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Mesosphere and Lower Thermosphere Dynamics Revealed by Radar Observations

by
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November 1991

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Abstract

In the past a few decades our knowledge on the middle atmosphere (10–100 km altitude) has extremely increased owing to various observational and theoretical studies. Dynamics of the middle atmosphere is now understood as it is governed by a complicated mechanism of the interactions between the atmospheric waves and mean winds, or wave-wave interactions. However, the upper part of the middle atmosphere, i.e. the mesosphere and lower thermosphere are still the regions where the whole dynamics are not well clarified, since the observations of these regions are insufficient because of the difficulties of the technique. Ground based radar measurement is a useful technique for the observational study of the atmospheric dynamics of the mesosphere and lower thermosphere. In the present thesis, we studied the dynamics of the mesosphere and lower thermosphere such as mean wind, gravity waves, atmospheric tides and planetary waves mainly using the observations of Kyoto meteor radar and the MU radar, at Shigaraki (35°N, 136°E), and comparing their observed results with those by two MF radars at Saskatoon (52°N, 107°W) and Adelaide (35°S, 138°E) in order to study the hemispheric and latitudinal differences of these waves.

After reviewing the observational techniques of the MST, meteor and MF radars and the experimental set up of the radars utilized in this thesis, mean winds at 60–110 km are presented in comparison with the updated CIRA 1986 model atmosphere as well as Adelaide MF radar observations. The hemispheric symmetry of the mean winds was basically found between the hemispheres, while some hemispheric difference was observed such that the eastward wind in late winter was weaker in the northern hemisphere and that the reversal height of the mesospheric westward wind into the eastward direction in July/January had large and small interannual variability in the northern and southern hemispheres, respectively. Correlation between the stratospheric sudden warming and the change of the mean wind direction from eastward to westward was observed at Shigaraki (35°N).

The atmospheric tides are also compared among the results observed with Kyoto meteor radar and two MF radars. The semidiurnal tide showed bimodal feature with short and long vertical wavelength at all three locations, but long vertical wavelength emerged in summer in each hemisphere. The closer comparison of the transition between summer-like and winter-like months indicates that the enhancement of another mode at equinoxes besides the summer and winter modes can explain the transition well. The diurnal tide are closely compared between Shigaraki and Adelaide, and significant asymmetries between hemispheres were observed. The amplitude of diurnal tide observed at Adelaide was much

larger than that at Shigaraki. This large asymmetry can be caused by the hemispheric difference of wave source, propagations through background wind fields or dissipation, among which the difference of dissipation has been the potential candidate to explain the tidal asymmetry to explain the observed characteristics of diurnal tides.

The two-day waves were observed simultaneously between Shigaraki and Adelaide, showing anti-symmetric phase structure in the southern summer, while amplitudes were much stronger in the southern hemisphere suggesting the distortion of the global normal mode. The annual mean amplitude was larger in the southern hemisphere by a factor of 2, again suggesting that two-day waves in the northern hemisphere are more strongly damped than in the southern hemisphere. Thus, the latter half of the thesis is devoted to the gravity wave study, because the dissipations are caused by the eddy diffusion due to turbulence, which are mainly produced by the gravity wave breaking.

Characteristics of gravity waves were observed with the MU radar. Frequency spectral analysis of the wind velocity fluctuations shows that the major part of the zonal momentum flux were found in the wave periods from 30 min to 2–8 hours. Characteristics of dominant gravity waves in the vertical profile of the wind velocity were also studied, finding that their preferred propagation was eastward except for the westward propagation in early winter. The momentum flux carried by the dominant gravity waves, observed in the vertical profiles with wavelengths of 5–15 km and intrinsic periods of 5–15 hours, were smaller than that carried by the waves with periods of 8 hours – 30 min component throughout a year except for in autumn when the momentum flux carried by the dominant gravity waves became more significant than that by the other component.

The gravity waves observed with the MU radar were compared with the result by the Saskatoon MF radar for a full year, and with that by the Adelaide MF radar for summer and winter campaigns. The variance of the gravity waves (120–10min) were larger at Shigaraki than at Saskatoon, while almost same between Shigaraki and Adelaide. This indicates the hemispheric difference of eddy diffusivity, which is mainly caused by the turbulence induced by gravity wave breaking, is small. Therefore, the reason of the large hemispheric asymmetry of tide is not explained by the difference of dissipations.

The propagation direction of the gravity waves are compared between the observations at Shigaraki and Adelaide, finding that at Shigaraki the propagation direction tends to be biased into eastward-westward directions, while at Adelaide the polarization into the northward-southward directions were often observed. This finding suggests the possibility that the gravity wave activity in the mesosphere is affected by the topography around the observation site, and hence the longitudinal variation of gravity wave activity might be large.

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Chapter 1

General Introduction

1.1 The Middle Atmosphere

The middle atmosphere has recently become an interesting and important region not only to the scientists but also to the general public. Issues of earth's environment such as Greenhouse effects, destruction of ozonospheres frequently appear in newspapers and broadcasts. The study of the dynamics of the middle atmosphere, therefore, has become a very important subject because the effects of the dynamic process are firmly concerning with these issues. Especially, the studies based on the observations are widely required to understand the complicated dynamics of the atmosphere. This thesis is devoted to clarifying the dynamics of the upper part of the middle atmosphere, i.e. the mesosphere and lower thermosphere, by the observations using radar technique.

The earth's atmosphere is classified according to the temperature structure. Figure 1.1 shows a typical height profile of temperature at 35°N in the mid-latitude region, which is composed of four regions of the troposphere, stratosphere, mesosphere and thermosphere.

The region from the ground up to approximately 15 km is called the troposphere, where the temperature decreases with height at a lapse rate of about 6.5°C/km. The main heat source for this region is the solar radiation absorbed by the surface of the earth. The atmosphere there becomes unstable by the heating around the ground, and thus convection is generated in the troposphere. The temperature of an air parcel moving upward decreases because of the adiabatic expansion, while water vapor in the air is cooled when it condenses. The temperature lapse rate in the troposphere is thus determined by the balance of these effects. The height of the upper boundary of troposphere, that is the tropopause, clearly shows latitudinal variation with maximum at equatorial latitude, also shows seasonal variation with summer maximum at mid- and high latitude.

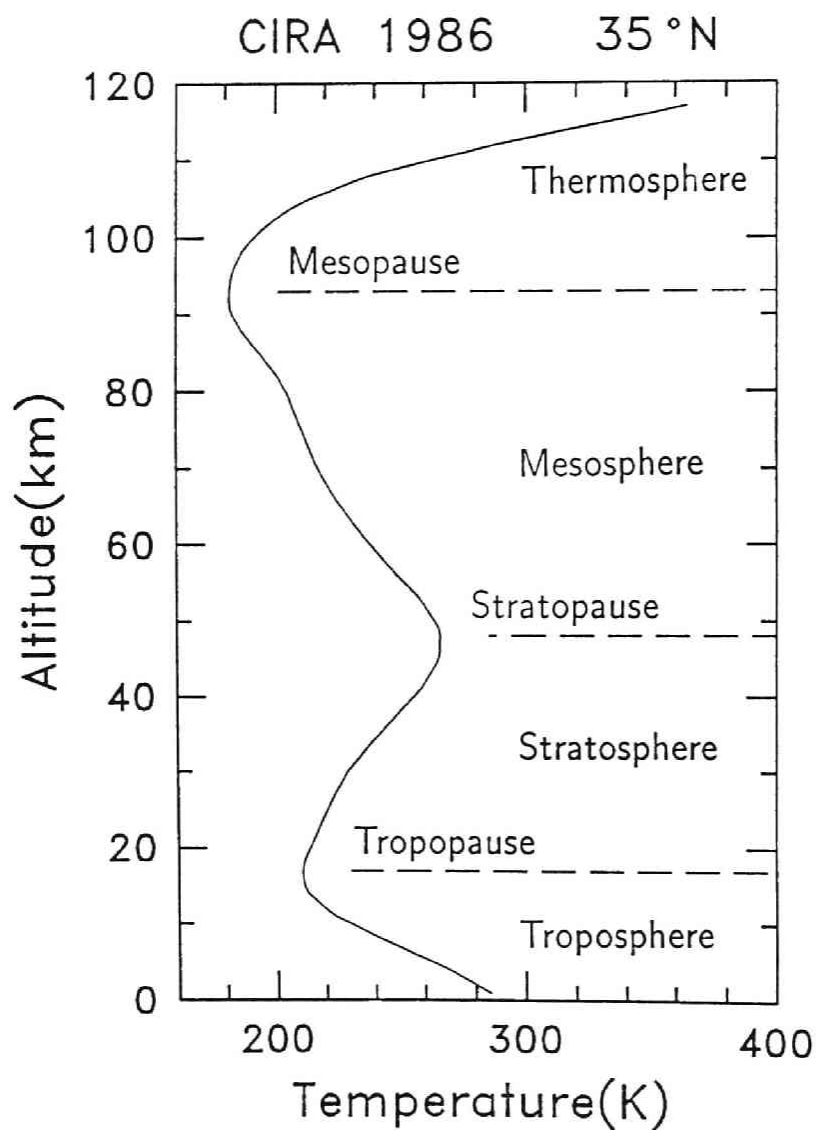


Figure 1.1: Annual mean temperature distribution and nomenclature of the earth's atmosphere at 35°N [CIRA, 1986].

The region above 15 km is called the stratosphere, where the temperature is almost constant with height up to 20 km, and then increases up to approximately 270 K at around 50 km. Atmosphere in this region is stable and convective motion is scarcely generated because of the structure of the temperature profile. The upper boundary of the stratosphere is called the stratopause.

Above the stratopause, the temperature again decreases with height, and it reaches minimum at around 90 km. This region is called the mesosphere and its upper boundary is called the mesopause. Since temperature lapse rate here is not so large as troposphere, atmosphere of this region are not unstable. Existence of strong turbulence is a characteristic feature of the mesosphere, which seems to be generated by the breaking of gravity waves. The heat source of the high temperature around the stratopause is the absorption of solar ultra violet (UV) radiation by ozone layer, which is a unique characteristic among the planets in our solar system.

The region above the mesopause is called the thermosphere, where the temperature rapidly and monotonously increases with height because constituents of the atmosphere are ionized and strongly heated by extreme ultraviolet (EUV) radiation and solar X-rays. The vertical transport due to macroscopic motions of the atmosphere becomes less important in this region than the molecular diffusion because the mean free path of the molecule increases according to the decreasing density.

Besides the classification by the temperature structure, the region from 10 to 100 km is also called the middle atmosphere, which consists of stratosphere, mesosphere and lower thermosphere. Middle atmosphere is a comparatively stable regions where the atmospheric motions are mainly driven by the energy of the solar UV radiation absorbed in the ozone (O_3) layer.

The radiative heating by ozone is a function of solar incident flux, hence the intensity shows large seasonal and latitudinal variation. The difference of the heating between the northern and southern hemispheres causes the general circulation of the middle atmosphere. The intensity of heating attains its maximum and minimum in the summer and winter polar regions, respectively. The primary factor to form the general circulation is the geostrophic balance between the latitudinal pressure gradient $\partial p_0 / \partial y$ and the Coriolis force, in addition to the hydrostatic equilibrium in the vertical direction (z). First, the hydrostatic equilibrium is given as

$$\frac{\partial p_0}{\partial z} = -\rho_0 g \quad (1.1)$$

where p , ρ and g are pressure, density and the gravitational accelerations. Subscript 0 denotes the static state. Note that coordinate system x, y and z in this thesis is aligned to the eastward, northward and upward directions, respectively.

The pressure, p_0 , is expressed by density, ρ_0 , and temperature, T_0 , through the equation of state for a perfect gas,

$$p_0 = \rho_0 R T_0 \quad (1.2)$$

where R is the gas constant per unit mass. Integration of Eq. (1.1) with the aid of Eq. (1.2) yields

$$p_0(z) = p_{00} \exp\left(-\int \frac{dz}{H}\right) \quad (1.3)$$

$$\rho_0(z) = \rho_{00} \frac{T_{00}}{T_0} \exp\left(-\int \frac{dz}{H}\right) \quad (1.4)$$

where p_{00} and ρ_{00} are the pressure and the density at a reference height, respectively, and H is the atmospheric scale height given by

$$H = \frac{RT_0}{g} \quad (1.5)$$

Then, the geostrophic balance is expressed as

$$u_0 = -\frac{1}{f\rho} \frac{\partial p_0}{\partial y} \quad (1.6)$$

$$v_0 = \frac{1}{f\rho} \frac{\partial p_0}{\partial x} \quad (1.7)$$

$$w_0 = 0 \quad (1.8)$$

where $f = 2\Omega \sin \phi$ is the Coriolis frequency determined by the earth's angular velocity Ω and the latitude ϕ . By differentiating Eqs. (1.6)–(1.8) with respect to z , and considering Eqs. (1.1) and (1.2), we can obtain the thermal wind equation [Beer, 1974] as

$$\frac{\partial}{\partial z} \left(\frac{\mathbf{U}_g}{T_0} \right) = -\frac{g}{fT_0} \left(\frac{\partial T_0}{\partial y}, -\frac{\partial T_0}{\partial x}, 0 \right) \quad (1.9)$$

where $\mathbf{U}_g = (u_0, v_0, w_0)$. We now consider the thermal wind driven by the latitudinal difference of the temperature due to the difference of the solar heating. The positive latitudinal temperature gradient in northern summer, for example, gives westward (easterly) wind for the northern hemisphere and eastward wind for southern hemisphere, and both of which increase with height. These wind directions are reversed in northern winter.

Figure 1.2 shows a latitudinal and height distribution the temperature of the CIRA (COSPAR International Reference Atmosphere) 1986 model in January and July, which is constructed by a large amount of observations by means

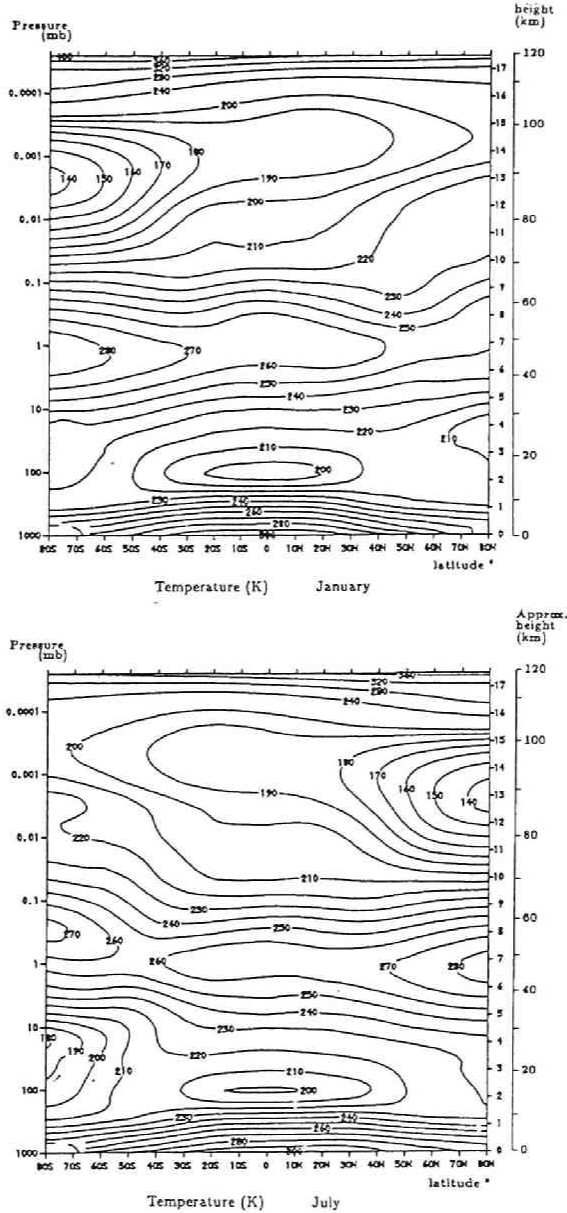


Figure 1.2: Latitude vs. pressure scale height cross sections for temperature in January (top) and July (bottom) [CIRA,1986]. The left ordinate is pressure (mb); the first right ordinate (0–17) is pressure scale height; the second right ordinate is approximate geometric height.

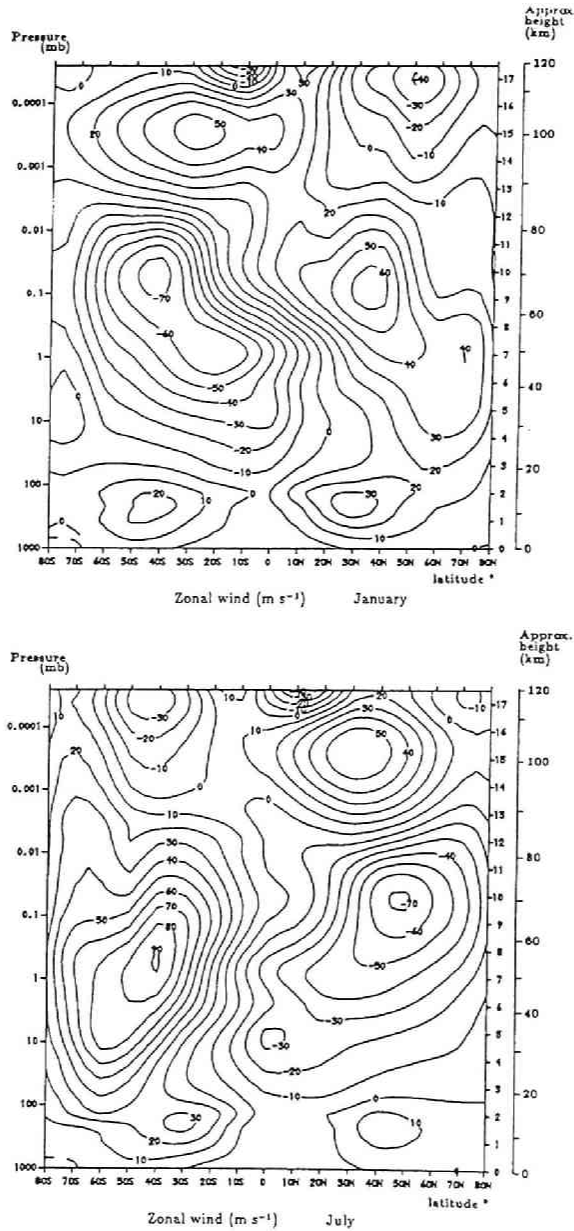


Figure 1.3: Latitude vs. pressure scale height cross sections for zonal mean wind in January (top) and July (bottom) [CIRA,1986]. Positive and negative values correspond to eastward (westerly) and westward (easterly) wind velocities. The ordinates are same as Figure 1.2.

of various methods. Negative and negative latitudinal gradient of temperature around 40–50 km altitudes are shown in northern winter (January) and in northern summer (July), respectively, where the absorption of the UV by ozone is large. Figure 1.3 shows a similar distribution of the zonal mean wind in the middle atmosphere in zonal average [CIRA, 1986]. It is found that eastward and westward winds at mid-latitude increase up to the height of 50–70 km which is explained by the thermal wind shown in Eq. 1.9. Another pronounced feature of the Figure 1.3 is the decrease of zonal wind above 70km and the reversal of the wind direction at around mesopause (80–100km). Furthermore, the temperature distribution in this height range in Figure 1.2, which shows that temperature in summer polar region is lower than that in the winter region, is not explained in terms of the radiative equilibrium. Hence, dynamical processes due to atmospheric waves are considered to brake the mean zonal wind and to heat (cool) the summer (winter) polar region. Now the role of the gravity waves in this braking is recognized to be the most important among the various dynamic processes, which has been studied by theoretical works [Lindzen, 1981, Matsuno, 1982], and observations [e.g., Fritts, 1984; Reid and Vincent, 1987; Meek and Manson, 1989; Tsuda et al., 1990a].

Another feature of the CIRA 1986 model shown in Figures. 1.2 and 1.3 is the asymmetry between northern and southern hemispheres. If the middle atmosphere is homogeneous and symmetrical between hemispheres, the latitudinal cross section in January should coincide with the figures of July if the horizontal axis is reversed. However, Figure 1.2 shows that temperature profiles at high latitude in the stratosphere and mesosphere are largely different between winters in the northern and the southern hemispheres. Also the mean zonal wind in Figure 1.3 shows remarkable difference of the winter eastward wind both in its amplitude and location between northern and southern winters; the maximum eastward wind occurs at lower height and higher latitude, and the peak in intensity is stronger in southern winter than in northern winter. The previous CIRA model [CIRA, 1972] did not describe these hemispheric differences but it showed a symmetrical model, i.e. profiles in the southern hemisphere is exactly same as that in northern hemisphere of six months' delay. The lack of observations was the primary reason of the difficulty in describing a symmetrical model. The updated CIRA, CIRA 1986, is expanded as a global model, which can express the hemispheric difference as shown above. Therefore, observational study to check and improve this model is highly desired. Moreover, to investigate the hemispheric symmetry of the atmospheric waves in the atmosphere, which significantly affects the general circulation, is the most important subject for the observational study.

1.2 Atmospheric Waves

The prominent characteristics of the middle atmosphere is that various atmospheric waves with various temporal and spatial scales are superimposed on the mean winds. These waves are categorized into planetary waves (> 1 day), atmospheric tides (24 hrs, 12 hrs,...) and gravity waves (21 hrs~5 min at 35°N) according to their typical wave periods. Turbulence is also an important process causing wind velocity fluctuations. Many of these atmospheric waves can propagate vertically, and if they propagates upward, their wave amplitudes tend to increase with height as the background density decreases when the wave energy conserves. The waves are saturated when their amplitudes become too large, and give their energy to the background wind fields, a part of which accelerates the mean wind, and other causes turbulence through either convective or dynamic instabilities. Therefore, the study of interactions of atmospheric waves with mean winds and waves become as important as the investigation of the excitation, propagation and dissipation of single waves.

Here, we briefly review the atmospheric waves. Atmospheric motions are described by the following set of equations as

$$\left(\frac{d\mathbf{v}}{dt}\right) + 2\boldsymbol{\Omega}_z \times \mathbf{v} = -\frac{1}{\rho}\nabla p + \mathbf{g} + \frac{1}{\rho}\mathbf{F} \quad (1.10)$$

$$\frac{d\rho}{dt} + \rho\nabla \cdot \mathbf{v} = 0 \quad (1.11)$$

$$\tilde{Q} = C_v \frac{dT}{dt} + p \frac{d}{dt} \left(\frac{1}{\rho} \right) \quad (1.12)$$

$$p = \rho RT \quad (1.13)$$

where $\mathbf{v} = (u, v, w)$ is the velocity vector, $\mathbf{g}$ is the gravitational acceleration and $d/dt = \partial/\partial t + \mathbf{v} \cdot \nabla$ is the total derivative following the atmospheric motion. The first equation is the equation of motion, in which $\mathbf{F}$ is external force such as frictional force. The term $2\rho\boldsymbol{\Omega}_z \times \mathbf{v}$ represents the Coriolis force, where $\boldsymbol{\Omega}_z$ is the vertical component of angular velocity of the earth $\boldsymbol{\Omega}$. The second equation is the equation of continuity which describes the conservation of mass in the air parcel. Eq. (1.12) is the energy conservation law where $\tilde{Q}$ and C_v are the heat input per unit mass and time and the specific heat capacity per unit mass at constant volume, respectively. The last equation is the equation of state for the ideal gas.

Eqs. (1.10)–(1.13) give a closed set of equation of $\mathbf{v}$, p , ρ and T , when $\mathbf{F} = 0$ and $\tilde{Q} = 0$, i.e., in an inviscid and adiabatic atmosphere. However, direct analytic solution for this set of equations is difficult to obtain, because they

are complicated nonlinear equations. Therefore, two simplifications are usually adopted to obtain the solution of this set. The first one is called the scale analysis, in which undesired small terms with respect to the time and spatial scales of the waves of interest are neglected [e.g., Holton, 1975]. The second one is the perturbation method, where atmospheric motions are expressed in terms of basic state and small perturbation such as $T = T_0 + T'$, and only the perturbation of the first order is treated. Thus, simplified linear equations are obtained and various kinds of waves are well approximated by these linear equations [e.g., Gossard and Hooke, 1975].

1.2.1 Atmospheric Tides

The atmospheric waves with a large horizontal scale, approximately ten thousands km, such as atmospheric tides and planetary waves have to be discussed considering the earth's sphericity. We here consider the waves of these kinds.

The basic equations of the atmospheric motion, Eqs. (1.10)–(1.13), are written in the form of the perturbation equation as

$$\frac{\partial v}{\partial t} + 2\Omega u \cos \theta = \frac{1}{a} \frac{\partial}{\partial \theta} \left(\frac{p}{\rho_0} \right) \quad (1.14)$$

$$\frac{\partial u}{\partial t} - 2\Omega v \cos \theta = -\frac{1}{a \sin \theta} \frac{\partial}{\partial \phi} \left(\frac{p}{\rho_0} \right) \quad (1.15)$$

$$\frac{\partial \rho}{\partial t} + w \frac{d\rho_0}{dz} + \rho_0 \chi = 0 \quad (1.16)$$

$$\frac{\partial p}{\partial t} - g\rho_0 w + \rho_0 C_s^2 \chi = (\gamma - 1) \bar{Q} \rho_0 \quad (1.17)$$

where a is the radius of the earth, θ denotes colatitude, and χ is the divergence of the wind velocity, i.e. $\chi = \nabla \cdot \mathbf{v}$.

The hydrostatic equation

$$\frac{\partial p}{\partial z} = -\rho g \quad (1.18)$$

is also included to this set of equations. Now we consider the Fourier component of each perturbation term and describe the component of frequency ω with a subscript of ω , for instance, χ_ω , $\bar{Q}_\omega$ etc. Then, Eqs. (1.14)–(1.18) are reduced to a second order partial differential equation with respect to θ , ϕ and z

$$\begin{aligned} C_s^2 \frac{\partial^2 \tilde{\chi}}{\partial z^2} + \left(\frac{dC_s^2}{dz} - \gamma g \right) \frac{\partial \tilde{\chi}_\omega}{\partial z} + \frac{g}{4a^2 \Omega^2} F \left[\left(g(1 - \gamma) - \frac{dC_s^2}{dz} \right) \tilde{\chi}_\omega \right] \\ = -\frac{g^2}{4a^2 \Omega^2} F \left[\frac{\bar{Q}_\omega}{C_p T_0} \right] \end{aligned} \quad (1.19)$$

where

$$\tilde{\chi} = \chi - \frac{\tilde{Q}}{C_p T_0} = -\frac{1}{\rho_0 C_s^2} \frac{dp}{dt} \quad (1.20)$$

$$\begin{aligned} F = & \frac{1}{\sin \theta} \frac{\partial}{\partial \theta} \left[\frac{\sin \theta \left(\frac{\partial}{\partial \theta} - i \left(\frac{\cot \theta}{f_0} \right) \left(\frac{\partial}{\partial \phi} \right) \right)}{f_0^2 - \cos^2 \theta} \right] \\ & + \frac{i \left(\frac{\cot \theta}{f_0} \right) \left(\frac{\partial^2}{\partial \theta \partial \phi} \right) + \left(\frac{1}{\sin^2 \theta} \right) \left(\frac{\partial^2}{\partial \phi^2} \right)}{(f_0^2 - \cos^2 \theta)} \end{aligned} \quad (1.21)$$

$$f_0 = \frac{\omega}{2\Omega} \quad (1.22)$$

We can solve Eq. (1.19) by the method of separation of variables by considering $\tilde{\chi}_\omega$ and $\tilde{Q}_\omega$ as

$$\tilde{\chi}_\omega = \sum_{m,n} Y_{\omega,m,n} \Theta_{\omega,m,n} \exp(im\phi) \quad (1.23)$$

$$\tilde{Q}_\omega = \sum_{m,n} q_{\omega,m,n} \Theta_{\omega,m,n} \exp(im\phi) \quad (1.24)$$

where m is the zonal wave number of the corresponding component. $Y_{\omega,m,n}$ is a solution of the following equation.

$$C_s^2 \frac{d^2 Y_{\omega,m,n}}{dz^2} + \left(\frac{dC_s^2}{dz} - \gamma g \right) \frac{dY_{\omega,m,n}}{dz} + \frac{Y_{\omega,m,n}}{h_{\omega,m,n}} \left[\frac{dC_s^2}{dz} + g(\gamma - 1) \right] = \frac{g q_{\omega,m,n}}{h_{\omega,m,n} C_p T_0} \quad (1.25)$$

which is called the vertical structure function. $\Theta_{\omega,m,n}$ is an eigenfunction known as a Hough function of the Laplace's tidal equation.

$$F_{\omega,m} \Theta + \frac{4a^2 \omega^2}{gh} \Theta = 0 \quad (1.26)$$

where $h_{\omega,m,n}$ is named as the equivalent depth. In order to obtain the atmospheric tidal modes, first we specify (ω, m) and determine $h_{\omega,m,n}$ and $\Theta_{\omega,m,n}$ which satisfies the boundary condition at $\theta = 0$ and $\pi/2$. Then from Eq. (1.25) vertical structure $Y_{\omega,m,n}$ is determined using the boundary condition at the upper and the lower ends.

Thus, the classical tidal theory was established mainly by Kato [1966] and Lindzen [1967]. The eigen-function of Laplace's tidal equation are fully solved by Longuet-Higgins [1968].

It is noteworthy for atmospheric tidal modes that equivalent depth of diurnal tides can take a negative value as well as a positive value [Kato, 1966]. The mode with positive and negative equivalent depth is called gravitational mode or

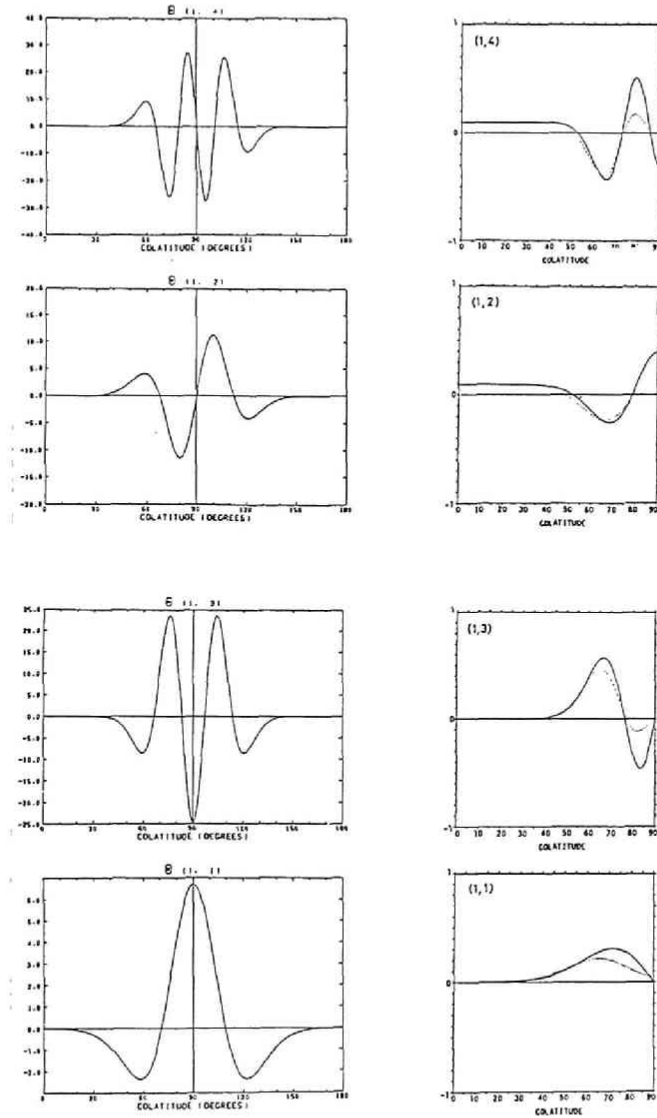


FIG. 1.2

Figure 1.4: Latitudinal distribution of Hough function Θ (left panels) and zonal (dotted in right panels) and meridional (solid in right panels) wind component of diurnal tides. Positive modes are plotted. (a) $S(1,4)$ mode, (b) $S(1,2)$ mode, (c) $S(1,3)$ mode, and (d) $S(1,1)$ modes. [Kato, 1980].

