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Vertical eddy diffusivity in the lower and middle atmosphere: a climatology based on the MU radar observations during 1986–1992

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Abstract—Wind velocity variance due to turbulence in the inertial subrange (spatial scale: 10^0–10^2 m) has been estimated from a seven-year (1986–1992) database of the echo power spectral width observed by a VHF (46.5 MHz) Doppler radar (the MU radar) in Japan. We study a climatology of the vertical eddy diffusivity K in the tropo-stratosphere (altitude: 5–20 km) and mesosphere (60–90 km) which has been calculated from the observed wind velocity variance. We confirm completely the seasonal variabilities of K suggested by Fukao *et al.* (1994) based on a three-year database: an annual variation with winter maximum in the tropo-stratosphere, and a semi-annual variation with solstice maxima (plus a weak annual variation with summer maximum) in the mesosphere. In this paper we further point out some interannual variabilities of K with a dominant period of about two years, and discuss an overestimation of the median value of K near the tropopausal jet stream.

1. INTRODUCTION

The vertical eddy diffusivity K is used in many chemical and dynamical models for the middle atmosphere, but its value has been left quite ambiguous (see for example, Hocking, 1991). Recently, MST radars have provided a powerful measurement technique for the determination of K over quite a broad altitude range, with far higher vertical and temporal resolutions than previously afforded with other techniques (Hocking, 1983, 1985, 1988). On the basis of observations with the MU (Middle and Upper atmosphere) radar in Japan (35°N, 136°E) during 1986–88, we have analysed seasonal variabilities of K in the upper troposphere, lower stratosphere and mesosphere (Fukao *et al.*, 1994).

The objective of the present study is to analyse seasonal and interannual variabilities of K based on the MU radar observation database accumulated for seven years from 1986 to 1992. A description on our observational technique is given in Section 2. Results for annual medians are described in Section 3. The seasonal and interannual variabilities are shown in Sections 4 and 5, respectively. Technical problems in analysing medians are discussed in Section 6.

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2. MU RADAR OBSERVATIONS

The MU radar is an MST radar, of which the operational frequency is 46.5 MHz. This radar is located at Shigaraki, Japan (35°N, 136°E) and detects radial Doppler velocity spectra associated with atmospheric turbulence with spatial scales from half the radar wavelength (3 m) to the radar reflection volume thickness (150 m for the tropo-stratosphere and 600 m for the mesosphere). This spatial scale range can be regarded approximately as the Kolmogorov's inertial subrange of three-dimensional turbulence (Hocking, 1985). Diffusion due to the inertial-subrange turbulence is (at least locally) isotropic, but is considered to be effective mainly for the vertical transport processes, because the horizontal transport processes may be governed by the effects of much larger-scale atmospheric dynamics. Thus, K due to such turbulence can be computed from the echo power spectral (half-power half) width observed by the MU radar.

The method of computation follows Hocking (1983, 1985, 1988) and Fukao *et al.* (1994)

$$K \approx 0.1\sigma^2/N, \qquad (1)$$

where N is the Brunt–Väisälä frequency, and σ is the half-power half width of the Doppler velocity spectrum (due to turbulence alone) obtained from (originally) a total of 100–200 samples of the spectral width $\sigma_{\rm OBS}$ observed by the MU radar during about 100 hours, after removing the contamination $\sigma_{\rm BRO}$ (due to beam broadening, vertical shear and transience) from $\sigma_{\rm OBS}$ through

$$\sigma^2 = \sigma^2_{\rm OBS} - \sigma^2_{\rm BRO} \quad (2)$$

(Hocking, 1983). However, if $\sigma_{\rm OBS}$ satisfied a condition that

$$\sigma_{\rm OBS} < \sigma_{\rm BRO}, \quad \text{that is} \quad \sigma^2 < 0 \quad \text{and} \quad K < 0, \quad (3)$$

then we omitted it from the original samples. It must be noted that, in the upper troposphere and lower stratosphere (altitude range 5–24 km), $\sigma_{\rm BRO}$ is almost proportional to the background horizontal velocity $V_{\rm H}$ (Fukao *et al.*, 1994), and that the accuracy (estimation error) $\Delta V_{\rm H}$, included in the spectral parameters $V_{\rm H}$ and σ, becomes worse with an increase of $V_{\rm H}$ (Yamamoto *et al.*, 1988). Based on the this fact, a sample of σ satisfying another condition

$$\sigma < \Delta V_{\rm H}, \quad \text{that is} \quad \sigma^2 < (\Delta V_{\rm H})^2 \quad (4)$$

was also excluded from the samples of median-value calculations. However these procedures may reject too many data in the case of strong background winds. This will be discussed again in Section 6.

