NAGW-826
NAG5-796
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P-24

CONSTRAINTS ON GRAVITY WAVE INDUCED DIFFUSION IN THE MIDDLE ATMOSPHERE

by

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February 1988

1. Abstract

A review of the important constraints on gravity wave induced diffusion of chemical tracers, heat, and momentum is given. Ground-based microwave spectroscopy measurements of H_2O and CO and rocket-based mass spectrometer measurements of Ar constrain the eddy diffusion coefficient for constituent transport (K_{zz}) to be (1-3)x 10^5 cm^2 s^{-1} in the upper mesosphere. Atomic oxygen data also limits K_{zz} to a comparable value at the mesopause. From the energy balance of the upper mesosphere the eddy diffusion coefficient for heat transport (D_H) is, at most, 6 x 10^5 cm^2 s^{-1} at the mesopause and decreasing substantially with decreasing altitude. The available evidence for mean wind deceleration and the corresponding eddy diffusion coefficient for momentum stresses (D_M) suggests that it is at least 1 x 10^6 cm^2 s^{-1}, in the upper mesosphere. Consequently the eddy Prandtl number for macroscopic scale lengths is > 3.

2. Introduction

The importance of gravity waves in the mesosphere and lower thermosphere has been known for more than twenty five years. The seminal paper by Hines (1960) opened up an area of research which unfortunately took over twenty years to realize the full significance of gravity waves in this region of the atmosphere. Research efforts by Pitteway and Hines (1963), Lindzen (1967, 1968), and Hodges (1969) provided the needed foundation for present day research on gravity wave induced diffusion, but did not then yield a parameterization of the gravity wave stresses on the mean circulation and structure of the middle atmosphere. It was not until 1981 when Lindzen (1981) published his perceptive recipe for parameterizing the gravity wave momentum stresses on the mean zonal flow that the full impact of gravity waves on the middle atmosphere was comprehended. An unfortunate result of this important paper was the almost sole emphasis on the momentum balance of the middle atmosphere.

Dynamical models were constructed which examined in great detail the deceleration effects of gravity waves on the zonal flow but omitted any thermodynamic effects on the global mean temperature structure (e.g. Holton, 1983; Garcia and Solomon, 1985). However in the Garcia and Solomon (1985) study diffusive effects on temperature departures from the global mean temperature were included. The consensus of these studies was that the eddy diffusion coefficient for momentum stresses (D_M) is ~ 10^6 cm^2 s^{-1} (Lindzen, 1981; Holton, 1982, 1983; Garcia and Solomon, 1985). It should be noted that this quantity is proportional to $(\bar{u}-c)^4$, where $\bar{u}$ is the mean zonal wind and c is the phase speed of the gravity wave. This extreme sensitivity of D_M to $(\bar{u}-c)$ makes an accurate calculation of its value very difficult from observed values of $\bar{u}$ and c. Because the mean circulation and associated transport of constituents and heat are also directly dependent on $D_M/(\bar{u}-c)$, theoretical prediction of their strengths is also subject to the uncertain magnitude of $(\bar{u}-c)^3$ through the uncertainities associated with the computed radiative drive and the input gravity wave spectrum of phase speeds.

Johnson and Wilkins (1965) noted that the observed lower thermospheric temperature gradient was inconsistent with molecular conduction of heat alone and concluded that eddy transport of potential temperature was required. In a more quantitative study Johnson and Gottlieb (1970) inferred from the globally averaged heat balance that the eddy diffusion coefficient for heat transport (D_H) was 1×10^5 cm^2 s^{-1} at 60 km and increased to 1×10^7 cm^2 s^{-1} at 120 km. Of more importance to us are the values in the 75-85 km region which they deduced to be $(4\text{-}7) \times 10^5$ cm^2 s^{-1}. Hunten (1974) argued that these analyses ignored an essential aspect of the physics, namely that dissipation of mechanical energy to generate turbulence might yield net heating rather than net cooling of the mesopause region. His inference of net heating depended crucially on the assumption that turbulence persists only if the Richardson number of the flow (Ri) remains at its critical, onset value, generally ≤ 0.2. Johnson (1975) noted that turbulence can persist in a flow with Ri as large as 1 and thus argued that steady-state turbulence would maintain the very stable lower thermosphere near Ri ~ 1, which implies net cooling by the action of turbulence on the mean circulation.

In a similar vein, chemical tracers bear signatures of transport effects and Colegrove et al. (1965, 1966) used the O/O_2 density ratio to infer the vertical eddy diffusion coefficient for tracer transport (K_{zz}). Their average value in the 80-120 km region was in the range of $(0.8\text{-}8) \times 10^6$ cm^2 s^{-1}, with a preferred value of 4×10^6 cm^2 s^{-1}. It is standard practice in 1D photochemical models with eddy and molecular diffusive transport to empirically deduce the eddy diffusion coefficient profile with a minor constituent whose density profile is extremely sensitive to the adopted values. In the one extreme, Hunten and Strobel (1974) used argon measurements by von Zahn (1970) to deduce a homopause value of $K_{zz} \sim 3 \times 10^5$ cm^2 s^{-1}. In the other extreme Keneshea and Zimmerman (1970) argued for highly structured profiles of K_{zz} with peak values in excess of 1×10^7 cm^2 s^{-1}. In the lower thermosphere the uncertainty in the heating and cooling rates at that time could not exclude such large values of the eddy diffusion. In a turbulent atmosphere, tracer and potential temperature transport by eddies should be described by the same coefficient, i.e., $D_H = K_{zz}$, in the limit of a chemically inert species and adiabatic motion. For chemically active species one must include chemical acceleration of vertical transport which arises physically because the mixing ratio perturbation for chemical species with finite chemical loss undergoes a phase shift relative to the wave velocity fields such that the eddy fluxes are nonzero rather than in quadrature, as in the limit of no chemical loss (Strobel, 1981). Strobel et al. (1987) in a study of mesospheric chemistry found this term to be potentially important for only odd oxygen and then in a subtle way.