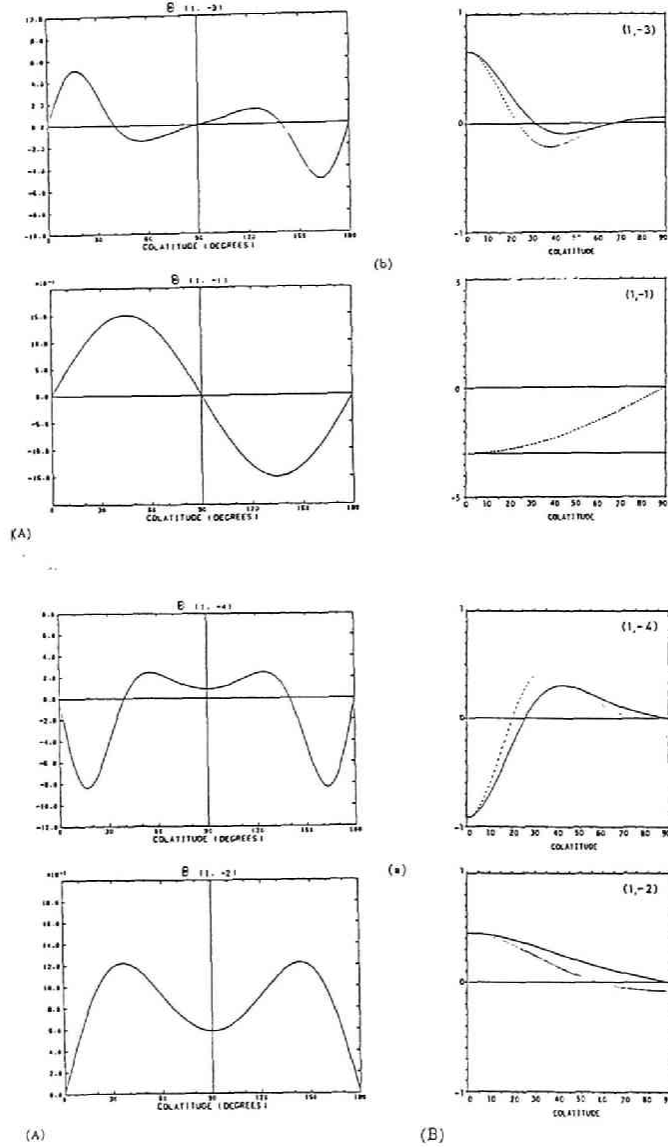


Figure 1.5: Latitudinal distribution of Hough function Θ (left panels) and zonal (dotted in right panels) and meridional (solid in right panels) wind component of diurnal tides. Negative modes are plotted. (a) $S(1, -3)$ mode, (b) $S(1, -1)$ mode, (c) $S(1, -4)$ mode, and (d) $S(1, -2)$ modes. [Kato, 1980].

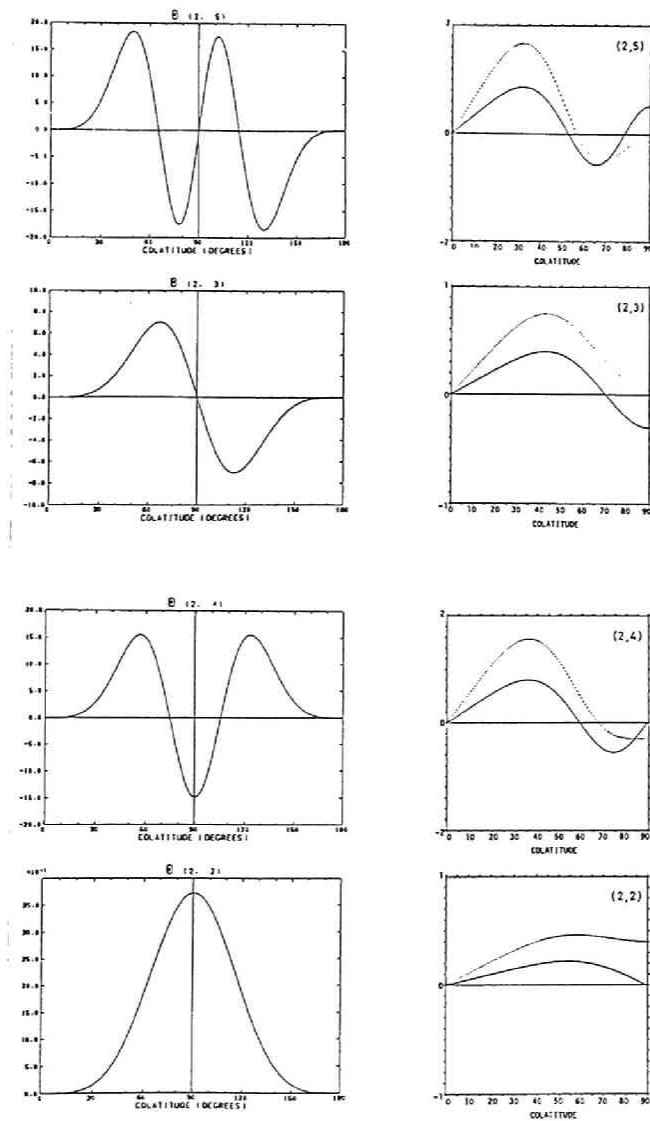


Figure 1.6: Latitudinal distribution of Hough function Θ (left panels) and zonal (dotted in right panels) and meridional (solid in right panels) wind component of semidiurnal tides. (a) $S(2,5)$ mode, (b) $S(2,3)$ mode, (c) $S(2,4)$ mode, and (d) $S(2,2)$ modes. [Kato, 1980].

positive mode and rotational mode or negative mode, respectively. The diurnal tide of a positive mode has a real wavenumber which means that the wave can propagate vertically, while the one of negative mode cannot propagate vertically but act as an external wave. Figure 1.4–1.6 shows latitudinal structure of Θ , u and v for several tidal mode. It is found from Figure 1.4 that the positive modes of diurnal tides have the major part of energy at low latitude region, while the negative modes show large amplitude at high latitude. This is recognized by the fact that since the Coriolis frequency $f = 2\Omega \cos\theta$ becomes equal to the frequency of diurnal tide at 30° latitude, at the lower latitudes diurnal tides can propagate as a gravity wave as discussed in the next subsection while at the higher latitude diurnal tides cannot propagate vertically, resulting the dominance of negative mode. On the contrary, semidiurnal tide can only take positive modes, i.e., all the semidiurnal tides can propagate vertically. This also corresponds to the fact that frequency of semidiurnal tide is always larger than the Coriolis frequency. As shown in Figure 1.6, semidiurnal tide can exist at high latitude as well as low latitude. Especially in the higher mode major part of energy are in the high latitude region.

Theoretical models are further improved from the classical tidal theory shown above by including molecular and eddy diffusions of heat and momentum, Newtonian cooling, electrodynamic forces, composition variations and interactions with background mean winds and meridional temperature gradients [Forbes, 1984]. The models without the diffusion were studied by Lindzen and Hong [1974] and Aso et al. [1981], etc. Forbes [1982a,b] included the eddy and molecular diffusions of heat and momentum in the model in order to properly address the structural modification of tides and their penetration into the upper atmosphere. Those theoretical models were mainly provided at the solstice and equinox conditions.

During the Middle Atmosphere Program (MAP) in 1982–1985, many observations on atmospheric tides were carried out. New CIRA model (CIRA, 1986) of the atmospheric temperature and zonal mean wind was also provided utilizing various observations carried out in MAP duration. Theoretical models of atmospheric tides utilizing these updated model atmosphere and improved informations of the dissipating process due to the eddy diffusion were recently proposed [Vial, 1986; Forbes and Hagan, 1988; Forbes and Vial, 1989], and the discrepancies with the observations has become smaller. Now the monthly mean climatologies comparatively agree better between the models and observations for semidiurnal tides [Forbes and Vial, 1989], however diurnal tides are not yet well described by the modelings.

1.2.2 Planetary Waves

The wind oscillations on time scale longer than 1 day are observed in the middle atmosphere. These waves are often global-scale waves with integer wave number in zonal direction, which are called planetary waves.

The long period waves with period of several days at mid-latitude are known as Rossby-wave, which is deduced by the approximation of the mid-latitude β plane, i.e., by treating the latitudinal difference of the Coriolis parameter. Madden and Julian [1972] showed the existence of global wave with period of 4–6 days and wavenumber of 1, by comparison among the global data of atmospheric pressure. Later, these waves were identified to symmetrical (1,1) mode of the free oscillation which is given by Laplace's tidal equation (Eq. (1.26)). This wave is named as 5 day wave. The satellite observation provided a number of data of the global feature of planetary waves in the height range up to the upper stratosphere. Many of them were found to be the normal mode of the Laplace's tidal equation. Besides 5 day waves, 4 day wave of symmetrical (2,1) mode, 10 day waves of anti-symmetrical (1,2) mode, 16 day wave of (1,3) mode were reported by Madden [1978], Hirota and Hirooka [1984] and Hirooka and Hirota [1985]. These are addressed as normal mode Rossby wave.

In the mesosphere and lower thermosphere, quasi-two day wave was observed by meteor and MF radars [e.g. Müller and Nelson, 1978; Craig et al, 1980] since 1970's, which was known to become large amplitude in summer season in both northern and southern hemispheres. The satellite observation of the upper stratosphere and mesosphere showed the global structure of the waves with zonal wave number 3, which was propagating toward westward direction [Rodgers and Prata, 1981]. This wave was also pointed out as a mixed Rossby-gravity wave of (3,0) derived from the tidal equation [Salby and Roper, 1980; Salby 1981b] as shown in Figure 1.7. However, Plumb [1983] showed the possibility of a baroclinic wave.

1.2.3 Gravity Waves

Now we apply the linear perturbation method to Eqs. (1.10)–(1.13). Hereafter, we use u , v , w , p and ρ to show the first order perturbations. Furthermore we transform the variables as

$$\begin{aligned}\tilde{u} &\equiv (\rho_0/\rho_s)^{1/2}u, \quad \tilde{v} \equiv (\rho_0/\rho_s)^{1/2}v, \quad \tilde{w} \equiv (\rho_0/\rho_s)^{1/2}w, \\ \tilde{p} &\equiv (\rho_0/\rho_s)^{-1/2}p, \quad \tilde{\rho} \equiv (\rho_0/\rho_s)^{-1/2}\rho\end{aligned}\tag{1.27}$$

where ρ_0 is the zero-order state of density, and ρ_s is the density at a reference height. The linear perturbation equations for Eqs. (1.10)–(1.13) are given as

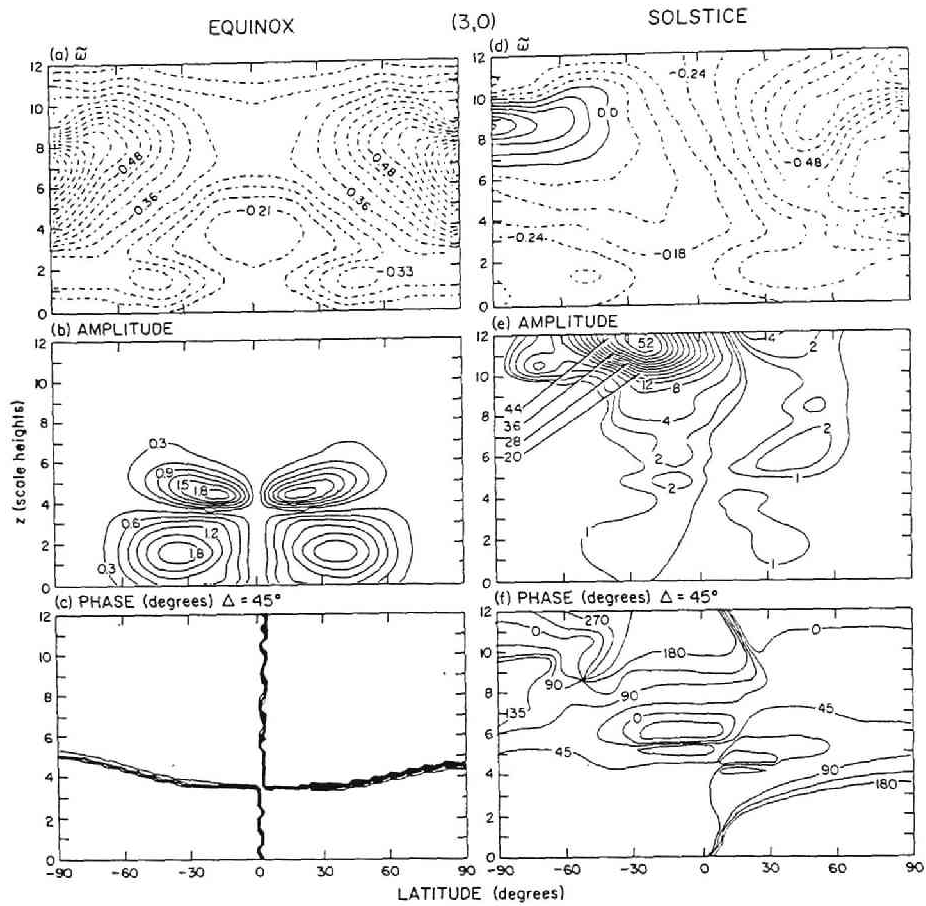


Figure 1.7: Latitudinal and height structure of the Rossby-gravity wave (3,0) in equinox condition: (a) intrinsic frequency $\tilde{\omega}$ (b) amplitude (c) phase, and in solstice condition: (d) intrinsic frequency $\tilde{\omega}$ (e) amplitude (f) phase [Salby, 1981b].

follows:

$$\frac{D\tilde{u}}{Dt} + \frac{1}{\rho_s} \frac{\partial \tilde{p}}{\partial x} - f\tilde{v} = 0 \quad (1.28)$$

$$\frac{D\tilde{v}}{Dt} + \frac{1}{\rho_s} \frac{\partial \tilde{p}}{\partial y} + f\tilde{u} = 0 \quad (1.29)$$

$$\frac{D\tilde{w}}{Dt} + \frac{1}{\rho_s} \left(\frac{\partial}{\partial z} - \Gamma \right) \tilde{p} + \frac{g}{\rho_s} \tilde{\rho} = 0 \quad (1.30)$$

$$\frac{1}{\rho_s c_s^2} \frac{D\tilde{p}}{Dt} + \frac{\partial \tilde{u}}{\partial x} + \frac{\partial \tilde{v}}{\partial y} + \left(\frac{\partial}{\partial z} - \Gamma \right) \tilde{w} = 0 \quad (1.31)$$

$$\left(\frac{D^2}{Dt^2} + N^2 \right) \tilde{w} + \frac{1}{\rho_s} \frac{D}{Dt} \left(\frac{\partial}{\partial z} + \Gamma \right) \tilde{p} = 0 \quad (1.32)$$

where c_s is the speed of sound given by $c_s^2 = \gamma RT_0$. Γ is the Eckert's coefficient given by

$$\Gamma = \frac{g}{2c_s^2} = \frac{1}{2\gamma H} \quad (1.33)$$

where $\gamma = C_p/C_v$, i.e., the ratio of the specific heat capacities at constant pressure (C_p) and at constant volume (C_v). N is called Brunt-Väisälä frequency, which is given by

$$N^2 = \frac{g}{\theta} \frac{\partial \theta}{\partial z} \quad (1.34)$$

where θ is the potential temperature of dry air parcel when the parcel is brought adiabatically to the reference level. D/Dt is the total derivative following the zero-order flow shown as

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + u_0 \frac{\partial}{\partial x} + v_0 \frac{\partial}{\partial y} \quad (1.35)$$

Next we investigate the characteristics of gravity waves by the equations derived above. We assume, hereafter, the solution of planar waves which propagate to the x -direction and that v_0 is 0. The solutions are described in the form of

$$\begin{aligned} \tilde{u} &= U \exp[i(kx + mz - \sigma t)] \\ \tilde{v} &= V \exp[i(kx + mz - \sigma t)] \\ \tilde{w} &= W \exp[i(kx + mz - \sigma t)] \\ \tilde{p} &= P \exp[i(kx + mz - \sigma t)] \\ \tilde{\rho} &= Q \exp[i(kx + mz - \sigma t)] \end{aligned} \quad (1.36)$$

for variables $\tilde{u}$, $\tilde{v}$, $\tilde{w}$, $\tilde{p}$ and $\tilde{\rho}$, where U , V , W , P and Q are the phasor (complex amplitude) of each variable, and k , m and σ are horizontal and vertical

wavenumber and frequency observed from the ground, respectively. If the perturbations are small enough to allow the linear assumption, and the coefficients ρ_s , c_s , N and Γ are constant, we obtain a system of homogeneous polynomials from Eqs. (1.28)–(1.32). The dispersion relation of the waves can be derived from the determinant of the coefficient matrix of the polynomials as follows

$$m^2 = \frac{N^2 - \omega^2}{\omega^2 - f^2} k^2 + \frac{1}{c_s^2} (\omega^2 - N_a^2) \quad (1.37)$$

where $N_a^2 = N^2 + c_s^2 \Gamma^2$. Note that ω in Eq. 1.37 is the frequency of the wave detected by an observer drifting with the zero-order flow, and is called the “intrinsic frequency”. The relationship between ω and σ is

$$\omega = \sigma - u_0 k \quad (1.38)$$

This equation describes the Doppler shift of the wave.

For vertically propagating waves, m should be real as shown in Eq. (1.36). Otherwise, the waves are attenuated exponentially with height (i.e., they are evanescent). Eq. (1.37) indicates that the propagating waves occur when $\omega^2 > N_a^2$ or $N^2 > \omega^2 > f^2$, which corresponds to acoustic waves and gravity waves, respectively. In case of gravity waves, the dispersion equation is given by

$$m^2 = \frac{N^2 - \omega^2}{\omega^2 - f^2} k^2 \quad (1.39)$$

It is noteworthy that in an isothermal atmosphere as derived from Eqs. (1.4) and (1.27), the amplitudes of gravity waves vary with vertical propagation, that is the amplitudes of u , v and w are proportional to $\exp(z/2H)$.

By using Eqs. (1.28)–(1.32) and (1.36), the following relationships among the variables of gravity waves are obtained as

$$V = -i \frac{f}{\omega} U \quad (1.40)$$

$$U = \frac{\omega k}{\rho_s(\omega^2 - f^2)} P \quad (1.41)$$

$$W = -\frac{\omega m'}{\rho_s(N^2 - \omega^2)} P \quad (1.42)$$

$$Q = -i \frac{N^2 m'}{g(N^2 - \omega^2)} P \quad (1.43)$$

where $m' \equiv m + i\Gamma$. Equation (1.40) shows that U and V are 90° out of phase and the ratio is proportional to f/ω , which means that the horizontal wind

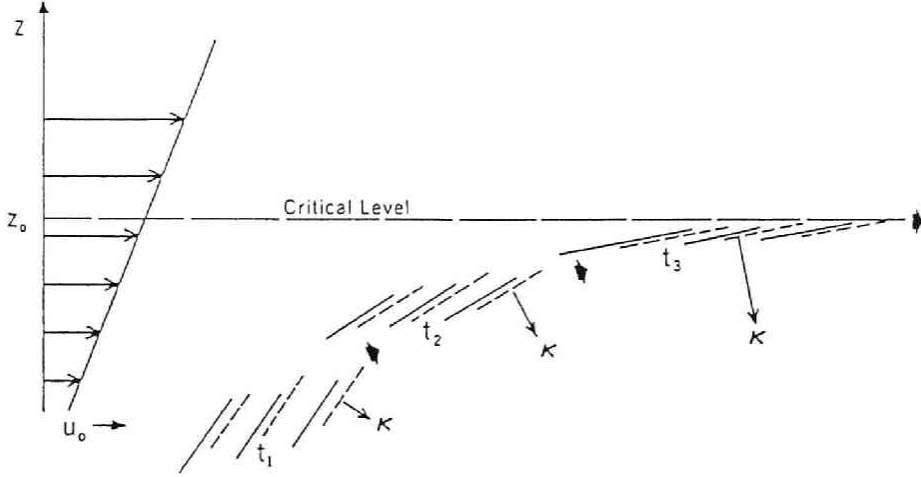


Figure 1.8: Schematic illustration of a wave packet propagating upward toward a critical level z_0 at successive times t_1 , t_2 and t_3 . Dashed lines represent new position of equiphase contours a moment after that represented by the solid lines [Gossard and Hooke, 1975].

vector of a gravity wave draws an ellipse when the variation in time or height is plotted into a hodograph. Assuming that the vertical wavelength of a gravity wave is less than the scale height H , we can assume $m' \approx m$. Thus, for typical gravity waves, we find from Eqs. (1.39) and (1.42) that

$$W = -\frac{k}{m}U \quad (1.44)$$

When a gravity wave propagates in the background wind $u_0(z)$ which is a function of height, the frequency σ observed from the ground is invariant, but the intrinsic frequency ω varies as

$$\omega = \sigma - ku_0(z) = k\{C - u_0(z)\} \quad (1.45)$$

where $C = \sigma/k$ is the horizontal phase velocity of the gravity wave. Figure 1.8 schematically illustrates a wave packet of the gravity wave which is propagating upward where the background wind is increasing with height by a small wind shear. Eq. (1.45) shows that the intrinsic frequency decreases when the mean wind, u_0 , approaches the phase velocity, C . This means that the vertical wavelength and the vertical phase velocity become smaller. The height where $C = u_0$ is called the “critical level”, which corresponds to $\omega = 0$, and gravity

waves cannot propagate across the critical level. In the real atmosphere, the wave is dissipated, and deposits the momentum flux $\rho_0 \overline{uw}$ to the mean flow. The acceleration of the mean flow is given by the following equation:

$$\frac{\partial u_0}{\partial t} = -\frac{1}{\rho_0} \frac{\partial}{\partial z} \rho_0 \overline{uw} \quad (1.46)$$

As shown in Fig. 1.3, in the 60–70 km region, the strong eastward and westward wind exists in the winter and summer hemisphere, respectively. These winds are primarily generated by the latitudinal difference of heating in the ozone layer, as shown in Section 1.1. Above this region, however, the zonal wind velocity decreases with altitude, and the wind direction is reversed around 80–100 km. Furthermore, the temperature at the mesopause is higher in winter than in summer as shown in Figure 1.2. To maintain these features some dynamic process is required to decelerate the mean zonal wind because the temperature distribution in the mesosphere is not in the radiative equilibrium, and the thermal wind equation describes just the balance between distributions of mean wind velocity and temperature. Now the momentum deposition by the gravity wave breaking is considered to result in the deceleration of the mean winds [Lindzen, 1981; Holton, 1982; Matsuno, 1982].

Fig. 1.9 shows a schematic illustration of the linear saturation theory proposed by Lindzen [1981]. The gravity wave becomes stable under the condition of

$$|u| \leq |C - u_0| \quad (1.47)$$

As shown in Fig. 1.9, we assume that gravity waves propagate upward until they reach saturation amplitudes at $z = z_s$, and above that height wave amplitudes are limited by the saturation conditions of Eq. (1.47). Then, saturation amplitudes of gravity waves are written as

$$|u_s| = |C - u_0| \quad (1.48)$$

These relations indicate that the horizontal phase velocity C and the mean wind u_0 is very important parameters to know the saturation and drag of the gravity waves, and hence they are intensively observed and discussed later in this thesis.

1.3 Scope of the Thesis

As described in this chapter, our knowledge on the dynamics of the middle atmosphere has rapidly increased in these two decades. Especially, the contribution of MAP (middle atmosphere program) studies was quite large. The improved

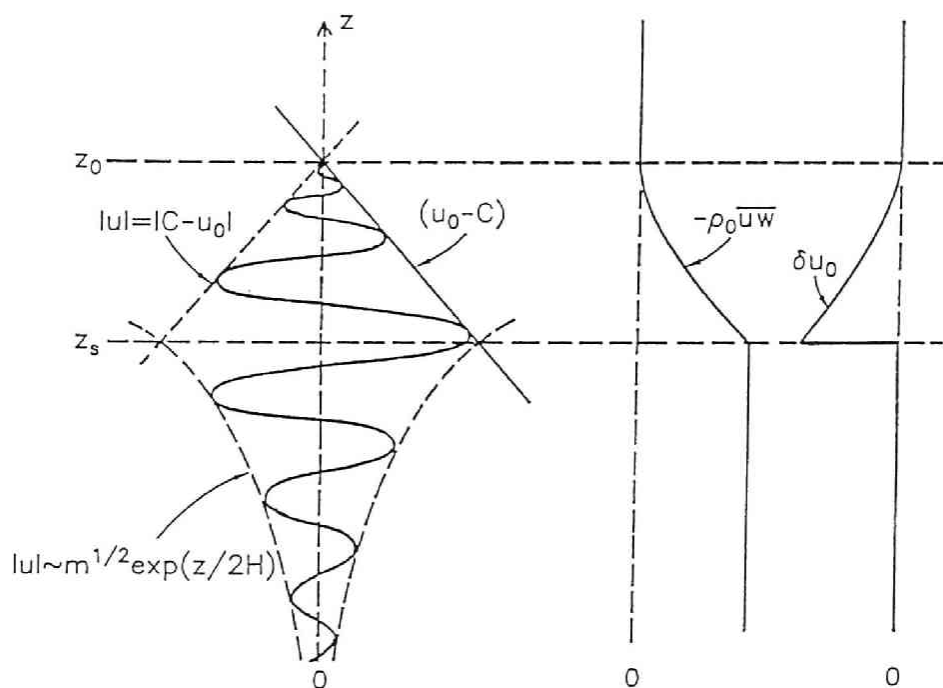


Figure 1.9: Schematic of the growth with height and saturation of a gravity wave due to convective instability. Wave damping produces both a divergence of the vertical flux of horizontal momentum and an acceleration of the mean flow δu_0 toward the phase speed of the wave. Deceleration and diffusion cease above the critical level $z = z_0$ in the linear theory [Fritts, 1984].

CIRA describes more realistic model of the middle atmosphere. However, the region of mesosphere and lower thermosphere is still the region where many uncertainty are left on dynamics, because we have few observational data in comparison with the complicated dynamics occurring there. The purpose of this thesis is to reveal the dynamics, especially atmospheric waves in the mesosphere and lower atmosphere, by using radar observations.

We here briefly summarize the general development in our understanding of the mesosphere and lower thermosphere dynamics concerning with this thesis. The soundings of the mesosphere and lower thermosphere at the beginning of the study were carried out in order to describe the mean wind profile at these altitudes, which were summarized by atmospheric models. Climatologies of the mean winds have been observed and studied by numbers of radar observations and other observation techniques, including the observation at Shigaraki of this work, and the asymmetry of the northern and southern hemispheres of the mean wind has been clarified. This has been reflected in the new model of CIRA 1986.

Next, the atmospheric tide was extensively observed by utilizing the radar techniques. By the comparisons of the long periods observation among several radars, our knowledge on the global structure of tide in the mesosphere and lower thermosphere has been greatly improved. They also stimulated the improvement of the tidal modeling. The observations between the northern and southern hemispheres such as Shigaraki and Adelaide [Vincent et al., 1988, 1989] have revealed a significant asymmetry in tidal structure, that is the conspicuously larger amplitude of diurnal tide in the southern hemisphere. The planetary waves were also studied by radar observations, while it is rather difficult to identify the waves because the wave periods are not clarified. Among the planetary waves, two-day waves were well observed at these height, and also large asymmetry in amplitude has been observed between the hemispheres.

One possible explanation of the hemispheric difference of the tides is the difference of the dissipation due to the eddy diffusion [Vincent et al., 1988], but the difference of sources and propagations through background mean winds can also be the reason of the asymmetry. Because the major part of these diffusions are now considered to be generated by the gravity wave breaking, the gravity wave amplitudes are expected to be larger in northern hemisphere where the tidal amplitudes are smaller. Therefore, the global observation of the gravity waves is very important to study the origin of the tidal asymmetry between the hemispheres.

The gravity waves are more recently studied because they needs high time resolution in observations. Because the gravity waves are considered to cause the significant momentum drag to the mesopause regions and composing the general

circulation of this altitude, many authors tried to observe them by radars and other techniques, and now their characteristics are being revealed.

In the present thesis, we studied the gravity waves closely in order to reveal their characteristics and to reveal the global structure of gravity wave activity, which will be deeply concerned with the asymmetry of the tides and planetary waves.

The chapters of this thesis are composed as below.

Chapter 2 describes the radar techniques and experimental set up of the radars used in this study. The MU radar as an MST radar, the Kyoto meteor radar as a meteor radar, and two MF radars of Adelaide and Saskatoon are described. The importance of the interferometry for the meteor radar observations is further discussed.

In Chapter 3, the mean wind observations are presented. The mean wind at Shigaraki (35°N) observed with the MU radar and Kyoto meteor radar is shown, and the difference from CIRA 1986 model is discussed. The result is still compared with Adelaide data and the hemispheric difference is discussed.

Chapter 4 presents the atmospheric tide and two-day waves by the observation of Kyoto meteor radar in 1983–1986. They are compared with the data observed at Adelaide and Saskatoon. Hemispheric symmetry and latitudinal structure of the tides are discussed.

Chapter 5 shows the gravity wave observation by using the MU radar. Four campaign observation in October 1986, June 1987, July 1990 and January/February 1991 are presented. Furthermore, results from monthly observation are also shown. The momentum flux due to gravity waves, and propagation direction of gravity waves are discussed.

In Chapter 6, gravity waves observed with the MU radar are compared with the observation at Saskatoon (52°N) for the whole year of 1989, and with Adelaide (35°S) for two campaigns of June 1987 and January/February 1991. The latitudinal and hemispheric difference of the gravity wave activity is discussed here.

The summary and conclusion of this thesis are presented in Chapter 7.

Chapter 2

Radar Observation Techniques

In the present thesis we utilize several kinds of radar equipment to observe the mesosphere and lower thermosphere. The observation at Shigaraki, Japan (35°N , 136°E) was carried out by the MU (Middle and Upper atmosphere) radar and Kyoto meteor radar. These two observations are closely analyzed, and then compared with other two radar observations; MF radar observation at Adelaide, Australia (35°S , 139°E), and at Saskatoon, Canada (52°N , 107°W) (Figure 2.1). It is noteworthy that Adelaide is located at the geographically conjugate point of Shigaraki, so that the comparative study between Shigaraki and Adelaide gives us invaluable information in discussing hemispheric symmetry of the dynamics of earth's atmosphere. The comparison with Shigaraki and Saskatoon will contribute the understanding of latitudinal difference between low latitude and middle latitude. In this chapter, we present techniques to observe the middle atmosphere, especially mesosphere and lower thermosphere region, and describe the radar systems utilized in this study.

