Combined MU radar and ozonesonde measurements of turbulence and ozone fluxes in the tropo-stratosphere over Shigaraki, Japan

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[1] Turbulent diffusivity and turbulent ozone fluxes in the tropo-stratosphere are studied employing simultaneous observations with the Middle and Upper (MU) Atmosphere radar and ozonesondes in Shigaraki, Japan during April 16–24, 1998. A broad region around the tropopause was dynamically active. Maxima of turbulent diffusivity were observed at 8–14 km altitude. Such maxima may produce vertical turbulent ozone fluxes across the tropopause with magnitudes comparable to those required for the global ozone budget. Mesometeorological ozone intrusions may enhance the fluxes.
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1. Introduction

[2] Major global-scale transport of ozone between the stratosphere and troposphere is believed to be produced by the Brewer-Dobson circulation [*Holton et al.*, 1995]. Local downward fluxes and mixing of atmospheric ozone through the tropopause due to synoptic and mesoscale motions and small-scale turbulence may also contribute to global-scale ozone transport [*Lamarque and Hess*, 2002]. But the role of turbulence is not clear and its understanding requires simultaneous observation of atmospheric turbulence and tracers, e.g., ozone. Aircraft and ozonesonde measurements of atmospheric dynamics and ozone transport were described by *Pavelin et al.* [2002] and *Whiteway et al.* [2003]. Also *Krishna Reddy et al.* [2003] compared measurements of surface ozone with the wind profiler data from Gadanki, India.

[3] In this paper we study characteristics of turbulence and turbulent ozone fluxes near the tropopause from the data of first simultaneous MU radar and ozonesonde observations in Shigaraki, Japan (35°N, 136°E) during April 16–24, 1998. For the first time, we have estimated vertical ozone fluxes and their variation in the tropo-stratosphere using MU radar data on turbulent diffusivity simultaneously with measurements of ozone mixing ratio.

2. Measurements and Data Analysis

[4] The MU radar has an antenna a two-way half-power beamwidth of 2.6°. We steered the beam to 5 directions: one vertical and four inclined 10° from vertical [*Fukao et al.*, 1994]. We used vertical height and time resolutions of 150 m and 2 min. During April 16–24, 1998, 21 balloon ozonesondes were launched near the MU radar [*Tomikawa et al.*, 2002]. They measured temperature and ozone concentration with height resolution of 50 m.

[5] The MU radar measures the Doppler velocity variance σ^2, which gives information about turbulence [*Hocking*, 1983, 1999]. Observed values σ^2_{obs} were corrected for spectrum broadening due to mean wind and shear using formulae of *Nastrom* [1997]. The turbulent energy dissipation rate, ε, and diffusivity, K, were calculated from corrected σ^2 in a manner similar to that used by *Fukao et al.* [1994] and *Hocking* [1999]:

$$\varepsilon = C_1\sigma^3/L; \qquad K = C_2\sigma L, \tag{1}$$

where C_1 and C_2 are constants, L is the scale of largest observed turbulent eddies, the latter is restricted by the scale of the atmospheric volume observed, L_r, or by the buoyancy scale

$$L_B = C_3\sigma/N, \tag{2}$$

where C_3 is a constant and N is the Brunt-Väisälä frequency. Following the practice of *Hocking* [1999] we used in (1) $L = L_r$, when $L_r < 0.24\ L_B$, $L = L_B$ at larger L_r, $C_1 = 3.3$, $C_2 = 0.02$ and $C_3 = 6.9$ in (1) and (2). At $L = L_B$ formulae (1) and (2) are equivalent to the ones used by *Fukao et al.* [1994]. Turbulence parameters were estimated for each 2-minute radar data sample and 30-minute median values were calculated. Vertical gradients of temperature and ozone mixing ratio were obtained from ozonesonde data using least-square fit approximations within vertical layers of 250 m thickness.

3. Results

[6] Measured distributions of temperature and wind during the experiment show the passage of an atmospheric front on April 19, 1998, when temperatures below and above altitude 10 km became, respectively, higher and lower than before. The front was followed by a decrease