The horizontal dimension of sampling volume is given by the nominal beam width (3.6°). It should be noted that K is dependent, not only on the turbulence intensity, but also on how fully the turbulence is developed in this volume, or how completely this volume is filled by turbulence. The latter is theoretically determined by a typical value of the flux Richardson number $R_{\rm f}$ in the sampled volume, and the factor 0.1 in (1) is derived in the case of $R_{\rm f} \approx 1/4$, which is close to its minimum value. If $R_{\rm f}$ is larger (in the case of stronger stability), the factor may be increased up to 0.2, but K can only double.

Observations of σ have been carried out (every 30 min for the tropo-stratosphere and 1 hour for the mesosphere) for about 100 h each month from January 1986 to December 1992 (see Table 1). We estimated K from an average of the values of σ for four oblique (northward, eastward, southward and westward) beams with a zenith angle of 10°. The small-scale variability of K is smoothed by taking a median of results each month, since altitudes of turbulence layers and patches distribute evenly all over the whole observed altitude range during the observational period of $\sim$100 h. Details of the MU radar observation and data calculation procedures have been described by Fukao *et al.* (1991, 1994).

3. VERTICAL DISTRIBUTION OF ANNUAL MEDIAN

Figure 1 shows the vertical distributions of the annual medians of K in the upper troposphere, lower stratosphere and mesosphere from 1986 to 1992. The annual median is calculated from a median value of 12 monthly median values. We confirm that the general characteristics of the vertical distributions of K during 1989–92 are quite similar to the earlier results shown in Fukao *et al.* (1994) based on the data during 1986–88.

The magnitude of K in the troposphere is smaller than the values required by chemical model studies (e.g. Massie and Hunten, 1981). It is considered by modellers that advection and diffusion due to synoptic and/or planetary-scale waves may be more important causes of mixing in the troposphere than small-scale eddy diffusion (see for example, Matsuno, 1980). Such a large-scale quasi-horizontal mixing is considered to take place along the isentropic surfaces, so that its effect becomes quite large in the mid-latitude troposphere in which the inclination of the isentropic sur-

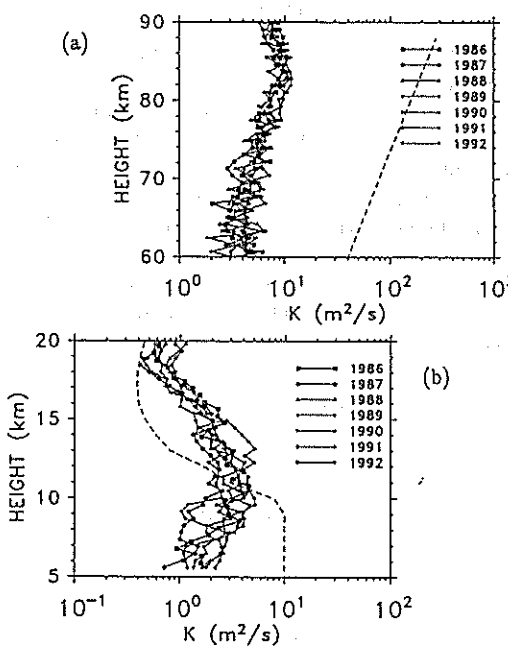

Fig. 1. Vertical profiles of the annual medians of vertical eddy diffusivity K observed by the MU radar during 1986–1992 in (a) the mesosphere and (b) the upper troposphere and lower stratosphere, compared with standard model profiles indicated by dashed curves (Ogawa and Shimazaki, 1975; Massie and Hunten, 1981).