2.1 Observation Techniques of the Middle Atmosphere

There are several methods to observe the physical parameters of the middle atmosphere, such as temperature, pressure, density and wind velocity, etc. Each of these measurements has its own observational height range. *In-situ* measurements such as rocket and radiosonde observations have been operated for long time, in which a small measuring instrument is carried by a rocket or balloon. Rocket observations are categorized into several kinds by means of the payload of the measuring system. Among them a falling sphere and datasonde are often launched in recent observations because these payloads are small and less ex-

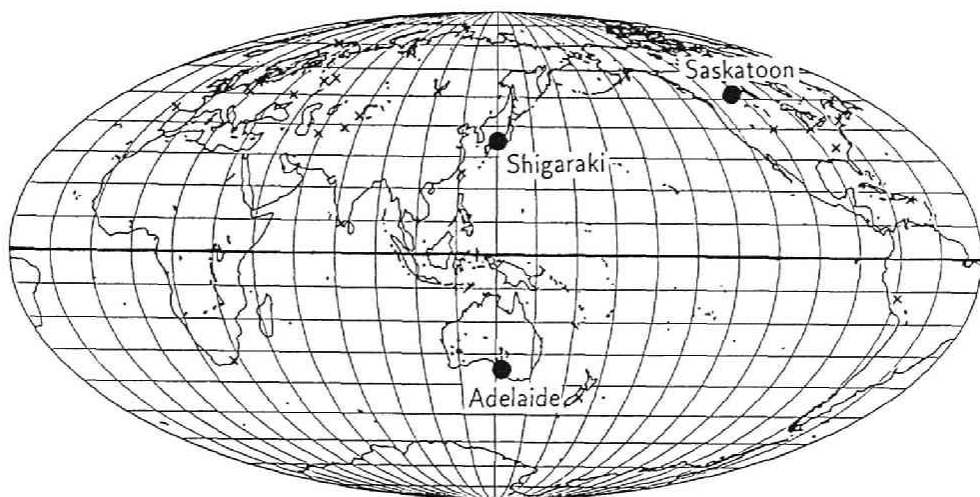


Figure 2.1: Location of the radar observation sites utilized in the present thesis; Shigaraki (35°N , 136°E), Adelaide (35°S , 138°E) and Saskatoon (52°N , 107°W).

pensive [Schmidlin, 1986]. Their observational height ranges are 30–90 km and 20–65 km, respectively. The falling sphere observation is carried out by tracking a sphere, which is launched and released by a rocket while descending using a primary radar system. On the other hand, the rocketsonde is a sounding system with a radio transmitter, which is released from the rocket and descends slowly by means of parachute. Now rocketsonde observation is carried out at many stations in the world periodically.

Radiosonde is the same as rocketsonde except that the instrument is launched from the ground by a balloon and the measurement is done while ascending. It can measure the atmosphere up to 35 km altitude. These *in-situ* measurements are good in height resolution. However, these observations give only one vertical profile for a parameter by a single launch, so that successive launching is necessary for studying atmospheric dynamics as a time series of data. Rocket and radiosonde observations are limited by 2–4 soundings per a day, so that these observations are appropriate only for the waves with long periods, such as planetary waves [e.g. Hirota, 1978]. On the other hand, the good height resolution of these soundings enables us to detect a gravity wave as a fluctuation in the vertical profile [Hirota, 1984; Tsuda et al., 1991]. The statistical analysis of the rocketsonde observation at many meteorological sites are revealing the characteristics of inertia gravity waves at 20–60 km [Eckermann et al., 1991].

On the contrary, remote sensing measurement using radio waves and optical rays provide more successive data in time and height. The satellite observation has now become one of the most powerful observation methods in atmospheric dynamics. It can obtain global data by turning round the earth. Their height resolution is the order of 2–8 km [Rodgers, 1984], and hence not enough to study the atmospheric waves with short vertical wavelength in the middle atmosphere, but rather suitable for observing global scale waves such as planetary waves [Hirota and Hirooka, 1984]. The gravity waves with long vertical wavelength ($>10\text{km}$) were also observed at 15–65 km altitude by Limb Infrared Monitor of the Stratosphere (LIMS) observations [Fetzer, 1990].

The ground based remote sensing system, such as radar and lidar, are characterized by the continuous monitoring of the atmosphere and high time and height resolution of 1–10 min and 100m–5km. Various radar measurements of the middle atmosphere have clarified characteristics of the atmospheric waves such as tides [e.g. Manson and Meek, 1985; Aso and Vincent 1982], planetary waves [e.g. Salby and Roper, 1980; Ito et al., 1984] and gravity waves [e.g. Vincent, 1984a; Manson et al., 1981; Tsuda et al., 1989]. Each radar technique is described in the following sections. The lidar is an observation method developed more recently. The Rayleigh lidar can observe the atmospheric density at 30–80km altitude, which can cover the blind height of radar observation (30–60km). It is revealing the gravity waves [Wilson et al, 1991], and tides [Gille et al., 1991] in those altitudes. Sodium lidar also can observe the density of 80–105km altitudes, which can observe gravity waves [Gardner and Voelz, 1987]. Doppler lidar system has also been developed and used for measuring the wind velocity.

2.2 MST Radar Observations

The MST (Mesosphere Stratosphere and Thermosphere) radars are a radar system with large antenna aperture and powerful transmitter using the VHF (30–300MHz) and UHF (300–3000MHz) radio waves to detect weak backscattering arisen by refractive index fluctuations in the mesosphere, stratosphere, troposphere and lower thermosphere [Balsley and Gage, 1980]. The MST radar technique was developed by the incoherent scatter radar which observes the ionosphere around 1970. Unexpected strong echoes from mesosphere and stratosphere were received by Jicamarca radar in Peru, and later it is clarified that these echoes are the scatter from the irregularity of refractive index caused by the turbulence drifting with the neutral winds [Woodman and Guillén, 1974]. This MST technique has become one of the most important radar techniques in observing the wind velocity and turbulence intensity in the middle atmosphere.

The scatterer of the MST radar observation is not a point target, but a distributed volume. The radar equation is written by using the scattering cross-section per unit scattering volume, i.e. "reflectivity" η , as

$$P_r = \frac{P_t A_e \pi \Delta r}{64 r^2} \eta \quad (2.1)$$

where P_t , A_e , r and Δr are transmitted power, effective area of antenna, range and range resolution, respectively [Probert-Jones, 1962].

The reflectivity η is expressed as

$$\eta = \frac{k^4}{\pi} \left\langle \left| \int_V \Delta n \exp(i2\mathbf{k} \cdot \mathbf{r}) dV \right|^2 \right\rangle \quad (2.2)$$

where k ($= 2\pi/\lambda$) is the radar wavenumber, $\mathbf{k}$ is the propagation vector and $\mathbf{r}$ is the radius vector to a point in the scatterer volume. $\langle \rangle$ denotes an ensemble average. We now consider the isotropic scatterer distributed uniformly in the radar sampling volumes caused by the turbulent fluctuations. Then the volume reflectivity η is expressed in terms of three-dimensional wavenumber spectrum in the reflective index $\Phi_n(\mathbf{K})$ as [Ottersten, 1969],

$$\eta = \frac{\pi^2}{2} K^4 \Phi_n(\mathbf{K}) \quad (2.3)$$

In the case of the backscatter by monostatic radar, $\mathbf{K}$ is equal to $2\mathbf{k}$. The magnitude of $\mathbf{K}$ is expressed in terms of the wavelength of the radio wave λ as

$$K = |\mathbf{K}| = 2 \left(\frac{2\pi}{\lambda} \right) = \frac{4\pi}{\lambda} \quad (2.4)$$

The radio wave is backscattered mainly by the $\lambda/2$ Fourier component of irregularities of atmospheric refractive index which are mainly caused by turbulence. The refractive index of the atmosphere is also known to be expressed in terms of pressure P [mb], temperature T [K], water vapor pressure e [mb], and electron density N_e [m^{-3}] as [e.g., Röttger, 1980]:

$$n = 1 + \frac{3.73 \times 10^{-1} e}{T^2} + \frac{7.76 \times 10^{-5} P}{T} - \frac{40.3 N_e}{f_r^2} \quad (2.5)$$

where f_r is radio wave frequency [Hz] of the radar.

The MST radar technique is capable of obtaining useful returns over the height range of 1–100 km. Figure 2.2 shows the relative echo power profiles for various MST radar facilities [Gage and Balsley, 1984], and indicates intense echo power in the lower stratosphere (10–30 km) and the mesosphere (60–90 km)

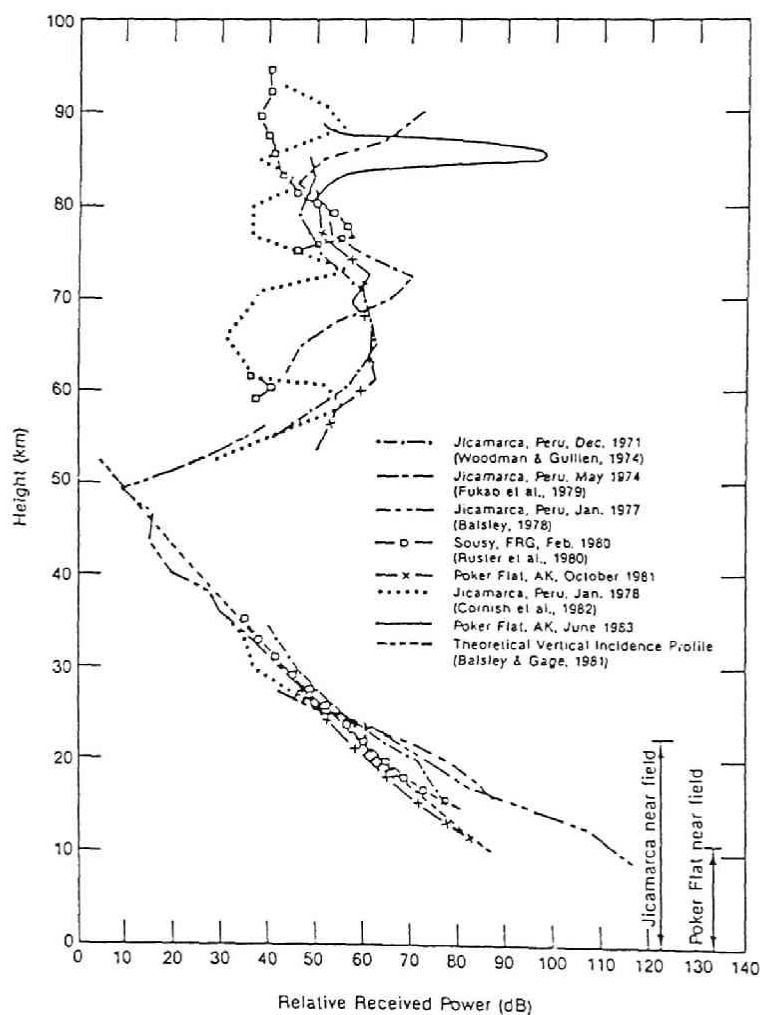


Figure 2.2: Relative echo power profiles for various MST radar facilities (curves normalized at lower stratospheric heights) [Gage and Balsley, 1984].

in the middle atmosphere. Echoes coming from heights below the stratopause ($\simeq 50$ km) arise primarily from refractive index fluctuations due to small-scale turbulence in the inertial subrange. Water vapor pressure, e , is usually the most important factor in the lower troposphere where the humidity is high. Above this region the second term, that is the density term, dominates up to the stratopause. Echoes returned from the mesosphere are normally observed only during daylight hours, and are much stronger than the level of incoherent scatter caused by free electrons. The mesospheric echoes arise from the neutral turbulence fluctuations which are enhanced by free electrons at these heights. The most difficult region to be observed by the MST radar technique is the region near the stratopause because of the combined effect of the exponentially decreasing atmospheric density and the lack of sufficient free electrons [Balsley and Gage, 1980]. The MU radar is an MST radar located at Shigaraki, Shiga prefecture. It also can be operated as an incoherent scatter radar [e.g. Sato et al., 1989]. The detail of the MU radar system is described by Fukao et al. [1985a, b], however we briefly summarize it in this subsection.

The MU radar is characterized by the active phased array antenna system and flexible design of the system using full computer control. The former enables us to steer the antenna beam pulse by pulse, i.e. 2,500 times per a second at the maximum. Thus, the tridimensional structure of the atmosphere is easily obtained. The latter means that the MU radar can be used as a various types of the radar system by changing the software parameters of the radar system. For example, the antenna array is easily divided into several groups and can carry out an SA (spaced antenna) measurement [Van Baelen, et al., 1990]. The special set up for observing meteor echoes are also developed by Nakamura et al. [1991a], which is described in another section.

Basic parameters of the MU radar are shown in Table 2.1 [Fukao et al., 1985a]. The MU radar is a monostatic pulse Doppler radar operating at a frequency of 46.5 MHz. The maximum peak and average transmit power are 1 MW and 50 kW, respectively. The antenna is a circular array with an aperture of 8330 m². The one way half-power full width of the main beam is 3.7°. The shortest pulse with 1- μ s width is available in the 1.65-MHz bandwidth reserved exclusively for the MU radar.

The MU radar has been collecting excellent data for the study of the dynamics in the lower and middle atmosphere (2km–25km, 60–90 km) and the ionosphere (100–500km), and is utilized by many researchers in and outside of Japan.

Fig. 2.3 shows the schematic diagram of the passive and active phased array antennas. In a conventional MST radar, a passive antenna array is used in which a high-power transmitter feeds all array elements via an appropriate feeding

Parameter	Value
Location	Shigaraki, Shiga, Japan (34°51'N, 136°06'E)
Radar system	monostatic pulse radar; active phased array system
Operational frequency	46.5 MHz
Antenna	circular array of 475 crossed yagis
aperture	8330 m ² (103 m in diameter)
beam width	3.7° (one way; half power full-width)
steerability	steering is completed in each IPP
beam directions	1657; 0°–30° off zenith angle
polarizations	linear and circular
Transmitter	475 solid state amplifiers (TR modules; each with output power of 2.4 kW peak and 120 W average)
peak power	1 MW (maximum)
average power	50 kW (maximum duty ratio is 5 %)
bandwidth	1.65 MHz (maximum)
IPP	400 μ s to 65 ms (variable)
Receiver	
bandwidth	1.65 MHz (maximum)
dynamic range	70 dB
IF	5 MHz
A/D converter	12 bits $\times$ 8 channels
Pulse compression	binary phase coding up to 32 elements; Barker and complementary codes

Table 2.1: Basic Parameters of the MU radar [Fukao et al., 1985a]

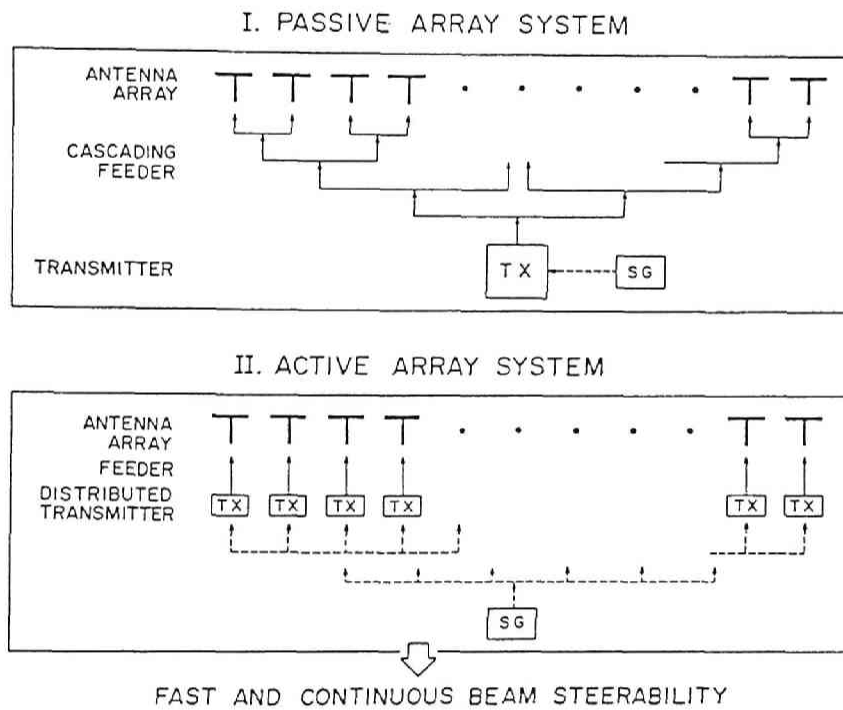


Figure 2.3: A schematic diagram of passive and active phased array systems [Fukao et al., 1985a].

network. On the other hand, antenna array of the MU radar is composed of an active phased array, in which each element of the antenna array is provided with an individual power amplifier, and all the amplifiers are driven by low-level pulsed carriers. This architecture enables the radar very fast beam steering by controlling the phase of radio waves emitted from each antenna.

Figure 2.4 shows a block diagram of the MU radar system [Fukao et al., 1985a]. The 475 transmitter-receiver (TR) modules are accommodated in six TR booths near the antenna array, each of which drives its own array element. The whole antenna and TR modules can be divided into 25 groups (i.e. one group consists of 19 array elements and 19 TR modules). A TR module is composed of a solid state transmitter, a receiver preamplifier, a T/R switch, and a digital phase shifter. Input of transmitting pulse and the output of the received signal of TR module are performed at the intermediate frequency (IF) of 5 MHz in order to improve the stability of the amplifiers.

An antenna element consists of two orthogonally crossed three-subelement yagis pointed toward the zenith direction. Linear polarizations aligned EW or NS directions and lefthand and righthand circulation polarizations can be selected in each TR modules. The array configuration is designed to be nearly circular with the diameter of 103 m in order to obtain almost symmetrical side-lobes with respect to azimuth angle. A triangular grid is used for arranging the array elements in order to attain a wider range of antenna beam steering than a rectangular grid with the same element density [Sato, 1981].

The whole arrangement of antenna array is shown in Fig. 2.5. The element spacing is 4.5 m, which corresponds to 0.698λ (λ is wavelength), for which no grating lobe is formed at beam positions within 40° away from the zenith. The first side lobe of the antenna beam is suppressed by more than 18 dB with respect to the main lobe. The radiation at elevation angles below 20° is reduced by more than 40 dB even when the antenna beam is tilted 30° from the zenith.

The overall operation of the MU radar is supervised by a radar controller which is composed of a engineering workstation HP9000/330. Timing signals to operate the whole system are generated according to instructions from the radar controller. A variety of flexible operations can be carried out by softwares in the radar controller. The received signals are decoded for pulse compression, and then coherently integrated in order to reduce the amount of data by the special hardware. A large amount of data can be processed in real time by using a minicomputer (VAX-11/750) and an array processor (MAP-300).

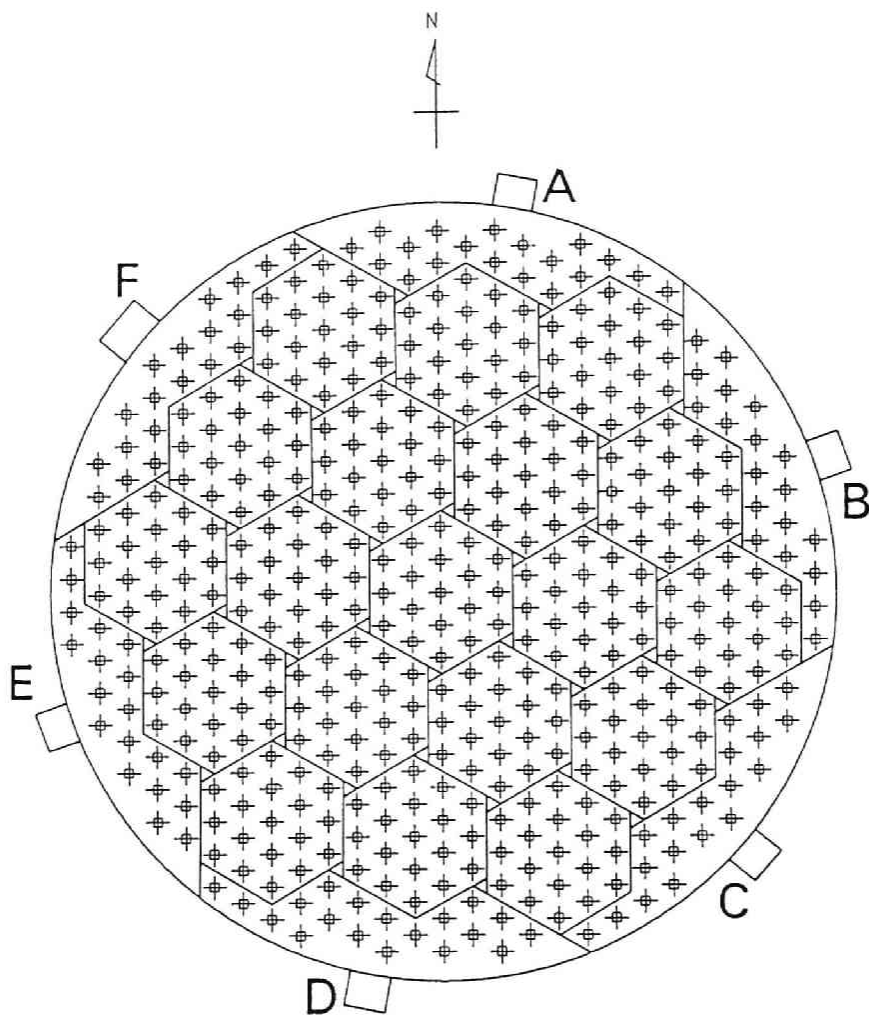


Figure 2.5: MU radar antenna array. Each cross represents a crossed three-subelement yagi. The small square at each array element position is a concrete block for the element foundation. The six boxes outside the antenna array, indicated as A–F, represent the booths which accommodate the TR modules. All booths except F accommodate TR modules for three hexagonal subarrays and one peripheral subarray. The TR modules for the central hexagon are housed in booth F [Fukao et al., 1985a].

2.3 Meteor Wind Radar Observations

Meteor is a phenomenon that a meteoric particle drifting in the interplanetary space is caught by the gravitational force of the earth, and impinges on the earth's atmosphere. Mean original mass of a meteor is as small as 0.1 mg for sporadic meteors [Verniani, 1973]. However, mean apparent velocity is as large as 34 km/s, and hence the intense ionization columns are formed at the height of mesosphere and lower thermosphere. These ionized meteor trails scatter the radio waves, before they diffuse away. The radar of detecting these meteor trails are first used to observe astronomical properties of the meteors [McKinley and Millman, 1949], while later neutral wind velocities around the meteor trail were observed by measuring Doppler shift of the meteor echoes due to the meteor trail drifting with the surrounding atmosphere [Manning et al., 1950]. More than forty meteor radars were constructed in the world, and observed the mean winds and atmospheric waves in the meteor height of 80–110 km altitudes.

The meteor echoes are divided into two kinds; underdense and overdense echoes. In the former case, the plasma frequency in the ionized column is larger than the radar frequency, and the radar wave can penetrate the trail and partially reflected by each electron. In the latter case, radio waves are totally reflected by the ionized column. The transition of those two echoes is about 10^{14}m^{-3} in line electron density for the VHF radio waves [McKinley, 1961]. For the meteor wind measurement, overdense echoes are not appropriate because the Doppler shift of an overdense echo includes the expansion velocity of the column surface as well as drift by the neutral wind. Thus, we discuss the scattering by underdense echo below. In the underdense meteor trail, radio wave is reflected by each electron independently at a scattering cross section of

$$\sigma_e = \frac{\mu_0^2 e^4}{16\pi^2 m^2} \quad (2.6)$$

where μ_0 , e and m are magnetic permeability of air, electron charge and mass, respectively. Then the received power by each electron ΔP_r is represented by a radar equation as

$$\Delta P_r = \frac{P_t A_e A'_e \sigma_e}{4\pi r^4} \quad (2.7)$$

where P_t , A_e , A'_e and r are transmitting power, effective aperture of transmitting and receiving antennas and range, respectively. The total received power is obtained by the summation of ΔP_r for the column volume and given by a form of Fresnel integration as

$$P_r = \frac{P_t A_e A'_e \sigma_e}{8\pi r^3 \lambda} q_e^2 \frac{C^2 + S^2}{2} \quad (2.8)$$

Parameter	Value
Location	Shigaraki, Shiga, Japan (34°51'N, 136°06'E)
Radar system	monostatic pulse radar; active phased array system
Items	Echo amplitude, range, Doppler shift, and arrival directions (azimuth, elevation)
Operational frequency	31.57 MHz
Antenna	5 element Yagi (1 for TX; 6 for RX)
Receiver	3-channel (for interferometer)
Power	10 kW (peak), 840 W (max average)
IPP	3ms to 50ms (computer controlled)
Range resolution	1.5 km
Pulse compression	28 bit quasi random code
Computer	HP2108A (small computer) radar control and real time data processing

Table 2.2: Basic parameters of Kyoto meteor radar [Tsuda, 1982]

where q_e is a line density of electron. This means that the meteor echo scattering is regarded as a specular reflection. The value of $(C^2 + S^2)/2$ becomes nearly 1 when the Equation 2.8 is integrated for a few Fresnel zones. After the formation of the ionized column, it expands radially by ambipolar diffusion, causing the echo intensity decay as [Herlofson, 1948],

$$P_r(t) = P_r(0) \exp\left(-\frac{32\pi^2 Dt}{\lambda^2}\right) \quad (2.9)$$

where D is the ambipolar diffusion coefficient of surrounding atmosphere.

Kyoto meteor radar is a compact economical radar system, which was constructed at Shigaraki (35°N, 136°E) in 1977. The characteristics of this system is the interferometer system which can determine the direction of each meteor echo, and the computer system which is adopted for the radar controlling and real time data processing [Tsuda et al., 1986]. This radar system worked until 1986, and in the last three years continuous observation were carried out by the automatic observing system for 24 hours and a large amount of observational data base of the meteor height of 80–110 km were stored.

The parameters of Kyoto meteor radar is summarized in Table 2.2. It is a phase coherent pulse Doppler radar with the radar frequency of 31.57 MHz, and the peak and average power of the transmitter is 10 kW and 840 W, respec-

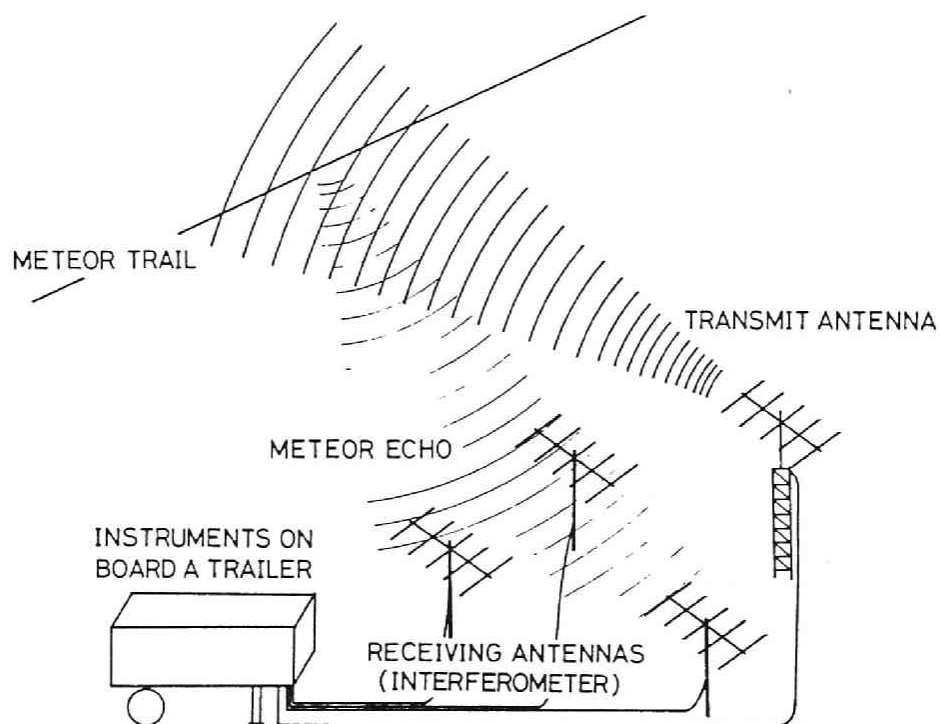


Figure 2.6: A perspective view of Kyoto meteor radar [Tsuda, 1982].

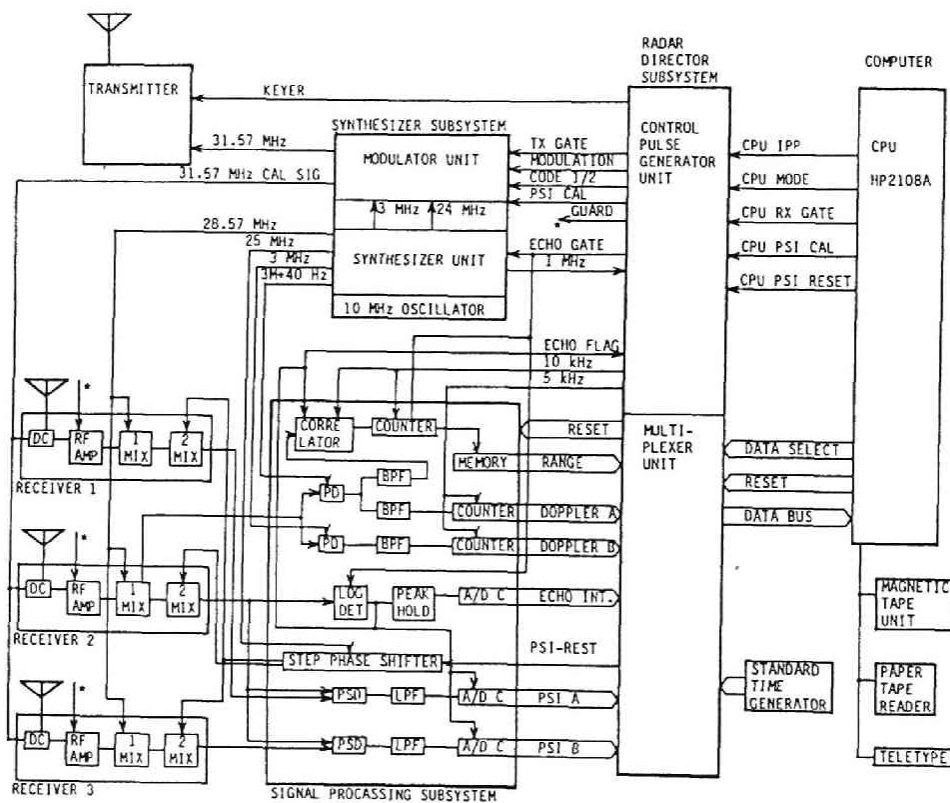


Figure 2.7: A block diagram of Kyoto meteor radar [Tsuda, 1982].

