The Saturation of Gravity Waves in the Middle Atmosphere. Part III: Formation of the Turbopause and of Turbulent Layers beneath It

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(Manuscript received 28 May 1990, in final form 21 December 1990)

ABSTRACT

A Doppler-spread theory for the "saturation" of middle-atmosphere gravity–wave spectra (in vertical wavenumber m) is presented in a companion paper. It includes a formula for the large-m limit that would be imposed by the onset of instability. At sufficiently great heights, however, this limit may not be attained because of dissipation imposed by molecular viscosity and conductivity. Here, the transition between the two regimes is taken to mark the turbopause—the level at which turbulence ceases—and relevant relations are obtained. These are shown to be consistent not only with observations at turbopause levels, but also, after extrapolation downward through five orders of magnitude in atmospheric density (with the use of the Doppler-spread theory), with similar observations in the middle stratosphere.

In a second application, the concepts behind the Doppler-spread theory are applied to circumstances that would be found on individual occasions (rather than in statistical ensembles, which the basic theory treats). Horizontally stratified layers of intensified turbulence are then found to be expected, perhaps in the stratosphere and more probably in the mesosphere, as is observed. The layers are often observed by medium-frequency, partial-reflection radar techniques, including in particular spaced-receiver "drift" techniques that are often taken to represent the ambient winds. Such an interpretation is confirmed to be appropriate to the present model.

1. Introduction

The irregular winds of the middle atmosphere are most commonly attributed to atmospheric gravity waves propagating upward from underlying sources, as is discussed in some detail in section 1 of Part I (Hines 1991a, henceforth H91a). From the very outset of this interpretation, these waves (and, subsequently, their longer-period tidal cousins) have been recognized as the likely generators, via instability, of turbulence in the same region (e.g., Transactions 1959).

A criterion for instability in a single, monochromatic gravity wave was established by Hodges (1967). This criterion is conveniently expressed in terms of the rms vertical shear of horizontal wind, σ_S, as $\sigma_S = 2^{-1/2} = 0.71$, the shear being nondimensionalized (here and henceforth) by division of the true shear by the buoyancy frequency N_0.

Recent attention has been focused on the case of a broad spectrum of gravity waves, rather than the monochromatic case, in an effort to explain the saturation of broad spectra observed in the middle atmosphere (see section 1 of H91a). For the purpose, the critical rms shear $\sigma_{S\mathrm{crit}}$ was taken to be of order unity by Dewan and Good (1986, henceforth DG86) and to be $2^{-1/2}$ once again by Smith et al. (1987, henceforth SFV87), but in neither case was a detailed determination of this critical value undertaken. The full development appropriate to broad spectra was pursued in my own critique of those two papers (H91a), where the probability of instability was found to increase smoothly with σ_S from values as low as zero. The rise was sharpest at values of σ_S near unity, consistent with the values proposed by DG86 and SFV87, and I suggested $\sigma_{S\mathrm{crit}} = 1$ as a convenient measure of the critical shear. Also in H91a (section 6), I established that, with a wave spectrum representative of those found in the middle atmosphere, shear was dominated by small-scale (in the vertical) wave modes whereas winds were dominated by large-scale modes.

In Part II (Hines 1991b, henceforth H91b), I have presented a new theory for the saturation of the gravity-wave spectrum. It was based upon the Doppler shifting of the smaller-scale members of the spectrum—primarily those members having vertical wavenumber m exceeding a certain characteristic value m_C that marks the small-m onset of the saturated large-m "tail"—toward even smaller scales (larger m) and, ultimately, to obliteration as they approached critical-layer conditions and their m values came to exceed some maximum, m_M. The Doppler shifting was imposed by the horizontal winds of the full wave spectrum, but primarily by those of relatively large vertical wavelength (small m).

With the adoption of a "modified Desaubies" spectral model (following VanZandt and Fritts 1989), this

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theory led to well specified vertical variations of the rms wind amplitude σ_T and of the characteristic vertical scale size m_C^{-1}, both parameters increasing with height with a scale height equal to four times the atmospheric density scale height H. The length of the saturated-tail spectrum (in m), as measured by m_M, was taken to be determined by the criterion for instability, the tail being just long enough to produce instability and no longer. Spectral components with larger m, if observed, were to be interpreted as representing the turbulence that the instability of the wave system would generate. The vertical scale size of the transition to turbulence, m_M^{-1}, was also found to increase with height in company with m_C^{-1}, with a scale height $4H$, if, as might be assumed, the appropriate criterion for instability was independent of height.

