining terms which look like a/az

(u'w'), which are derivatives of the so-called Reynold's stress terms. These terms provide
forcing on the mean state. In some senses, they can be thought of as a sort of "viscous drag"
acting at scales greater than the scales at which molecular processes are important. Because of
their complexity, and because of this rather weak analogy with viscous drag, it is common to
include all such terms in a few simple quantities which look like K •Ý2(u)/•Ýz2, (actually •Ý/•Ýz
(K •Ýu/•Ýz)), thereby giving rise to a new set of equations

(2)

where p0 represents the mean density, D/Dt = •Ý/•Ýt + u.•Þ, and F' represents forcing of the
mean state. [K] is in fact a tensor, not a scalar. For more details about this transformation,
see for example HOUGHTON (1977). The details are not the main point of this work, but rather
it is the concept which we wish to address here. The "K" terms contain ("hide") all the
Reynolds stress terms. They are called "turbulent diffusion coefficients", because of the
analogy with viscous drag, but as already seen they represent a simplification of some
potentially quite complicated terms, and represent perhaps the coarsest visualization of

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turbulence. No heed is paid to the time scales involved in mixing processes, and K represents
simply a coefficient which determines how effective the smaller scales are at causing mean-
state forcing and large scale mixing. In some applications, such as planetary wave studies,
even the fluctuating effects of gravity waves are absorbed into such terms! One especially
important component of this tensor is often denoted Kzz, and represents the rate of vertical
diffusion. In one-dimensional models, this term actually accommodates both true vertical
small scale diffusion and "vertical" diffusion due to the effects of quasi-horizontal motions
along tilted isopleths. The term KzZ is in fact often loosely denoted as K. Despite its obvious
limitations, for many workers this simple visualization seems adequate.
For many studies of small scale turbulence, the term 2i x u is often ignored, an
assumption which is valid provided that typical scales L are small enough the Eckman
number EK = K/(2ƒ¶ L2) >> 1 (e.g. CHARNEY, 1973).
An only slightly more sophisticated approach involves comparing the mixing of a
turbulent field with the mixing which occurs at a molecular level in a gas, and a "diffusion
coefficient" D is defined through relations like

(3)

Here,ƒÐZ2 represents the mean square vertical displacement of an originally compact array of
parcels which spreads out over time t. The analogy with molecular processes is quite poor,
however, and if indeed ƒÐz2 does vary proportionally to t it is more fortuitous than rigorous.
Nevertheless, this representation does at least attempt to consider the time scales associated
with the turbulent mixing, something that (2) does not.
More refined consideration of the temporal variation of ƒÐz2 with respect to time shows
that (3) is incorrect at small scales, because as the particles diffuse apart, larger scale motions
become effective, hastening the process. In fact

(4)

more accurately describes the diffusive process, although the power of t depends largely on the spectrum of scales within the turbulent region. Only at scales greater than the largest turbulent scales can (3) be applied.

3. The "Gross" Approach

These types of "gross overview" descriptions, in which only the consequences of turbulence are important, and its detail is not, shall be referred to as "gross" views. Many attempts have been made to make estimates of "K" in the middle atmosphere, largely by utilizing equations like (2). For example, one can generate equations which describe momentum and fluid motions in the atmosphere, leaving "K" as an "unknown" variable, and then proceed to use measurements of other known quantities like wind speeds, temperature distribution, and gravity wave fluxes to infer values for K. One of the more recent and best known attempts at this is the model due to GARCIA and SOLOMON (1985). It is also possible to include chemistry in these models, and use measurements of concentrations of atomic and molecular species to infer values for K. Early attempts at this include papers by JOHNSON and WILKINS (1965), COLGROVE et al. (1965) and others. These