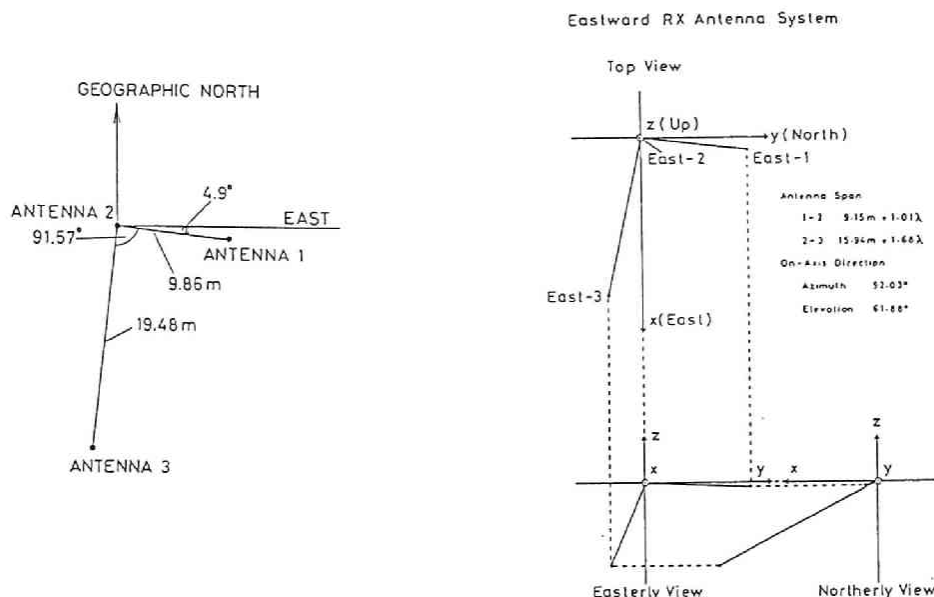


Figure 2.8: Arrangement of six receiving antennas utilized in Kyoto meteor radar.

tively. The general view of Kyoto meteor radar is drawn in Figure 2.6 and the block diagram of the Kyoto meteor radar is shown in Figure 2.7. The in-house equipment was taken on board a trailer in order to obtain mobility.

The 5-element Yagi antennas are used for transmitting and receiving antennas. Figure 2.8 shows the location of 6 receiving antennas. Three out of six antennas are facing northward, and the rest are toward eastward. Each of the RX antennas is connected to the individual receiver channel, which construct a phase sequenced interferometer system [Rudman et al., 1970]. The half power beam width of 5-element Yagi antenna is about 60° , so that the received meteor echoes are assumed to be in this main lobe. Then the elevation and azimuth are determined by the interferometry. If a unique direction cannot be deduced in the antenna illuminating area, the direction of that echo is not determined. The long baseline along the bore site comparative with the other baseline causes that the elevation error is smaller than the azimuth error in the direction measurement.

The northward and eastward sets of antennas can be switched and the transmitting antenna can be turned the azimuth by an electric rotator, so that we can operate the northward and eastward observations alternately. The narrow antenna beam will collect meteor echoes of a limited azimuth area, and therefore the measurement of the wind component with the perpendicular azimuth to the

antenna boresight become less accurate. However, in the continuous observation carried out in 1983–1986, antenna beam is fixed to the eastward in order to observe zonal wind component closely. Therefore, wind data observed with Kyoto meteor radar used in this thesis have better accuracy in zonal component than the meridional component.

2.4 Meteor Wind Observation with an MST Radar

Although the MST radar is a powerful radar that can observe the whole middle atmosphere with good height and time resolutions except for the blind height of 30–60 km, the radar echoes from the mesosphere (60–90km) are received only in the daytime. The scatterer of MST radars in the mesosphere is the fluctuation of refractive index which is at this altitude governed by the electron densities. The electron density in the mesosphere is significantly affected by the incident of solar radiation, the electron density in the mesosphere at night is too low to scatter radio waves enough to be observed by MST radars. On the contrary, meteor echoes are available throughout a day because numbers of meteors are always impinging into the atmosphere, although in the evening the meteor rate becomes smaller because the observation site becomes behind of the earth to the apex. Therefore, even for MST radars a meteor scatter is an important target to observe a wind for a long time in order to observe long period waves, such as tides and planetary waves as well as mean winds.

However, meteor echoes usually last not more than 1 sec. Thus, special set-up or adherent instrument is necessary to extract meteor echoes efficiently. Avery et al. [1983] have developed a meteor echo observation by the Poker Flat MST radar, Alaska (65°N), and Wang et al. [1988], Avery et al. [1990] further developed a data-acquisition sub-system that can be used with an MST(/ST) radar. However, those studies do not determine the arrival direction of each meteor echo, but rather assume the meteor echoes are received in a pencil-beam of the antenna. Because the dynamic range of meteor echo intensity is so large, significant number of meteor echoes are considered to be received in the sidelobes of the antenna. So that the wind velocity analyzed by the meteor data without measuring the arrival direction can include a significant error.

One of the characteristics of the MU radar is the versatility by means of computer controlled system, which enables us to realize the experimental set up suitable for meteor observation. Figure 2.9 shows the example of antenna usage for the meteor observation mode by the MU radar. For transmission 21 group out of 25 group of antenna subarray are used, and four Yagi antennas, which

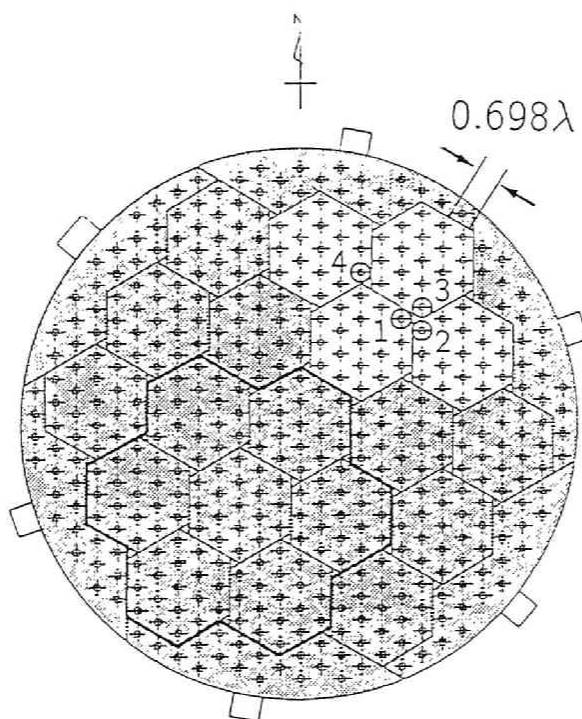


Figure 2.9: Antenna configuration for meteor observation with the MU radar. Circles with numbers are receiving antenna with the channel number to be connected. Shaded area shows the transmitting antenna [Nakamura et al., 1991a].

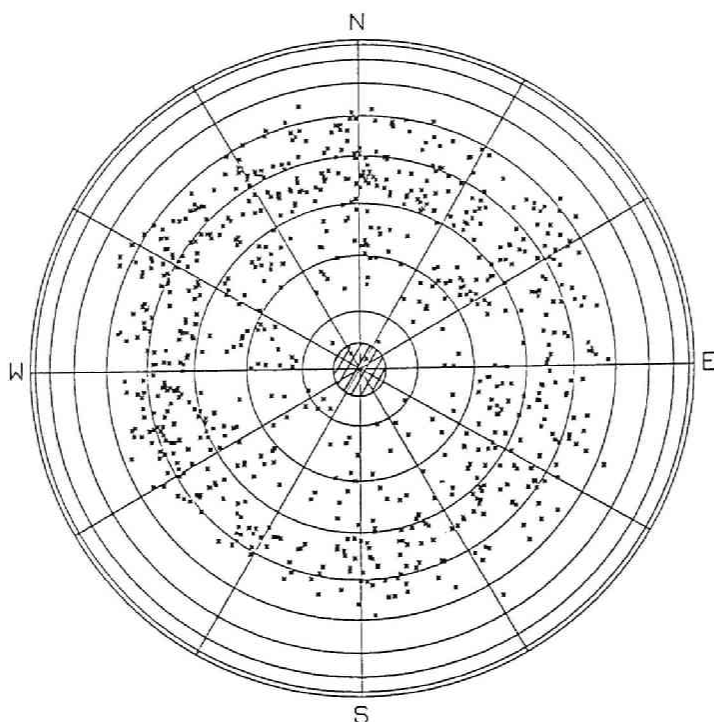


Figure 2.10: The arrival angle distribution of meteor echoes observed on 10–11 December, 1989, where a concentric circle is plotted as $\sin(\text{zenith angle})$ for every 10° . The hatched area corresponds to the width of the main lobe of the transmitting antenna with 21 subarrays [Nakamura et al., 1991a].

are numbered 1 to 4, are connected to individual receiver channel, composing a interferometer to determine echo arrival directions.

It is noteworthy that since the shortest baseline ($0.697 \lambda \times \cos 30^\circ$) of the interferometer is longer than 0.5λ , the echo direction cannot be uniquely determined. That is, the echoes with zenith angle larger than 56.5° can be contaminated by the grating lobe of interferometer. This ambiguity, however, can be removed by assuming that meteor echoes appears at the altitude higher than 70km. By limiting the maximum sample range to 127km ($=70\text{km}/\cos 56.5^\circ$) all the meteor echoes should be received by the first lobe of the interferometer. Data taking software in VAX-11 is also specially adjusted to detect meteor echoes quickly in real time and store only the data of ranges with significant meteor echoes. By these set up of the MU radar, 3000–4000 meteor echoes are

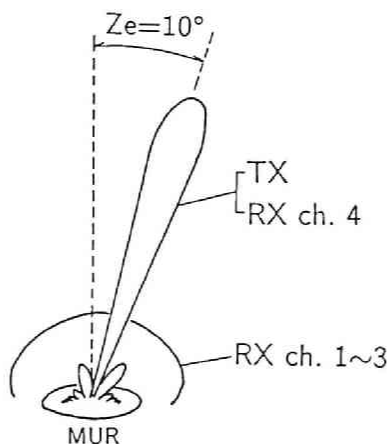


Figure 2.11: Schematic diagrams of the antenna patterns to determine the meteor echo distribution with oblique pencil-like beam (see test) [Nakamura et al., 1991a].

received per a day, which is more than 10 times as many echoes as those by the interferometer system of Kyoto meteor radar, which caused the better time and height resolution in determining the wind velocity. Figure 2.10 shows an example of the distribution of the arrival direction of meteor echo observed with the MU radar utilizing a vertical beam for transmission. The meteor echoes exist up to 50° in zenith angle.

We next describe the validity of the assumption that meteor echoes are detected in the sharp main lobe in MST radar measurement. For the usual setup of MST radar observation, both transmitting and receiving antenna was narrow pencil-beam antenna, so the number of meteor echoes received in the sidelobes is expected to be smaller than the experiment shown in Figure 2.10.

An experiment were carried out for 14 hours on 16–17 March, 1990, using the obliquely pointed pencil-like antenna shown in Figure 2.11, where the zenith angle of the main lobe was set at 10° . Receiver channel 4, which is connected to the pencil-like beam, detected 2027 meteor echoes, out of which 956 were simultaneously received by the other three interferometer channels.

The left panel in Figure 2.12 shows the height distribution of the 956 echoes, where the height was calculated by multiplying the range with the cosine of the zenith angle determined with the interferometer. Meteor echoes are mainly distributed between 80 and 100 km, which is similar to a Gaussian distribution with a mean height and standard deviation of 88 km and 6.6 km, respectively. This agrees well with the meteor echo distribution by Kyoto meteor radar [Aso

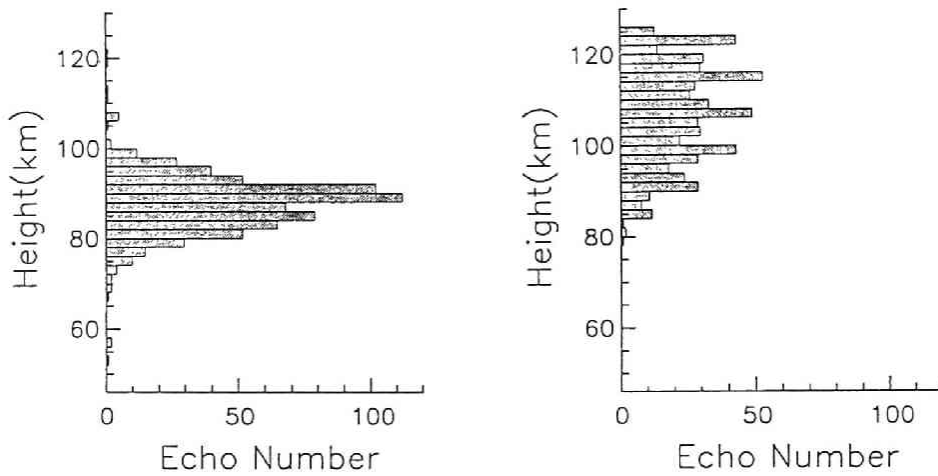


Figure 2.12: The height distribution of the meteor echoes observed on 16–17 March, 1990. The left panel shows a histogram corresponding to the distribution determined from the interferometer measurements, while the right panel shows the height calculated by assuming that a meteor echo is received in the direction of the main lobe of the transmitting antenna [Nakamura et al., 1991a].

et al., 1980] while the distribution is shifted by about 6 km for Kyoto meteor radar because of the lower radar frequency (31.57MHz) than that of the MU radar (46.5MHz).

While, assuming that the zenith angle of a meteor echo is the same as the main lobe direction, i.e., 10° , another height distribution is plotted in the right panel of Figure 2.12 for the meteor echoes detected by channel 4. This distribution extends up to 120 km without a clear peak. The most probable explanation for the big difference between the two height distributions is that a major part of the meteor echoes observed by the pencil-like beam channel were detected by the side lobes of the pencil-like beam.

We here present another test for the height distribution of meteor echoes for the data obtained by the present experiment. From the Eq. (2.9), the ambipolar diffusion coefficient D is related to a decay constant of the meteor echo intensity, T , by an equation

$$D = \frac{\lambda^2}{32\pi^2 T} \quad (2.10)$$

where λ is the wavelength of the radio wave [Herlofson, 1948]. By using this relation, a decay height h_D is defined as

$$h_D = C * \ln D + h_0 \quad (2.11)$$

where C and h_0 are constants [Greenhow and Neufeld, 1955], which are here estimated to be $C = 11.533$ and $h_0 = 76.916\text{km}$ from the meteor echoes detected by the interferometer.

Figure 2.13 (a) shows a correlation of the decay height versus the measured heights by means of the interferometer, h_I , in which data points are distributed near a line, $h_D = h_I$, with a standard deviation of $\sigma = 5.508\text{km}$. The number of echoes within 2σ of the distribution becomes 889 (93%). We here separate the distribution into four areas: area I where the discrepancy, $h_D - h_I$, is larger than $+2\sigma$, area II within $+2\sigma$, area III within -2σ and IV larger than -2σ .

Figure 2.13 (b) is a similar distribution for the echoes received by channel 4 using a pencil-like beam, where the height, h_M , is measured by a relation $R \cos \theta$, where R and θ are range and zenith angle of the main beam, respectively. Because the h_D is not affected by the method of the height delineation, a linear relation found in Figure 2.13 (a) could also be expected in Figure 2.13, if h_M were correctly determined. More than half of the echoes are shown in the area IV, which suggests that h_M for these echoes are significantly higher than the actual height, i.e. h_I . Since the numbers of echoes are uneven between the areas II and III, the echoes in area III seem to also be contaminated by the erroneous height determination.

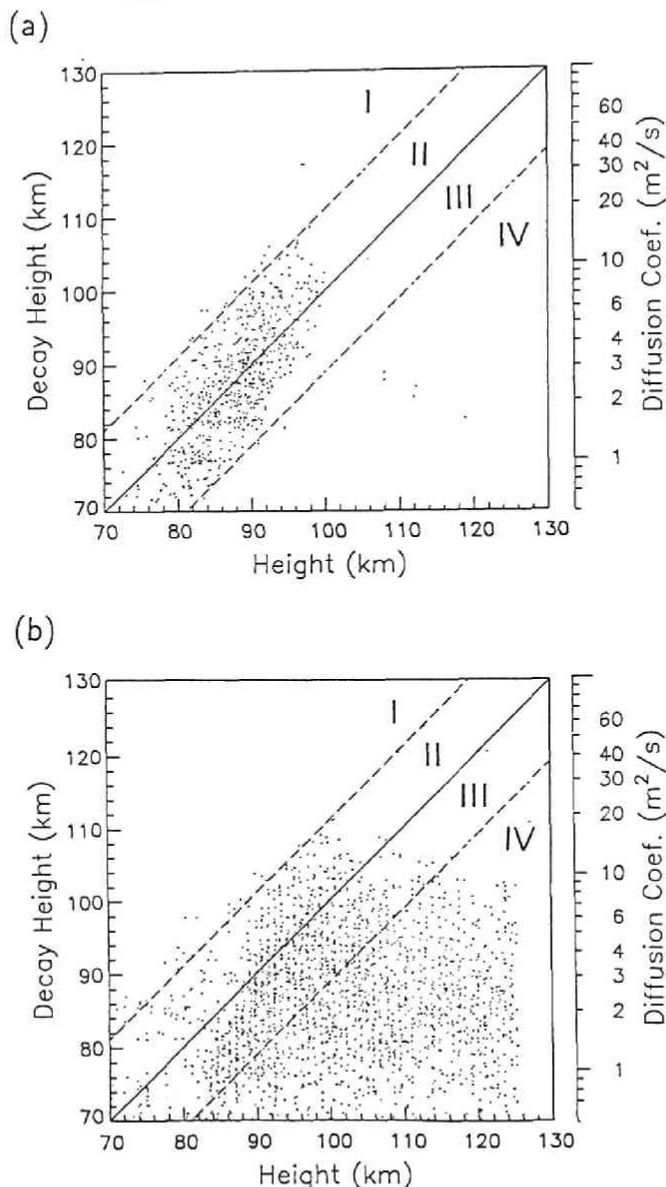


Figure 2.13: Scatter diagrams between the decay height, h_D , and the measured heights, h_I and h_M , by using two different techniques: (a) h_D vs h_I and (b) h_D vs h_M (see text). Dashed lines indicate a range of 2σ , where σ is the standard deviation of the distribution in (a). Numbers of echoes in areas I, II, III and IV are; 35, 436, 453, 32 in (a) and 50, 244, 610, 1123 in (b), respectively [Nakamura et al., 1991a].

Parameter	MU/MST observation	MU/Meteor observation
Height	60–90 km	80–100 km
Time Resolution	1 min	30 min
Height Resolution	0.6km	2km
Observation duration	Daytime	24 hours

Table 2.3: Comparison of the parameters of mesosphere and lower thermosphere observation by MST mode and meteor mode of the MU radar.

It seems appropriate to assume that the number of echoes in the area II corresponds to a half of that detected near the main lobe direction. As a result, we can estimate the number of the meteor echoes with $h_M \simeq h_I$ is 488, which is only 24 % of the total echoes received by the pencil-like beam. Moreover, we still cannot discriminate meteor echoes detected by near-axis side-lobes with similar zenith angles, the number of echoes received by the main lobe could become less than 1/4 of all the echoes. As a conclusion, it is not adequate to assume that majority of meteor echoes received by an MST radar in the VHF band are located within the narrow main lobe of an antenna.

When the zenith angle is scaled by the inverse of the antenna diameter, the side lobe patterns of, for example, square or circular antennas are the same regardless of their diameters. So that even if the antenna diameter becomes larger, the side lobes near the main lobe retain enough sensitivity to detect meteor echoes. Moreover, the narrower beam width of the main lobe of a larger aperture gives a smaller illuminating volume, which makes it difficult to increase the number of meteor echoes detected by the main lobe. So the use of an antenna with a large aperture is not effective as to improvement of the ratio of the meteor echo rate with the main lobe to that with side lobes.

Thus, it is essentially important to use an interferometer to determine the arrival angles of meteor echoes. This is also true not only for the MST radars but also for the observation by meteor radar. So that, in the present thesis all the wind data obtained by Kyoto meteor radar also determined using interferometry system.

The comparison between normal MST observation and meteor echo observation are shown in Table 2.3. Although the time and height resolution is better for MST observation, the potential of the continuous observation is appropriate to observe long period waves such as tides and planetary waves. Figure 2.14 shows an example of the comparison of the wind measurement by the turbulent scatter and meteor wind by the alternating observations with the MU radar.

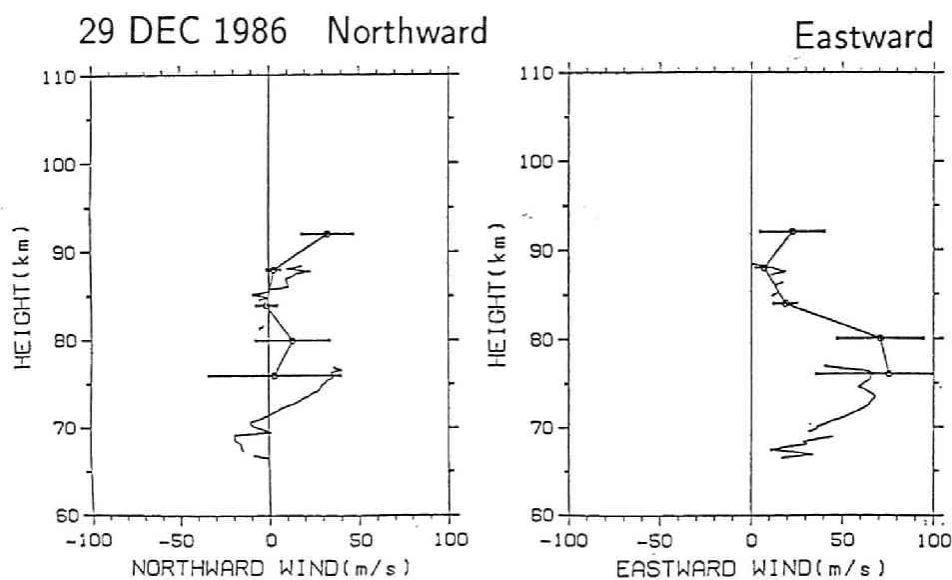


Figure 2.14: Mean wind profiles derived from the MU radar observations using meteor echoes (dots with an error bar) at 0–11 LT and turbulence echoes at 9–10 LT. The results on 29 December, 1986 are plotted [Nakamura et al., 1991a].

It is recognized that two observations agree well suggesting the meteor wind measurement by our interferometry system is working correctly.

2.5 MF Radar Observations

Weak partial reflections are observed from the 60–110 km by MF/HF radar at 2–6 MHz. Initially, these echoes were used for observing electron density and its irregularity, which was first carried out by Gardner and Pawsey [1953]. These partial reflection echoes were applied to measure the mesospheric winds using spaced antenna (SA) method at Christchurch, N.Z. [Fraser, 1965, 1968]. Later several MF radar measurements were carried out, and improvement of the technique and system has been made. Now about 10 MF radars are operating or under construction. Typical MF radars operate at the frequency of the 2–3 MHz with peak power of the order of 50 kW [Vincent, 1984b]. Their typical pulse length is 20–30 μ s, resulting the height resolution of 3–4 km.

Adelaide MF radar is situated at 35°S, 138°E using the frequency of 1.98 MHz. The characteristics of their system are the large circular array antenna with 1km diameter [Briggs et al., 1969], and therefore, the Doppler measurement of the wind velocity by pencil beams are capable as well as spaced antenna measurement [Vincent and Reid, 1983]. The antenna configuration is shown in Figure 2.15. Although in the same location antennas of 6MHz and 50MHz system are situated, these are not utilized in this thesis. The transmitting antenna consists of four folded dipole giving a beam width of about 40° half-width half-maximum. Peak power and pulse width are 25 kW and 25 μ s, respectively. Typical IPP is 50 ms. The height resolution, therefore, is about 4 km, while the data are oversampled at every 2 km.

The returned signal is received by the large array as shown in Figure 2.15, which consists of 2 sets of 89 half-wave dipoles aligned north-south and east-west. For the usual spaced antenna measurement, a part of the whole array is used (Figure 2.16), that is, four dipoles are connected together to form one antenna and then three sets are used for spaced antenna analysis. For Doppler measurement, the whole antenna array is at first divided into 11 rows and in each row signals are combined to one. Then the combined signal is divided into 2 channels equally, finally the output of these rows are added with some phase delay to compose an off-vertical beam. Consequently, 2 beams of 2 channel are formed, which are utilized for momentum flux measurement of the gravity waves [Vincent and Reid, 1983].

Saskatoon MF radar is located at 52°N, 107°W. Operating Frequency is 2.219MHz. The antenna configuration of Saskatoon MF radar is shown in Figure 2.17. The transmitting antenna is a vertical pointing 4 $\times$ 4 array of folded

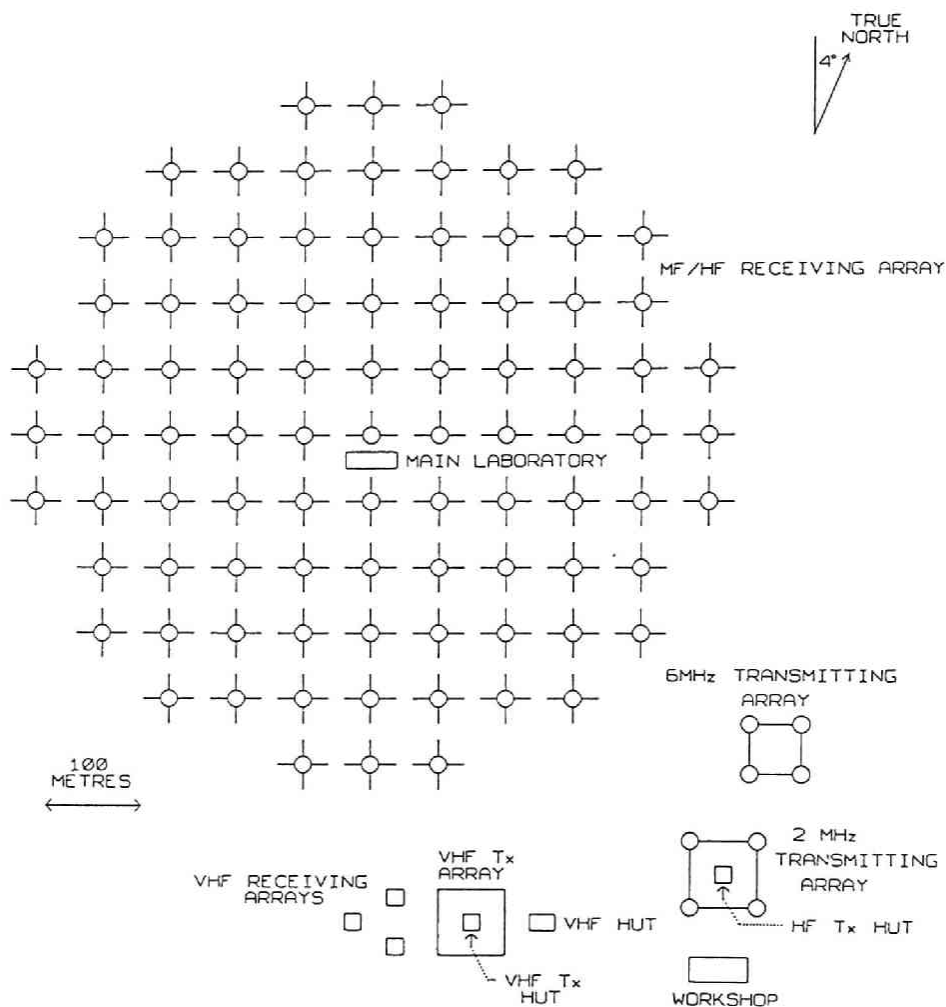


Figure 2.15: Antenna configuration of the Adelaide MF radar [Murphy, 1990].

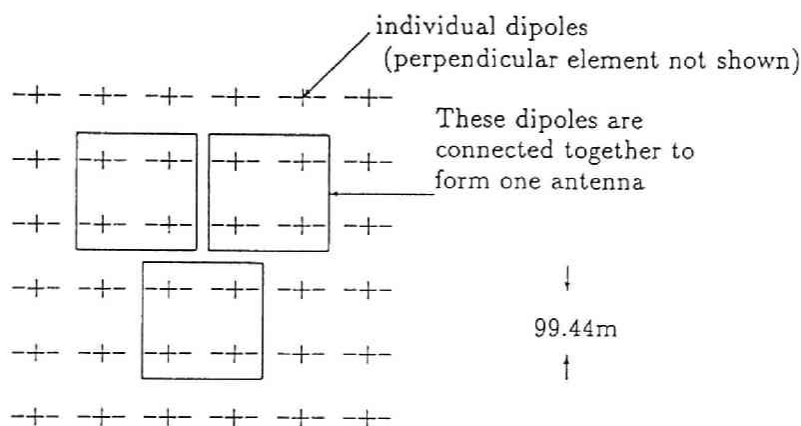


Figure 2.16: The antenna configuration used for spaced antenna measurement at Adelaide MF radar [Murphy, 1990].

half-wave dipoles with beam pattern of half power width of 26° . Transmitting power is 50 kW at peak, and pulse width is $20 \mu\text{s}$ corresponding to the height resolution of 3km [Meek and Manson, 1990]. IPP is 16.7 ms. The receiving antenna is composed by four half-wave dipoles, and three antennas of this kind are separated by 2λ as to form a triangle. The receiver system was improved in 1988, which is called COHRTW (Coherent Real Time Wind) system. The schematic diagram of COHRTW is shown in Figure 2.18. The received signals are coherently integrated, resulting the improvement in S/N and reduction of radio interference [Meek et al., 1989].