The picture thus presented leads now in a natural way to a mechanism for the formation of the turbopause, at heights typically taken as 100–110 km, where the turbulence of the lower levels gives way to the nonturbulent regime of the upper atmosphere. The mechanism results from an increase in the importance of diffusive molecular processes—viscosity and heat conduction—as height and mean-free-path increase. These processes introduce their own maximum permissible vertical wavenumber $m_{M\mathrm{mol}}$ on the wave spectrum, a value that decreases with increase of height more rapidly than does the corresponding $m_{M\mathrm{inst}}$ derived from instability. At heights such that $m_{M\mathrm{mol}} < m_{M\mathrm{inst}}$, molecular dissipation limits the length of the tail to values too small to produce instability: turbulence is no longer generated, and the turbopause has been reached and passed.

A rough analysis of this mechanism of turbopause formation is conducted and applied to observations here in section 2. [The mechanism itself was proposed by Hines (1968), in the spirit of earlier, vague suggestions (Transactions, 1959), but only now has a model spectrum suitable for its testing become available.] Parameters adopted as representative of turbopause heights are shown to yield, via the theory, values for other parameters that are consistent with turbopause observations, thereby giving some support to the proposed mechanism. Moreover, with the use of the Doppler-spread model of H91b the deduced parameters are shown to extrapolate downward to the middle stratosphere, through five orders of magnitude change in atmospheric density, yielding corresponding values that are closely compatible with observations made there. The results of section 2 therefore provide further support for the Doppler-spread theory in its own right, in addition to their tendency to confirm the proposed mechanism of turbopause formation.

The basic analysis of H91b was statistical only, in that the spectrum of wave-induced winds was specified by a probability distribution only. The resultant key parameters, such as σ_S, σ_T, m_C and m_M, varied only smoothly with height, as would the intensity of the associated turbulence below the turbopause. But in actuality, real-life observations are made with a single realization of the wind field at a time, however thoroughly these observations may be mixed subsequently. This fact, ignored in H91b, leads to conclusions that differ in some respects from those implicit in the basic analysis by itself. These differences are encountered initially in section 2, where it is recognized that turbulent layers may be formed on occasion at heights above the normal turbopause. The same differences are likely to arise at somewhat lower levels as well, producing layers of enhanced turbulence sandwiched between layers of diminished turbulence. This layering, which may well account for the observed layering of radar returns from the middle atmosphere, is discussed briefly in section 3 along with associated implications relating to partial-reflection observations in the mesosphere.

The overall thrust of this series of three companion papers is summarized in a concluding discussion in section 4.

2. Formation of the turbopause

For purposes of this section, I adopt the modified Desaubies spectral model represented by (3.7)–(3.14) of H91b, including its height variations, and take it to be applicable from the turbopause all the way down to the tropopause.

The turbopause may be expected to lie where dissipative molecular processes come to dominate instability processes in removing wave energy at large vertical wavenumbers. They would do so, presumably, by quenching the large-m end of the wave spectrum at values of m less than those required for the generation of instability, $m_{M\mathrm{inst}}$. The latter is given by (4.1) of H91b:

$$m_{M\mathrm{inst}} = m_C \exp(\pi\sigma_{S\mathrm{crit}}^2) = [N_0/2\sigma_T] \exp(\pi\sigma_{S\mathrm{crit}}^2), \qquad (2.1)$$

N_0 being the buoyancy frequency and σ_T the rms horizontal wind speed (in an azimuthally isotropic spectrum); σ_T varies with height z as $\exp(z/4H)$, hence $m_{M\mathrm{inst}}$ as $\exp(-z/4H)$.

For the proposed process of turbopause formation to be feasible, this m_M would have to come to exceed the m_M imposed by molecular processes, $m_{M\mathrm{mol}}$. Atmospheric gravity waves are quenched by those processes when the molecular kinematic viscosity η exceeds something like

$$\eta_{\mathrm{crit}} = \Omega^3/2\pi N_0{}^2 h^2, \qquad (2.2)$$

rewritten in present notation from (49) of Hines (1960) and valid under present approximations (which are taken from H91a,b), with h the horizontal wavenumber and Ω the intrinsic frequency. This equation may be rewritten, by elimination of Ω through the use of the approximate dispersion relation $\Omega/h = N_0/m$, as a condition specifying $m_{M\mathrm{mol}}$:

$$m_{M\mathrm{mol}} = [N_0 h/2\pi\eta]^{1/3}. \qquad (2.3)$$

Since η varies with height roughly as $\exp(z/H)$, $m_{M\mathrm{mol}}$ decreases with increase of height as $\exp(-z/3H)$, which is more rapid than the decrease of $m_{M\mathrm{inst}}$. Accordingly, it is indeed possible for the decreasing $m_{M\mathrm{mol}}$ to "overtake" the decreasing $m_{M\mathrm{inst}}$ and produce the termination of turbulence in the manner proposed.

