On the large wind shear and fast meridional transport above the mesopause

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[1] Unexpectedly strong winds/shears and fast meridional transport above the mesopause have been revealed from previous rocket and satellite observations. Using Richardson number criteria for dynamic stability, we estimate the maximum wind shears allowable by the background static stability, which peaks above the mesopause. These maximum shears are in general agreement with the large wind shears inferred from the rocket measurements at low and mid-latitudes, indicating the close relationship between the latter and the stability constraint set by the background atmosphere. Diagnostic calculations also indicate that the meridional transport in this region may not be well understood solely by examining the mean meridional circulation, and large amplitude tides/planetary waves can play an important role in the bulk transport of tracers. Strong stochastic winds, presumably due to gravity waves, do not seem to significantly change the large scale pattern of the transport but may extend the range of the tracer movement. Citation: Liu, H.-L. (2007), On the large wind shear and fast meridional transport above the mesopause, Geophys. Res. Lett., 34, L08815, doi:10.1029/2006GL028789.

1. Introduction

[2] Very large winds and wind shears have been consistently observed for decades in the mesosphere and lower thermosphere (MLT) at low and mid-latitudes through sounding rocket measurements at various locations, which cover a wide range of longitudes, seasons, and local times [Larsen, 2002]. The winds often exceed 100 ms$^{-1}$ between 95–115 km, with the maximum reaching 200 ms$^{-1}$. The profiles of the shears peak around the turbopause (100–110 km) with magnitudes often reaching 70 ms$^{-1}$ km$^{-1}$ and as large as 100 ms$^{-1}$ km$^{-1}$. At high latitudes, wind gradients of similar magnitude were observed, along with large temperature gradients, above the summer mesopause (85–90 km) during polar summer mesopause campaigns [e.g., Fritts et al., 1988, 2004]. These winds and wind shears are much larger than those from empirical models and global-scale theoretical models, and thus have been thought to be caused by gravity wave (GW) perturbations and/or GW breaking [Fritts et al., 2004]. It was also suggested that there might be a link between the large winds and the fast meridional transport of water plumes from shuttle exhaust [e.g., Stevens et al., 2005a, and references therein].

[3] In this study, the profiles of the maximum shears are related to the background static stability of the atmosphere by arguing that the magnitude of the horizontal wind shear is constrained by the dynamic stability condition, which depends on the background static stability. We also investigate the roles of mean meridional circulation, tides/planetary waves, and large amplitude GWs in tracer transport in this region to examine the possible link between the large winds and the fast meridional transport proposed in previous studies.

2. Analysis and Discussion

2.1. Attainable Wind Shear and Background Static Stability

[4] From the mesopause to the lower thermosphere, the atmospheric temperature increases rapidly due to heating from radiative absorption and exothermic reactions. Consequently the atmospheric static stability, $N^2 = g/T(dT/dz + g/c_p)$, reaches a maximum in this region, with the temperature gradient being small at the mesopause and the temperature large in the thermosphere. It is straightforward to show that the maximum $N^2$ approximately coincides with the local maximum value of $d^2T/dz^2$ if the variation of $N^2$ dominates over the lapse rate. Figures 1a and 1b shows the Brunt-Väisälä frequency at a specific longitude from the NCAR thermosphere-ionosphere-mesosphere-electrodynamics general circulation model (TIME-GCM) [Roble, 2000, and references therein] simulation under year 2004 equinox and solstice conditions (results similar for other years). The model winds are in general agreement with the UARS climatology [McLandress et al., 1996]. The model resolution in these simulations is $5^\circ \times 5^\circ \times 0.5$ scale height. It can be seen that $N$ peaks near the turbopause height (100–110 km) at most latitudes except in the summer polar region, where the peak $N$ is at $\sim 90$ km due to the lower altitudes of the polar summer mesopause. The maximum values of $N$ are between 0.035–0.04 s$^{-1}$.

[5] One implication of the maximum background static stability at the turbopause height is that large wind shears could be sustained in this region before the atmosphere undergoes dynamic instability. Using $R_i = N^2/S_{max}^2 = 1/4$ as the threshold for dynamic instability (Richardson number, subscript $c$ denoting the critical value, and $S_{max}$ the maximum vertical shear magnitudes of the horizontal wind achievable under dynamic stability), we may take $S_{max} = 2N$ as a proxy for the magnitude of the maximum wind shear attainable prior to the onset of the instability.