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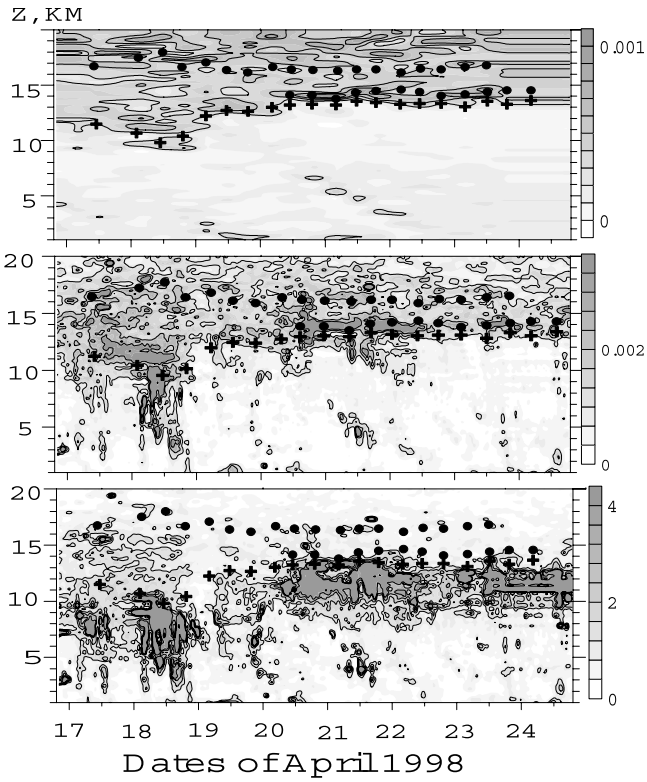

Figure 1. Height-time distributions of (top) Brunt-Väisälä frequency squared, (middle) turbulent energy dissipation rate and (bottom) turbulent diffusivity. Long ticks on horizontal axis correspond to 00LT of the respective day. Contour intervals for plots from top to bottom are 3×10^{-4} s^{-2}, 1×10^{-3} m^2s^{-3}, and 0.8 m^2s^{-1}, respectively. Dots and crosses denote the cold point and temperature gradient tropopause, respectively.

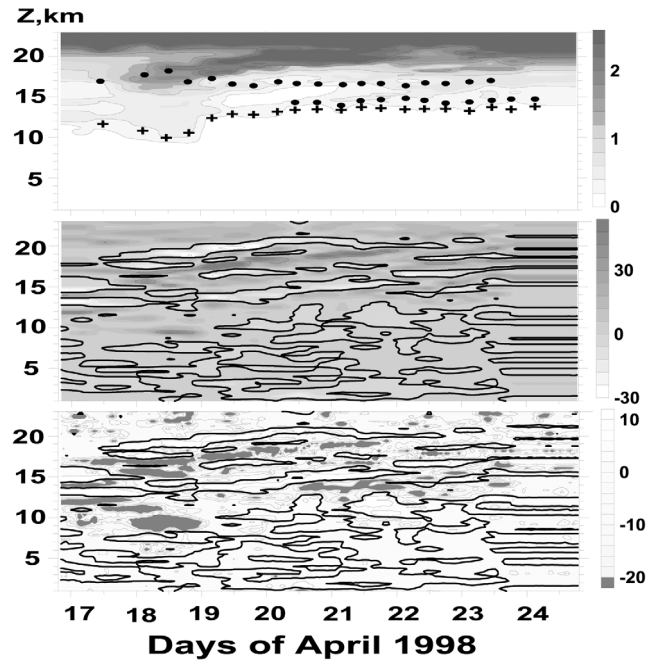

Figure 2. Same as Figure 1, but for (top) ozone mixing ratio in ppbv, (middle) effective vertical gradient of ozone number density in units of 10^{18} m^{-4} and (bottom) vertical turbulent ozone flux in units of 10^{14} $m^{-2}s^{-1}$. Contour intervals at top plot are 0.6 ppv. Thick lines at middle and bottom plots show zero value contours. Dots and crosses see Figure 1.

in zonal wind velocity and a reverse of the meridional wind from northward to southward.

[7] The Brunt-Väisälä frequency squared, N^2, is shown in Figure 1 (top). It was larger above 10–13 km than below. The lower boundary of this increase corresponded to the "temperature gradient tropopause" [*Dameris*, 2002]. Before April 19, a temperature minimum existed at 16–18 km altitude. After the front passed, the lower boundary of the tropopause region moved from 10–11 km up to 12–13 km altitude and a second temperature minimum appeared (see Figure 1) forming a thick tropopause transition region.

[8] Calculated values of ε and K, also shown in Figure 1, have their maxima at 8–14 km altitude, near the temperature gradient tropopause, located within the upper tropospheric jet stream. Since the large vertical shear and small static stability just below the tropopause minimize the Richardson number, an increase in turbulent activity might be produced by instabilities of the jet stream [*Joseph et al.*, 2003]. Another reason for an increase in turbulent activity may be an increase in amplitude and instability of atmospheric waves propagating from below due to a sharp increase in N^2 in the tropopause region [*VanZandt and Fritts*, 1989; *Gavrilov and Fukao*, 2004]. During the passage of the front on April 18–19, the temperature gradient tropopause descended and the maximum of turbulent activity shifted to 5–10 km altitude (Figure 1) due to instabilities in the frontal zone.