The focus of this paper is on gravity wave induced diffusion over macroscopic vertical length scales on the order of the atmospheric scale height, H. There is a significant body of literature on measurements and interpretation of smaller scale diffusion observed by rocket experiments and radars and reviewed in depth by Hocking (1985). Turbulent layers observed by these techniques tend to be "sub-macroscopic". There is the difficult question on how to relate diffusion on these smaller scales to the larger scales of interest here. The reader should keep in mind that additional constraints may be supplied by this broad data base, although the extrapolation to larger scales is not obvious and straightforward.

3. Constraints on K_{zz}

The momentum stresses of saturated gravity waves are largest in the upper mesosphere and it is there that independent constraints on gravity wave diffusion are most important. Investigations by Allen et al. (1981) and Strobel et al. (1987) identified H_2O, CO, O_3, and O as the chemical species most diagnostic of tracer transport in the upper mesosphere. Of these species Strobel et al. (1987) argued that the vertical H_2O profile was best suited to infer the correct magnitude of vertical mixing in the mesosphere.

In their study they derived an approximate expression for the vertical eddy diffusion coefficient in terms of the eddy diffusion coefficient for heat transport and chemical acceleration of vertical transport

$$K_{zz}(i) = D_H + 2x10^{10} L_i A \qquad (1)$$

with the numerical value in cgs units, A is a measure of gravity wave amplitude and equal to 1 at the saturation limit, and L_i is the chemical loss rate of specie i. According to recent measurements by Vincent and Fritts (1987), gravity wave amplitudes vary throughout the year in the range of A = 0.6-2. With the exception of odd oxygen, all other species which undergo significant transport in the mesosphere have $L_iA \ll 5 \times 10^{-9}$ D_H and thus $K_{zz}(i) = D_H$.

In Fig. 1 theoretical H_2O mixing ratio profiles from an 1D eddy diffusion, photochemical model (Allen et al., 1981 with updates discussed in Strobel et al., 1987) are compared with observed values. The corresponding vertical eddy diffusion coefficient (D_H) profiles are illustrated in Fig. 2. The mesospheric water vapor measurements were obtained by ground-based microwave spectroscopy from the Jet Propulsion Laboratory in Pasadena, California during the time period April to June 1984 (Bevilacqua et al., 1983). At this time vertical motions associated with the mean meridional circulation would be expected to be upward (Garcia and Solomon, 1985) and the inferred values of D_H ($=K_{zz}$) from the observed H_2O mixing ratio profiles should be upper limits. The water vapor is most sensitive to transport in the 70-80 km region and there the observations clearly constrain D_H to a range of (1-2) x 10^5 cm^2 s^{-1} as Models B, C, and D show in comparison to Model A. Model A was preferred by Allen et al. (1981, 1984) in order to satisfy the water vapor measurement by H. Trinks (private communication, 1979) at 90 km. Ground-based microwave spectroscopy measurements imply that the H_2O mixing ratio at 80 km is comparable to the Trinks' value at 90 km. The rapid dissociative loss of H_2O in the 80-90 km region requires a basic incompatibility between these measurements. If one examines the microwave signature of H_2O, the signal to noise increases with increasing altitude with the highest quality data in the

70-80 km region (Bevilacqua et al., 1987a). The associated error bars clearly exclude the observation of a water vapor profile even remotely close in mixing ratio to Model A in Fig. 1.

Ground-based microwave spectroscopy measurements of CO (Bevilacqua et al., 1985) confirm the deduction of slow vertical mixing in the mesosphere (Strobel et al., 1987). But it must be kept in mind that the large vertical scale height of the CO density profile makes it somewhat insensitive to diffusive transport and hence more susceptible to advective transport. This follows from a comparison of the respective time constants (H^2/K_{zz}) and ($H/\bar{w}$), where $\bar{w}$ is the zonally averaged vertical wind. In addition the Solar Mesosphere Explorer (SME) near-infrared spectrometer measurements of absolute O_3 concentrations are best understood with slow vertical mixing which yields low H_2O mixing ratios and odd hydrogen densities. This produces reduced catalytic destruction of O_3 by odd hydrogen and hence high ozone mixing ratios in better agreement with SME measurements. Brasseur and Offermann (1986) analyzed O concentration measurements and concluded that the vertical eddy diffusion coefficient is about 10^5 cm^2 s^{-1} at the mesopause, also consistent with the above results. Note that O unlike CO has a very small scale height at and below the mesopause which renders it extremely sensitive to diffusive transport. It is interesting to note that the argon measurements that guided Hunten and Strobel (1974) to adopt low vertical mixing in the mesopause region are in excellent agreement with inferences from other species. Evidence from absolute concentrations of chemical tracers of mesospheric transport thus suggest that K_{zz} ($=D_H$, for all tracers but O and O_3) is ~ $(1\text{-}3)\times10^5$ cm^2 s^{-1} in the upper mesosphere.