If the transition is taken to occur at the height for which $m_{M\mathrm{mol}}$ and $m_{M\mathrm{inst}}$ become equal, as suggested above, the resulting equality of (2.1) and (2.3) implies

$$\sigma^2_{S\mathrm{crit}} = \pi^{-1} \ln(2\sigma_T N_0^{-2/3} \lambda_h^{-1/3} \eta^{-1/3}), \qquad (2.4)$$

where λ_h is the horizontal wavelength, $2\pi h^{-1}$. (The dependence on λ_h is unfortunate, but it mirrors the fact that a single m_M was imposed on the model spectrum at an earlier stage; it will have to be dealt with by adopting a representative value for λ_h.) One would like to solve this equation for the height at which it is satisfied, but the appropriate value of $\sigma_{S\mathrm{crit}}$ is as yet unknown to the accuracy required for a meaningful conclusion. (The equation, if rewritten to yield an expression for the height of the turbopause above the 100 km level, for example, gives it as a multiple of $12H$, with the multiplier being the difference of two terms each typically of order unity.) I shall use the equation in the opposite sense, as written, in order to estimate $\sigma_{S\mathrm{crit}}$.

The turbopause is typically taken to lie at a height of 100–110 km, where $\eta \approx 100\ \mathrm{m^2\ s^{-1}}$ and $N_0 \approx 0.02\ \mathrm{s^{-1}}$. If I adopt a representative $\lambda_h \approx 50$ km and a representative $\sigma_T = 30\ \mathrm{m\ s^{-1}}$, then (2.4) yields the critical value $\sigma_{S\mathrm{crit}} = 0.70$, almost exactly the value chosen by SFV87. This seems to be a perfectly reasonable value of σ_S to accept as a condition for the termination of instability (see fourth paragraph below), even if a somewhat larger value, such as 1.0, might be thought to apply for the consistent maintenance of instability at lower altitudes, as suggested in section 1.

In the Doppler-spread theory as developed in H91b, the characteristic vertical wavenumber m_C is given by $N_0/2\sigma_T$ as in (2.1). The foregoing estimates of N_0 and σ_T for the turbopause imply $m_C = 3.3 \times 10^{-4}\ \mathrm{m^{-1}}$ there. In the (modified Desaubies) model being employed here, the peak of the wind power spectrum lies at $m_p = 3^{-1/4} m_C = 0.76\ m_C$, hence at $2.5 \times 10^{-4}\ \mathrm{m^{-1}}$, and the corresponding vertical wavelength is then $\lambda_p = 25$ km. This seems to be a perfectly reasonable estimate for the dominant vertical wavelength near the turbopause. Indeed, something like 25 km is frequently reported observationally to be the dominant vertical wavelength there, usually accompanied by the interpretation that it represents the so-called 1,1 mode of the diurnal tide (even when the winds are found to be varying on a time scale of only a few hours). This tidal interpretation is now seen to be suspect, in the absence of supportive measurements of temporal and/or latitudinal variation.

Incidentally, the estimate $\lambda_p = 25$ km comes close to violating the condition $m^2 H^2 \gg 1$, and the estimate $\lambda_h/\lambda_p \approx 2$ comes close to violating the condition $\Omega^2/N_0^2 \ll 1$, on which the approximations of this paper (and its companions) are based. Evidently further application of this type of analysis at turbopause heights may require more complete theoretical forms.

The inferred minimum vertical wavelength for waves at the turbopause is found to be about 4.0 km, a similarly reasonable value, though this value depends only on the estimates made and on quenching theory, not on the Doppler-spread theory. It is, of course, consistent with other such estimates made on the same basis in earlier work.