Table 1. List of the GRATMAC standard observations

Case	Year/Date/LST	Case	Year/Date/LST
JAN86	1986/ Jan 6/1109 – Jan 10/1303	JAN90	1990/ Jan 16/1122 – Jan 19/1633
FEB86	1986/ Feb 10/1304 – Feb 14/1547	FEB90	1990/ Feb 5/1135 – Feb 9/1700
MAR86	1986/ Mar 17/0800 – Mar 21/1542	MAR90	1990/ Mar 12/1758 – Mar 16/1613
APR86	1986/ Apr 7/1134 – Apr 11/1528	APR90	1990/ Apr 9/0804 – Apr 13/1636
MAY86	1986/ May 6/1123 – May 9/1535	MAY90	1990/ May 14/1228 – May 18/1631
JUN86	1986/ Jun 2/1200 – Jun 6/1426	JUN90	1990/ Jun 11/1154 – Jun 15/1601
JUL86	1986/ Jul 7/1156 – Jul 11/1329	JUL90	1990/ Jul 2/1133 – Jul 6/1754
AUG86	1986/ Aug 18/1017 – Aug 22/1416		
SEP86	1986/ Sep 1/1201 – Sep 5/1639	SEP90	1990/ Sep 10/0839 – Sep 14/1729
OCT86	1986/ Oct 13/1120 – Oct 17/1535	OCT90	1990/ Oct 15/1159 – Oct 19/1551
NOV86	1986/ Nov 10/2357 – Nov 14/1536	NOV90	1990/ Nov 5/1125 – Nov 9/1628
		DEC90	1990/ Dec 10/1202 – Dec 14/1631
JAN87	1987/ Jan 5/1316 – Jan 9/1530	JAN91	1991/ Jan 21/1036 – Jan 25/1635
FEB87	1987/ Feb 3/0053 – Feb 6/1529	FEB91	1991/ Feb 4/1146 – Feb 8/1706
MAR87	1987/ Mar 2/1201 – Mar 6/1536	MAR91	1991/ Mar 11/1150 Mar 15/1435
APR87	1987/ Apr 6/1104 – Apr 10/1536	APR91	1991/ Apr 23/0830 – Apr 26/1711
MAY87	1987/ May 11/1344 – May 15/1536	MAY91	1991/ May 13/1148 – May 17/1634
JUN87	1987/ Jun 22/1200 – Jun 26/1200	JUN91	1991/ Jun 17/1131 – Jun 21/1900
JUL87	1987/ Jul 6/1723 – Jul 11/0544	JUL91	1991/ Jul 3/0502 – Jul 7/1902
AUG87	1987/ Aug 3/1217 – Aug 7/1220	AUG91	1991/ Aug 5/1152 – Aug 9/1632
SEP87	1987/ Sep 7/1151 – Sep 11/1634	SEP91	1991/ Sep 17/1106 – Sep 20/1433
OCT87	1987/ Oct 5/1207 – Oct 9/1454	OCT91	1991/ Oct 21/1143 – Oct 25/1636
NOV87	1987/ Nov 26/1135 – Nov 30/1505	NOV91	1991/ Nov 11/1058 – Nov 15/1606
DEC87	1987/ Dec 7/1037 – Dec 11/1517	DEC91	1991/ Dec 16/0619 – Dec 20/1652
JAN88	1988/ Jan 5/0818 – Jan 8/1434	JAN92	1992/ Jan 20/1026 – Jan 24/1700
FEB88	1988/ Feb 15/1214 – Feb 19/1517	FEB92	1992/ Feb 17/1147 – Feb 21/1639
MAR88	1988/ Mar 7/0823 – Mar 11/1537	MAR92	1992/ Mar 16/1145 – Mar 20/1602
APR88	1988/ Apr 18/0915 – Apr 22/1530	APR92	1992/ Apr 6/1123 – Apr 10/1800
MAY88	1988/ May 9/0847 – May 13/1521	MAY92	1992/ May 18/1332 – May 22/1624
JUN88	1988/ Jun 6/1156 – Jun 10/1641	JUN92	1992/ Jun 15/1802 – Jun 19/1704
JUL88	1988/ Jul 18/1559 – Jul 22/1637	JUL92	1992/ Jul 6/1155 – Jul 10/1701
AUG88	1988/ Aug 8/1137 – Aug 12/1628	AUG92	1992/ Aug 24/1101 – Aug 28/1607
SEP88	1988/ Sep 29/1237 – Sep 2/1532	SEP92	1992/ Sep 7/1209 – Sep 11/1641
OCT88	1988/ Oct 24/1146 – Oct 28/1527	OCT92	1992/ Oct 12/1209 – Oct 16/1700
NOV88	1988/ Nov 14/1501 – Nov 18/1527	NOV92	1992/ Nov 9/1553 – Nov 13/1700
DEC88	1988/ Dec 19/0814 – Dec 23/1600	DEC92	1992/ Dec 14/1155 – Dec 18/1700
FEB89	1989/ Feb 6/1048 – Feb 11/1030		
APR89	1989/ Apr 3/0822 – Apr 7/1648		
MAY89	1989/ May 15/1156 – May 19/1700		
JUN89	1989/ Jun 26/1153 – Jun 30/1900		
JUL89	1989/ Jul 17/1041 – Jul 21/1459		
AUG89	1989/ Aug 21/1044 – Aug 25/1714		
SEP89	1989/ Sep 11/1121 – Sep 14/1602		
OCT89	1989/ Oct 16/1153 – Oct 20/1556		
NOV89	1989/ Nov 6/1141 – Nov 10/1701		
DEC89	1989/ Dec 18/1046 – Dec 22/1156		