[6] It is seen from Figure 1 that the maximum shear could be as large as $\sim 70$ ms$^{-1}$ km$^{-1}$ at turbopause heights at low to mid-latitudes during both equinox and solstice. These
[7] $N$ and $S_{\text{max}}$ vary with longitude due to tides. The vertical profiles of $S_{\text{max}}$ at 18 longitudes (20° apart) and 2.5°N on day 355 are given in Figure 1c. The values of $S_{\text{max}}$ increase from 35–40 m s$^{-1}$ km$^{-1}$ at 80 km to a maximum value of $\sim$73 m s$^{-1}$ km$^{-1}$ at $\sim$105 km, and decrease quite rapidly above, to 30–45 m s$^{-1}$ km$^{-1}$ at 130 km. Comparing these profiles with those of the shear magnitudes at low and middle latitudes [Larsen, 2002, Figure 11] reveals general agreement with the envelope of the majority of the profiles (judged by the density of the profiles). Therefore, this envelope may be interpreted as a measure of the maximum shear allowed by dynamic stability, and thus proportional to the static stability of the background atmosphere (including tides). On the other hand, the perturbative lapse rates due to GWs can modify the large-scale stability threshold, so that the actual maximum shears can fluctuate around the large-scale $S_{\text{max}}$. Further, because $Ri < Ri_c$ is a necessary but insufficient condition for dynamic instability and derived under inviscid assumptions, the actual shear could be even larger, especially as the molecular viscosity becomes increasingly important at the turbopause.

[8] The large vertical shear may arise from GWs and their superposition with tides and the background atmosphere. For upward propagating GWs, the region with maximum background $N$ is where the shear due to wave perturbation becomes large, because the vertical wavelength of a GW is approximately inversely proportional to $N$ according to linear theory. As well as contributing to large wind shears (in the limit of large static stability), some GWs can become unstable and break in this region [VanZandt and Fritts, 1989; Fritts et al., 2004]. VanZandt and Fritts [1989] also demonstrated that the acceleration rate due to a breaking GW is proportional to the vertical gradient of log $N$. Therefore, extensive GW breaking, as well as wave-mean flow interaction and turbulent mixing accompanying the wave breaking, may occur around the altitudes where $N$ maximizes. These processes, however, may not be well parameterized in current global models because of the rapid changes of $N$ and winds in a relatively thin layer (2–3 vertical grids between 100–110 km in current TIME-GCM). This may lead to the underestimation of the circulation strength, as will be discussed in the next section. The large shears and the wave breaking can lead to enhanced turbulence and variability as well as temperature inversion in this altitude range [Larsen et al., 2005].

[9] Because $N$ reaches a local maximum above the tropopause for exactly the same reason, large wind shear can in principle be sustained there too, though the values would be about 20 m s$^{-1}$ km$^{-1}$ smaller than those above the mesopause. Its implication for GW breaking, secondary GW generation, and turbulent mixing between the upper troposphere and lower stratosphere (UTLS) merits further investigation [VanZandt and Fritts, 1989].

2.2. Meridional Transport

[10] It is known that momentum deposition and mean flow acceleration/deceleration will accompany GW breaking, and that the momentum deposition is responsible for the zonal wind reversal at the mesopause. Above the primary wind reversal at the mesopause, there is a secondary wind reversal. The secondary reversal in the summer hemisphere is quite strong and is located between 100–
[11] The GW forcing responsible for the primary jet reversal near the mesopause induces a summer-to-winter circulation through the Coriolis force. The wave forcing causing the secondary jet reversal leads to a meridional circulation in the opposite direction, which can be regarded as a branch of the return flow above the mesopause circulation. Because the wave forcing associated with the secondary jet reversal is stronger in the summer hemisphere, a stronger return flow is expected above the summer mesopause. The residual circulation pattern during boreal winter is illustrated in Figure 2a by normalized residual circulation vectors. Figures 2b and 2c shows the meridional component of the residual circulation, \( \mathbf{v}^* \), for days 1–360 from the same year-long TIME-GCM simulation mentioned above at two pressure levels (around 83 km and 105 km, respectively). At \( \sim 83 \) km the circulation is uniformly summer-to-winter for most of the year. At \( \sim 105 \) km, the winter-to-summer return flow is most evident in the summer hemisphere, but it can extend to the low or mid-latitudes in the winter hemisphere. Although the exact latitude where the meridional circulation changes direction is variable, the area with robust return flow seems to extend further into the winter hemisphere later in the summer/winter season. A relatively fast transition is seen around equinox, when the GW forcing changes in response to the wind change in the stratosphere and mesosphere [Liu and Roble, 2004].