4. Constraints on D_H

As alluded to in the Introduction it is more difficult to obtain powerful constraints on the eddy diffusion coefficient for heat transport because it involves extracting a small residual from large terms in approximate balance in the thermodynamic heat equation. The importance of diffusive transport of potential temperature by breaking gravity waves was demonstrated by Schoeberl et al. (1983) in a numerical study of gravity wave breaking and stress in the mesosphere. Chao and Schoeberl (1984) emphasized that it is the turbulence created by the breaking gravity wave that transports potential temperature rather than coherent, $\overline{w'T'}$ heat transport by the gravity wave. Apruzese et al. (1984) in a study of the globally averaged temperature of the mesosphere and lower thermosphere concluded that D_H must be less than 10^6 cm^2 s^{-1}. Strobel et al. (1985) reiterated the arguments of Apruzese et al. (1984) and argued that the eddy Prandtl number ($P_r = D_M/D_H$) over macroscopic scale lengths must be large if D_M exceeds 10^6 cm^2 s^{-1}, because $D_H \leq 6\times10^5 cm^2 s^{-1}$.

To quantify this discussion, let us examine the thermodynamic heat equation for the globally averaged temperature field, $<T>$, with < > denoting global average and overbar zonal average

$$\frac{\partial <T>}{\partial t} = \frac{1}{\rho c_p} < Q_{UV} - Q_{IR} > + \frac{1}{\rho c_p}\frac{\partial}{\partial z} < \lambda \frac{\partial T}{\partial z} > - \frac{1}{\rho c_p}\left(\frac{\partial}{\partial z}\right) < \rho \bar{w} T > + < H_g - C_g > \qquad (2)$$

where Q_{UV} is the solar heating rate due to O_2 and O_3 absorption, Q_{IR} is infrared cooling rate, ρ is the mass density of the atmosphere, λ is the molecular heat conductivity, c_p is the specific heat at constant pressure, $\bar{w}$ is the zonally averaged vertical velocity,

$$H_g = \epsilon \frac{N^2 D_M}{2 c_p}\left(1 + P_r^{-1}\right) \qquad (3)$$

is the conversion rate of wave energy to heat with efficiency ϵ, and

$$C_g = \frac{1}{\rho c_p P_r}\frac{\partial}{\partial z}\left(\rho c_p D_M\left(\frac{\partial T}{\partial z} + \frac{g}{c_p}\right)\right) \qquad (4)$$

is the cooling rate due to the divergence of the downward turbulent or eddy heat flux. Here

$$N^2 = \frac{g}{T}\left(\frac{\partial T}{\partial z} + \frac{g}{c_p}\right) \qquad (5)$$

is the buoyancy frequency. Note that when $D_w^{-1}\partial D_w/\partial z$ is less than H^{-1} the heat flux is divergent and there is cooling, whereas if it is greater than H^{-1} the heat flux is convergent and heating occurs. The expressions for H_g and C_g were derived by Schoeberl et al. (1983) under the assumption of Lindzen (1981) that when gravity waves break or saturate their amplitudes remain constant with altitude.

Equation (2) has been solved for the globally averaged, steady-state temperature without the term $<\rho\bar{w}T>$ which represents the convergence of the downward heat flux associated with the anticorrelation of the zonal mean vertical velocity (a result of gravity wave breaking) and the zonally averaged temperature. For a Prandtl number of 1, Fig. 3 shows illustrative globally averaged temperature profiles with the corresponding vertical eddy diffusion coefficient profiles in Fig. 4. These results led Apruzese et al. (1984) and Strobel et al. (1985) to deduce an upper limit on D_H of 6 x 10 ^{5} cm^2 s^{-1} from thermodynamic considerations only. Model A in Fig. 4 is the appropriate height dependent upper limit on D_H. A comparison of these results with earlier work of Johnson and Gottlieb (1970) based on the Kuhn and London (1969) IR cooling rates indicates agreement in the critical mesopause region to within 50%, the Johnson and Gottlieb D_H values being larger. It should be noted that they did not include gravity wave heating in their calculation and other input quantities such as solar flux, IR cooling rates, O recombination heating rates, etc., were different from values used by Apruzese et al. (1984) and Strobel et al. (1985). But the Johnson and Gottlieb (1970) results are still an accurate representation of heat balance constraints on gravity wave induced diffusion. The gravity wave heating term (H_g, Eq.2) is at most 25% of the total solar UV and O recombination heating rate in the calculations of Apruzese et al. (1984) and Strobel et al. (1985).

To further illustrate aspects of the above discussion, the analytic expressions of Strobel et al. (1985) are adopted for the gravity wave terms in Eq. (2) with $\epsilon=1$

$$H_g - C_g = 1.5\left(\frac{7}{P_r}\frac{H}{H_D} + 1 - \frac{6}{P_r}\right)D_M\left(10^6 cm^2 s^{-1}\right) \qquad units\left(K \quad d^{-1}\right) \qquad (6)$$

where

$$H_D^{-1} = \frac{1}{D_M}\frac{\partial D_M}{\partial z}$$

When $P_r = 1$ and D_M is constant, then $H_g - C_g = -7.5$ and -22 K d^{-1} for $D_M = 10^6$ and 3×10^6 $cm^2\ s^{-1}$, respectively. These values should be compared to the total, globally averaged, solar heating rate including O recombination of Apruzese et al. (1984),

z(km)	= 65	70	75	80	85	90	95	100
Q_{UV}(K d^{-1})	= 3.2	1.6	1.7	5.0	9.2	8.9	13	18

The upper mesosphere (70-80 km) is thus seen to be a critical region where the solar heating rate is low. With constant D_M (=D_H) a value as small as 2×10^5 $cm^2\ s^{-1}$ is sufficient to balance solar heating at 75 km. In competition with solar heating is CO_2 infrared cooling primarily in the 15 μ bands. The most accurate calculations of this cooling are by Dickinson (1984) who included as accurately as possible non-LTE effects. His globally averaged cooling rates are

z(km)	= 65	70	75	80	85	90	95	100
Q_{IR}(K d^{-1})	= 3.0	1.0	0.5	1.1	3.0	4.4	7.4	11.5

A comparison of Q_{UV} and Q_{IR} below the mesopause suggests that gravity waves are not driving the globally averaged structure of the mesosphere far from radiative equilibrium. If one deduces values of D_M from a balance of Q_{UV} - Q_{IR} with H_g - C_g given by Eq. (6) one would infer mesospheric values similar to Model A in Fig. 4. These values up to 80 km would be consistent with the constraints on K_{zz} (=D_H) derived from ground-based microwave spectroscopic data discussed above.