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authors looked in particular at the relative concentration of atomic oxygen and molecular oxygen at 90 and 120km to infer information about K. More recently GROVES (1986) has done similar things. When this type of approach is adopted, profiles of K as a function of height can be deduced. Values for K have been obtained which generally lie in the range between a few tens of m2s-1 up to a few hundred. More exact profiles will be presented towards the end of the paper.
When one begins to consider these "diffusion" coefficients a little more carefully, however, it becomes clear that it is necessary to distinguish different types of K values. The diffusion coefficient for momentum is different from that for temperature, for example, and one needs to define Km and KT as separate entities. The ratio between them, Pr = Km/KT is called the Prandtl number. The estimates made by GARCIA and SOLOMON, for example, represent estimates of Km. However, recently STROBEL et al. (1987) and BEVILACQUA et al. (1990) have carefully constructed a computer model using detailed atmospheric chemistry schemes. They assumed a variety of forms for KT and compared resultant profiles of minor species to measured values. The values they obtained for KT are less, by a factor of 3 or more times, than the values for Km inferred by Garcia and Solomon. These authors have taken this to infer that the Prandtl number in the middle atmosphere is greater than 3, a significant result. JusTus (1967) also made experimental determinations of the Prandtl number using rocket studies, and obtained values of around 2 and more. It should nevertheless be emphasized that the determinations of Strobel et al. apply mainly below 80 km altitude; their model is not very accurate above that height. Earlier estimates of temperature diffusion coefficients above 80km (e.g. JOHNSON and WILKINS, 1965) give values more like 100 m2s-1, and are more compatible with the momentum diffusion coefficients above 85 km. There has always been some uncertainty as to whether Justus had adequate resolution to trust his estimates of the Prandtl number. Nevertheless the results are suggestive and should be borne in mind whenever typical "turbulent diffusion coefficients" for the middle atmosphere are discussed.

4. A Closer Look

Often this gross view is inadequate however. Even for workers who only need to know
an estimate of K, it is useful to look closer at the turbulence causing the diffusion. To some, turbulence is visualized as a homogeneous process acting everywhere, something like the process in the atmospheric boundary layer, and diffusion is seen to occur simply because of this. Inertial range turbulence theory is adequate for such a model. In such a turbulent patch, it is possible to show that an individual cloud of particles which are initially close together diffuse apart in a manner

(5)

as already discussed. One can then develop models around such a visualization. Unfortunately, however, the middle atmosphere is not so simple! (In fact, turbulence generally is not.) Rather than being a homogeneous process acting uniformly throughout the region, it has been clearly shown through the period of the recent "Middle Atmosphere Program" that turbulence is both temporally and spatially intermittent. It occurs only for short periods of time at varying spatial locations. Examples of this intermittency have been

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illustrated particularly well with radar studies (e.g. CZECHOWSKY et al., 1979; ROETTGER et al, 1979; WOODMAN 1980; SATO et al., 1985, amongst many). Turbulent layers also have varying depths and spatial extents. An example is shown in Fig. 1(a).
It appears that most of the turbulence is due to gravity waves and tides, and especially due to the superposition of gravity waves. The waves add up in such a way that the Richardson number R; drops below 0.25, and even 0.0, producing instability. FRITTS and RASTOGI (1985) have shown that convective breakdown seems to be the major cause, especially at high frequencies. DESAUBIES and SMITH (1982) have modelled the results of an ensemble of gravity waves adding together, and do indeed find that a random distribution of

Fig. 1. (a) Typical plot of turbulent layering observed with a VHF radar. Note the intermittency both in time and space (from CZECHOWSKY et al., 1979). (b) Profiles of Brunt-Vaisala frequency squared, wind shear squared and Richardson number for an ensemble of gravity waves in the atmosphere (adapted from simulations using oceanic internal gravity waves performed by DESAUBIES and SMITH, 1982). Absolute scales are different for the atmosphere and the oceans, so scalesare not shown for the first
two graphs). (c) Histogram of typical layer thickness distributions, again adapted from DESAUBIES and SMITH (1982). Scales are not shown precisely because they are different for the oceanic and atmospheric cases, but approximate scales are shown based on experimental experience. (d) Distribution of layer separations, again from DESAUBIES and SMITH (1982). Similar comments apply with respect to scales as for (c).