face is in general, large. It must be noted that the values near the tropopause jet stream (altitude: 10–16 km), in particular in winter, might be somewhat overestimated, because many samples with smaller values are excluded by excessive beam broadening in the data analysis procedure used here (see Section 6). For this reason, in this section we do not discuss K in this limited altitude range.

The magnitude of K in the lower stratosphere is quite close to that required by modelling studies. This implies that the vertical transport process in the lower stratosphere is mainly due to turbulence with scales

10^0–10^2 m observed here by the MU radar. Such turbulence appears frequently as thin sporadic turbulence layers, and is quite different from the patchy turbulence observed in the troposphere. With regard to this turbulence behaviour, the suitability of the K parameterization might be controversial (cf. Dewan, 1981; Weinstock, 1990; Hocking, 1991). The results described here are based on statistic analysis of sufficient numbers of observed data, and we regard them as practically useful results.

K becomes larger in the mesosphere (reaching $\sim 10^1$ m^2 s^{-1}), increasing gradually with height, but the annual median values are smaller than 1/10 of the values required by chemical model studies (e.g. Ogawa and Shimazaki, 1975). We found in Section 2 that ambiguity in the evaluation of the factor 0.1 in (1) makes K only double at the maximum, which cannot explain the discrepancy. The maximum value observed at each altitude each month is close to the chemical model value (see Fukao *et al.*, 1994), and such a rare value of a large K, possibly induced by a strong gravity wave, might be essential to describe the vertical transport processes. A large gradient of the physical quantities induced just before the breakdown of the strong wave also contributes to making the diffusion transport more effective. It must be noted that recent studies on short-lifetime constituent transport models (e.g. Strobel *et al.*, 1987) and *in situ* measurements (e.g. Lübken *et al.*, 1987, 1993) suggest small values of K as estimated in this study. Since very large eddies and secondary circulations may be generated by breaking gravity waves with a very long (vertical) wavelength ($\sim 10^3$–10^4 m) in the mesosphere, their advection and/or large-scale diffusion might explain the discrepancy, as discussed by Fukao *et al.* (1994). In this meaning the K parameterization itself might again be controversial (cf. Hocking, 1991; Lübken *et al.*, 1993).

Above an altitude ($\sim$84 km) near the mesopause, K seems to decrease upward, which may suggest effects of strong molecular viscosity: suppression of turbulence generation and/or dissipation of eddies with a half radar-wavelength scale. It must be noted that the MU radar observations (using the meteor trail echoes) also provide variabilities of the molecular diffusivity (Tsutsumi *et al.*, 1994). A comparison between the variabilities of both eddy and molecular diffusivities in the upper mesosphere is very interesting in order to discuss the control of the homopause level, but this is beyond the scope of this study.

4. SEASONAL VARIABILITY

Figure 2 shows the variabilities of the monthly medians of K in the mesosphere and the tropo-stratosphere. In general, the results are quite similar to those described in Fukao *et al.* (1994) based on observations during 1986–88. The monthly median of K shows a maximum reaching 10^0 m^2 s^{-1} near the tropopause jet stream altitude, and has a minimum less than 10^0 m^2 s^{-1} in the lower stratosphere. K has a striking annual variability with a maximum in winter and an amplitude reaching about an order of magnitude. Although the median values near the tropopause jet stream in winter are less reliable, as mentioned earlier (and also to be mentioned in Section 6), the winter maximum seems to be reasonable, since in this season the mean vertical shear is strongest and the mean dynamic stability takes the minimum. In recent climatological analysis of wind fluctuations, based on the same database by the MU radar observations, gravity wave activities have similar seasonal variabilities (Sato, 1994; Murayama *et al.*, 1994).