One possibility for strong gravity activity in the mesosphere without substantial effects on the transport of constituents and heat is for the induced diffusion over macroscopic scale lengths to have a large effective Prandtl number, a point advocated by Strobel et al. (1985,1987) with strong theoretical support from Fritts and Dunkerton (1985). Fritts and Dunkerton (1985) examined constituent and heat fluxes driven by localized gravity wave breaking where the breaking zones are small in vertical extent in comparison to the vertical wavelength. In their analysis localization of turbulence yields an eddy Prandtl number greater than 2 between saturation ($A = 1$ for saturation amplitude) and modest supersaturation ($A \sim 1.3$). Only for large supersaturation ($A \sim 2$) does P_r approach 1.

For constant D_M in Eq.(6), $P_r = 6$ yields $H_g - C_g = 0$, i.e. no net heating or cooling from gravity waves. High effective eddy Prandtl number turbulence would allow significant deceleration of the mean zonal winds without a substantial diffusive signature in the thermal structure. The physical reason for this result is that the conversion of the gravity wave's kinetic energy still produces significant heating of the atmosphere although conversion of internal energy is reduced. High Prandtl number turbulence leads to sluggish eddy transport of heat and large reductions in the associated divergence of the eddy heat flux. As a consequence the conversion of gravity wave energy into heat can approximately balance the divergence of the eddy heat flux.

It is also worthy to note that Justus (1967) determined P_r to be ~ 3 from photographic tracking of rocket released chemical clouds and analysis of turbulent wind data in the 90-110 km region. There was considerable scatter in his data points but according to his error bars P_r was at least 2.2.

Net gravity wave heating occurs also when D_M increases sufficiently rapidly with height (small H_D); but at some altitude it must level off with a consequent large divergence in the eddy heat flux. The model results of Apruzese et al. (1984) and Strobel et al. (1985) yield gravity wave heating by this circumstance of, at most, 0.4 K d^{-1} and limited to below 80 km.

In Eq. (2) the term involving the divergence of $<\rho\bar{\omega}T>$ has been omitted in the above quantitative discussions. The only model results available to consistently ascertain its importance are from Garcia and Solomon (1985). For solstitial conditions at the mesopause with a mean circulation driven by gravity wave breaking, their globally averaged values are $D_M = D_H = 1.5 \times 10^6$ $cm^2\ s^{-1}$, and $<\bar{\omega}T> = 16K\ cm\ s^{-1}$. The $\bar{\omega}$ and T fields have an approximate height dependence of $\sqrt{\rho}$. The convergence of this heat flux is spread over at least 2-3 scales heights and has a magnitude of ~ 1 K d^{-1}, substantially less than the value of $H_g - C_g$ ~ 11 K d^{-1} for this high diffusivity, $P_r = 1$ model of the middle atmosphere. If the convergence of the $<\rho\bar{\omega}T>$ flux were this large for values of D_H ~ 10^5 $cm^2\ s^{-1}$, then this downward heat flux would be have to be evaluated more carefully in globally averaged heat budgets. The fact that this heat flux is not large is not surprising as $\bar{\omega}$ and T have large amplitudes only at polar latitudes.

5. Constraints on D_M

Of all the eddy diffusion coefficients discussed so far this is the most difficult to evaluate. The fundamental effect is the deceleration of the mean zonal and meridional winds. Theoretical models of the mean circulation of the middle atmosphere are sensitive to a number of input parameters. The radiative drive depends on the net imbalance of Q_{UV} and Q_{IR}. Whereas the former can be reasonably accurately calculated in a large dynamical model, the latter cannot be computed at the level of detail and accuracy of Dickinson's (1984) non-LTE model with present day computer resources. The infrared cooling acts also to damp temperature perturbations associated with the mean circulation and accurate simulation of the temperature and wind fields depends critically on the radiative damping.

The deceleration of the mean zonal wind by gravity wave stresses is written as

$$\frac{\partial \bar{u}}{\partial t} + \ldots = -\frac{1}{\rho}\frac{\partial}{\partial z}(\rho \overline{u'w'}) \approx -\frac{N^2 D_M}{\bar{u}-c} \qquad (7)$$

Thus an approximately accurate value of D_M can only be inferred from the deceleration if independent knowledge of the phase speeds of the gravity waves is available. Furthermore the right hand side of expression (7) holds only if a saturated gravity wave maintains constant amplitude with height, as Lindzen (1981) originally hypothesized. Recently this hypothesis has been questioned; Schoeberl (1988) argues that the amplitude of a saturated gravity wave can still grow with height. Only when the convection zone is comparable in vertical extent to the vertical wavelength will wave growth be significantly attenuated.