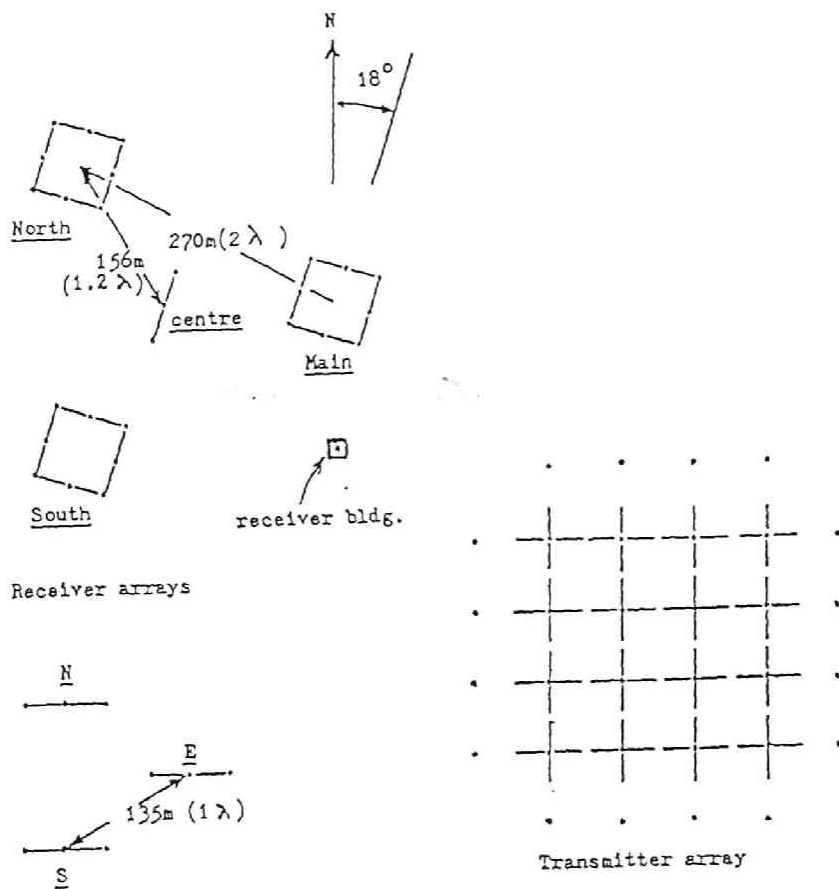


Figure 2.17: Antenna configuration of Saskatoon MF radar [Meek et al., 1989]

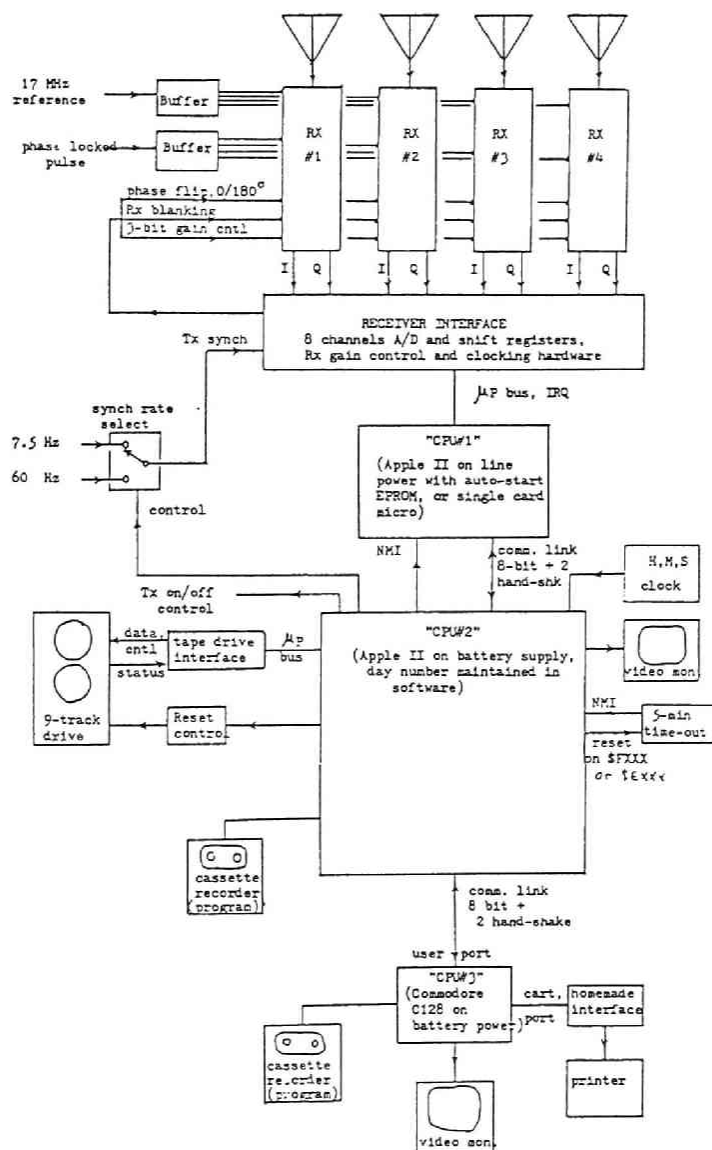


Figure 2.18: Schematic diagram of COHRTW system of Saskatoon MF radar.

Chapter 3

Mean Winds

This chapter is concerned with the seasonal and height variations of the mean winds at 35° latitude. We present the mean winds observed at Shigaraki (35°N , 136°E) by the Kyoto meteor radar (80–110km) and the MU radar (60–90km), and compare them with CIRA model atmosphere. The results are further compared with Adelaide MF radar observations (35°S , 138°E) at 60–100km to discuss the hemispheric difference.

It is important to describe characteristics of mean winds in order to understand interactions of the mean winds with various atmospheric waves, and study the dynamical system of the middle atmosphere together with the underlying physical mechanism. However, the observations of the upper middle atmosphere, i.e., the mesosphere and lower thermosphere are not enough to expose the whole behavior of the mean winds. One of the reasons of the lack of observation is the difficulty of the satellite observation in the upper middle atmosphere. Although the satellite observation can provide the global data of the temperature, the vertical resolution is insufficient, and furthermore, the geostrophic wind calculated from the temperature fields are different from the real wind fields because of the drag of the atmospheric waves above 60 km altitude [Barnett and Corney, 1985]. Therefore, the measurement of the mean wind by radar observations is important to clarify the general circulation of these regions.

3.1 Observation Techniques

Figure 3.1 shows the observation periods of the Kyoto meteor radar and the MU radar used in the present chapter. The Kyoto meteor radar continuously monitored the wind fields in 80–110 km region from May 1983 to March 1986. However, data were sometimes not collected for about 10 days mainly because of the malfunction of a transmitter [Tsuda, et al., 1987]. Antennas were pointed

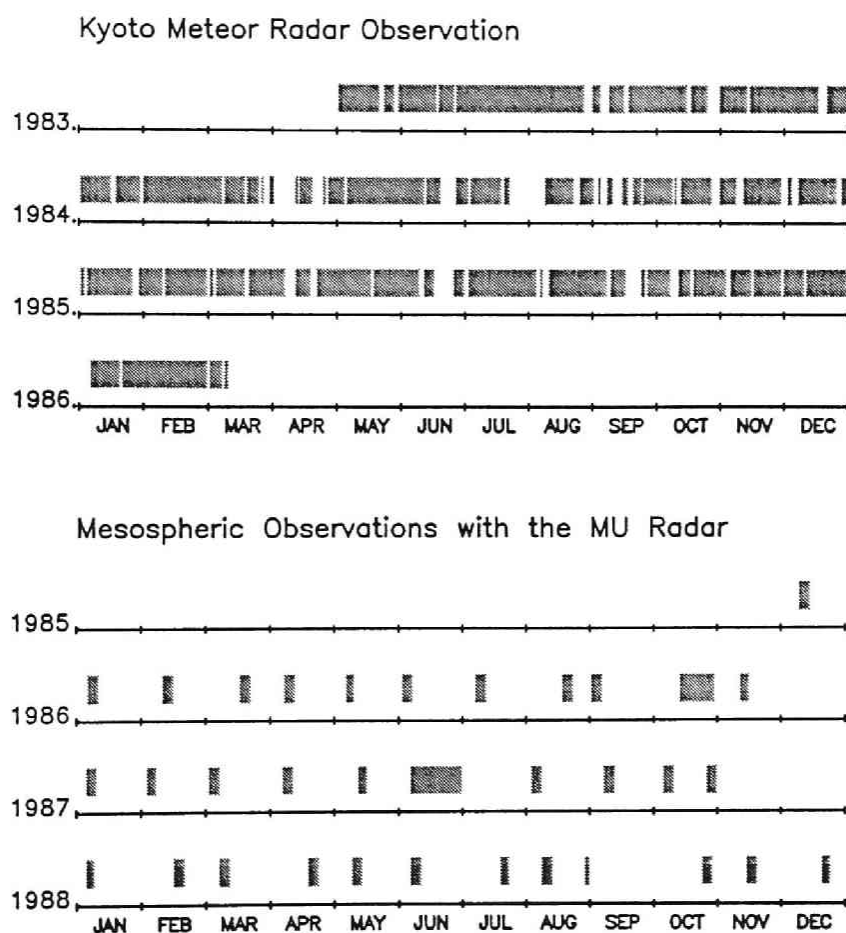


Figure 3.1: Observation duration of Kyoto meteor radar in 1983–1986 (top) and the MU radar (bottom) in 1985–1988.

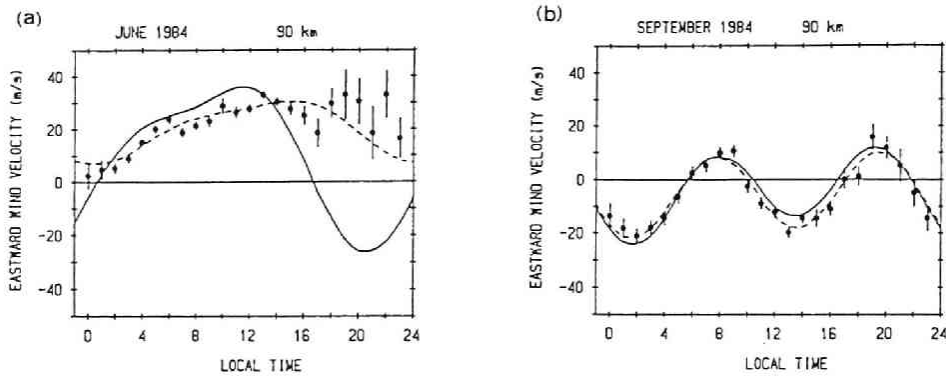


Figure 3.2: Eastward component of wind fields observed at 88–92 km altitude in (a) June and (b) September 1984 [Tsuda et al., 1987]. A circle corresponds to wind velocity fitted by least squares in a $1 \text{ h} \times 4 \text{ km}$ bin. Solid and broken lines correspond to determinations by old and new harmonic analysis, respectively (see text).

to the eastward direction throughout the observation period as described in Chapter 2.

Since the meteor echo rate is high in the morning and low in the afternoon, simple average of the measured wind velocities does not always give the realistic mean wind but often biased to some extent, which is mainly due to the diurnal tidal oscillation. Therefore, a harmonic analysis is adopted for determining mean winds.

For each observation period meteor echoes detected at the same local time of a day are accumulated and collected into a time-height bin of $1 \text{ h} \times 4 \text{ km}$. Thirty days and ten days of observational data are accumulated and utilized to determine the mean wind velocities in order to get the seasonal variation of mean winds and short-term variations during stratospheric sudden warmings, respectively. Zonal and meridional wind velocities are assumed to be constant in each bin, neglecting the vertical wind component. Radial wind velocities obtained by meteor echoes observed in a bin are used to determine zonal and meridional wind components by applying the least square fit. Thus, we get wind fields at several heights as function of local time, which can further be fitted by mean wind, semidiurnal and diurnal sinusoidal components at the same time. Note that the statistical error in zonal wind determination is smaller than that in meridional wind, because the antennas are pointed to the eastward direction [Tsuda et al., 1987].

Groves [1959] presented the data analysis algorithm for tides and mean winds

of meteor wind observation, which was widely applied to meteor radar observation [e.g., Aso et al., 1980]. From a meteor echo we can determine the arrival direction of radio echo, radial wind velocity, range of the reflection point and observational time. The radial wind velocity, V_r , is expressed as

$$V_r = ul + vm + wn \quad (3.1)$$

where u , v and w are eastward, northward and vertical component of the wind, and (l, m, n) is the directional vector of the arrival angle.

We can neglect the w component because the periods of the wind fluctuation of our interest by the meteor observation is usually longer than 1–2 hours and w component of these frequencies is small. Then, Eq. (3.1) became as

$$V_r = ul + vm \quad (3.2)$$

Then the wind fields u and v are assumed to be composed of the mean wind, and the harmonics of diurnal tides, which are further described by a function of height z . For example, u can be written as,

$$\begin{aligned} u = & u_0(z) + u_1(z) \sin(\omega t) + u_2(z) \sin(2\omega t) + \dots + u_p \sin(p\omega t) \\ & + u_1^*(z) \cos(\omega t) + u_2^*(z) \cos(2\omega t) + \dots + u_p^* \cos(p\omega t) \end{aligned} \quad (3.3)$$

where ω is the angular velocity of the diurnal oscillation. Each coefficient such as $u_0(z)$, $u_1(z)$... are expressed by polynomials. By substituting Eq. (3.3) for u and the corresponding equation for v , an equation between the quantities measured by meteor observation and the wind parameters can be obtained. By the least square method for minimizing the summation of the squared error of V_r for meteor echoes collected in the observation periods, wind parameters such as mean wind, diurnal tide, semidiurnal tide... are determined.

It is noteworthy that when the Groves' algorithm is applied to the data analysis of Kyoto meteor radar, wind velocity obtained by each echo was weighted equally in the determination of wind fields.

Tsuda et al. [1987] compared the Groves' analysis with the harmonic analysis, and found that the Groves' analysis sometimes gives misleading results. The comparison of two methods are shown in Figure 3.2. Note that the present analysis of Groves' algorithm is its particular case, in that we separated all the echoes into 4 km height bin and in each bin wind parameters are assumed to be invariant with height. Hereafter, we call it as the earlier analysis. Figure 3.2 (a) and (b) show examples of fitting of eastward wind component in June and September 1984. Error bars of hourly wind become large in the evening in June, while smaller in the evening in September. This is because meteor echo number

in the evening become small, but in September the echo number in the evening is larger than those in June due to larger echo rate in September [Kato and Tsuda, 1987]. Since in the earlier analysis each echo has the same weight, the wind data in the morning, where echo rate is large, are fitted better than that in the afternoon. Consequently, the earlier analysis gives a fitted curve with large discrepancy from the hourly winds in the afternoon. However the new analysis provides a reasonable fit to the hourly wind throughout a day. Because the error in least square fit is not only due to the random error in each wind measurement, but also due to the wind fluctuation other than diurnal and semidiurnal tides. The latter factor is not reduced by increase of meteor echoes. Therefore, simple weighting by the number of echo is not appropriate in determining mean wind and tidal component.

In case of Figure 3.2 (b), two analyses show similar results. This is because semidiurnal tide is so dominant in September that the wind fields in the morning could represent the total daily variation. Therefore, in this thesis we utilize the harmonic analysis by Tsuda et al. [1987] for the Kyoto meteor radar observations in determining the mean wind and tides.

The MU radar observations of the mesosphere (60–90km) in December 1985–December 1988 are also used to determine the mean wind at this height region, which are slightly lower than that of Kyoto meteor radar. It is noteworthy that the MST radar observation of the mesosphere is limited during the daylight hours because of the low electron density at night. Thus, the mesospheric observation with the MU radar in 1985–1988 was basically carried out from 8LT to 16LT for 8 hours, which means that the correct estimation of mean wind considering tidal effect is difficult for the MST radar observation. In the present study, we calculate the simple average of the wind field for the whole observation hours in each day. Although the estimated mean winds should be biased by the fluctuations which are synchronous to the local time such as atmospheric tides. This kind of bias should be smaller at the height of the MU radar observation than at the meteor height by Kyoto meteor radar, because the amplitude of atmospheric tide is smaller in the lower height [Vincent et al., 1988].

Figure 3.3 shows the difference of the daytime average (8LT–16LT) of the wind and mean wind deduced by harmonic analysis discussed above estimated from the diurnal and semidiurnal wind fluctuations determined by Kyoto meteor radar observation, which will be discussed as atmospheric tides in the later chapter. For eastward component, the difference is generally smaller than 5 m/s below 90 km, except for in May–June and above 80 km in early August. In total, daytime average at 80–90km gives larger eastward wind by as large as 5 m/s in summer, while the difference is negligible in the other seasons. The meridional component shows that below 85 km the discrepancy is smaller than 5 m/s.

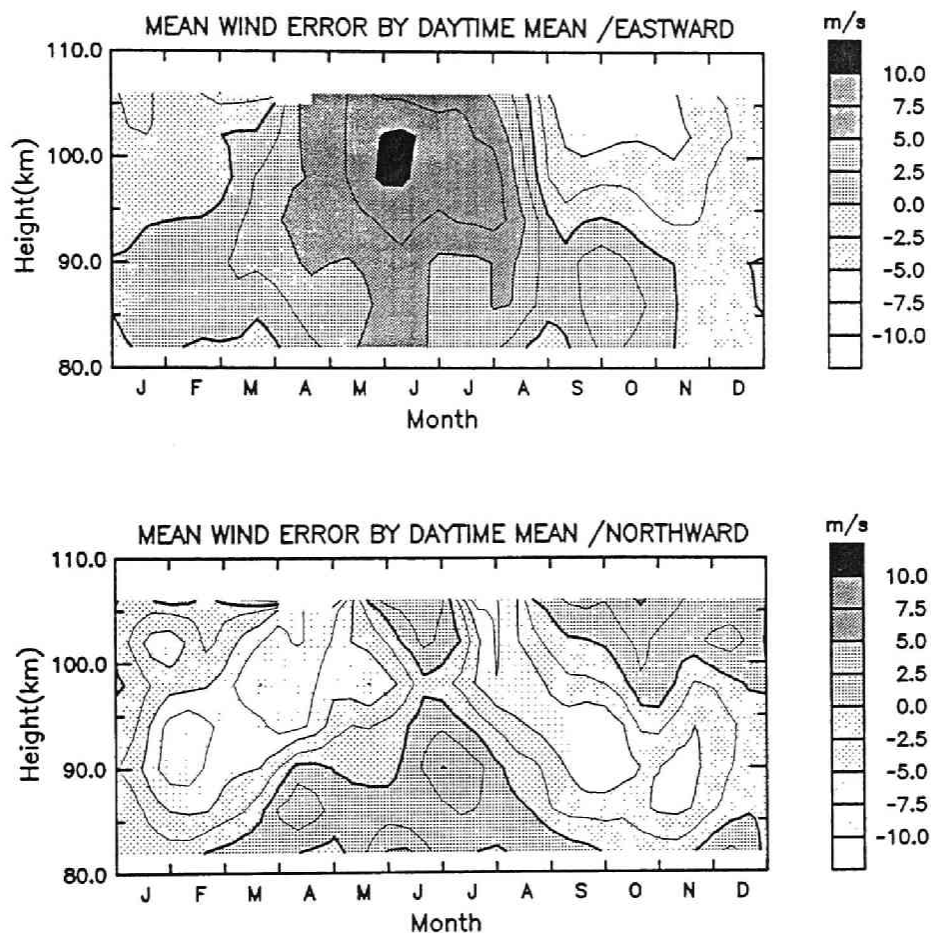


Figure 3.3: The estimation of the difference between simple daytime averaged wind (8LT–16LT) and the mean wind deduced by harmonic analysis of mean wind and diurnal and semidiurnal tides by the data of Kyoto meteor radar observation in 1983–1986. The top and bottom panels correspond to mean zonal and meridional winds, respectively. Positive values correspond to the eastward and northward biases generated by the day-time average.

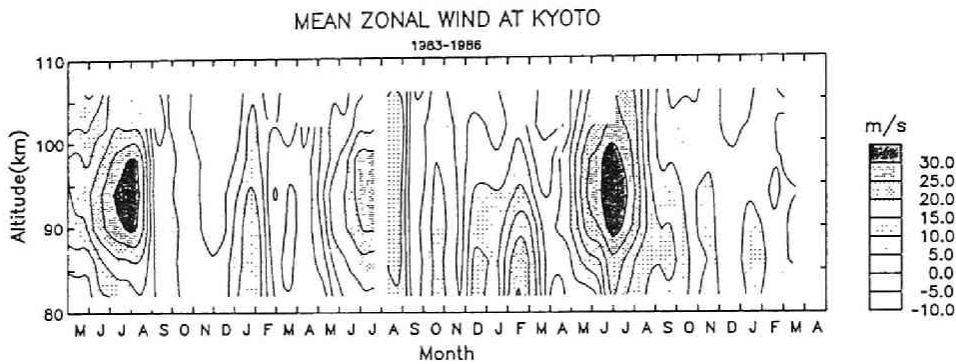


Figure 3.4: Eastward wind component of mean winds. Variations shorter than 25 days are filtered out. Shaded area corresponds to the eastward component.

These differences become significant only above 100 km for both components. Thus, the simple average of daytime gives a good estimation of mean wind for all the year below 85 km in northward component and below 90 km for eastward component. The mean winds are available by the MU radar observation without significant biases as discussed above.

Another comment for the mean winds observed with the MU radar is that the observation is limited to successive 4 or 5 days per a month except for two campaigns of October 1986 and June 1987. As shown in Chapter 5 later, the mean wind at 65–80 km height shows the day to day variation of more than 10 m/s in amplitude. The variation also shows a periodic change with period longer than several days which can be interpreted as a planetary wave. Therefore, mean winds deduced by averaging for successive 4 or 5 days do not always represent the seasonal variation of the mean wind of a single year. Thus, in this study we averaged over the data collected with the MU radar in each month during the three years from December 1985 to December 1988 in order to reduce such variability, and deducing one year cycle variation, smoothing by the 2-dimensional low pass filter with the cut-offs of 25 days in time and 5 km in height.

3.2 Mean Winds at Shigaraki (35°N)

In this section, we present the mean wind observed with Kyoto meteor radar and the MU radar. Mean zonal and meridional winds observed with Kyoto meteor radar are shown in Figures 3.4 and 3.5, respectively, which are basically

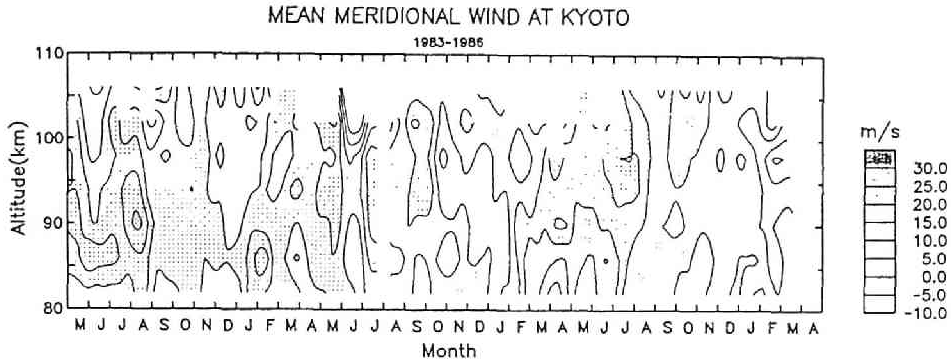


Figure 3.5: Northward wind component of mean winds. Variations shorter than 25 days are filtered out. Shaded area corresponds to the southward component.

the same as shown in Tsuda et al. [1987] except that the observation period was extended to March 1986. Small scale variation with period less than 25 days was filtered out. The eastward component of mean winds was enhanced in summer at around 95 km with peak value of more than 30 m/s in July. The amplitude began to increase in May and quickly decreased in the late August and the beginning of September. The amplitudes of the maximum eastward wind were 30–35 m/s in 1983 and 1985, while they were slightly smaller as not more than 30 m/s in 1984, because observations were lacking in the late July and early August when eastward winds are normally maximized.

In January and February eastward winds with amplitudes of 10–20 m/s was found below 95 km. The strong eastward wind in winter stratosphere and lower mesosphere came up into meteor height significantly. The interannual variation in winter seems larger than that in summer, which could be attributed to the effect of stratospheric sudden warmings, as discussed later. Eastward wind below 95 km in winter emerged from late December 1983 to the middle of February 1984 with the peak amplitude of about 10 m/s, while in the next year it began in late November 1984 and lasted till March 1985 with the peak amplitude exceeding 25 m/s. In the next winter of 1985 to 1986, eastward wind was weaker than in the previous two winters, and did not exceed 10 m/s for all the winter duration. In autumn and spring the mean zonal winds were small in the whole observed altitude range, as well as high altitude in winter.

The mean meridional winds shown in Figure 3.5 were smaller than zonal mean winds and were generally southward i.e. equatorward, with amplitude occasionally exceeding 10 m/s in summer. The center of the southward wind in summer was located at 90–95 km altitude in July–August. The southward wind

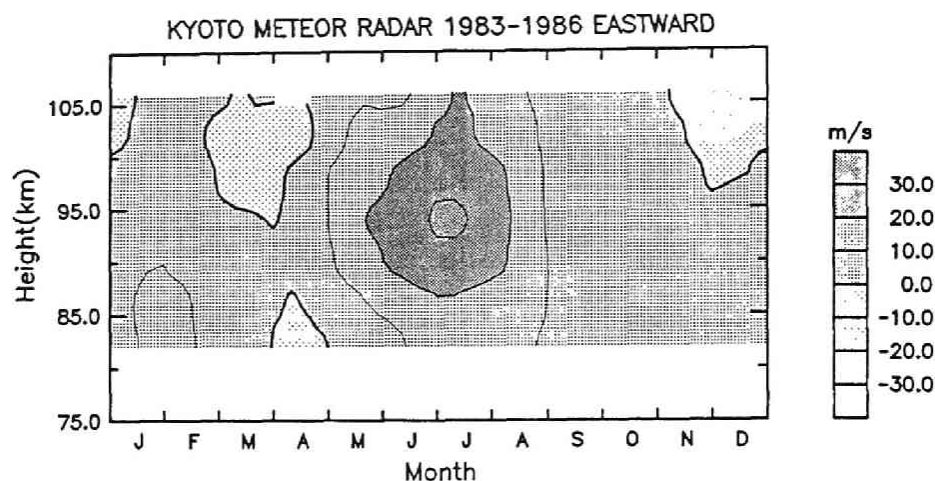


Figure 3.6: Annual variation of the eastward wind component of mean winds observed by Kyoto meteor radar in 1983–1986. Shaded area corresponds to the eastward component, while the thick solid lines show the zero wind lines.

in summer at 90 km altitude was consistent with the general characteristics of meridional wind in the mid- and high latitude of northern hemisphere in June [Nastrom et al., 1982]. However, northward mean wind were detected in May–June 1983 and June 1984 at altitudes higher than 97 km. In November–January the mean meridional wind became weakly northward in the major part of the observed region.

Figures 3.6 and 3.7 are annual variation of mean zonal and meridional winds, which are deduced by the Kyoto meteor radar observations shown in Figures 3.4 and 3.5 by averaging the mean winds of the same date of the year together. Although fluctuations with small time scales were filtered out and the amplitude of the variation become smaller by averaging, the main feature described with regards to Figures 3.4 and 3.5 is still recognized well. The eastward mean wind attained its maximum in July at 95 km with the amplitude exceeding 30 m/s. Eastward wind with the amplitude over 10 m/s was also found in the middle of January to the middle of February below 90 km. Furthermore, the westward wind in March and April above 100 km and below 87 km, and in November and December above 100 km were clearly shown in Figure. 3.6 by the elimination of the small scale variation. In brief, the zonal wind between 80–110 km was generally eastward.

The meridional winds shown in Figure. 3.7 show the summer southward wind

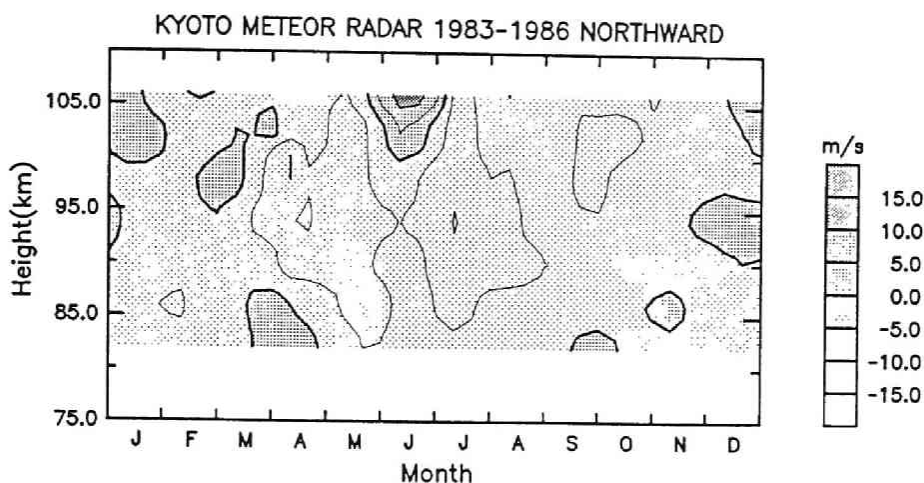


Figure 3.7: The same as Figure. 3.6 except for northward wind component. Shaded area corresponds to the northward component.

over 5 m/s from April to May and June to August at around 90 km altitude. The wind direction was generally southward, while weak northward wind with the amplitude less than 5 m/s were found in April below 87km, September below 85km, December and January above 90km etc. Above 100km in June the northward wind over 5 m/s occurred.

The mean zonal and meridional winds at 60–90 km altitude observed with the MU radar are shown in Figures 3.8 and 3.9. The daytime averaged wind of the observation in December 1985 to December 1988 are collected and filtered with the cut-offs of 5km in height and 25 days in time, respectively.

The mean zonal wind shows clear annual variation. In summer, from middle of April to the end of August the zonal mean wind was westward below 80 km. It reached maximum in June at 65–70 km altitude with the peak amplitude of about 40 m/s. Above about 85 km in summer the westward wind turned the direction into eastward. From September to the middle of April, the eastward winds were found for the whole time and altitudes generally, except for at 80 km in late September. The maximum amplitude of the eastward wind occurred in the end of November below 70 km with maximum amplitude up to 60 m/s. Compared with summer westward wind, eastward wind in winter shows large variability. The strong eastward winds exceeding 40 m/s emerged from the middle of November to the middle of December at the height below 78 km, while the next region of eastward wind was found in the middle to the end of January at above 70 km. Furthermore, in February eastward wind over 40 m/s

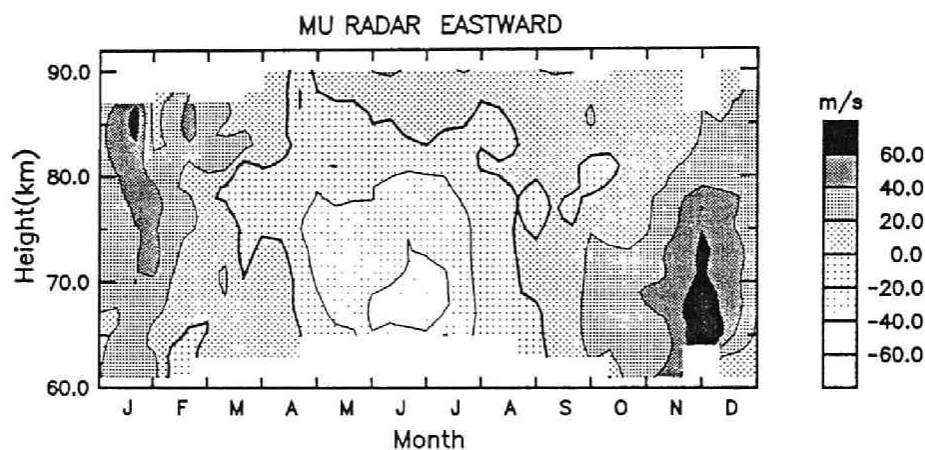


Figure 3.8: Eastward wind component of mean winds by the MU radar in 1985 December–1988 December. Variations shorter than 25 days and 5km are filtered out. Shaded area corresponds to the eastward component.

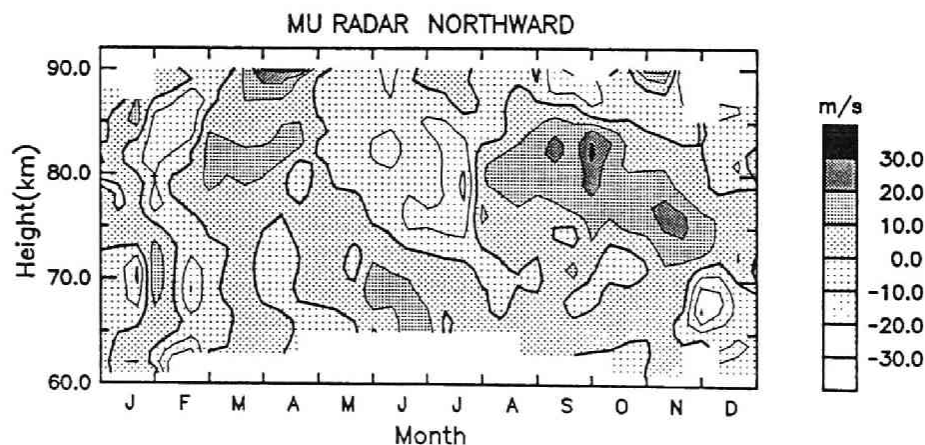


Figure 3.9: The same as Figure. 3.8 except for northward wind component of mean winds. Shaded area corresponds to the northward component.

occurred above 83 km altitude. The complicated structure in winter could be due to the stratospheric warming which was not well filtered out by the averaging for three years.