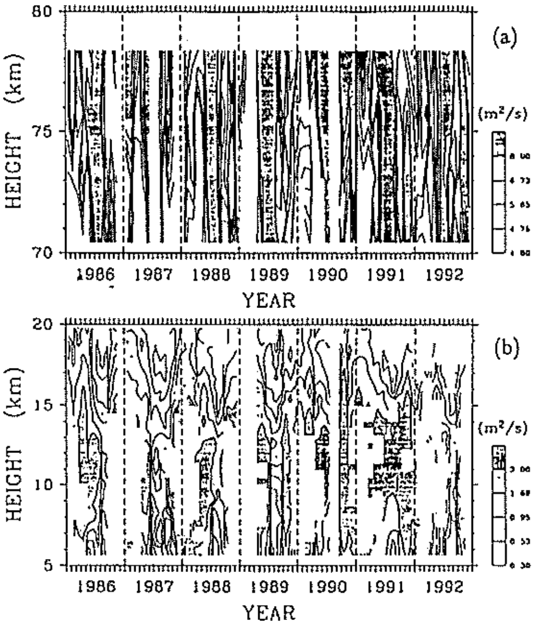

Fig. 2. Vertical–temporal plots of the monthly medians of vertical eddy diffusivity K observed by the MU radar during 1986–1992 in (a) the mesosphere and (b) the upper troposphere and lower stratosphere.

A semi-annual variation, reaching approximately an order of magnitude, is dominantly detected in the mesosphere. The annual maximum of K exists in summer owing to a weaker annual variation superposed on the semi-annual variation. This seasonal variability is consistent with a similar one observed for gravity wave activity (Tsuda *et al.*, 1990), and another calculated in a chemical–dynamical coupling model including the gravity-wave breaking effect (Garcia and Solomon, 1985).

It is quite interesting that the seasonal variabilities of K are coincident with those of gravity wave activities in the whole altitude range analysed here. This may give a concrete proof to the hypothesis that turbulence is mainly generated by gravity wave breaking. It is strongly expected that some model studies should explain quantitatively the relationship among the mean field, gravity waves and turbulence, although this is also beyond the scope of this paper (cf. Yamanaka, 1992; Yamanaka and Fukao, 1994).

5. INTERANNUAL VARIABILITY

In this study, we analyse the interannual variability of K, using the seven-year database, which is clearly improved from our earlier study (Fukao *et al.*, 1994). Figure 3 shows the variabilities of 12-month running-mean data of the monthly medians of K (that is, the interannual variabilities) in the mesosphere and the tropo-stratosphere. The running mean used here was calculated for $\log K$, because the distributions of $\log K$ are approximately Gaussian-shaped but those of K are not so. We find that the interannual variability is, in general, quite small. This result is consistent with the fact found in Section 3 (Fig. 1) that the annual median profiles are quite similar for all seven years.

However, we have found, (although the detailed examinations are beyond the scope of this paper) a 2–3 years variation in the lower stratosphere (Figs 3(b) and 4). Variations of a 2–3 year period, with downward phase progression, are also seen in zonal wind data (Fig. 5), which may be affected more or less by the quasi-biennial oscillation (QBO) in the equatorial lower stratosphere. Variations of K and of the zonal wind seem to be in phase, and we consider that the latter may induce the former through variations of vertical shear or of the stratification stability. Furthermore, variations with a longer time scale (~4 years) are also suggested, which might be related to the El Niño-southern oscillation (ENSO) in the tropical troposphere. Anyhow, such interannual variations of K must directly affect the transport processes and hence the global distributions of atmospheric constituents, which are quite interesting in view of the global environment issue.