On the basis of available estimates for D_M from Lindzen (1981), Holton (1982, 1983), and Garcia and Solomon (1985), it would appear that deceleration of the mesospheric zonal winds require that D_M exceed 10^6 cm^2 s^{-1} on a globally averaged basis in the upper mesosphere, a value definitely in excess of K_{zz} and D_H. For example, a typically deceleration rate of 100 m s^{-1} d^{-1} (Holton, 1983), would for $(\bar{u}-c) = 40 ms^{-1}$ yield $D_M \sim 1.3 \times 10^6$ cm^2 s^{-1}.

6. Concluding Remarks

The most stringent constraint on gravity wave induced diffusion is obtained from ground-based microwave spectroscopy measurements of H_2O with support from microwave measurements of CO and rocket-based mass spectrometer measurements of Ar (von Zahn, 1970). These data constrain K_{zz} ($= D_H$) to be (1-3) x 10^5 cm^2 s^{-1} in the upper mesosphere. Atomic oxygen data also limit K_{zz} to $\sim 10^5$ cm^2 s^{-1} at the mesopause, but indicate an order of magnitude increase in K_{zz} in the first 10 km of the lower thermosphere (Brasseur and Offermann, 1986).

The evidence cited above, which leads to the conclusion that K_{zz} is low in the mesosphere, is based on the absolute concentrations of the selected chemical tracers. An alternate point of view can be advanced, based on the seasonal variations of O and O_3. The green line radiated by O shows a semiannual variation which Garcia and Solomon (1985) argued can only be explained by a semiannual variation in the gravity wave induced diffusive transport. The equinoctal periods are characterized by weak zonal winds and hence, by the Lindzen (1981) parameterization in their model, weak gravity wave induced diffusion. Soltistial periods have characteristically strong zonal winds and are accompanied by strong diffusion, thus creating the semiannual variation in diffusive transport. This variation in diffusive transport should also produce a semiannual component in the seasonal variation of the mesospheric H_2O mixing ratio profile. But Bevilacqua et al. (1987b) found from ground-based microwave measurements of water vapor no obvious semiannual component, only a pronounced annual component. Ozone mixing ratios inferred from SME data exhibit a pronounced semiannual variation particularly in the spring of the first two years of data acquisition (1982-1983) at 80 km. Data from later years do not contain such a distinctive component. The existence of this semiannual component may depend to some degree on vertical displacements of the ozone profile as it contains a minimum value in mixing ratio at ~ 80 km, which if displaced by a few kilometers could create the appearance of a time varying component at a fixed height with the periodicity of the vertical displacement.

Consideration of the energy balance of the upper mesosphere indicates weak departures from radiative equilibrium, consistent with $D_H < 6 \times 10^5$ cm^2 s^{-1} at the mesopause and decreasing substantially with decreasing altitude. The uncertainty attached to the independent calculations of solar UV heating and atmospheric IR cooling cannot exclude the possibility of the upper mesosphere being in radiative equilibrium and the inferred values of D_H being upper limits. Similarly the uncertainty cannot exclude a value of D_H as large as 10^6 cm^2 s^{-1}.

The available calculations to date suggest that D_M is at least 10^6 cm^2 s^{-1}. The consequences of an order of magnitude smaller momentum diffusion coefficient due to more modest reductions in zonal wind deceleration and in $(\bar{u}-c)$ have not been adequately explored, although we note that Schoeberl (1988) advocates reduced wave stress on the mean flow in the mesosphere. It is worth citing the lidar studies of the nighttime Na layer over Urbana, IL by Gardner and Voelz (1986). They found average values for mean flow deceleration of -27.2 m s^{-1} d^{-1} and a corresponding eddy diffusion coefficient, presumably D_M, of 1.8×10^5 cm^2 s^{-1}. Individual measurements yielded deceleration rates up to -200 m s^{-1} d^{-1}, but the long term averages were an order of magnitude less. But Reid and Vincent (1987) found the mean zonal flow deceleration to be typically between -50 and -80 m s^{-1}, with occasional values as large as -190 m s^{-1}. Also they measured comparable mean meridional deceleration rates. Likewise Meek et al. (1985) obtained comparable deceleration rates. Clearly a global climatology mean wind deceleration needs to be constructed from the available, but sparse data base and more precise altitude information is needed to determine where the deceleration actually occurs. For the present we are in somewhat of a dilemma; depending on what evidence is adopted, a case can be made for the Prandtl number (D_M/D_H) applicable to turbulent mixing over macroscopic scale lengths to be anywhere between 1 and 10, although model values of D_M clearly suggest a large Prandtl number. In this author's judgment the available evidence suggests that the Prandtl number is > 3.

Most of the inferences of eddy diffusion coefficients for constituent and heat transport were accomplished with 1D models. The exclusion of mean circulation transport of constituents and heat may render the inferred values of gravity wave induced diffusion, in many instances, to be only upper limits. The low values deduced for K_{zz} and D_H, in fact, suggest that the mesosphere is advectively controlled rather than diffusively controlled.

Acknowledgments: This work was supported by NASA Grants NAGW-826 and NAG5-796 at The Johns Hopkins University.