The turbopause transition from instability to stability will probably be determined, on any given occasion and in any given location, by local circumstances such as the local small-scale spectrum and its relation to the local horizontal wind. The transition could then occur sharply (as has been reported on occasion), not gradually as might be inferred from the smooth variation of probability of instability with variation of σ_S (Fig. 3 of H91a). Indeed, from the known fact that the overlying F region reveals strong gravity waves almost exclusively in individual bursts (TIDs), it seems quite possible that the spectrum of upgoing waves will have been so depleted at turbopause heights that the statistical treatment given here is no longer applicable. It might well be that the spectrum at those heights has been narrowed to something approximating a single wave, in which case the value $\sigma_{S\mathrm{crit}} = 0.70$ just obtained can be readily understood on the basis of Hodges' early analysis. Alternatively, the inferred transition at the turbopause might really be a transition between a highly probable and a highly improbable occurrence of specific wave spectra that lead to instability. With suitable small-scale wave spectra and large-scale wind oscillations on a given occasion, instability could in fact terminate at one level only to occur again, over a limited height range, at an overlying level (as has sometimes been reported).

With the model spectrum taken to be applicable down to the tropopause, it can now be employed to estimate the wave spectrum in the middle stratosphere (ca. 25 km, where again $N_0 \approx 0.02\ \mathrm{s^{-1}}$), for which there are some good data (e.g., Dewan et al. 1984). There, both λ_p and σ_T will be decreased by the fourth root of the midstratosphere–turbopause density ratio. I take that ratio to be 10^5 and its fourth root is then 18. This implies that, in the middle stratosphere, the horizontal-wind power spectrum would peak at $\lambda_p = 1.4$ km and the rms horizontal wind would be 1.7 $\mathrm{m\ s^{-1}}$. If $m_{M\mathrm{inst}}$ is similarly scaled, it yields a transition from waves to turbulence at a vertical wavelength of about 220 m, whereas if $\sigma_{S\mathrm{crit}}$ is raised to 1 before the scaling is done it yields a vertical wavelength of 46 m for the transition. The observations in fact are said to exhibit a tail with log–log slope of −3.0 extending from about 1 km down to about 200 m in vertical wavelength, with some curvature (leading to a mean slope of −2.7) at smaller wavelengths, down to about 40 m. The observations are said to be unreliable outside this range, but one can read into them a possible peaking at larger wavelengths and a possible change of nature, hinted at by the onset of strong fluctuations of spectral density, at smaller wavelengths. (The curvature, incidentally, is consistent with that of the theoretical curves in Fig. 4 of H91b.) The mean power spectral density at a vertical wavelength of 1 km was found to be 3.42 $(\mathrm{m\ s}^2)/(\mathrm{c\ m}^{-1})$, which converts into present units as giving a tail spectrum of $0.306\ N_0{}^2/m^3$. This is to be compared with $N_0{}^2/\pi m^3 = 0.318\ N_0{}^2/m^3$ in the present theoretical model.

Despite its admittedly crude level of development to date, one could hardly ask more of the Doppler-spread theory (in conjunction with the modified Desaubies spectral form and the present model of turbopause formation) than these comparisons provide.

It should be specially noted that, in the Doppler-spread theory above the turbopause, molecular dissipation simply replaces instability as the mechanism of removal of the large-m waves. The tail is shortened in consequence, but its form and intensity remain unchanged. These tail properties, and their universality, then have nothing whatever to do with the occurrence or non-occurrence of instability. This fact indicates a complete departure of Doppler-spread theory from the linear-instability theory of saturation as it has been presented in the past, for example by DG86 and SFV87, despite the similarity of the model spectra developed by the two.

3. Formation of turbulent layers

In the basic Doppler-spread theory as developed in H91b, the wave spectrum was azimuthally isotropic (except in appendix C) and governing parameters such as σ_S, σ_T, m_C and m_M varied smoothly with increase of height as the winds of the large-scale waves increased in amplitude. In actuality, however, the wave spectrum will not be azimuthally isotropic nor will the listed parameters vary smoothly. Indeed, the listed parameters have only statistical meaning and so will be inapplicable to any single realization of the ensemble of observational states. The concepts underlying the Doppler-spread theory must then be applied (mentally) if the facts attending a single observation are to be understood.