In Figure 3.9 mean meridional wind shows more complicated variation. Note that the interval of contour level is smaller in Figure 3.9 than in Figure 3.8 because of small amplitude of meridional wind component. From May to December meridional winds above 80–85 km altitude were generally southward. On the other hand, the major part of the other heights and months shows northward winds, except for the winter when northward and southward wind occurred alternatively with period of 1–2 months from late December to March, suggesting the effects of sudden warming or planetary wave in winter not filtered well by the average of 3 years observations.

It is noteworthy that the wind velocity sometimes exceeds 20 m/s; the northward wind in late September at around 80 km, the middle November at about 75 km, and early April at near 90km, and the southward wind in the end of November at 68 km altitude. Tsuda et al. [1990b] showed that the averaged profile of the daily mean meridional winds for 21 days in October 1986 still showed a wave-like feature which reaches 20 m/s at its maximum amplitude with height. And he pointed out that these profiles were due to the gravity waves which were not well filtered out by the daytime average of the wind velocities. However, 3 years average profile shown in Figure 3.9 seems a systematic variation of these wavelike feature, for instance, northward wind of May to July below 75 km and northward wind from August to November between 75 and 85 km in altitude shows downward phase progression. Further investigation is necessary to reveal the details of these profiles.

Now we compare the mean winds observed by Kyoto meteor radar (80–110km) and the MU radar (60–90km). The zonal winds between 80 and 90 km of Figures. 3.8 and 3.6 showed similar seasonal variation. That is, eastward wind enhancement in summer, weak eastward wind from September to November, eastward wind again in January to February, and spring westward wind in April were observed. However, the amplitude of the wind on each season sometimes showed discrepancy. In summer, the 0 m/s line observed with the MU radar and the contour line of 10 m/s of eastward wind observed by Kyoto meteor radar are found at the same altitude in May and June, so that the eastward amplitude by meteor radar in summer was larger by about 10 m/s. On the other hand, in winter eastward wind by meteor radar was about 10 m/s while the wind by the MU radar exceed 40 m/s, although the comparison between the different years was somewhat difficult in winter for its interannual variability. The westward wind in April and the eastward wind in September to November seems to agree well.

The difference of the two observations is interpreted as below. First, the MU radar data are expected to include the effect of tides which were not averaged out by the daytime mean, as described in the previous section. The tidal effect are estimated as less than 5 m/s at 80–90 km except for eastward component in May to June and northward component above 85 km in autumn and winter, where the tidal effect becomes as large as 5–10 m/s. Next, the wind measurements at near 80 km altitude by Kyoto meteor radar and near 90 km by the MU radar are less accurate than at the central height of their observation region; 95 km by Kyoto meteor radar and 75 km by the MU radar, respectively. In the former case, meteor echo rate at low altitude is small, while in the latter case turbulent echo is weak at around 90 km.

Furthermore, the meteor radar winds are deduced for each height bin of 4 km by assuming the homogeneousness of the wind. The meteor echo number shows a Gaussian distribution in height with the mean height around 94 km [Aso et al., 1980], therefore the meteor echoes in the height bin of 80–84 km, which is the lowest height of the meteor radar, are distributed mainly at the higher altitude than the central height of 82 km. Therefore, the meteor wind at 82 km is expected to be affected by the wind above 82 km, which means the positive shift of the eastward wind observed in summer. However, those effect cannot fully explain the difference up to 10m/s between 80 and 90 km, but instead some interannual variation is considered to exist. This variability are discussed in the following section.

The meridional winds by Kyoto meteor radar and the MU radar at 80–90 km altitude also provide a similar situation. The northward wind in April and September below 85 km agree well between the two, while the amplitude is larger by the MU radar observation which sometimes exceeds 20 m/s. Generally the wind direction is southward for both observations, while the wind observed by Kyoto meteor radar seems to be biased to southward.

Thus, both wind observation of the Kyoto meteor radar and the MU radar include the uncertainty in determining the mean wind between 80–90 km. We connected both results by the weighted averaging between 80–90 km using the weighting function of

$$\begin{aligned} W_{MU}(h) &= (h - 80)/10 \quad (80 \leq h \leq 90) \\ W_{MWR}(h) &= 1 - (h - 80)/10 \quad (80 \leq h \leq 90) \end{aligned} \quad (3.4)$$

where $W_{MU}(h)$ and $W_{MWR}(h)$ denote the weighting function of the MU and Kyoto meteor radars' data, respectively, and h is the altitude in km. The resultant profiles are shown in Figures. 3.10 and 3.11.

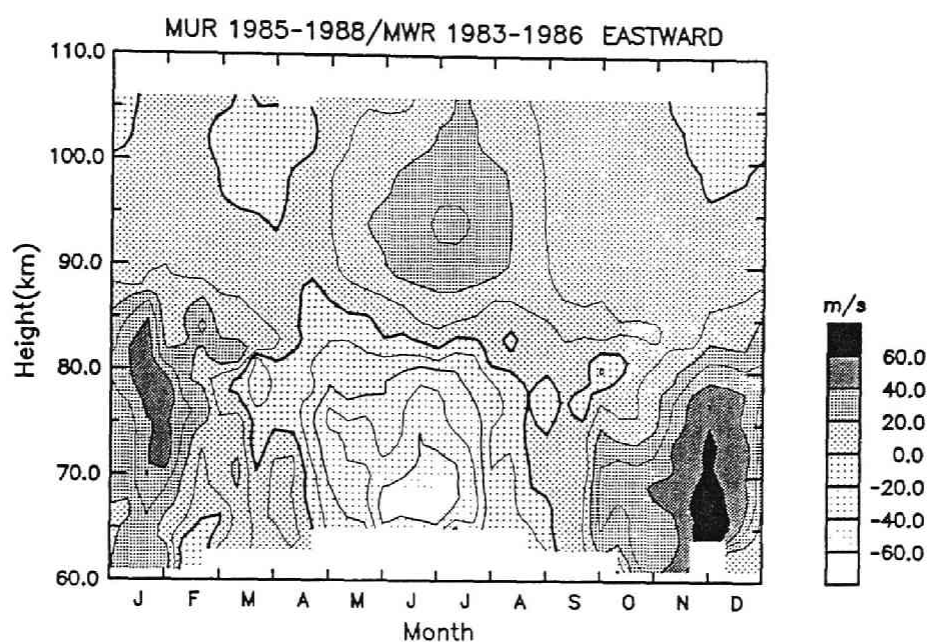


Figure 3.10: Eastward wind component over Shigaraki (35°N) by means of Kyoto meteor radar in 1983-1986 and the MU radar in 1985-1988. Shaded area corresponds to the eastward component, while the thick solid lines show the zero wind lines.

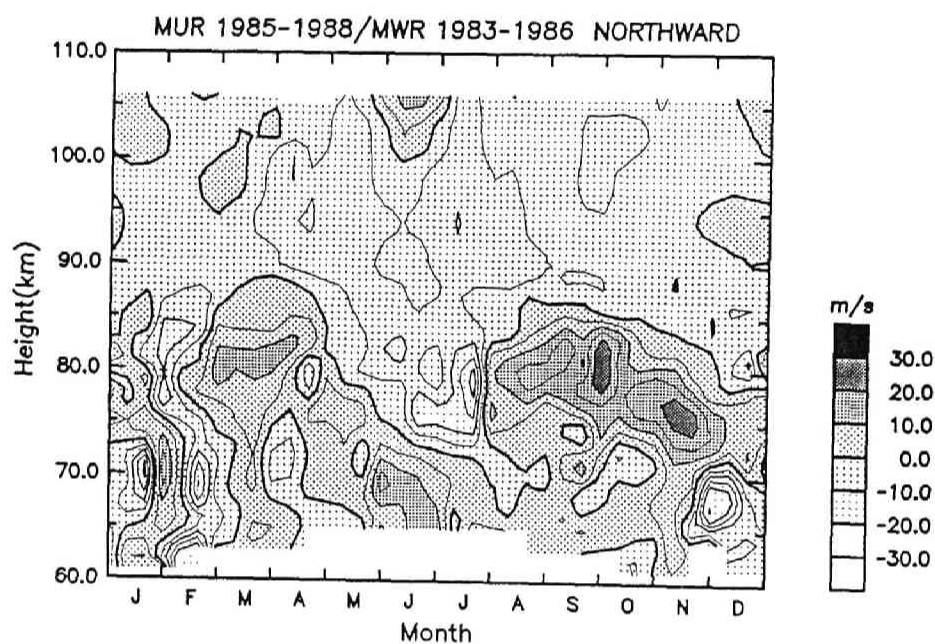


Figure 3.11: The same as Figure 3.10 except for northward wind component. Shaded area corresponds to the northward component.

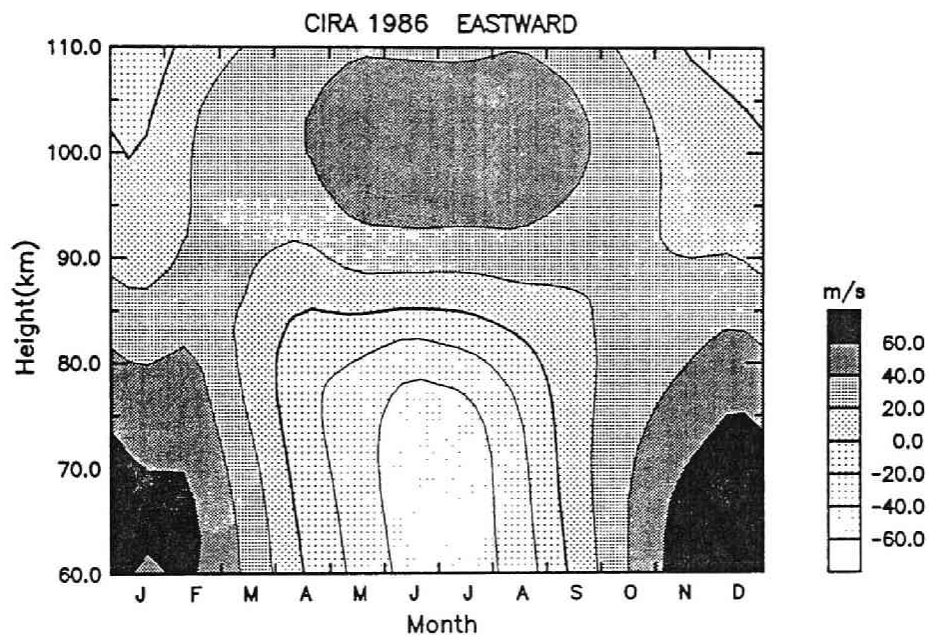


Figure 3.12: Mean zonal wind at (35°N) of CIRA 1986. Shaded area corresponds to the eastward component, while the thick solid lines show the zero wind lines.

Figure 3.12 shows the zonal mean wind at 35°N by CIRA 1986 in the same format as Figure 3.10. Now we compare this profile with the observed zonal wind shown in Figure 3.10. In summer the westward winds below about 80 km are quite similar between CIRA and the observation. The westward peak of CIRA occurs in June and July at 70 km altitude with the amplitude of 50 m/s, which agrees well with the observation while the maximum time occurs in June for the observation. On the other hand, the eastward wind in summer above 80 km in CIRA is significantly different from the observations. The eastward wind in CIRA attains the maximum in June and July at 100 km altitude with the amplitude of over 50 m/s, while it is observed that maximum eastward wind is about 30–35 m/s.

The period of strong eastward wind at around 100 km altitude in CIRA is as long as from March to October which shows the amplitude more than 20 m/s, while for the observational wind the eastward wind with the amplitude over 10 m/s is observed only from May to the end of August, and moreover in the end of August the eastward wind rapidly decrease for all the altitude range from 80 to 100 km. Thus, the CIRA model overestimates the amplitude and duration of the summer eastward wind at around mesopause.

The reversal height of the mean zonal wind, i.e., the height where the summer westward winds in the upper stratosphere and the lower mesosphere turned the direction into eastward, is a useful index to discuss the comparison of the mean wind around the mesopause [Manson et al., 1987a]. Figure 3.6 of the Kyoto meteor radar observations shows the reversal height in 1983–1985 was lower than 82 km, and the MU radar results in 1986–1988 shown in Figure 3.8 indicates the summer reversal height as 83–85 km, and then the average profile in Figure 3.10 showed it as 82–83 km. The reversal height of the eastward wind in the mesosphere into the westward direction in summer in CIRA is 83–84 km, which agree well with the observational height although it is still noteworthy that there were discrepancies between the meteor radar and MU observations of about 10 m/s.

In winter, the observation shows the eastward maximum in the end of November at around 65 km with the amplitude of over 60 m/s which did not last long, and the second peak occurs at the higher altitude at about 85 km in January with about 40 m/s, while the eastward wind shows maximum at 60–65 km with the amplitude over 80 m/s by the CIRA 1986 model and strong eastward with over 40 m/s lasts from October to March.

In February and March, the eastward wind below 75 km are observed to be already weak probably due to the stratospheric warming, while the CIRA model shows stronger eastward wind again in February at the same height region. On the other hand, the weak westward wind above 100 km from November to

January agrees well between CIRA and observation. In brief, the eastward wind in winter at 60–90 km in CIRA are larger in amplitude and longer in duration than the observed wind. The observed wind shows strong eastward wind up to 60 m/s only in late November and early December. The westward wind above 100 km in winter agrees well between CIRA and the observation.

3.3 Hemispheric Difference between Shigaraki and Adelaide

In this section we compare the mean winds observed at Shigaraki (35°N , 136°E) with the result of Adelaide (35°S , 138°E) MF radar, which is located at the geographical conjugate of Shigaraki. The comparison of winds at these two locations gives us important information on the symmetry of the global atmosphere in the mesosphere and lower thermosphere.

For the comparison, we use the combined data observed with Kyoto meteor radar in 1983–1986 and with MU radar in 1985–1988, which are shown in Figures 3.10 and 3.11 and replotted in Figure 3.13 as the winds at Shigaraki. Mean winds at Adelaide, shown in Figure 3.13, is taken from Manson et al. [1990], which were observed in 1984–1986. While the earlier observation at Adelaide in 1978–1983 is often used for the comparison of mean winds [e.g., Manson et al., 1987a], the updated results of mean winds at Adelaide are better in coverage of height, which are extended as low as near 60 km. Therefore, we selected this new version of mean winds at Adelaide for comparison with those at Shigaraki.

In summer in each hemisphere, the westward wind below 80–85 km began April/October and ended in the end of August/February at Shigaraki/Adelaide, showing 6 months delay between the hemispheres, exactly. The maximum westward wind at Shigaraki occurs in late June while at Adelaide in early January at around 70 km, which also shows 6 months lag, while the maximum amplitude was about 40 m/s at Shigaraki and over 50 m/s at Adelaide, indicating the stronger westward wind in the southern hemisphere. The reversal height of the westward wind at Adelaide was about 85 km, which was higher by a few km than that at Shigaraki in Figure 3.10.

Manson et al. [1987a] intensively studied the reversal height in summer from the radar data of 10 stations, including Kyoto meteor radar and Adelaide MF radar, and found the significant hemispheric asymmetry in summer reversal height as shown in Figure 3.14. That is, in southern hemisphere the reversal height is at about 85 km altitude in the latitudinal range from 20°S to 44°S , while in northern hemisphere, the reversal height in the latitudinal range from 44°N to 34°N was lower than 80 km but at the higher (52°N) and lower (18°N)

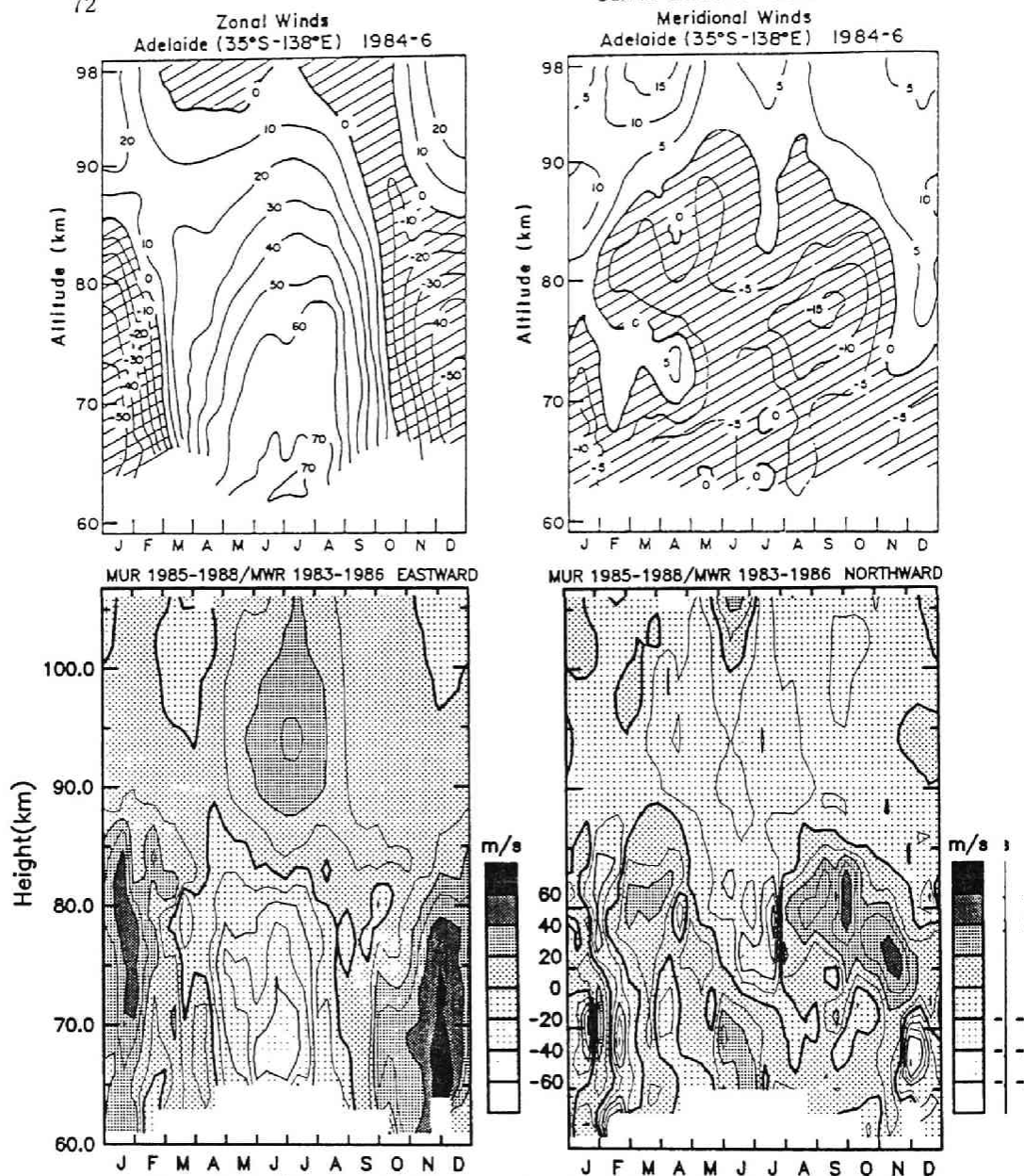


Figure 3.13: (Top panels) Mean zonal (left) and meridional (right) winds at 35°S observed with the Adelaide MF radar in 1984-1986. Shaded area corresponds to the westward and southward components. Numbers on the contour lines shows eastward and northward wind amplitude in m/s [Manson et al., 1991]. (Bottom panels) Corresponding mean zonal (left) and meridional (right) winds at 35°N observed by the MU radar in 1985-1988 and Kyoto meteor radar in 1983-1986, which are replots of Figures 3.10 and 3.11.

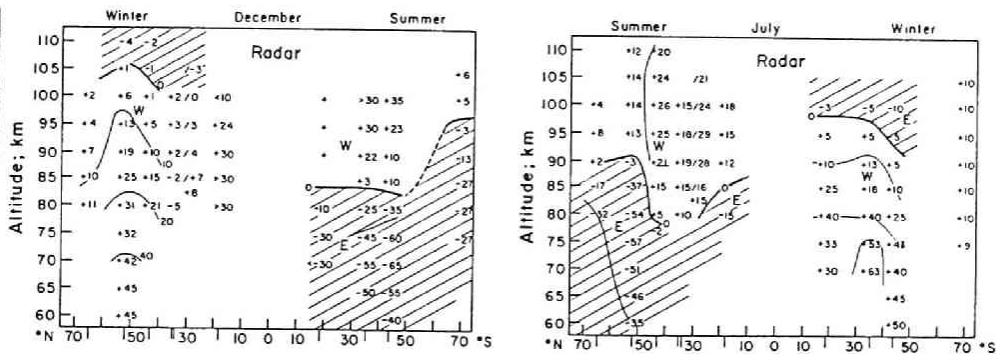


Figure 3.14: Latitudinal distribution of the zonal mean wind observed by MF and meteor radars in December (left) and June/July (right). Solid lines indicate reversal height of zonal mean wind [Manson et al., 1987a].

latitude the height is about 90 km and 85 km, respectively. Thus, the summer reversal at lower mid-latitude in the northern hemisphere (44°N – 34°N) was low comparing the other latitude in the same hemisphere and the southern hemisphere. The mean wind at around 35°N is reported that it has the interannual variability showing like mid-latitude and tropical year by year [Manson et al., 1987a]. We also have found the difference of the summer reversal height observed by Kyoto meteor radar in 1983–1985 and the MU radar in 1986–1988; below 80 km and at about 85 km, respectively. If we assume that this difference was due to the interannual variability, the shift of the reversal height to high altitude at the latter observation periods at 35°N can be explained by the two situations. The one is that the latitudinal distribution of low reversal height between 44 – 34°N latitudes was shifted towards the higher latitude. The other is that the low reversal height region between 44 – 34°N disappeared for the latter observation periods. The data obtained at Adelaide did not show a clear difference in the reversal height between the earlier and updated plots. This can be the interannual variability in summer reversal height shows large asymmetry at 35° latitude between the hemispheres.

The eastward wind above 80–85 km began in April and May at Shigaraki and November at Adelaide. The eastward peak occurred in early July at 95 km altitude with the amplitude of exceeding 30 m/s at Shigaraki, while the maximum eastward wind existed during December and January above 90 km with over 20 m/s at Adelaide where the peak is not clearly displayed. The

earlier Adelaide observation in 1978–1983 showed that eastward wind above 95km became over 30 m/s in summer [Manson et al., 1987a], therefore the amplitude of eastward wind in summer were similar in both hemisphere. At both locations the summer eastward wind above 80–85 km lasted until autumn and early winter, while the strong eastward wind with amplitude of over 20 m/s became small in early August and in the end of January at Shigaraki and Adelaide, respectively. In brief, mean zonal winds in summer are similar between hemispheres, but the westward wind maximum is found to be larger in southern hemisphere by about 10 m/s, and also large interannual variability of reversal height existed in northern hemisphere.

In winter eastward winds, covering almost whole observation height, began in September/March and end in April/October at Shigaraki/Adelaide, while the distribution of height and month of strong eastward winds were significantly different between these two locations. The maximum eastward wind occurred at around 65 km altitude and with the amplitude of 60–70 m/s at both locations, while the time of the eastward maximum was in the end of November at Shigaraki and between June to August at Adelaide. At Shigaraki, below 75 km the eastward wind was mainly strong in the former half of the winter i.e. from October to January, while at the height between 75km the eastward wind from December to February was stronger than before. On the other hand, at Adelaide the period of strong eastward wind was longer than that at Shigaraki, and the wind was stronger in the latter half at almost all the altitude, which seems to have the maximum in early August.

During the stratospheric warming which often occurs in the late winter, the eastward wind in the stratosphere and mesosphere are reported to be significantly weakened and often become westward [Labitzke et al., 1987]. The observed differences of the winter eastward winds here was probably due to the effect of stratospheric warming in the northern hemisphere. The relation between the stratospheric warming and the mean wind is presented in the following section. The westward wind at high altitude in early winter was observed from November to January above 98km at Shigaraki, and from March to June above 95 km at Adelaide. In total, winter eastward wind at Adelaide was more steady and stronger in the latter half, while at Shigaraki strong eastward wind did not last long having the peak at the former half of the winter, and weak in the latter half especially below 75km.

The meridional component shows generally poleward wind below 85 km and equatorward wind above that height in the both profiles obtained at Shigaraki and Adelaide. It is noteworthy that in the both profiles obtained at Shigaraki and Adelaide, contour of meridional wind shows many small fluctuations that means meridional component often show a wave-like structure in a vertical pro-

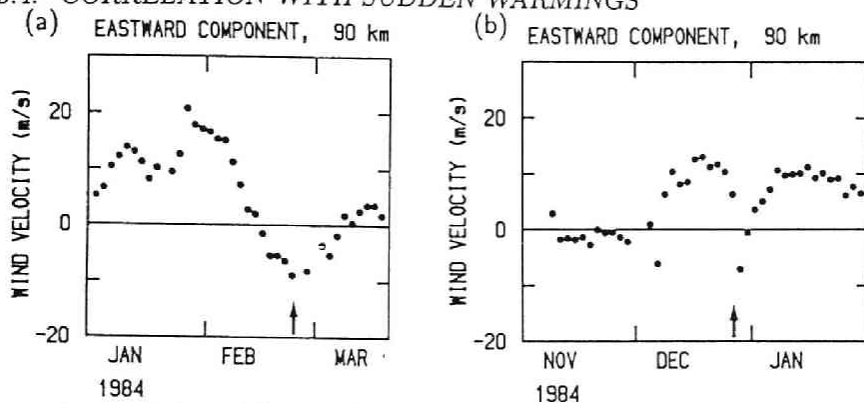


Figure 3.15: (a) Eastward wind velocity at 90 km altitude detected in January–March 1984. Wind velocities are determined by 10 days of observations. An arrow shows the date of maximum of the stratospheric sudden warming in February 1984. (b) The same as (a) except for a period from November 1984–January 1985. An arrow shows the start of sudden warmings occurring in late December 1984. [Tsuda et al., 1987]

file. It is noteworthy that the intervals of contour lines are smaller for the meridional component than for the zonal component because the annual variation of mean winds are smaller for meridional component. However, the amplitude of fluctuating structure is still larger for the meridional component.

In summer, the region of equatorward wind were expanded as low as 70 km from May/November to July/January in Shigaraki/Adelaide, respectively. Furthermore, the equatorward wind at 70 km is likely to continue till October/April. The poleward wind shows maximum in early October at around 80 km with amplitude of 20 m/s at Shigaraki, while in September at 77km with amplitude of 15 m/s at Adelaide, showing significant asymmetry. The variability in winter, from January to March, at Shigaraki also differed from that at Adelaide. The equatorward wind became as large as 10 m/s in January at 80–90 km and 15m/s in March at 98km at Adelaide, while it became about 20 m/s in July at 80 km but not more than 10 m/s above 85 km at Shigaraki. In total, significant difference could not be seen in mean meridional wind except for the winter variability at Shigaraki.

3.4 Correlation with Sudden Warmings

The large interannual variation of the mean winds at Shigaraki were observed by Kyoto meteor radar as shown in Figure 3.5. In this section we investigate the correlation between the stratospheric sudden warming and the mean zonal

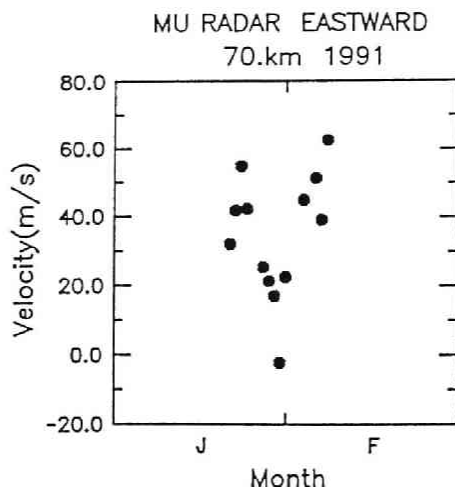


Figure 3.16: Daily mean eastward wind velocity at 70 km altitude detected in January 21–February 8, 1991. Wind velocities are determined by daytime average of each day.

wind.

Tsuda et al. [1987] showed a correlation of the mean zonal wind at 90 km and the stratospheric sudden warmings in the northern hemisphere. A stratospheric sudden warming reached its maximum on February 23, 1984 [Naujokat and Labitzke, 1984]. The radar observation at Saskatoon (52°N, 107°W), Kühlungsborn (54°N, 12°E) and Collm (51°N, 13°E) and satellite observations at 60°N showed that zonal wind velocity became westward in the 45–90 km altitudes at 50°–60°N and then turned to the westward in March. Figure 3.15 (a) shows the eastward wind velocity at 90 km altitude observed by Kyoto meteor radar in January–March 1984 [Tsuda et al., 1987]. The eastward wind velocity decreased rapidly in the middle of February and then reversed the direction into westward for about 20 days. In March the wind again became eastward. The maximum westward velocity occurred simultaneously with the peak of the sudden warming.

Another stratospheric warming was reported by Labitzke et al. [1985]. They found that the warming began on December 26, 1984 and developed strongly within 5 days. Figure 3.15 (b) shows the eastward wind velocity by Kyoto meteor radar in November 1984–January 1985 at 90 km altitude [Tsuda et al, 1987]. The zonal wind velocity again became westward when the warming occurred.

A similar comparison between the mean zonal wind and sudden warming can be done with the results of the MU radar, although the mesospheric observation is less frequently carried out. We next show an instance of the comparison by the MU radar. Figure 3.16 shows the daily mean zonal wind at 70 km

altitude averaged over daylight hours observed with the MU radar in January 21–February 8, 1991. Note that the low pass filtering with time are not applied in this case, so that each plot is independent of each other. This is the longest campaign of the mesospheric observations which has been carried out in winter with the MU radar. In January 21–25 and February 4–8, the eastward winds range from 40–60 m/s. Then, at the end of January; January 28–February 1, the eastward wind became weak with the values of around 20 m/s. On January 31, the decrease of the eastward wind became maximum and the wind velocity was almost zero. Naujokat et al. [1991] reported that a strong stratospheric sudden warming occurred from the end of January to the former half of February in 1991. The rapid decrease of the eastward wind at 70 km observed with the MU radar is considered to be correlated with this major warming, and the sudden warming seems to affect the mean wind as low as 70 km at 35°N latitude.

Thus we can demonstrate three cases of observation that the stratospheric sudden warming in the northern winter was accompanied by the change of the eastward wind into westward direction at as low as 35°N at 70–90 km altitude. These results suggest that the interannual variability of the mean wind in winter is closely correlated with the stratospheric sudden warmings even at 35° latitude.

3.5 Concluding Remarks

In this chapter the mean winds are deduced from Kyoto meteor radar observation (80–110 km) in 1983–1986 and the MU radar observation (60–90 km) in 1985–1988. From meteor radar observations, we clarified that the Groves' [1959] algorithm is not adequate for meteor radar data with low echo rate in the evening, while a harmonic analysis by Tsuda et al. [1987], which uses the 1 h $\times$ 4 km time-height bin and deduces the mean wind and tide by least square fitting, can be reasonably applied to estimate the tides and mean winds. The wind velocities observed with the MU radar are simply averaged for the daytime (8LT–16LT), which gives comparatively good estimation with less than 5 m/s bias due to the effect of diurnal wind variation such as tides below 85–90km. The mean wind by Kyoto meteor radar showed that the summer eastward wind enhanced at around 95 km with peak value more than 30 m/s in July. In January and February eastward wind again enhanced below 95 km. Mean meridional wind was generally southward and enhanced up to 10 m/s in summer. The MU radar data showed that westward wind in summer below 80 km reached maximum in June at 65–75 km with the amplitude of about 40 m/s. In winter, eastward peak was in the end of November at 70 km with the amplitude of 60 m/s. The eastward wind in winter shows significant variability suggesting the effect of stratospheric warmings.