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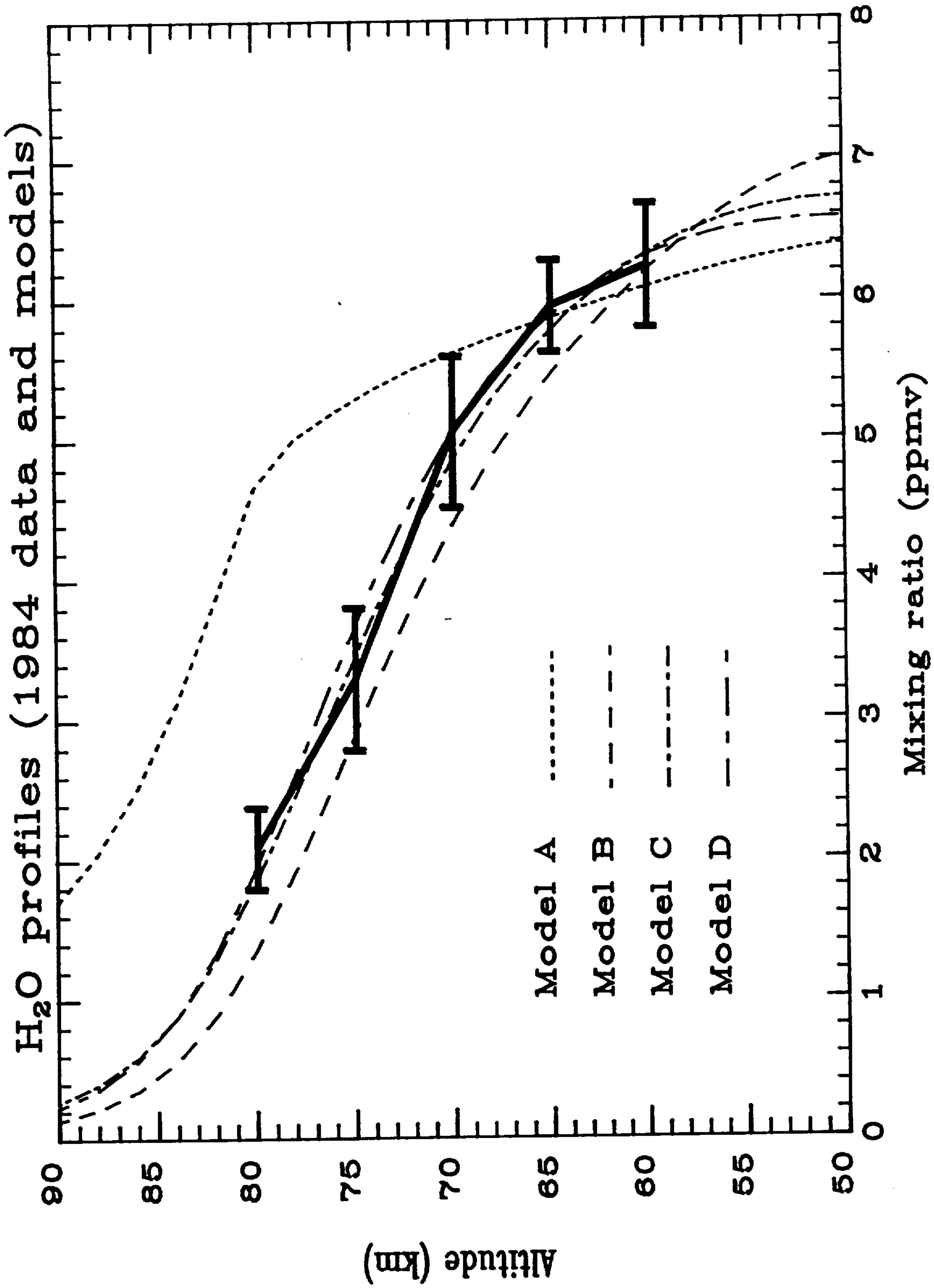
8. Figure Captions

Fig.1. Comparison of ground-based, observed H_2O mixing ratios with 1 σ error bars (Bevilacqua et al., 1985) and model results for the D_H profiles given in Figure 2 and with chemical acceleration for O and O_3. After Strobel et al. (1987). Copyright by the American Geophysical Union.

Fig.2. Vertical eddy diffusion coefficient (D_H) profiles for indicated models in Figure 1. After Strobel et al. (1987). Copyright by the American Geophysical Union.