The most basic concept is this: that instability (and hence turbulence) will be produced primarily by certain members of the large-m end of the wave spectrum as they are Doppler shifted, primarily by winds of the small-m end of the wave spectrum (plus any nonwave, truly background, winds), toward critical-layer conditions. One might picture, as an extreme departure from the smoothly varying spectra of the statistical description, a situation in which a single large-scale wave, with winds oscillating in the east–west direction only, dominates the wind spectrum, whereas an ensemble of waves of varying short vertical wavelength and varying azimuth of propagation dominates the shear spectrum. The amplitude of all will tend to increase with height, except for those members of the ensemble that approach critical-layer conditions too closely and are obliterated.

The wind-dominant wave will exhibit, at any instant, a sinusoidal wind profile amplifying with height. On one swing to the eastward wind direction (with increase of height), it will Doppler shift toward larger vertical wavenumbers those members of the shear ensemble that have azimuths of propagation into the eastward half-space. Some of those members may be shifted so far—i.e., to m values exceeding the appropriate $m_{M\mathrm{inst}}$—as to become obliterated as waves in consequence of the instability produced by all, and turbulence will result. More and more members will suffer this fate as the eastward wind increases to its local maximum amplitude and then reverses, whereafter the remaining, previously "stressed", members of the eastward ensemble will relax again toward smaller m values and safety; instability and the production of turbulence are then suppressed.

In the next vertical half-wavelength, the wind-dominant wave will produce winds to the west and the process will be repeated, now with westbound waves enhancing the instability and being subjected, in part, to obliteration. In the next vertical half-wavelength again, the process of obliteration is repeated with eastbound waves, but not until the eastward wind comes close to its previous maximum eastward value (since the waves that would be rendered critical at smaller values have already been obliterated). The wind need not attain quite that previous maximum before producing instability, since the residual small-scale, shear-important waves have larger amplitudes here than they did below and need not be brought quite as close to critical conditions as previously, before they generate instability once again. The eastward wind will now carry a further portion of the shear-important waves into near-critical conditions, since its amplitude will be greater at this greater height, and then it will relax once again and the sequence will be repeated.

This picture provides automatically a layering of the unstable regions and so of turbulence production. This layering can be removed, as in the statistical analysis of H91b, only by the inclusion of wind-important waves propagating into other azimuths an so by the continuous provision of some new portion of the shear-important waves—a portion that previously missed obliteration but now comes to face it. This broad range of azimuths may well be available low in the stratosphere, but much of it is likely to have been filtered out by the background winds of the middle atmosphere before mesospheric heights are reached (e.g., Hines and Reddy 1967). The layering of turbulence is therefore more likely at mesospheric than at stratospheric heights, in this picture, but it can occur in both (as is observed on occasion).

The picture itself is by no means new—it is very similar to that proposed by Holton and Lindzen (1972) in their model of the equatorial quasi-biennial oscillation, for example—but it may now be developed with new ends in mind and with a new understanding of the key elements as they apply to the production of turbulence.

As an example, consider a wind profile dominated by a large-scale east–west oscillation as before, but no longer constrained to be sinusoidal and possibly containing true background winds. A first look at such a profile might lead to the expectation that the heights of strongest large-scale shear should be unstable, the small-scale shears perhaps being unresolved observationally. These heights would lie roughly midway between wind maxima in opposite directions and so would be separated roughly by a representative large-scale vertical half-wavelength. The analysis of Hodges (1967) might then be invoked to argue that, instead, convective destabilization is more likely than shear to produce turbulence, and that it will occur close to a maximum of wind (rather than wind shear), and even that only for the wind in one direction: the layering should be at a spacing of a full vertical wavelength. The present picture leads to quite a different conclusion again: that the production of instability and turbulence depends primarily on small-scale waves that may not even be resolved in the wind data. It depends on these waves as they change within the changing large-scale wind field, but in a complicated fashion. The instability should set in as the large-scale wind speed is approaching and passing its own previous record maximum (so to catch new waves), hence in a region of large-scale shear once again, but the shear must now be in the proper sense, with eastward speed being both comparable to or greater than at lower heights and increasing with height, for example. (The requirement "comparable to or greater than" can be circumvented, on occasion, by smaller-scale wave packets entering the observed region along suitable oblique trajectories.)

In this picture, there is reason to suppose that the intensity of turbulence will tend to increase with increase of the large-scale shear, but not because the shear of the large-scale winds is a major contributor to destabilization. Instead, it would do so because more of the small-scale spectrum will be forced into critical conditions in a given height range when the large-scale shear is strong, and so will deposit more wave energy per unit volume into the turbulence spectrum. But again, a strong shear of the large-scale winds is of no relevance if there are no small-scale waves available to be rendered unstable.