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breaking heights, times, and layer depths result. An example of their work is shown in Fig.
1(b), which shows plots of the windshears, Brunt-Vaisala frequency and Richardson number
as a function of depth in the ocean. Whilst Fig. 1(b) applies for internal waves in the oceans,
similar concepts nevertheless apply in the atmosphere.
Also shown in Figs. 1(c) and 1(d) are plots of the distributions of layer thicknesses and
layer separations produced in the analysis of DESAUBIES and SMITH (1982). Again, the scales
used by DESAUBIES and SMITH (1982) do not apply in the atmosphere, but the distributions
are undoubtedly quite representative in form. The scales have been roughly adjusted to match
atmospheric conditions, although it should be emphasized that the rescaling is only
approximate. Partial support for the rescaling used is provided by experimental measurements
of turbulent layer thicknesses presented by CRANE (1980), who showed stratospheric layer
thicknesses varying from a few tens of metres out to a kilometre or so. In the mesosphere,
layer thicknesses tend to be greater than in the stratosphere. With respect to horizontal
extents, a first order estimate can be obtained by assuming that the large scale structure of the
"envelope" of the turbulent patches is controlled to some degree by rotational effects of the
Earth, in which case the ratio of the vertical scale H to the horizontal scale L is
approximately f to N, where f is the inertial frequency and N the Vaisala-Brunt frequency. L
corresponds to Rossby's deformation radius for H. The ratio might be typically 1:100 in mid-
latitudes. However, f/N probably represents an upper limit to H/L because other processes
may tend to break up the horizontal coherence of an extended layer of this type.
It is not always true that the distribution of turbulence is completely random, however.
At times a single large quasi-monochromatic wave can dominate the fluctuating part of the
wind field, and layers of turbulence can appear at quasi-regular height intervals (e.g. FRITTS et
al., 1988; MURAOKA et al., 1988). Even then, however, the layers of strongest turbulence are
separated in space.
The consequences of this intermittency are important. They mean, for example, that we
must revisualize how large-scale turbulent diffusion takes place. An important proposal due
to DEWAN (1981) and WOODMAN and RASTOGI (1984) suggested that the random occurrence of
layers acts like a Monte Carlo process gradually causing diffusion, as first one layer forms,
causing diffusion, and later another forms to cause transport over the depth of that layer. Thus
the factors which control the large-scale diffusion are not the rates of diffusion across
individual layers, but the frequency of occurrence and depth of individual layers (Fig. 2). Any
determinations of effective diffusion coefficients must take this into account.
Other consequences of the intermittency of turbulence include the possibility that the
average rates of diffusivity of momentum and heat may be different, and that the Prandtl
number may exceed 1, and perhaps be in the range 1 to 3 (FRITTS and DUNKERTON, 1985).
This is to say that if one parameterizes the rate of heat transport as a KT (•Ý ƒÆ /•Ýz), where

•ÝƒÆ /•Ýz is the mean temperature gradient, ignoring the effects of the wave on the mean, then

the effective diffusive coefficient which must be used to describe the rate of diffusion is less

than it would be if we properly included the effect of the wave in •ÝƒÆ/•Ýz. This is not so for
momentum diffusion, because u' and w' are not in phase quadrature. FRITTS and DUNKERTON
(1985) have proposed this process as a mechanism to explain the conclusions of STROBEL et
al. (1987) about the turbulent Prandtl number in the atmosphere.

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Fig. 2. Illustration of the process of atmospheric vertical diffusion proposed by DEWAN (1981) and WOODMAN and RASTOGI, (1984). (a) A layer of turbulence forms at time t1 and constituents diffuse against their background gradient. The small graph on the left-hand side shows a representative
profile of the mean density of a constituent, and in this case diffusion will be upwards across the layer. (b) Constituents could only diffuse across the layer in (a), and no further, because the layer has
a limited depth. The layer in time dies out, and this decayed layer is shown as the broken line. Eventually, at some later time t2, a new layer forms above the present one, and is then able to allow further diffusion of the constituents upward. The overall rate of large scale diffusion therefore depends
at least as much on the frequency of formation of these layers as it does on the actual turbulent intensity within the layers.

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. Measurements of Turbulence.