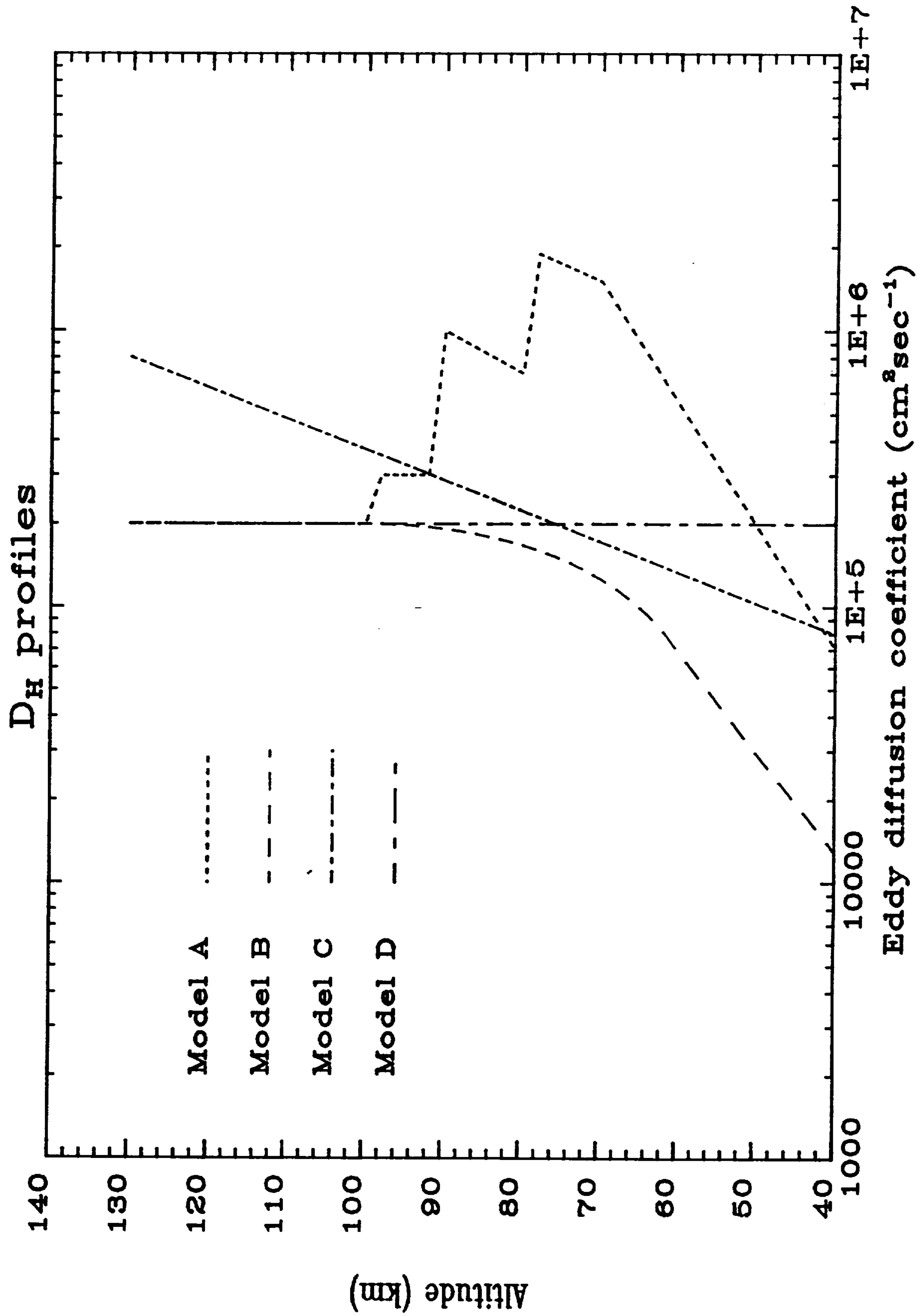
Fig.3. Calculated globally averaged temperature profiles for the D_H profiles given in Figure 4, with $P_r = 1$, and compared with the CIRA 1972 temperature profile. After Strobel et al. (1985). Copyright by the American Geophysical Union.

Fig. 4. Vertical eddy diffusion coefficient (D_H) profiles for indicated models in Figure 3. After Strobel et al. (1985). Copyright by the American Geophysical Union.



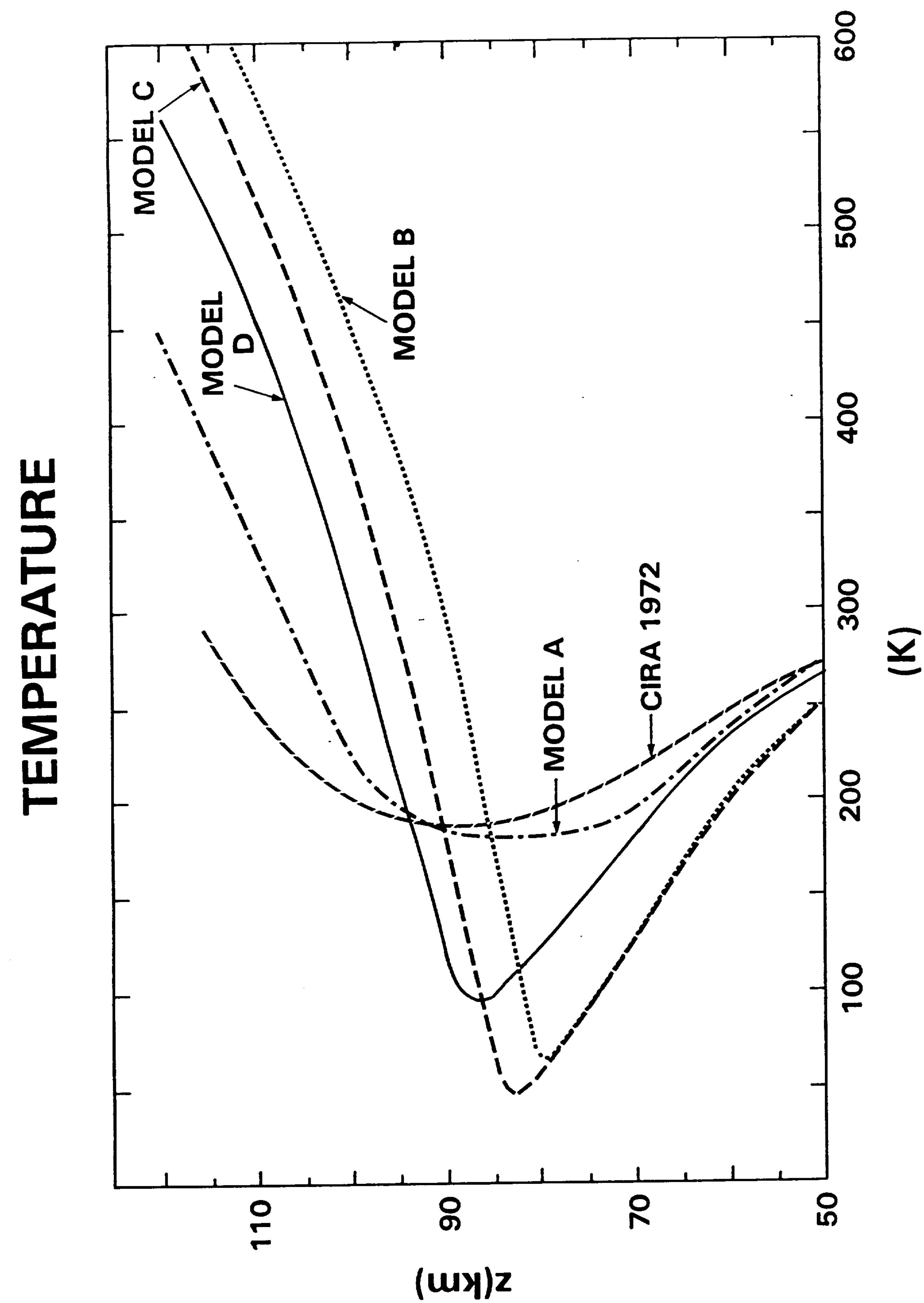
H2O profiles (1984 data and models)
Altitude (km)
Mixing ratio (ppmv)
Model A
Model B
Model C
Model D
90
85
80
75
70
65
60
55
50
0
1
2
3
4
5
6
7
8