[9] The measured ozone mixing ratio, c_3, plotted in Figure 2 (top), shows elevated values down to 8–10 km altitude. Above 15 km, this increase is believed to be connected with ozone lamina advected by atmospheric circulation [*Tomikawa et al.*, 2002]. Vertical gradients of

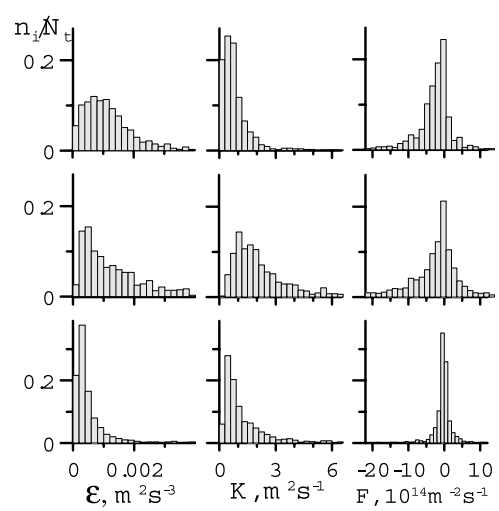

Figure 3. Histograms of observed (left) turbulent energy dissipation rate, (middle) turbulent diffusivities and (right) vertical ozone fluxes for altitude regions of (top) 16–24 km, (middle) 8–16 km and (bottom) 1–8 km.

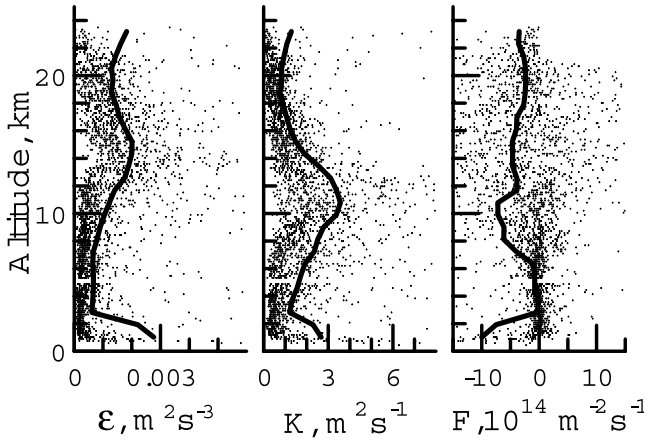

Figure 4. Vertical distributions of (left) turbulent energy dissipation rate, (middle) turbulent diffusivity and (right) vertical ozone flux derived for all ozonesonde launches on April 16–24, 1998 (dots) and their average values (solid lines).

the ozone number density, dn_3/dz, may produce local vertical turbulent ozone fluxes

$$F = -Kdn_3/dz, \quad dn_3/dz = d(nc_3)/dz \tag{3}$$

where n is the number density of the atmosphere. Distributions of dn_3/dz obtained from ozonesonde data during the experiment are shown in Figure 2 (middle). The gradients have very complicated altitude-time structure leading to a highly variable distribution of local turbulent ozone fluxes estimated from (3) and shown in Figure 2 (bottom). During the meso-meteorological events on April 18–20 increased magnitudes of local gradients and ozone fluxes were observed at lower altitudes up to 5–10 km in Figure 2.

[10] Histograms of computed values of ε, K and F for different altitude layers are shown in Figure 3. One can see maxima of occurrence at small values of all parameters. At higher altitudes (Figure 3, middle and top) the histograms are wider and have longer tails at large magnitudes of ε, K and negative F.

[11] Figure 4 shows all data of turbulent parameters and turbulent ozone fluxes obtained during April 16–24, 1998. Solid lines show 4-km running averages. Most individual dots in Figure 4 have smaller values than the average ones. Rare cases of strong turbulence may make substantial impact on atmospheric dynamics and composition and the average curves in Figure 4 may indicate this impact. Average ε maximizes $\sim 2 \times 10^{-3}$ m^{2}s^{-3} at 14–16 km altitude. Average K maximizes at $\sim$3–4 m^{2}s^{-1} at 10–12 km altitude. Average F maximizes at $\sim -(6-8) \times 10^{14}$ m^{-2}s^{-1} at 8–11 km altitude.