The accuracy of the winds between 80 and 90 km altitude by Kyoto meteor radar and the MU radar is carefully discussed, and the results of the two radars are combined into the mean wind from 60 km to 110 km. The reversal height of the summer westward wind is then determined as 83 km altitude, while it seems that large interannual variability exists; below 82 km in 1983–1985 by Kyoto meteor radar and about 85 km by the MU radar in 1986–1988. This mean wind profile is compared with the mean wind obtained by Adelaide MF radar in 1984–1986. The reversal height in summer at Shigaraki was slightly lower than at Adelaide, while the interannual variation was larger at Shigaraki than at Adelaide. The total mean wind characteristics are basically similar between the hemispheres except for the winter variability observed in northern hemisphere, and the variability in summer reversal height in the northern hemisphere.

Three cases are reported that the stratospheric sudden warming was accompanied with the change of the eastward wind into westward at 70–90 km at Shigaraki, suggesting close relation between sudden warmings and winter variability in mean wind at Shigaraki.

Chapter 4

Atmospheric Tides and Two-day Wave

4.1 Historical Background

This chapter is devoted to presenting the observations of atmospheric tides and two-day wave in the mesosphere and lower thermosphere.

Various observations utilizing radar and optical remote sensings clarified the basic characteristics of atmospheric tides [e.g. Kato, 1980]. Theoretical models including interactions between tides and mean winds [e.g., Lindzen and Hong, 1974] and improved global distribution of excitation sources and dissipative processes [e.g., Forbes, 1982a, b] greatly advanced our understanding of tides. The improvement of the global model of the temperature and mean wind distribution, which are summarized as an updated CIRA (CIRA 1986), further improved the tidal models [Vial, 1986; Forbes and Hagan, 1988], which utilized improved dissipating process as well as a new background model of temperature and mean wind. However, in order to understand the behavior of atmospheric tides more closely, coordinated simultaneous observations of tides are inevitable.

The Atmospheric Tides in the Middle atmosphere Program (ATMAP) campaign observations were carried out including more than 10 radar stations of the medium frequency (MF), meteor, and MST radars [Forbes, 1985]. Forbes [1985] showed that the observations generally agreed with the models, but found that the short duration of campaigns (4 days) often affected by large day-to-day variation of tides, which makes it difficult to compare with the models. Comparisons of profiles of tides between the northern and southern hemispheres were carried out between Adelaide (35°S, 138°E) and Kyoto (Shigaraki) (35°N, 136°E) [Aso and Vincent, 1982], and Saskatoon (52°N, 107°W) and Christchurch (44°S, 173°E) [Manson et al., 1985], which suggested the importance of antisymmetric

tidal mode. Comparisons of the semidiurnal tides among the observations at midlatitudes (Saskatoon, 52°N; Christchurch, 44°S; Monpazier, 45°N) [Manson et al., 1985, 1987b] and high-latitudes (Poker Flat, 65°N) [Manson et al., 1988], have clarified a regular tidal morphology.

Continuous operation of the Kyoto meteor radar (35°N) was carried out from May 1983 to March 1986, and the tidal climatology was compared with the observations by Adelaide MF radar (35°S) collected in 1983–1985 [Vincent et al., 1988, 1989], and also with the observations by Saskatoon MF radar (52°N) in 1979–1983 [Tsuda et al., 1988a] for semidiurnal tides. In this chapter we summarize these results together and discuss the comparison of the tides at these three locations, and further compare the result with recent tidal models, by adding some new analysis of the observations.

Planetary waves with periods longer than 1 day in the mesosphere and lower thermosphere, have also been observed with meteor and MF radars, [Salby and Roper, 1980; Massebeuf et al., 1981; Ito et al., 1984]. Among these the two-day wave was reported by many authors [e.g. Müller and Nelson, 1978] being enhanced in summer months with dominant period of around 51h in the northern hemisphere and 48h in the southern hemisphere [Vincent, 1984b]. Cooperative radar observations determined that this wave propagates westward and has a zonal wavenumber of three [Müller and Nelson, 1978, Craig et al., 1980]. Craig et al. [1983] reported a large asymmetry between northern and southern hemispheres. Theoretical study was carried out by Salby [1981a] who showed the two-day wave was considered to be a mixed Rossby wave with a zonal wavenumber of three, and determined its period as 53h (2.25 days). Plumb [1983] showed that the two-day waves can be generated by a baroclinic instabilities in the stratospheric easterly wind. Plumb et al. [1987] also reported the importance of the two-day wave in accelerating the mean zonal circulation.

In this chapter we present two-day wave observed at Shigaraki and comparison of the event of the two-day wave between Shigaraki and Adelaide according to the result of Tsuda et al. [1988b], and show some further comparison and considerations.

4.2 Semidiurnal Tide

4.2.1 Climatology at Shigaraki

Figure 4.1 shows time-height cross section of the amplitudes of the zonal component of semidiurnal tides by Kyoto meteor radar observation at Shigaraki (35°N) in 1983–1985 [Tsuda et al., 1988a]. Semidiurnal component were determined by means of a harmonic analysis shown in Chapter 3, using 30 days of observations.

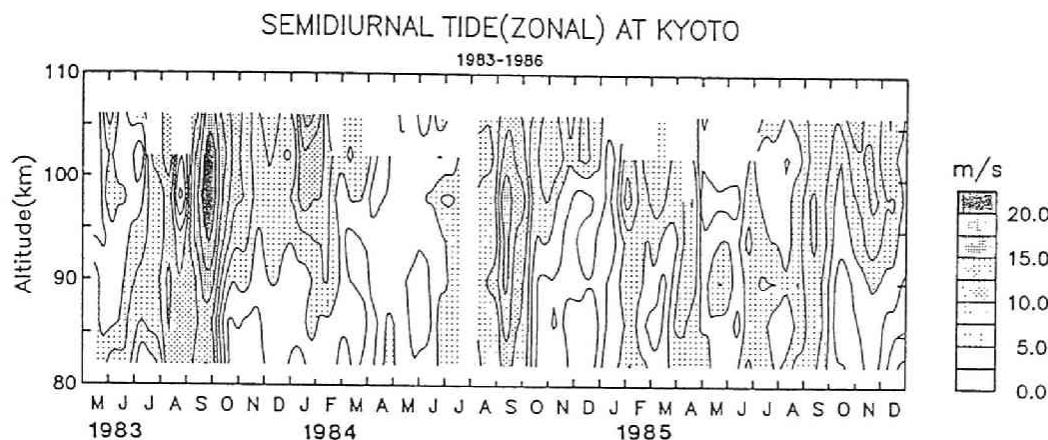


Figure 4.1: Contour plot of the amplitude of the zonal component of semidiurnal tides at Kyoto. Harmonic analysis using 30-day observation intervals is applied. Fluctuations with periods shorter than 18 days are filtered out [Tsuda et al., 1988a]

Low pass filtering with a cutoff period of 18 days was further applied to the fitted parameters. It is found that the large amplitudes were recognized at 90- to 100-km altitudes with amplitude more than 10–20 m/s in September, although there were significant interannual variations. Fairly large amplitudes were also observed in January and November 1984 and in February and November 1985 at altitudes above 95 km.

Figure. 4.2 shows amplitudes and phases of the semidiurnal tides at 90 km for eastward and northward components. Phases of the eastward component were almost constant around 0700 LT during the period around summer, that is from March–April to the middle of October for the successive three years of 1983–1985. Phases of the northward component also seems to be centered around 0400–0500 LT from March–April to the middle of October, although considerable fluctuations were found especially in 1985. The phase difference between the eastward and northward component was 3 hours (90°).

On the contrary, the phases of the eastward component in winter months of 1983–1985, from November to February, were approximately constant at 0000–0100 LT, indicating almost out of phase with the summer data. The phases of northward component was less stable in winter than those of eastward component because the amplitude of the northward component were small in winter months. However, the average of the phase value was determined to be approximately 0900 LT in November, December, and February, which was later than the summer phases by 4–5 hours. The phase difference between the eastward and

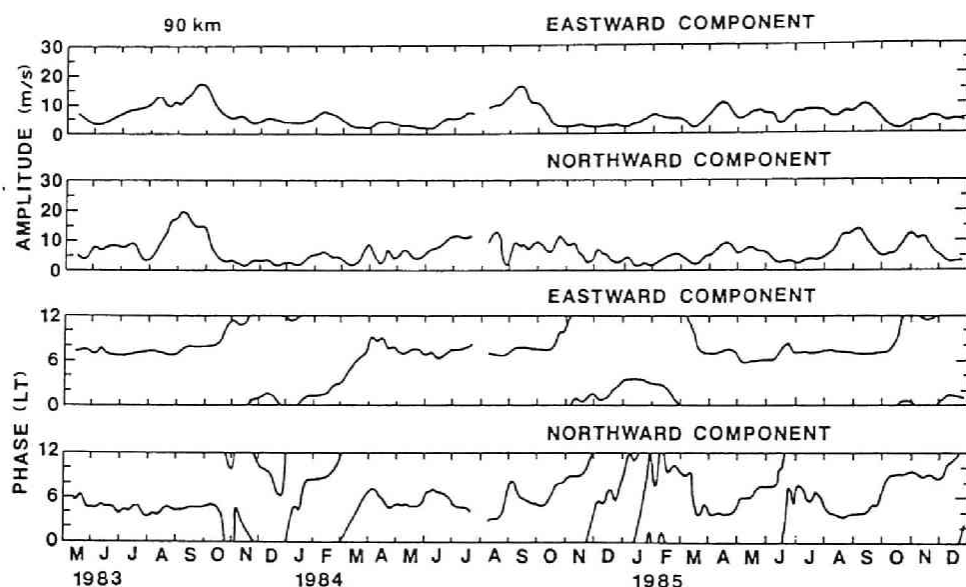


Figure 4.2: Amplitudes and phases of semidiurnal tides from Kyoto meteor radar observation at 90 km altitude [Tsuda et al., 1988a].

northward components in winter was also about 90° as recognized in summer.

Thus, the phase of the semidiurnal tide of both eastward and northward component can be divided into two states corresponding to the months around summer and winter. Hereafter, we call these two durations of the states as summer-like months and winter-like months. Transitions between the summer-like and winter-like months were observed in near October and March. In Figure. 4.2 these transitions seem to be gradual. However, it is partly because of the long duration of data analysis for 30 days and the cut off of short periods fluctuation due to the low pass filter. The detailed feature of transition between summer-like and winter-like months is described later.

Figure. 4.3 shows vertical profile of monthly mean amplitudes and phases of the eastward component of semidiurnal tides observed in July and December in 1983-1985. In July, amplitudes were 5 to 10 m/s and did not always increase with the altitude. The phase values lay around 0700 LT at almost all the altitudes.

In December, amplitudes slightly increased with the altitude, while they were less than 10 m/s in the whole altitude region and smaller than those in July. Phases were about 0200 LT at 82 km and 0900-1100 LT at 106 km. As is shown in Figure 4.2, phases in December at 90 km were almost out of phase with those in July.

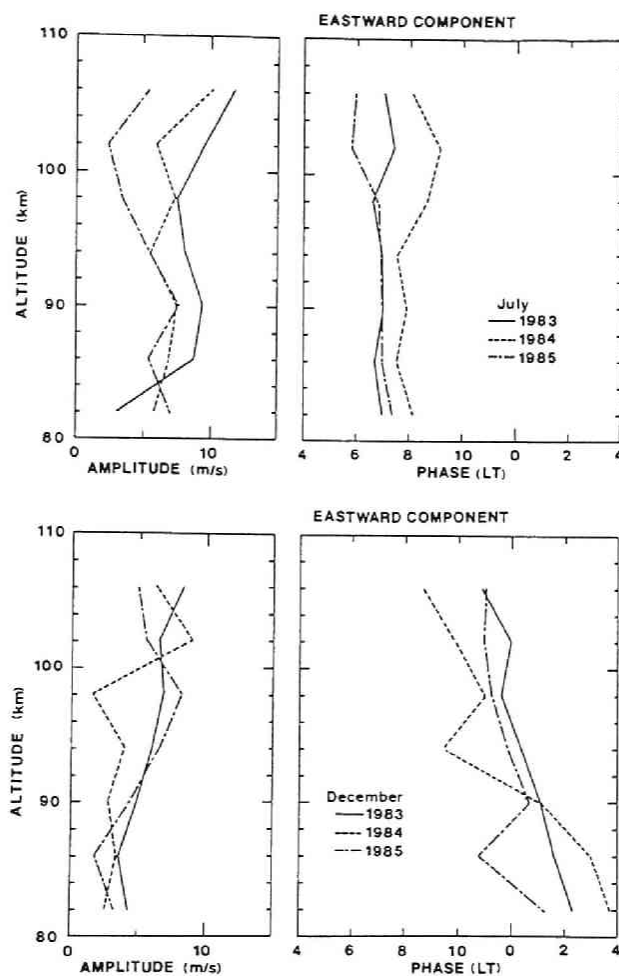


Figure 4.3: Monthly mean profiles of amplitudes and phases of semidiurnal tides observed at Kyoto (top) in July and (bottom) in December in 1983–1985 [Tsuda et al., 1988a].

By taking a linear fitting at around 90 km altitude, vertical wavelengths are estimated as longer than 200 km and approximately 70 km in July and December, respectively. Wavelengths were generally long in summer-like months and became relatively short in winter-like months, as described in a later section.

4.2.2 Latitudinal Variations

In this subsection we compare the semidiurnal tide observed at Shigaraki (35°N) with the result of Saskatoon (52°N) and Adelaide (35°S). The comparison are mainly focused on the altitude of 90 km, because it is the height of good data confidence for all the three stations.

Figures. 4.4 and 4.5 show monthly mean values of amplitudes and phases of semidiurnal tides at 90 km over Shigaraki and Saskatoon [Tsuda et al., 1988a]. The amplitudes and phases at Saskatoon were average of 10-day harmonic fits [Manson and Meek, 1986], while the values of Shigaraki were deduced by 30 days fitting. Mean values of the eastward component observed at Saskatoon in 1978-1982 are also plotted. It is noteworthy that the amplitudes are arithmetic means of daily tidal fits for the data of 1978-1982 and, hence larger than those based upon 10-day fits because of day-to-day fluctuations[e.g., Manson and Meek, 1985]. Error bars are not plotted for clarity in Figures. 4.4 and 4.5, but typical values are a few meters per second in amplitude and 0.2-0.5 hours in phase for monthly determinations at both Shigaraki and Saskatoon.

The phase of the semidiurnal tide at Shigaraki in April to September shows good repeatability for both eastward and northward components, while slight interannual variations were recognized in winter. Amplitudes at Shigaraki were 5 to 15 m/s for the whole observation period and attained a maximum in late summer and autumn (August-September). The phases at Shigaraki were summarized as followings, using the phases of the northward component in winter months inferred from Figure. 4.2: for summer-like months (April-September) the phases of the eastward and northward components were 0700 LT and 0400-0600 LT, respectively; for winter-like months (November-February), the phases of eastward component and northward components were 0100 LT, and 0800-1000 LT, respectively.

On the other hand, the amplitudes of both eastward and northward component observed at Saskatoon shown in Figures 4.4 and 4.5 were 8 to 23 m/s and generally larger than those at Shigaraki. The seasonal variation of the amplitudes were more complicated than that at Shigaraki, indicating three maxima in December-January, in May and in September with summer minimum. The peak of the amplitudes in September seems to coincide with that at Shigaraki, but in winter no large amplitude was recognized at Shigaraki. The phases of

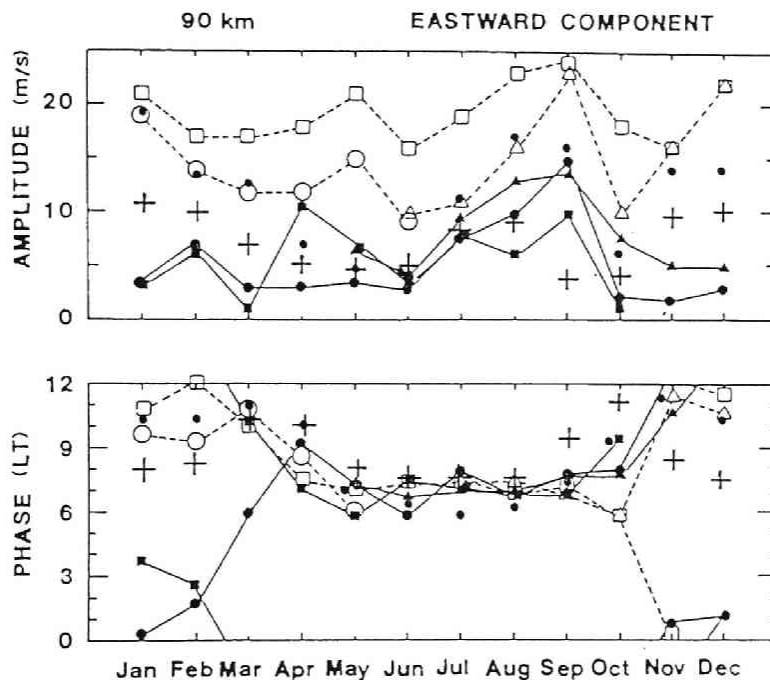


Figure 4.4: Monthly mean values of amplitudes and phases of the eastward component of semidiurnal tide at 90 km altitude observed by the Kyoto meteor radar and the Saskatoon MF radar.

Solid triangle, circle and square symbols show results at Shigaraki in 1983, 1984 and 1985, respectively. Open triangle and circle symbols correspond to measurements at Saskatoon in 1983 and 1984, respectively, and open square symbols to averages in 1978–1982. The 1978–1982 values from Saskatoon were arithmetic means of daily tidal fits and were therefore larger, especially in summer (see text) [Tsuda et al., 1988a]. The dots and crosses are the model calculation by Forbes and Vial [1989] for 52°N and 34°N, respectively.

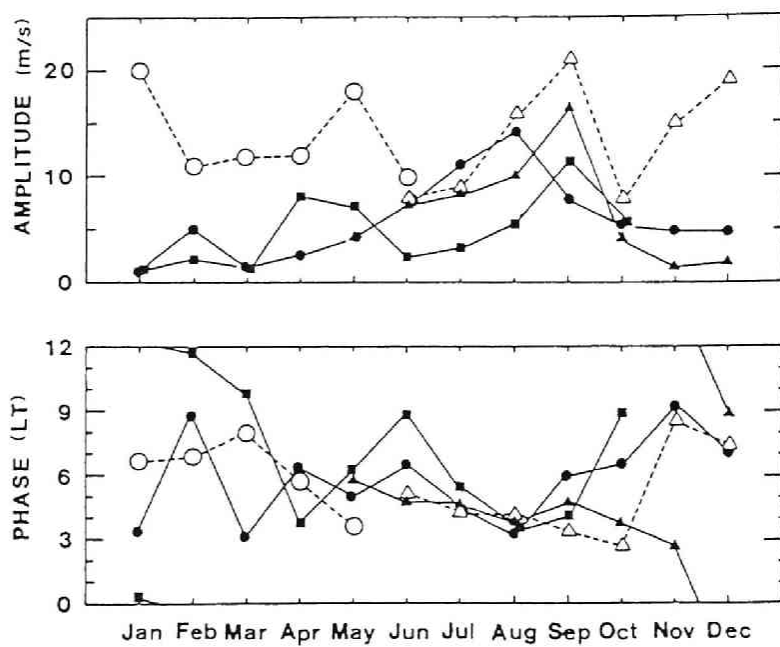


Figure 4.5: The same as Figure. 4.4, except for the northward component [Tsuda et al., 1988a].

the semidiurnal tide observed at Saskatoon were 0700 LT and 0400–0500 LT for eastward and northward components, respectively, in summer-like months (May–October), and 1000–1200 LT and 0700–0800 LT for eastward and northward components, respectively, in winter-like months (late November–March).

Thus, the overall seasonal variations at Saskatoon agree quite well with those at Shigaraki. Transitions between the summer-like and winter-like months were observed in March–April and October–November at Saskatoon, which agree again with the transitions at Shigaraki in a rough comparison. The detailed comparison between the transitions at both locations are discussed later.

Next, we compare the semidiurnal tides observed at Shigaraki and Adelaide. Figures 4.6 and 4.7 shows the monthly mean amplitudes and phases at 90 km observed at Shigaraki and Adelaide [Vincent et al., 1988]. The phases were defined as the local times of maximum eastward and northward wind velocities for both Shigaraki and Adelaide. Error bars were not indicated in the figures for clarity, but typical values were a few m/s and 0.2–0.5 hours for amplitude and phase, respectively. The amplitudes of semidiurnal tide of both northward and eastward components range from a few m/s to 15–20 m/s at both Adelaide and Shigaraki. However, the durations with the amplitudes larger than 5 m/s are longer at Adelaide than at Shigaraki, and hence the months with larger amplitude at Adelaide than at Shigaraki were more frequently recognized. The seasonal variation of the amplitudes at Adelaide showed maximum in summer and a secondary peak in early winter for both eastward and northward components. On the contrary, the amplitude at Shigaraki had a maximum in late summer and autumn. Appreciable interannual variability was recognized in both the amplitudes at the maxima and the months of their occurrence.

On the other hand, the phase variations were more repeatable among different years than those of amplitudes. The mean phases at Adelaide were 0000 LT and 0400–0500 LT for eastward and northward components in winter (March–September), and 0700 LT and 1000–1100 LT for eastward and northward components in summer (November–February), respectively. It is noteworthy that the variation of the phase at 90 km at Adelaide again can be divided into two states of months; winter-like and summer-like months. Furthermore, the duration of winter-like and summer-like months at Adelaide were almost the same as that of summer-like and winter-like months at Shigaraki and Saskatoon, resulting long summer-like months at Shigaraki and Saskatoon and short summer-like months at Adelaide. The phases of zonal components observed at Adelaide and Shigaraki were approximately antiphase throughout a year, and the phases of the meridional components were almost always in phase, although the phase varied largely between the summer-like and winter-like months at each location.

Vincent et al. [1988] explained this situation by the dominance of antisym-

SEMIDIURNAL TIDE (1983–1985) EASTWARD COMPONENT at 90 km

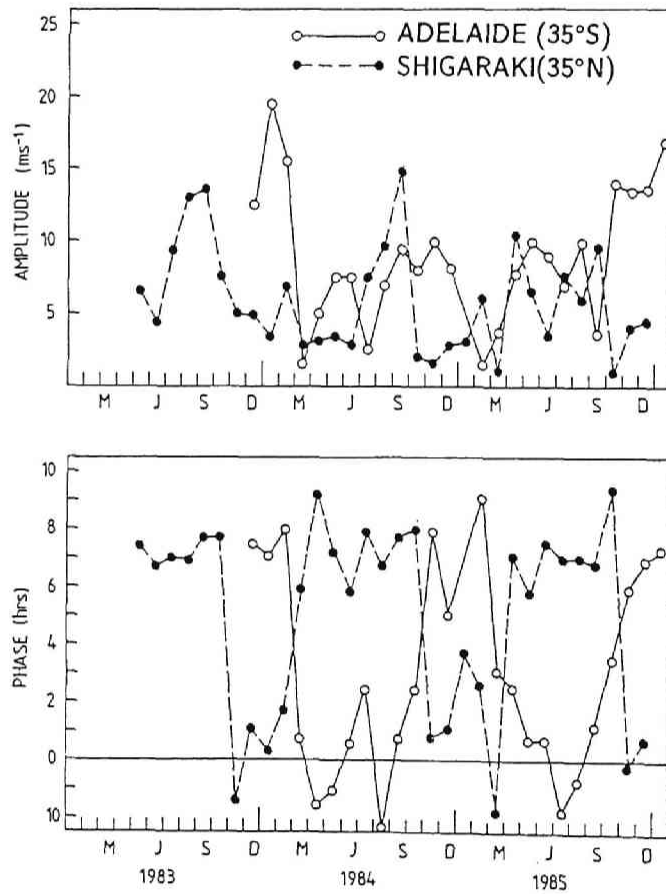


Figure 4.6: Amplitudes and phases of the zonal component of the semidiurnal tide at 90 km (monthly mean). The phase means the local time of eastward wind maximum [Vincent et al., 1988].

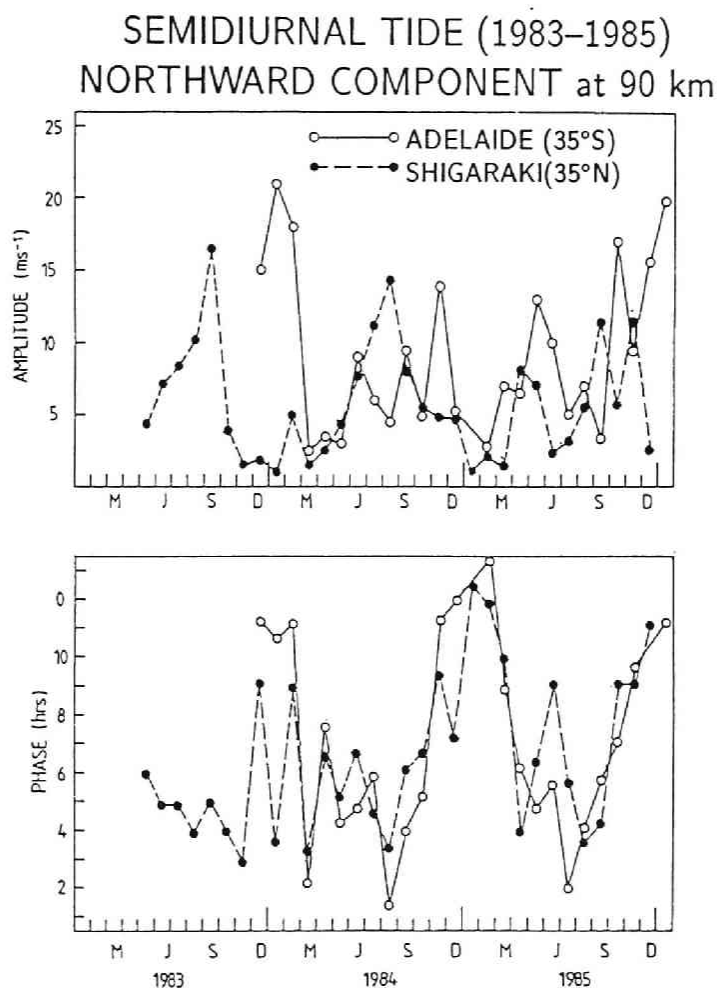


Figure 4.7: The same as Figure. 4.6, except for the meridional component. the phase is the local time of northward wind maximum [Vincent et al., 1988].

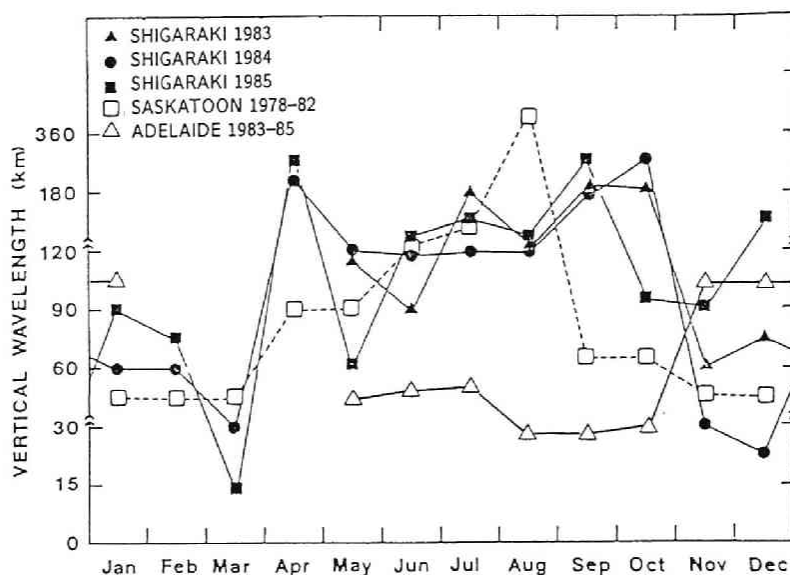


Figure 4.8: Monthly mean value of vertical wavelengths. Kyoto meteor radar results are inferred from the phase progression of eastward component at around 90 km; Saskatoon results are an average of both components [Tsuda et al., 1988a]. Adelaide results are deduced from Vincent et al. [1989].

metric mode, especially (2,3) mode because the component of (2,5) mode has nulls near 35° latitude. The equinoctial change between summer-like and winter-like months was more conspicuous and rapid at Shigaraki than at Adelaide. These transitions are discussed in detail in the later subsection. Vincent et al. [1988] further pointed out that there is a global asymmetry in the behavior of the semidiurnal tide. The comparison of semidiurnal tides between Saskatoon and Christchurch (44°S) [Manson et al., 1985] also showed that northward components were in phase, again favoring the dominance of antisymmetric modes.

Figure 4.8 shows the seasonal variations of vertical wavelengths observed at Shigaraki, Saskatoon and Adelaide. Vertical wavelength at Shigaraki was delineated at 90 km altitude by the same method described concerning with Figure 4.3 [Tsuda et al., 1988a]. Vertical wavelengths for Saskatoon observations were of 80–100 km altitudes in 1978–1982 [Manson and Meek, 1985]. Note that vertical wavelength at Adelaide was derived from Vincent et al. [1989] by the gradient of contour of time height cross section of the phase at around 90 km. The variations of vertical wavelength at Shigaraki agreed well among determinations in successive 3 years, indicating vertical wavelength longer than 100

km in summer-like months (April–October), and became 15–90 km in winter-like months (November–March). The vertical wavelengths at Saskatoon were longer than 100 km in summer-like months (June–August) and were approximately 45 km in winter-like months (November–March). Seasonal variation of vertical wavelengths therefore generally agreed with the results at Shigaraki. However, the duration of long vertical wavelength in summer was shorter at Saskatoon than at Shigaraki, and in equinoctial months such as April–May and September–October, the wavelength became shorter than 100 km and close to winter values at Saskatoon, although in these months the wavelength were larger than 150 km at Shigaraki.

At Adelaide, the vertical wavelength were as large as 100 km in summer-like months (December–January) and 25–50 km in winter-like months (May–October). The wavelength was not determined from February to April due to the complicated phase structure. These variations again agreed well with those in Shigaraki and Saskatoon, although the winter-like months lasted longer at Adelaide. These results suggest the excitation of antisymmetric higher-order modes in the winter hemisphere.

Forbes and Vial [1989] developed a numerical model for monthly profile of semidiurnal tide using improved background values of heating rates, wind, temperature, pressure and eddy viscosity. The results were compared with the observation of 52°N (Saskatoon), 35°N (Shigaraki) and 44°N, and showed fairly good agreement with those observations. The simulated amplitude and phase at 90 km were plotted in Figure 4.4 for the data of 52 °N and 35°N. At 52 °N, both simulated amplitude and phase agreed well with Saskatoon observation, while at 35°N, discrepancies are found in both amplitude and phase.

In summer (April – August), the simulated amplitude and phase agreed well with observation at Shigaraki, while the maximum amplitude at late summer/early autumn did not emerge for simulation. In winter (November – February), discrepancy is evident for both amplitude and phase. The simulation showed large amplitude of about 10 m/s and phase value of about 0800 LT, while the observation showed 5 m/s and 0000–0300 LT, respectively. Forbes and Vial [1989] discussed this large discrepancy of phase is due to the large error in phase determination of observation. However, the phase determination was applied for the 30 days of observation and so the error is expected to be as small as 1.5–2 hours, and therefore smaller than the difference between the simulated and observed phases shown in Figure 4.4. Thus, the simulated amplitude and phase of semidiurnal tide showed significant difference at 35°N, although in summer at the same latitude and in all seasons at 52 ° N the simulated semidiurnal tides generally agreed well with observation.