At this level of understanding, we have sufficient knowledge to make useful measurements of turbulence intensities. The three most direct techniques for measurements of turbulence energy dissipation rates used have been based on balloons (stratosphere), rockets and radar experiments. There are many subtleties associated with these techniques which we do not wish to dwell on here. They include such aspects as radar beam-broading, layer thickness, rocket probe response times and biases such as the fact that balloons can only be launched in fair weather. Nevertheless, useful measurements have been made. When high resolution measurements are recorded, one expects (classically) a spectral analysis to reveal a k-5/3 power law, and the structure function should be proportional to r2/3. (It is not the

purpose of this article to explain turbulence at an elementary level, and the reader is referred to some of the more well known texts on turbulence for discussions about the reasons for these laws. We will concentrate here on results of observations.) BARAT (1982) and colleagues measured detailed temperature and velocity fluctuation in the stratosphere with balloon-borne anemometers. They found that the resultant structure function showed at least some ranges in which an r2/3 law held (Fig. 3(a)) and used their data to infer turbulent energy dissipation rates in the stratosphere. Rocket measurements have been able to produce detailed spectra of ion density fluctuations, (e.g. Fig. 3(b), taken from THRANE et al. (1987)), as well as mass spectrometer measurements of fluctuations in various atmospheric species (LUEBKEN

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et al., 1987). Whilst not always showing the classic k-5/3 Kolmogroff law, most spectra hovered around the mark, and inferences about turbulent energy dissipation rate were possible. Large variability was found, and values were generally in the range 10-4 to 10-1 Wkg-1 above 80 km.

Mesospheric radar measurements have also been made, particularly at Adelaide, Australia (HOCKING, 1988). The measurements reported from there represent the largest data set yet obtained at any one site, with data being recorded between 80 and 95 km continuously at 2-h intervals for 3 years (Fig. 4). In this case, spectral width determinations were used to infer estimates of the turbulent energy dissipation rates (HOCKING, 1983a, b).
Figure 5 shows a plot of the distributions of eddy diffusion coefficients in the atmosphere between ground level and 120km altitude, taken from a large collection of references, and inferred both from large scale modelling studies of the type performed by GARCIA and SOLOMON (1985), JOHNSON and WILKINS (1965) etc., as well as direct experimental observations. There is large scatter, but the diagram at least shows'the general

Fig. 3. (a) Typical longitudinal velocity structure function in the inertial subrange for stratospheric altitudes (BARAT, 1982). (b) "Typical" best-case spectrum of electron density fluctuations in the mesosphere, illustrating the inertial (k-5/3) and viscous (k-7) subranges. (from THRANE et al., 1987).

Fig. 4. Seasonal variations of the turbulent energy dissipation rates measured with a 2 MHz radar at a mid-latitude site (Adelaide, Australia).

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Fig. 5. Collection of many different measurements of turbulent eddy diffusion coefficient in the atmosphere. The symbols refer to different authors, but there is no need to list these in this article.
The solid black dots are the medians at each height, and the vertical "borders" show the 16th and
84th percentiles. Above 80 km the data are due to a mixture of measurements of Km and Kt, but in the region 60 to 80 km they are dominated by Km measurements. Thus the Kt results of STROBEL et al.
(1987) are shown separately here; it is the fact that these are much less than Km in this range which
led STROBEL et al. (1987) to propose that the large scale Prandtl number in the atmosphere is greater
than unity.

trends, and gives a fairly reliable range of values which can be used for future modelling studies, and for determining the "drag" forces which waves propagating through the region might experience. Nevertheless, the warnings made earlier about this simple "K" parameter should always be borne in mind in such applications.
Seasonal variations have not been presented in Fig. 5, because at this stage it is too difficult to be certain just what the month to month variation actually is. Figure 4 showed

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seasonal variations for Adelaide, Australia, but it is not clear just how representative this example is. Evidence from other sites suggests it is at least typical of mid-latitudes.
One feature which is not very clear in Fig. 5 is the tendency for a region of very small values of energy dissipation rate (a local minimum) to exist around the menopause (e.g. THRANE et al., 1987; LUEBKEN et al., 1987). Figure 5 does not show this as clearly as individual profiles can, because variations in height of the menopause, both as a function of latitude and season, tend to mask it. Many individual studies show that such a minimum does exist, however.