The interannual variability in the mesosphere (Fig. 3(a)) is larger than in the lower stratosphere, as speculated by Hocking (1988) based on a two-year (1985–

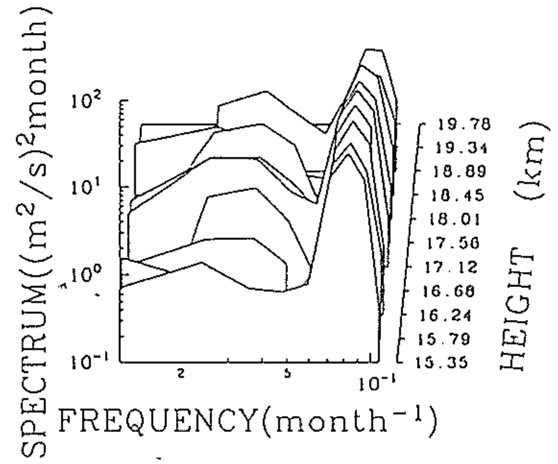

Fig. 4. Power spectra of vertical eddy diffusivity K observed by the MU radar during 1986–1992 in the lower stratosphere.

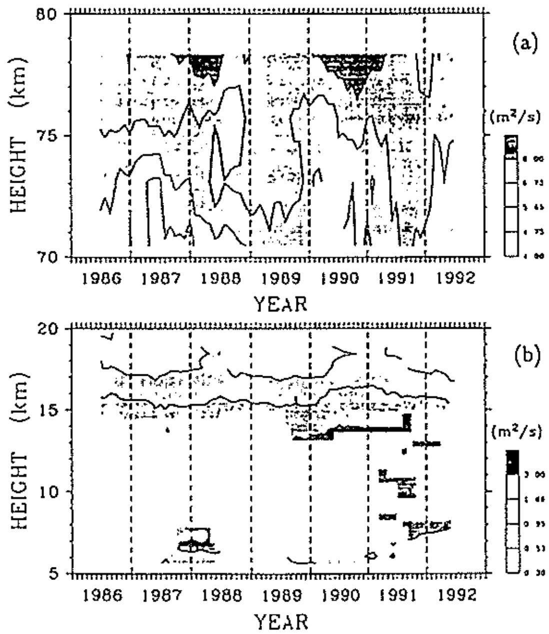

Fig. 3. Same as Fig. 2 but for 12-month running-averaged data.

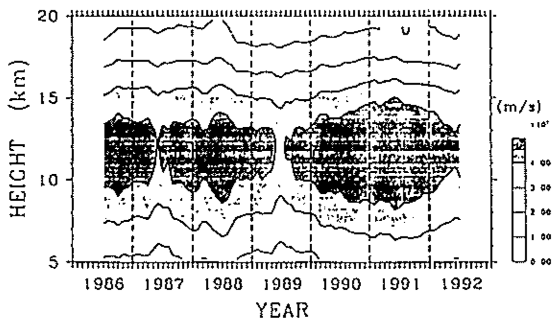

Fig. 5. Vertical–temporal plots of the 12-month running-averaged monthly medians of zonal wind velocity observed by the MU radar during 1986–1992 in the upper troposphere and lower stratosphere.

86) database for the mesopause region obtained with the Adelaide MF radar, Australia. Variations with a time scale of 2–3 years are found, and they seem to have an anti-phase structure between the upper mesosphere (altitude > 75 km) and the lower mesosphere (<70 km). There seems likely to be a positive correlation between the upper mesosphere and lower stratosphere, but detailed studies on such a stratosphere–mesosphere relationship (possibly through upward propagating gravity waves) is beyond the scope of this paper.

6. DISCUSSION ON ESTIMATION OF THE MEDIAN

Throughout the earlier sections we have used the monthly median values of K, each of which was calculated by (1) using a median value of σ after omitting data satisfying conditions (3) and (4). Thus, in the earlier estimations, the median value of σ was calculated only from samples with relatively larger values satisfying $\sigma^2 \geqslant (\Delta V_H)^2$. The number of samples excluded in the procedures mentioned above becomes quite large in the case of $V_H > 40$ m s^{-1}, which often appears near the tropopause jet stream, in particular in winter. Thus, the median values of K as so far shown, might be overestimated in such a case. (There are no problems in the mesosphere, where $V_H > 40$ m s^{-1} sometimes appears, but $\sigma \sim \sigma_{OBS} \gg \sigma_{BRO}$ holds in general, because turbulence is stronger than in the tropostratosphere.)