[12] It is assumed that the main global-scale mechanism of tropo-stratospheric exchange is the general atmospheric circulation ascending in the equatorial zone and descending at middle and high latitudes [*Holton et al.*, 1995]. Estimations by *Ebel et al.* [1993] give the mean downward ozone flux produced by the general atmospheric circulation of about $F \sim -7 \times 10^{14}$ m^{-2}s^{-1}. Also the analysis of tropospheric ozone balance requires the ozone fluxes from the stratosphere to the troposphere on the order of $-(4-8) \times 10^{14}$ m^{-2} s^{-1} [e.g., *Crutzen*, 1988]. Comparison of these values with the observed local turbulent ozone fluxes in Figures 3 and 4 shows their comparability over Shigaraki. Therefore, in some regions and at some times turbulence enhanced in the tropopause region may produce noticeable local transport of ozone from the stratosphere to the troposphere.

[13] Using aircraft measurements over England, *Pavelin et al.* [2002] and *Whiteway et al.* [2003] obtained values of $K \sim 1-2$ m^{2}s^{-1} near the tropopause, smaller than K obtained over Shigaraki discussed above. One reason for the difference may be a much larger speed of the upper tropospheric jet stream over Japan (up to 60 ms^{-1} during the experiment), than that observed over Europe. Also the MU radar is located in a mountain region, while the aircraft measurements of *Pavelin et al.* [2002] and *Whiteway et al.* [2003] were made over plane surface. Differences in orography and dynamical activity of the atmosphere should lead to local differences of IGW intensity, turbulence and ozone fluxes over different regions.

[14] Another reason for discrepancies between aircraft and MST radar measurements might be a possible overestimation of radar values for σ because of Doppler spectrum broadening caused by the mean wind and its vertical gradient, which is not fully corrected by analytical formulae [i.e., *Hocking*, 1985; *Nastrom*, 1997]. Recently, *VanZandt et al.* [2002] estimated σ^2 for MST radars using a dual-beamwidth method, which for a wind speed >30 m/s gives σ^2 about 25% smaller than that obtained with the one-beamwidth method used by *Fukao et al.* [1994] and *Kurosaki et al.* [1996].

[15] One more problem in MST radar turbulence measurements is the estimation of effective turbulent diffusivity from measured σ. *Hocking* [1999] stressed the role of spatial and temporal intermittency of turbulence. *Dewan* [1981] and *Woodman and Rastogi* [1984] suggested that the random turbulent layers may produce a random process of intermittent diffusion and an effective turbulent diffusivity of this ensemble should be introduced. *Fritts and Dunkerton* [1985] and *Gavrilov and Yudin* [1992] showed that turbulence intermittency produced by breaking gravity waves may lead to a difference in diffusivity of momentum and heat, increasing the effective Prandtl number from 1 up to 3.

[16] High resolution balloon temperature measurements over Shigaraki [*Luce et al.*, 2002] showed frequent appearance of thin turbulent layers in the tropo-stratosphere. Hence turbulent transport of a particle may occur only within such a turbulent layer until the layer dies out. Then the particle remains nearly stationary because of negligible molecular diffusion. When another turbulent layer appears around the particle, further transport over the depth of the layer is possible. Therefore, climatologically, the transport of atmospheric species would depend on an effective diffusivity averaged over substantial time periods rather than on local turbulent diffusivities measured by a particular experiment. Taking account of this and other aspects of the problem, *Hocking* [1999] supposed that the relation between turbulent diffusivity and energy dissipation rate measured by MST radars may be not as simple as it is usually assumed.

[17] Figure 4 shows that turbulence is very intermittent and K is very variable in the tropo-stratosphere. This may produce substantial variability of local turbulent fluxes of ozone and other atmospheric species (see Figures 2–4). Substantial difference in turbulent activity in the tropo-stratosphere over different regions may exist. Further observations would be necessary to clarify this effect on the global scale.

4. Conclusion

[18] An analysis of simultaneous MU radar and ozonesonde measurements in Shigaraki, Japan in April 1998 is presented in this paper. It shows the existence of a quite thick transition tropopause region rather than a definite surface. Instabilities of the jet stream and atmospheric waves may produce increased turbulence in a broad region near the tropopause. This may enhance local downward turbulent ozone fluxes up to the same order of magnitude as ozone transport by the Brewer-Dobson circulation. Vertical and temporal structures of turbulence parameters and ozone fluxes are very complicated (see Figures 1–4). Mesometeorological intrusion of stratospheric ozone may increase local mixing ratio gradients and turbulent ozone fluxes. Further studies of ozone diffusion through intermittent turbulence in different regions are required.

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