Figure 1



D_H profiles
Model A
Model B
Model C
Model D
Altitude (km)
140
130
120
110
100
90
80
70
60
50
40
1000
10000
1E+5
1E+6
1E+7
Eddy diffusion coefficient (cm^{2}sec^{-1})

Figure 2



TEMPERATURE
MODEL C
MODEL B
MODEL D
MODEL A
CIRA 1972
(K)
z(km)
100
200
300
400
500
600
110
90
70
50

Figure 3

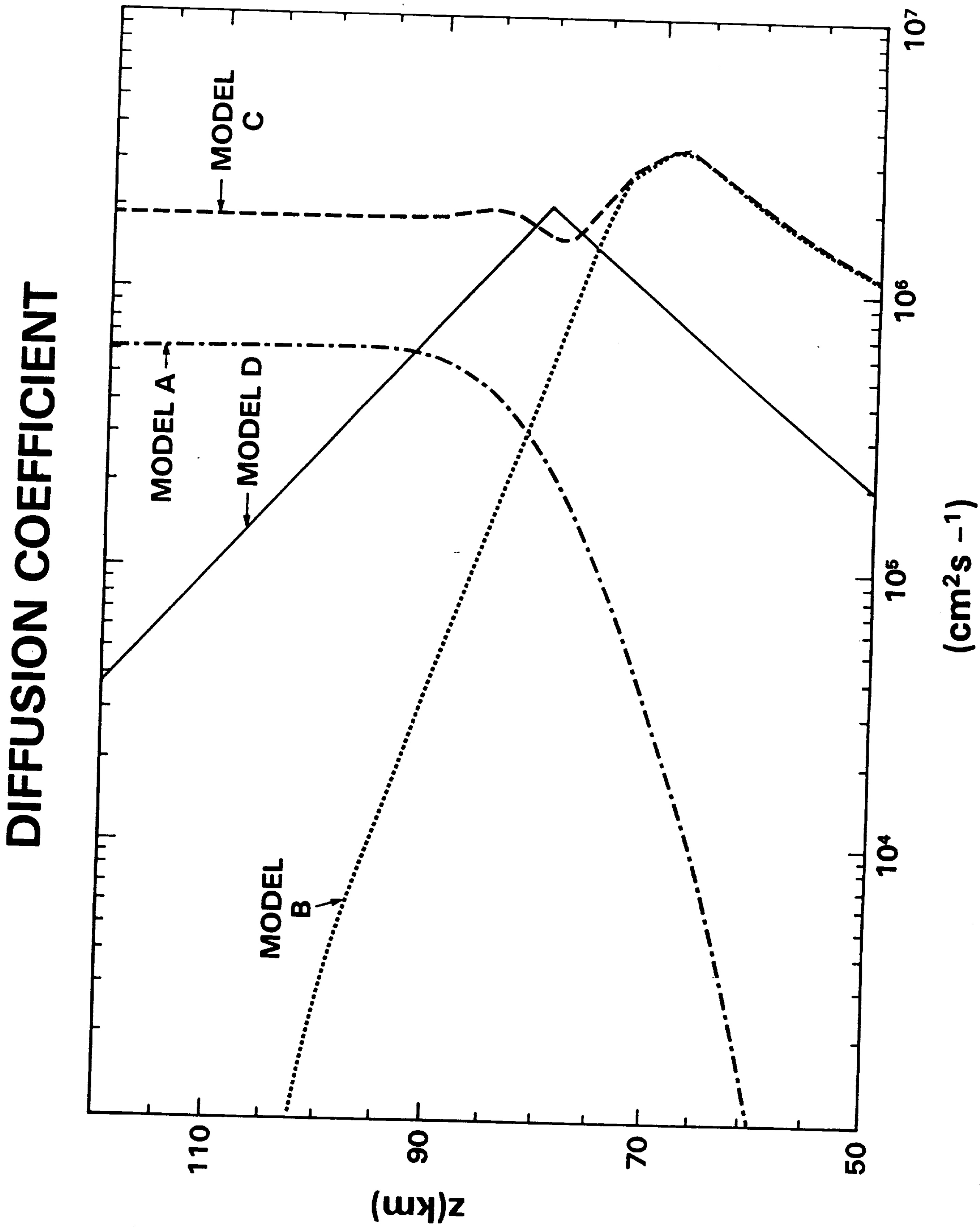

DIFFUSION COEFFICIENT
MODEL C
MODEL A
MODEL D
MODEL B
z(km)
110
90
70
50
10^4
10^5
10^6
10^7
(cm^2s^{-1})

Figure 4