One of the most widely used techniques for mesospheric observations employs medium-frequency radars operating in a partial-reflection (PR) mode, whereby signals are scattered back to the observer from irregularities of ionization of a vertical scale size roughly equal to the radio half-wavelength, typically 100 m or so. Since its earliest days (e.g., Gregory 1956, 1961), this technique has been found to yield echoes from vertically stacked, horizontally layered structures, and these structures have most often been taken to represent fields of enhanced turbulence. Such an interpretation seems quite consistent with the present picture of turbulence generation, with regard both to its layering and to the requisite scale size (since 100 m $\leqslant m_M^{-1}$ as may be estimated for the mesosphere by interpolation between turbopausal and stratospheric numbers in section 2).

Given a turbulence interpretation, it is common to argue that the turbulence will be windborne, and hence that PR spaced-antenna observations of "drift" motions will represent the winds themselves. This conclusion is, in general, open to the counterargument that, while the turbulent irregularities themselves may move with the ambient wind, the radio observations do not reveal the motions of individual irregularities but rather the motion of irregular distributions of irregularities. The latter could differ markedly from the former if, for example, the turbulence was being generated primarily by instability in a single large-scale wave or a collection of large-scale waves: patterns of turbulence would then shift and change in a fashion determined by the waves' propagation, not by any frozen-in motion with the wind. In such circumstances, the "drift" measurement could depart widely from the true wind of the region observed. (See, for example, Hines and Raghava Rao 1968.)

A conclusion reached in section 6 of H91a is important in this connection. There it was found that the regions of instability will be determined largely by the waves having small vertical scales, which are also the waves having small intrinsic horizontal trace speeds Ω/h. For example, half the requisite shear variance derives from waves with trace speeds less than $N_0(m_C m_M)^{-1/2}$. With the use of interpolated values from section 2, this implies speeds less than about 3 m s^{-1} in the mesosphere—a speed rather small in comparison with the rms wind speed and comparable to or less than typical observational errors, hence indistinguishable from zero speed. In such circumstances, the argument of Hines (1968) becomes relevant: a superposition of such waves will lead to a nearly unchanging pattern of irregularity that appears to move horizontally with the ambient wind, even though the waves themselves may be propagating into a wide variety of azimuths. The potential dichotomy between the motion of turbulent irregularities and the motion of the regions of occurrence of turbulence will then no longer occur: both motions closely match the ambient wind velocity, and the PR spaced-antenna observations should indeed reveal that velocity as advocates of the technique have maintained all along.

(This conclusion does not automatically carry over to total-reflection "drift" observations made at greater heights, typically with higher radio frequencies. For them, no corresponding case for small intrinsic trace speeds has ever been made, and it seems somewhat unlikely that one can be.)

3. Discussion

This three-part study began as an attempt to understand the saturation of the middle-atmosphere gravity-wave spectrum on the basis of linear-instability theory. That theory, as developed to date, was found in H91a to be missing an important physical attribute—the extremely slow trace speeds of the waves of the saturated portion of the spectrum relative to the rms wind field in which they found themselves propagating.

That attribute was added in H91b, in an admittedly approximate fashion, and a rudimentary Doppler-spread theory of saturation emerged. It led in due course to a more tightly specified spectral form, one that included (for a particular model spectrum) well-defined height variations, for example. In retrospect, the Doppler-spread theory might be said to have added only physics, since the model spectra it produces are essentially the same as those that the linear-instability theory might well have been made to produce empirically. In particular, it provided a physical basis for the conclusion already reached by SFV87, that the universality of the m^{-3} tail implied a constant value for $m_C \sigma_T$ (in present notation)—a value argued in H91b to be about $N_0/2$. But, in adding the physics, it simultaneously altered the conceptual understanding of the processes at work, introducing causes where none had been present before (as in the derivation of $m_{M\mathrm{inst}}$) and even reversing cause and effect at one point. To some extent, the relationship between the linear-instability theory and the Doppler-spread theory is then analogous to that between kinematics and dynamics: the kinematics may express the relationships adequately, while the dynamics provide the physical insight behind them.

In this third part, two unanticipated spinoffs of the new theory have been examined briefly: the formation of the turbopause and the formation of layered turbulence beneath it. Both have been found to be consistent with observation. Perhaps there are more such spinoffs still to come.

Acknowledgments. The Arecibo Observatory is funded by the National Science Foundation through a cooperative agreement with Cornell University.

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