6. The Turbopause Region

One of the least understood regions of the Earth's atmosphere is undoubtedly the region between 90 and 150km altitude. Below 90km, ion and electron motions are largely driven by neutral motions, whilst above 150km electric and magnetic fields more and more strongly dominate the plasma motions. It appears that the wave motions and tides which are observed at altitudes below about 90-100km generally have their origins in the troposphere and stratosphere, from which they propagate upwards. Substantial damping of these wave motions often occurs in the region 70 to 120km altitude. Above about 120km, locally generated wave motions can be important; for example, electrojet motions, and Lorentz forcing and Joule heating can generate waves. This is especially so in auroral regions. The high conductivity may even allow magnetospheric motions to influence ion motions at 120 km altitude. The region between 90 and 120km therefore represents an important transition region between (almost) complete neutral control at the lower altitudes and "plasma" control of atmospheric motions at the higher altitudes. It is also in this region that turbulence begins to be heavily damped, but because of the critical nature of the region, the very existence of turbulence at these heights can have important ramifications for atmospheric motions there. Studies of turbulence in this region are the hardest of all to make. Many of the tracer techniques which can be used at lower altitudes rely on measurements of ion or electron motions, and assume that these are carried by the neutral turbulence. Such assumptions are not valid above about 100km. Rocket releases of chemo-luminescent tracers, and chaff clouds, have been utilized, and were used especially in the late 1960's and early 1970's. For a long time between about 1975 and 1983 however they were not used much at all. Chaff has been re-introduced in some recent campaigns (e.g. THRANE et al., 1987), but mainly for measurements below about 80km altitude. Measurements of turbulence in the region above 100km are therefore limited in number. However, we do know that whilst turbulence is less frequent at these heights, it does from time to time occur; evidence for small patches of turbulence have been found up to 115km altitude (REEs et al., 1972).
There are at least two ways to define the turbopause. Firstly, it can be the height above which there is no real evidence of turbulence in say an individual rocket flight. Alternatively, one can measure a "longer-term" tracer like say the relative number densities of inert gaseous elements like Helium and Argon. This relative ratio is a constant at the lowest heights, because turbulence keeps all the atmospheric species well mixed. But above this height the densities of the respective species decay with their own scale heights, dependent on their respective molecular weights, so that their relative densities change as a function of height. The point where the ratio departs from a constant can be used as an alternative measure of the turbopause height. Sometimes this alternative definition is called the "homopause". Both definitions have been used in the past, but they are not necessarily the same physical height.

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The first definition is in some ways less rigorous; the temporal and spatial intermittency of turbulence means that the "turbopause height" determined on an individual rocket flight will depend on whether the rocket happens to pass through a high altitude patch of turbulence. However, even long term statistical studies of such data may not give the same "turbopause" heights; even though turbulence might be detected at high altitude, it may not be present often enough to keep the atmosphere uniformly mixed. VON ZAHN et al. (1990) have recently made simultaneous studies of both the mean Ar to N2 ratio and also the occurrence of small scale turbulence, and comparisons between these two different "definitions" are now at last possible. Reviews and some recent detailed studies of the height of the turbopause have been reported by DANILOV (1984) and DANILOV et al. (1991).
The height of the turbopause has implications for ionospheric studies, since its height determines the relative ratios of inert gas densities as a function of height. That is to say that the gas number densities all decay exponentially with increasing height in the higher regions, but the exact ratio of their densities at any one height depends on the point where they were originally equal (i.e. the turbopause, or more specifically the homopause). In fact this principle has been used in the past to make crude estimates of the height of the turbopause. By measuring the ratio of densities of inert species with satellites, and then tracking back down (assuming that the number densities of each species were governed by independent exponential decay rates dependent only on their molecular weights) attempts were made to determine the location of the turbopause by determining where the curves for each species intersected. BLUM and SCHUCHARDT (1978) and BLUM et al. (1978) tried to explain seasonal variations of the ratios of such densities by invoking seasonal variations of the height of the turbopause. Whilst it now seems that these seasonal variations are not related to the height of the turbopause, but rather to other larger scale advective processes, this does serve as an example of the potential way in which turbulence in the upper mesosphere and lower thermosphere could have implications for the ionosphere up to heights substantially higher than the turbopause. Another well known example in which turbulent diffusion across the menopause has been invoked is to explain the annual variation of mesospheric ozone (THOMAS et al., 1984). It has been proposed in this model that turbulent diffusion of odd oxygen down from the thermosphere plays a role in determining the mesospheric ozone density. Again the explanation has yet to be fully verified, but it serves as another example of the ways in which mesosphere-thermosphere coupling can be important.