In order to recover this overestimation, we have re-calculated medians of K without excluding the samples satisfying (3) or (4). As shown in Fig. 6, the annual medians of K become $\sim 1/2$ in the lower stratosphere of the earlier estimations (shown in Fig. 1(b)), and $\sim 1/10$ in the upper troposphere. The re-calculated results become closer to the chemical model values in the altitude range higher than ~ 13 km. Nevertheless, the profiles have maxima at an altitude 12–13 km and minima at an altitude 17–18 km, which are similar to those in the earlier estimations. Although the annual variability with the maximum in winter and the minimum in summer becomes weaker (compare Fig. 7(a) with Fig. 2(b)), the quasi-biennial variability still appears (compare Fig. 7(b) with Fig. 3(b)).

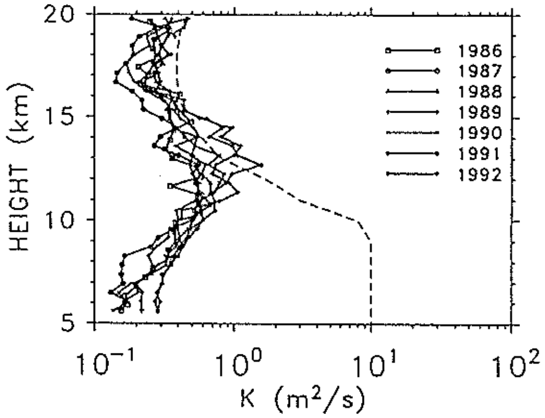

Fig. 6. Same as Fig. 1(b) but including samples of σ^2 with values that are negative or smaller than the accuracy.

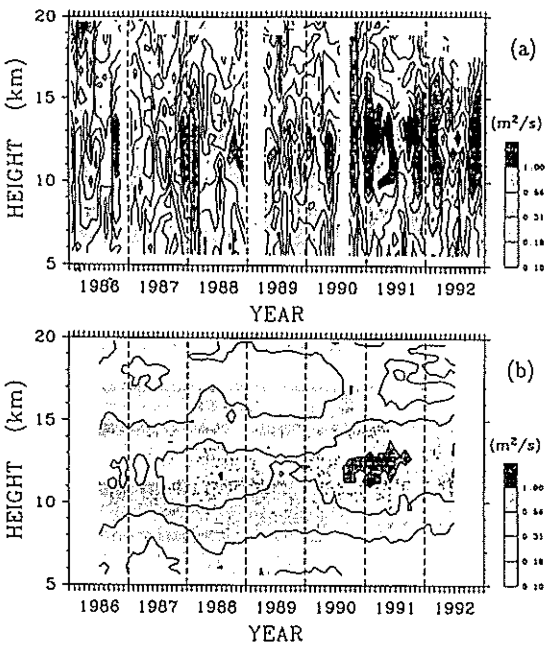

Fig. 7. Same as (a) Fig. 2(b); and (b) Fig. 3(b); but including samples of σ^2 with values that are negative or smaller than the accuracy.

We consider that the re-calculated results shown in this section are more appropriate to discuss the most common features of turbulence. However, from an efficiency on the transport process point of view of, the rare feature of a strong turbulence inducing K larger than the median value re-calculated here, might be much more important, as mentioned in Section 3. We need detailed analysis on the occurrence frequency of such larger values of K in order to obtain an effective value of K for modelling studies.

7. CONCLUSIONS

We conclude that the middle-atmospheric eddy transport due to the inertial-subrange turbulence has seasonal and interannual variabilities. The annual variability of K is predominant in the lower stratosphere, and the variability in the mesosphere includes semi-annual and relatively weaker annual components. The interannual variability of the lower stratosphere includes a quasi-biennial mode reaching 0.3–0.8 $m^2 s^{-1}$, and a similar variability of the mesosphere reaches 1–5 $m^2 s^{-1}$. These variabilities must directly affect the vertical transport of energy, momentum and constituents in the middle atmosphere.

We are now quantitatively examining a hypothesis that the middle-atmospheric vertical eddy diffusivity is mainly governed by breaking gravity waves. Based on this examination, we will explain the seasonal and interannual variabilities of K observed here, and also calculate those of energy, momentum and constituent distributions in the lower and middle atmosphere.

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