[12] The mean meridional circulation from the model can be compared against the observed transport of water plumes from shuttle exhaust. The transport of the water plumes has been found to be mainly southward on day of year (DOY) 16, 2003 (STS-107 [Stevens et al., 2005b] and northward on DOY 307/1994, 219/1997, and 204/1999 ([Stevens et al., 2002, 2003, 2005a] STS-66, STS-85, STS-93, respectively.). Between 100–115 km, where about half of the main engine exhaust from each shuttle is released, the latitudes of the shuttles were between 30°–40°N [e.g., Stevens et al. 2005b]. During boreal summer (days 204 and 219), the release latitudes (marked by arrows in Figure 2) were well within the region of the above mentioned return flow and would be subject to the winter-to-summer circulation. For boreal winter (days 16 and 307) the release latitudes would sometimes be where the meridional circulation splits (\( \mathbf{v}^* = 0 \)), and it would seem that the plumes could be transported either northward or southward over the vertical range between 100–115 km. The maximum residual mean meridional wind (10–15 ms \(^{-1} \)) from the model is \( \sim 1/3–1/4 \) of the average speed of the plume transport (30–40 ms \(^{-1} \)), which could result from inadequate representation of GW forcing in this relatively thin region.

[13] Apart from the mean circulation, tides, planetary waves, and GWs can also play important roles in the bulk movement of tracers in the MLT. Here we perform off-line kinematic calculations using the winds from the TIME-GCM simulation (DOY 16–18 and 219–221 are taken as examples) to advect tracers released at various locations and to examine their roles in tracer transport. In the calculations, the tracers are tracked at each time step and the zonal, meridional and vertical velocities are interpolated from model grids to the tracer positions to advect the tracers in the following time step. Figure 3 shows the movement of tracers “released” at 16UT on day 16 and 15UT on day 219 from various longitudes (20° apart) at 32.5°N and 105 km.
The movement is recorded at one hour intervals for 48 hours. It is seen in Figure 3a that the tracer displacements in the meridional direction are less than 15° in both directions and the wave number 1 pattern is primarily due to tidal winds. We then repeat the calculation with the model horizontal winds quadrupled (Figure 3b). For days 16–18, the tracers released in the western hemisphere show very prominent movement toward the summer hemisphere (southward), with those released around 100°W reaching 40–50°S after 48 hours. Tracers released from some eastern hemisphere sites move rapidly northward and reach almost 80°N after one day before moving back to lower latitudes. For days 219–221, all tracers are transported northward and after 48 hours they are located between 60 to 90°N. To illustrate the role of tidal transport, the kinematic calculation is repeated with only the eddy winds quadrupled (Figure 3c) while the original zonal mean horizontal winds are retained. From the plot, it is seen that the pattern of the tracer transport is quite similar to that in Figure 3b, though the strength of the transport is not as strong. On the other hand, the pattern of the transport is quite different if the mean horizontal winds are quadrupled while the eddy winds remain unchanged (not shown), with the tracers transported in the opposite directions to those shown in Figures 3b–3c.

These comparisons thus indicate that the global scale waves (tides dominate in this region) and the mean meridional winds affect the tracer transport in different ways. This is further elucidated in Figure 4, which shows the total residual mean meridional wind \( \bar{v}_e \) and the eddy contribution to the residual meridional wind \( \bar{v}_e \) for days 16 and 219. Note that \( \bar{v}_e = \bar{v}^{\text{ZM}} + \bar{v}_e \), where \( \bar{v}^{\text{ZM}} \) is the zonal mean meridional wind. On day 16, \( \bar{v}_e \) is northward poleward of 30–40°N between 95–110 km, opposite to \( \bar{v}_e \) in the same region. Between 30°N and the equator, both \( \bar{v}_e \) and \( \bar{v}_e \) are southward though the latter is much weaker. This explains why the transport patterns are different when \( \bar{v}^{\text{ZM}} \) and \( \bar{v}_e \) are adjusted differently. Large waves can play an important role in the bulk transport of tracers near where \( \bar{v}_e = 0 \); they will move tracers in the northward (southward) phase of the wave(s) into the domain where the northward (southward) \( \bar{v}_e \) becomes increasingly large. This leads to the enhanced meridional transport in Figure 3b compared with Figures 3a and 3c. We can use exactly the same argument to understand the transport patterns on day 219. From these numerical experiments, it is found that the pattern and speed of the water plume transport cannot be achieved solely by increasing the zonal mean meridional wind; it is necessary to also increase the bulk transport by tides and planetary waves. This also suggests that the plume transport is dependent on the local time/longitude of the release. From Figure 3, for example, it appears unlikely that shuttle plume injected at this altitude would have reached Antarctica if it were injected 12 hours later on day 16. Further, the plume could be transported both northward and southward near \( \bar{v}_e = 0 \) if tidal meridional wind changes sign between 100–115 km, which may explain why Lyman \( \alpha \) emission is also visible between 30–60°N on day 18 [Stevens et al., 2005b]. It should also be noted that the southward transport of the tracers on day 16 is reduced if the vertical wind perturbation is quadrupled, which may imply overestimation of tidal vertical wavelength by the model. The significance of tides in plume transport was also suggested by Siskind et al. [2003]. It is worth noting that multiplying the horizontal winds by a factor of 4 in the diagnostic calculation produces the same gain in \( \bar{v}_e \) as doubling the eddy amplitudes. This is because \( \bar{v}_e \) is proportional to \( \bar{v}' \), the eddy amplitudes of the meridional wind and potential temperature.