7. Fine Structure

To really understand turbulence properly, however, requires better knowledge of the fine-scale structure. Although we call it isotropic, individual density structures are far from isotropic. In a stably stratified environment, intense mixing produces a mixture of irregular shapes, and as the driving source of the turbulence becomes less effective, the density inhomogeneities tend to settle out into a horizontally layered structure of interleaved fine scales. PAO (1973) has made high resolution studies of turbulence in highly stratified environments, as has MCEWAN (1983a, b), and both show the tendency for the turbulence to stretch out horizontally and "laminate" as the turbulence dies.
Direct observations also suggest that turbulence in the middle atmosphere is anisotropic.
The left hand side section of Fig. 6 shows typical wind and temperature fluctuations through a turbulent layer, made in this case by an instrumented balloon in the stratosphere (BARAT, 1982). Note the tendency for larger amplitude temperature oscillations at the top and bottom

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Fig. 6. Collection of diagrams from several sources, all showing evidence in some way of anisotropic turbulence. The left hand side diagrams show experimentally determined velocity and temperature fluctuations through a turbulent layer in the stratosphere, from DALAUDIER and SIDI (1987). The middle section shows a proposal (see text) describing the expected mean degree of turbulent anisotropy within a turbulent layer, and shows greater anisotropy at the edges. The right hand side profile shows
the shape of the mean temperature profile in a turbulent layer at two stages of evolution, as predicted
by the model of KLAASSEN and PELTIER, (1985a, b), and adapted for comparison with the experimental results shown on the left. The broken line in this last graph shows an initial temperature inversion, and the solid line the profile expected after dynamic turbulence has existed for some time. The straight solid line shows an adiabatic temperature profile for reference. The agreement with experiment is quite remarkable, even in some of the fine detail. The occurrence of very stable temperature gradients at the top and bottom is very pronounced, both in the experimental results and
the prediction, and it is in this region that the most anisotropic turbulence is expected. One would
also expect the strongest radio-wave backscatter from these sections at the top and bottom, where the fluctuations in temperature as a function of height are the strongest.

of the layer, especially in the temperature fluctuations. HOCKING et al. (1984), HOCKING

(1985) and WOODMAN and CHu (1989) have proposed that turbulent layers tend to comprise quasi-isotropic "eddies" in their centres, but elongated horizontal "blini" towards the edges. The latter exist particularly in the stable regions at the top and bottom of the turbulent layer, as illustrated in the centre of Fig. 6. Numerical work by KLAASSEN and PELTIER (1985a, b) has also shown a tendency for oscillations as a function of height to develop during the evolution of a turbulent layer (RHS of Fig. 6), which partly explains why the extended blini develop towards the edges of the layers. The agreement between the model of KLAASSEN and PELTIER (1985a, b) and the experimental data is quite remarkable, even in the fine details of the resultant temperature profile, and serves as some support for the model. The comparison is not of course conclusive, expecially since the balloon data shows only a one-dimensional traverse through the region. Other simulations (e.g. TANAKA, 1975) have shown the formation of Kelvin-Helmholtz billows, which develop temperature profiles something like those shown in Fig. 6, and which do not necessarily lead to regions of anisotropic turbulence, but the details of the final breakdown of the instability can not usually be followed due to computer limitations in such models. Hence it is never clear in such models

just what the structure of the resultant turbulence is. Nevertheless this author believes that

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Fig. 6 does truly represent a reasonable model of turbulent layers in the middle atmosphere for a significant fraction of occasions.
Theoretical modeling also shows a tendency for the generation of horizontal interleaved structures. For example, a closure model employed by SYKEs and LLEWELLEN (1982) has shown the tendency for stratified structures to develop. Simulations of K-H vortices by KLAASSEN and PELTIER (1985a, b) also have shown a tendency for interleaved structures in temperature to develop as a result of the turbulence.
That stratified structures and turbulence do coexist has been shown by radar measurements. WOODMAN et al. (1981) presented evidence for this, and more recently HOCKING (1987) has shown that this can occur using the Adelaide 2 MHz radar. Certainly turbulence seems important as a diffusive mechanism, but it is not yet clear just how important it is from an energetic viewpoint. It is not clear whether turbulence can force the background state to alter substantially, although observations of turbulent layers with adiabatic temperature gradients have been made. It is possible that the turbulence could have driven the atmosphere towards such a state, but it is also possible that the turbulence only exists there because the atmosphere was already adiabatic. In general, turbulence must act for some time to be able to drive a layer to an adiabatic state. Studies of such questions of energetics are still needed.

8. New Perspectives

There are always new ways of visualizing turbulence, and new aspects to its
interpretation. One recent advance (WALTERSCHEID and HOCKING, 1991) concerns the process
of large-scale diffusion in the atmosphere. We have already seen earlier proposals to explain
how large-scale diffusion occurs in the atmosphere, but an important new model has also
been developed. A gravity wave carries a parcel of air in an elliptical orbit and returns it
(almost) to its start position after one cycle. But detailed analysis shows that it does not in
fact return exactly to its start position. In fact the particle drifts slightly from its start
position, and this constitutes a so-called "Stokes drift". This is true even for nondissipating
waves, but more so for dissipating ones. Figure 7 shows a typical particle "orbit" due to the
influence of several waves. When one has a spectrum of waves, the "Stokes drifts" of the
waves add in a random-like manner to produce a sort of random walk. This results in
dispersion of an initially compact cloud of particles. Computer calculations have shown that
this process gives effective diffusion coefficients (for whatever defining such a parameter is
worth!) of •`100-200 m2s-l. This process may be a major means of large-scale diffusion in
the middle atmosphere.
Other recent advances include studies by KELLEY et al. (1987, 1988) which have shown
that the electron density inertial range may extend to scales much smaller than those for
neutral fluctuations, at least in the case of high Schmidt number. (The Schmidt number in
the ionosphere is the ratio of the diffusion coefficient of neutral particles divided by that of
the ions. When it is large, it means that eddies of ions are less easily destroyed by diffusive
processes, since the ion diffusion coefficient is small, and hence inertial range turbulence
extends to much smaller scales in the ion fluctuations than in the neutrals.) This type of
study may not have direct relevance to ionospheric work just yet, but it does serve to
illustrate very clearly that turbulence in the middle atmosphere and lower thermosphere still
requires considerable study.

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Fig. 7. Lagrangian trajectory calculations of parcel displacement as a function of time, for a parcel acting under the influence of a spectrum of gravity waves in the atmosphere (from WALTERSCHEID and HOCKING, 1991).

9. Conclusions

The importance of turbulence in the middle atmosphere for ionospheric studies rests mainly in two main areas. Firstly, the mesosphere acts as a filter for upward propagating

gravity waves and tides, and turbulence acts as a "damping agent" for such waves. The waves themselves provide much of the coupling between the lower and upper atmosphere. Thus the strength and distribution of turbulence in the middle atmosphere is important because of the way it affects the "filter characteristics" of this region. Secondly, the turbulence which occurs in the region around 95-120km altitude has direct impact on the dynamics of the atmosphere in the important coupling region where the "neutrally controlled" part of the atmosphere meets the "plasma-controlled" altitudes.
The nature of turbulence in the middle atmosphere has been outlined, and typical "eddy diffusion coefficients" have been presented in Fig. 5. Nevertheless, warnings about the

physical significance of such "diffusion coefficients" have been given, and one should always be aware of the assumptions and limitations implicit in such "K" values.

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