To reflect the local effect of GWs, including the large acceleration due to wave dissipation, transience and breaking and the temporal and spatial irregularities they induce on winds in the mesosphere [Shepherd et al., 2000], stochastic perturbations are added to the winds at each time.
step in the diagnostic calculation. The horizontal components are between 0–100 ms$^{-1}$, in line with the large wind perturbations observed in the chemical release experiments, and the vertical component is between 0–0.01 ms$^{-1}$. By comparing Figures 3c and 3d, it is found that the stochastic winds do not significantly change the particle transport in the mean sense, but they can disperse the tracers to regions with large residual mean winds and extend the range of the transport.

3. Conclusion

16] As a result of the large background static stability above the mesopause, large values of wind shear can be achieved before the atmosphere becomes dynamically unstable. Using the necessary condition for dynamic instability ($R_i < 1/4$), the maximum attainable wind shear is estimated from the Brunt-Väisälä frequency. This enables us to relate the large observed shears, found to be consistently situated above the mesopause, to the atmospheric background state.

17] The region with the largest static stability is also the region where GWs are more likely to break due to the large perturbative shear and the large perturbative lapse rate, with the vertical wavelengths of GWs decreasing with increasing background static stability. This may explain the maximum variability near the mesopause identified from TIMED/SABER temperature observations (“wave turbopause” [Offermann et al., 2006]). Because $N$ and the winds change rapidly in this relatively thin region ($<10$ km), the impact of the wave breaking may be poorly represented in global models which could result in a weaker residual mean circulation.

18] According to satellite observations and model results, the mean meridional wind splits above the mesopause. Above the summer mesopause the flow is winter-to-summer, opposite to the summer-to-winter circulation at the mesopause and also to the flow above the winter mesopause. The latitude where the meridional wind splits varies with season and extends farther into the winter hemisphere later in the summer/winter season. The direction of transport of water plumes from shuttle exhaust agrees with the residual mean circulation calculated by the TIME-GCM during boreal summer, but the speed of the observed transport is much larger. Diagnostic calculations show that tidal/planetary waves, with amplitudes larger than the tides from TIME-GCM, can also play a key role in the bulk transport of tracers, suggesting dependence of the transport on the local time/longitude of the release. The contribution to the residual mean circulation by these waves could either enhance or weaken the zonal mean meridional flow depending on the latitude and the season, and increases with the square of the wave amplitudes. The waves can also enhance the bulk transport of tracers from regions where the residual circulation is weak to regions where it is strong. This could be especially important for tracers in the vicinity of the latitude where the meridional wind splits, as in the boreal winter cases when the tracers released between 100–115 km could be transported both north and south if the tidal meridional wind changes sign over that altitude range. The effect of dissipative and transient GWs is also assessed in the diagnostic calculations by introducing stochastic winds with random amplitude and directions. It is shown that they do not qualitatively change the transport patterns on large scales, but they can disperse the tracers to regions where the residual mean is large and extend the range of the transport.

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Figure 4. Residual mean meridional winds: (a and c) total and (b and d) eddy contribution from TIME-GCM for days 16 (Figures 4a and 4b) and 219 (Figures 4c and 4d). Contour interval, 2.5 ms$^{-1}$. Solid line, northward. Grey lines, the latitudes range (30–40°N) of the plume release.
References


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