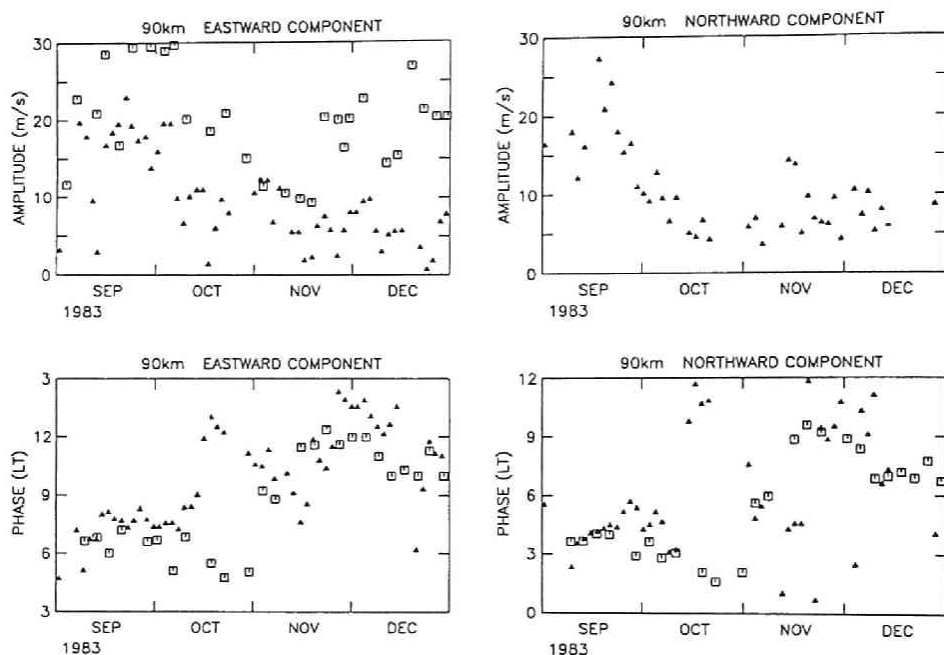


Figure 4.9: Fine structure of phase and amplitude variation in September–December 1983, detected at Shigaraki (triangles) and Saskatoon (square). For Saskatoon the amplitudes shown are means of zonal and meridional values, as their ratio is always close to the unity [Manson and Meek, 1986]. Note that the vertical axis for the phase of eastward component is shifted by 3 hours [Tsuda, et al., 1988a].

4.2.3 Transition of Semidiurnal Tides in Equinoxes

So far we showed that seasonal variations of semidiurnal tides can be separated into two different stable states of summer-like and winter-like months, with equinoctial transitions. Seasonal variations of semidiurnal tides at Shigaraki and Saskatoon agreed fairly well, and a large asymmetry was found in the seasonal variations between those locations in the northern hemisphere and Adelaide. In this subsection we describe more closely the behavior of semidiurnal tides during the transitions between summer-like and winter-like months among the three locations. The comparison between Saskatoon and Shigaraki was made by Tsuda et al. [1988a]. We first review their result, and then we further compare the transitions of Shigaraki and Adelaide, as well. Amplitudes and phases of the semidiurnal tides are determined from 4 days of observations at Saskatoon [Manson and Meek, 1985], and from 5 days of observations at Shigaraki and

Adelaide by a harmonic analysis described in Chapter 3, respectively.

Figure 4.9 shows the amplitudes and phases of the semidiurnal tides at 90 km at Shigaraki (triangles) and Saskatoon (squares). In September, phases of the eastward and northward components were distributed around 0600-0800 LT, and 0300-0500 LT, respectively, for both Shigaraki and Saskatoon data, indicating the characteristics of summer-like months. In the end of November and during December, phases became around the values of winter-like months.

Rapid phase increase by about 5 and 6 hours occurred in the middle of October at Shigaraki. In the middle of November they return to the values of summer-like months, and then became those of winter-like months by the end of November. Phases of both eastward and northward components observed at Saskatoon became earlier by about 2 hours in the middle of October, then jumped to the values of winter-like months in early November. Thus, the fine structure of the transition between summer-like and winter-like months did not necessarily agree with each other. The amplitude of the semidiurnal tide generally decreased from September to December, but at the rapid jump of phase in the middle October amplitudes also seems to have local minima, which is clearer for eastward components. Amplitudes are also found to fluctuate during the transition periods. Those features suggest interference among two waves. Amplitudes observed at Saskatoon were almost always larger than those at Shigaraki; however, after the September/October maximum values also decreased strongly in November during the phase transition to the winter-months, and then they became as strong as those in summer-like months again, which is different from the results of Shigaraki.

The features like interference of two single modes are also seen in the vertical profiles of the amplitudes and phases observed at Shigaraki in the transition duration of October 1983, as shown in Figure 4.10 [Tsuda et al., 1988a]. It is clearly found in the eastward component that the vertical structure abruptly changed in the middle of October with the vertical wavelength varying from 150 to 300 km to 30 to 60 km, which is corresponding to the phase shift at 90 km from 0700-0800 LT to 0000 LT. After this change in the middle of October, phases became almost out of phase with the values before the change at the altitudes lower than 90km, while at higher altitudes (higher than 95 km) the phase difference was smaller. The amplitudes of the eastward component at 95-100 km altitudes in the beginning of October were as large as 20 m/s, while they became as small as $5-10 \text{ ms}^{-1}$ by the end of October. The phase profile of northward component were more complicated. However, their general structure was similar to that shown in the eastward component.

Tsuda et al. [1988a] reported that these characteristics were repeated in 1984 and 1985. That is, the transitions from summer-like months to winter-like

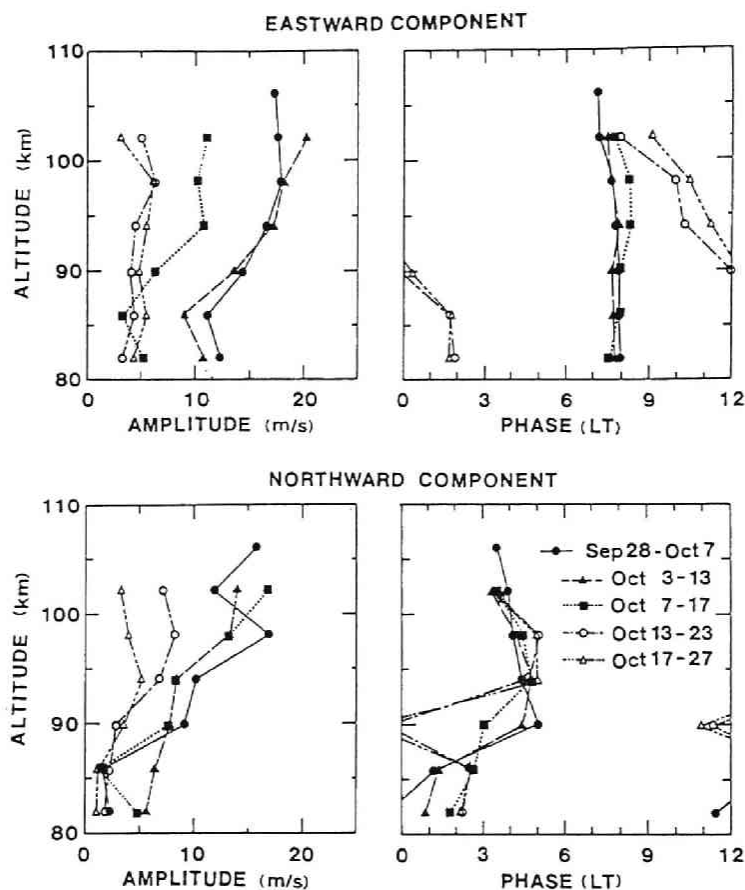


Figure 4.10: Height profiles of amplitudes and phases of eastward and northward components observed in October 1983 at Shigaraki. Profiles are determined every 10 days [Tsuda et al., 1988a].

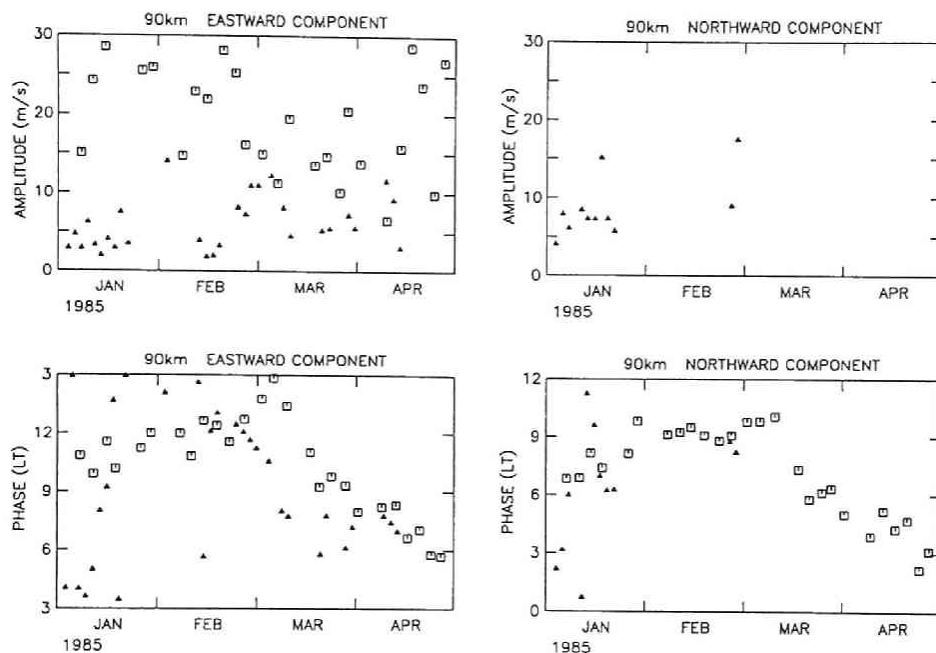


Figure 4.11: The same as Figure. 4.9, except for the results in January–April 1985 [Tsuda et al., 1988a].

months were in early October and early November at Shigaraki and Saskatoon, respectively.

Figure. 4.11 shows results for the spring transition at both locations. It is noteworthy that the northward components were not fully determined because of the low meteor echo rate in this season of the year. In January and February the phase of eastward component at Shigaraki gradually advanced and it abruptly varied in the end of February, and became around 0700 LT in the beginning of March, which is the value of summer-like months.

The phases of both northward and eastward components at Saskatoon began to advance in the first half of March and by the end of April they became the values of summer-like months. These transitions are similar to the results of other years [Manson et al., 1983; Manson and Meek, 1986]. Amplitudes became small during the transition, but again large amplitudes emerged in the end of April. It is found that before the rapid transition the phase delayed by 2–3 hours in the end of February and the beginning of March.

Figure 4.12 shows the transition of the phase of semidiurnal tide observed by Kyoto meteor radar and Adelaide MF radar in September–December 1984.

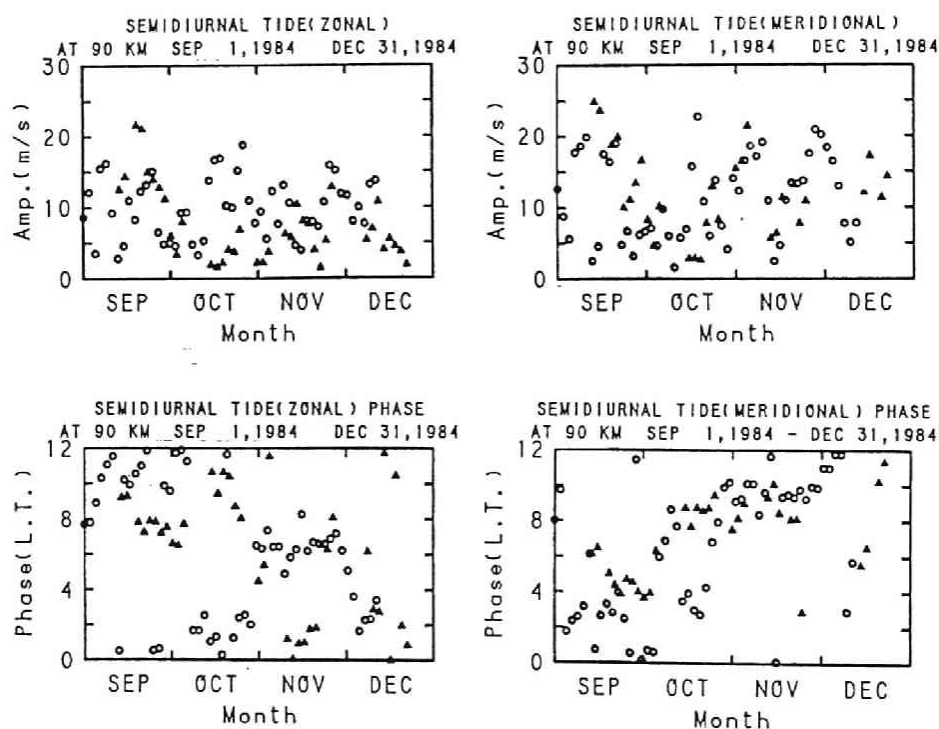


Figure 4.12: Fine structure of phase and amplitude variation in September–December 1984, detected at Shigaraki (triangles) and Adelaide (open circles).

The phase of the northward component shows a rapid jump in the first half of October for both at Shigaraki and Adelaide from around 0400 LT to 0800LT, and in November the phase became around 1000 LT, which is identical to that of winter-like (Shigaraki) and summer-like (Adelaide) months. The eastward component seems to show a similar variation at the same time, while the phases observed at Shigaraki were earlier by 4–5 hours than those at Adelaide. It is remarkable that in the first half of October, the amplitudes of both northward and eastward component at Shigaraki and Adelaide became small, showing that the interference of the two modes occurred simultaneously. At Adelaide the pulsation of the amplitude were observed in October and November as well as the data at Shigaraki. The profile of the next year in 1985 showed a similar variation found as in 1984.

The fine structure of transition around March, however, was not simultaneously observed because of the less frequent observations in 1985 at Adelaide and in 1984 and 1986 at Shigaraki. Therefore, we compare the profiles in 1984–1986 between the results at Adelaide and Shigaraki as shown in Figure 4.13. The phase of the eastward component observed at Adelaide changed rapidly in the end of February in both 1984 and 1986, being accompanied with the rapid decrease of the wave amplitude. The phase after the transition seems variable probably because of the small amplitude, while the center of the phase seems to be 0000–0400 LT. The phase of the eastward component at Shigaraki in 1985 also showed the rapid change in the end of February as described before, thus the change around the equinox near March is considered to coincide between Adelaide and Shigaraki, if the interannual variations are not significant.

In summary, the transitions from summer-like months to winter-like months at Shigaraki occurred in the beginning of October and ended in less than one month. The periods of transition agreed well between Shigaraki and Adelaide, but were different between Saskatoon and Shigaraki. At Saskatoon the transition occurred in the first half of November. This delay is considered as a latitudinal difference of tidal structure between 52° and 35° . The transition at Saskatoon and Poker Flat (65°N) were reported as they occurred at the same time [Manson et al., 1988]. The transition from winter-like months to summer-like months also showed a rapid change for both Shigaraki and Saskatoon, but the transition was earlier at Shigaraki than at Saskatoon, that is, the end of February and the beginning of March for the former, and the middle of March for the latter. Again the periods of transition coincide between Adelaide and Shigaraki.

If we interpret these time differences of transition as the transition of two single modes for summer-like and winter-like months, the difference can be occurred by the latitudinal difference of the amplitude of a single Hough mode. Now we assume two modes; one is dominant in summer-like months and has

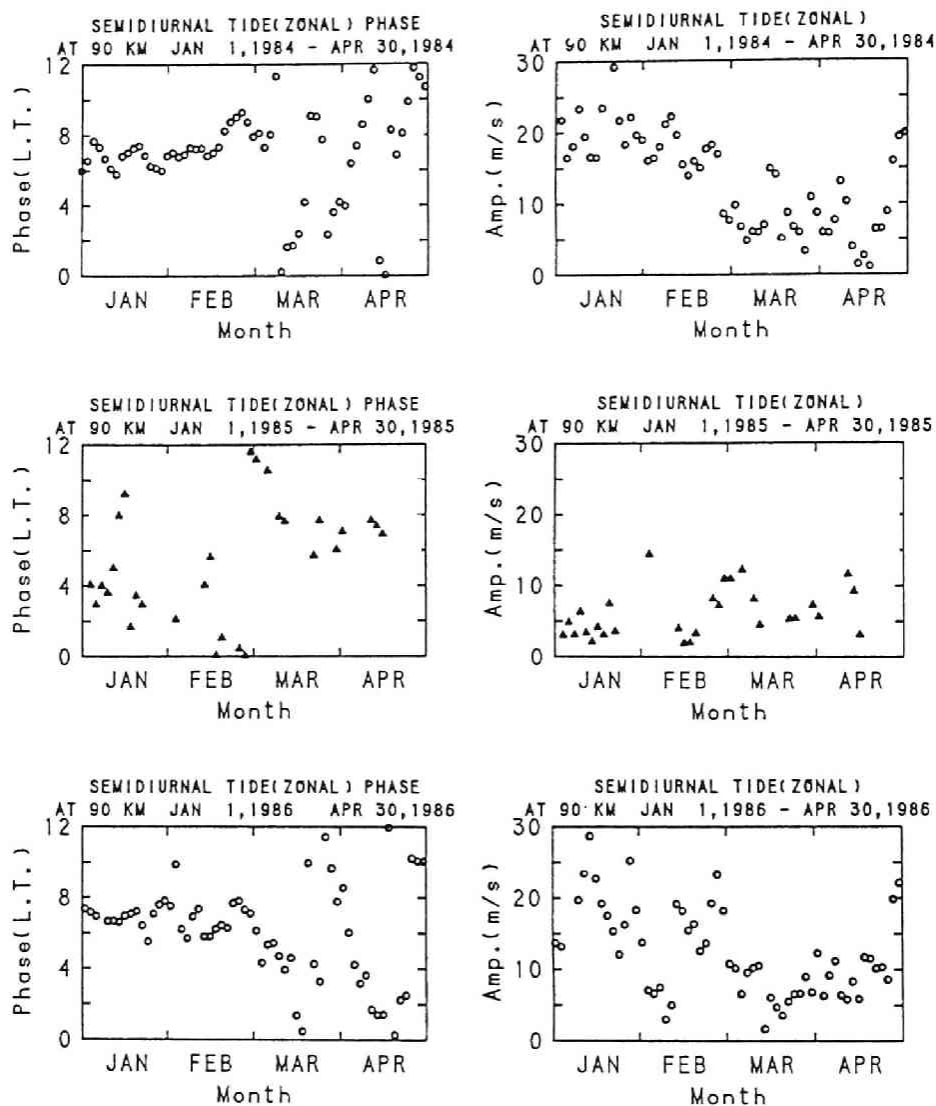


Figure 4.13: Fine structure of phase and amplitude variation of the eastward component in January–April in 1984 (top) at Adelaide, 1985 (middle) at Shigaraki and 1986 (bottom) at Adelaide. The left and right panels correspond to the phases and amplitudes, respectively.

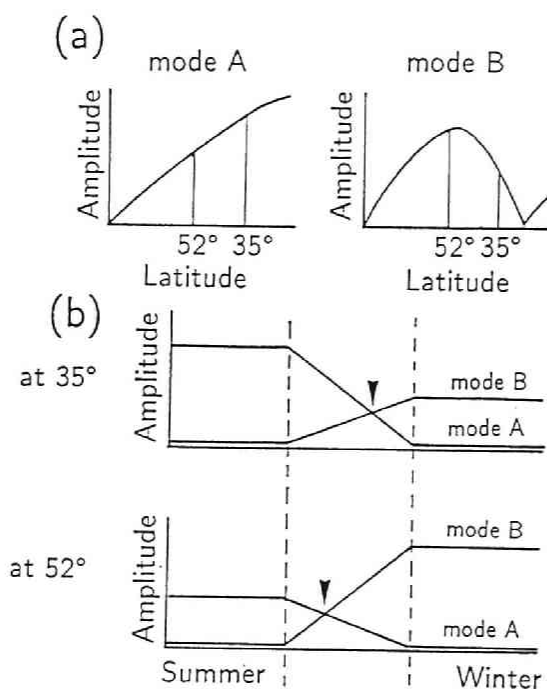


Figure 4.14: An illustration of a transition between two single tidal modes.
 (a) Latitudinal distribution of the amplitude of the wind of each mode.
 (b) The amplitude of each tidal mode as a function of time.

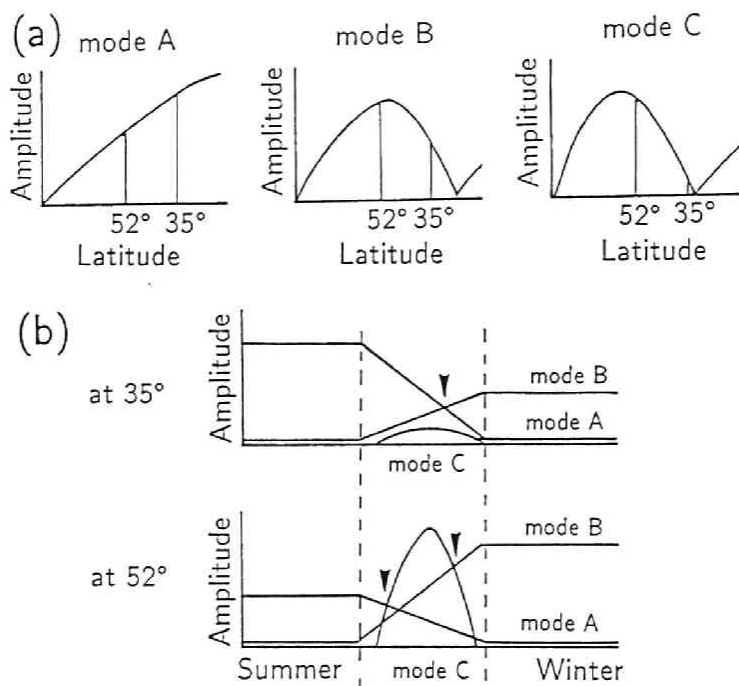


Figure 4.15: An illustration of a transition among three single tidal modes. (a) Latitudinal distribution of the amplitude of the wind of each mode. (b) The amplitude of each tidal mode as a function of time.

large vertical wavelength (i.e. basic mode) and the other is dominant in winter-like months and has comparatively small wavelength (i.e. higher mode). The amplitude of the higher mode semidiurnal tide usually has a larger amplitude at 52°N than at 35°N , as shown in chapter 1. Then, at the transition from summer-like months to winter-like months, the phase of semidiurnal tide can become winter-like earlier at 52°N than 35°N , as illustrated in Figure 4.14 (b). This situation does not agree with the observational results shown above; earlier transition at Shigaraki. Furthermore, in spring the transition was again earlier at Shigaraki, and earlier transition at both winter to summer and summer to winter transition cannot be explained by the exchange of two single modes with amplitude difference at different latitudes.

Another feature of the transition periods is that in September and October the vertical wavelength already became shorter than 60 km and close to the winter value at Saskatoon as shown in Figure 4.8, which was smaller than that in June to August. On the contrary, wavelength at Shigaraki showed longer than 100 km in September–October and April and May, showing similar wavelength as in mid-summer. These profiles suggest that in equinoxes the higher order mode already emerged around 90 km, while the phase at 90 km did not change from the value of summer-like months. The large vertical wavelength at Shigaraki at equinox is probably due to the small amplitude of Hough mode at 35°N which is enhanced in the equinox.

This situation is illustrated in Figure 4.15. In this case we assume the mode C with shorter wavelength than the basic mode A, which has the minimum amplitude at around 35° latitudes. Then, we further assume that this mode becomes significant only in the equinoctial months between summer and winter. Then the first transition at 52°N from mode A to mode C is occurred earlier than the transition at 35°N , and the second transition appears later than the transition at 35°N . If the phase of mode C is close to that of mode A at 90 km altitude, the transition from mode A to mode C is not apparent at the phase plot at 90 km. Moreover, this situation can naturally explain the short vertical wavelength in September and October and phase advance by two hours before the major transition in November. Furthermore, the situation of the three modes shown in Figure 4.15 can explain the vernal transition, as well, if the mode C has the phase value closer to mode B at 90 km for vernal equinox. Then, the earlier phase transition at Shigaraki than at Saskatoon in spring can occur (see Figure 4.15 (b) from right side to left side).

Forbes and Vial [1989] showed the Hough function decomposition of the simulated semidiurnal tide as shown in Figure 4.16. As well as fundamental (2,2) mode, many higher modes of symmetric and antisymmetric modes exist. In the equinoxes, (2,4) mode attains the large amplitude in March–April and

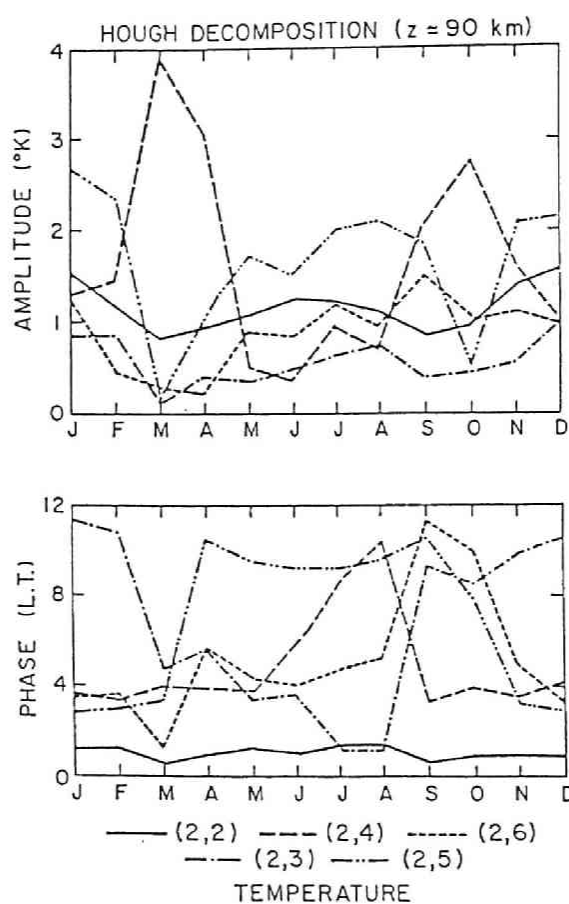


Figure 4.16: Amplitude (top) and phase (bottom) vs. month corresponding to Hough components of calculated semidiurnal temperature at 90 km [Forbes and Vial, 1989].

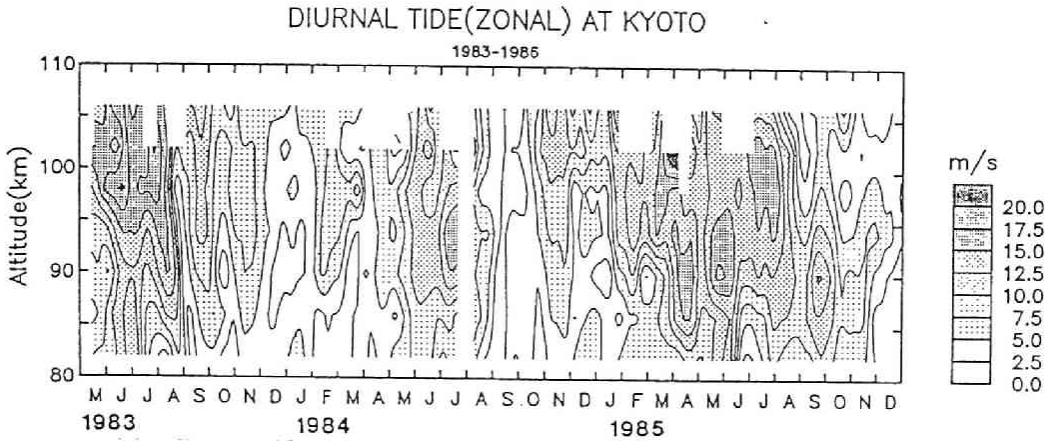


Figure 4.17: Contour plot of the amplitude of the zonal component of diurnal tides observed by Kyoto meteor radar. Harmonic analysis using 30-day observation intervals is applied. Fluctuations with periods shorter than 18 days are filtered out.

in September–October. The amplitude of (2,4) mode is larger at 52°N than at 35°N and so that the effect of the equinoctial enhancement of this mode does not affect the phase profiles at 35° latitude. However, the model by Forbes and Vial did not express the summer and winter difference of the vertical wave length enough, and neither the winter phase of 35° latitude. The modes enhanced in winter are similar to those enhanced in summer months, which can be the reason for that discrepancy.

The closer discussion of the transition of the winter-like and summer-like months observed between Saskatoon and Shigaraki suggests that if another or several higher modes can be considered in the equinoctial months rather than two single mode transition, the observed transition at Saskatoon and Shigaraki can be explained. The dominance of higher mode at equinox is also suggested by the Forbes and Vial model, and this could support the three mode model as described in association with Figure 4.15.

4.3 Diurnal Tide

4.3.1 Climatology at Shigaraki

We first present the climatology of diurnal tide observed at Shigaraki by the same way as the previous section. Figure 4.17 shows time-height cross section of

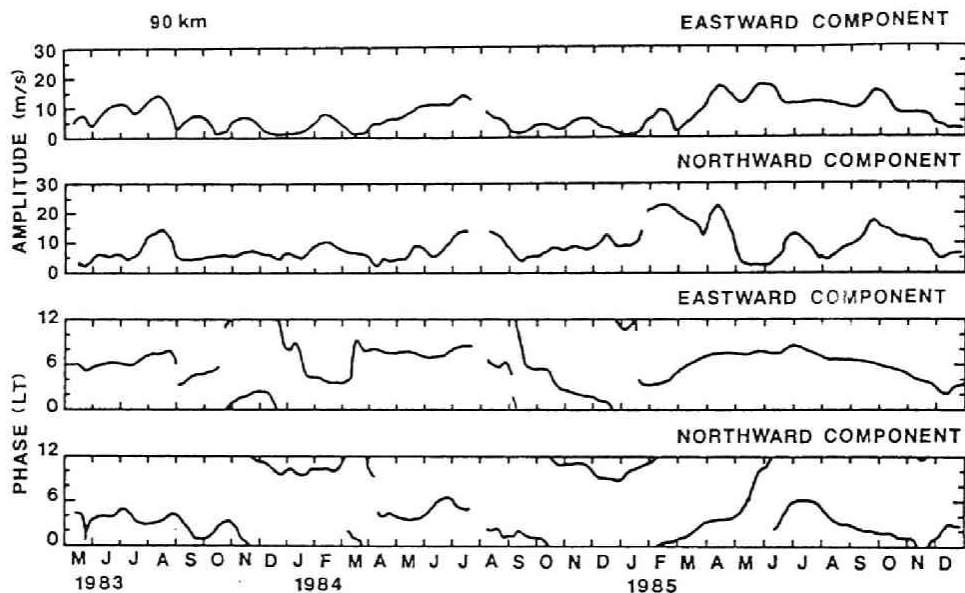


Figure 4.18: Amplitudes and phases of semidiurnal tides from Kyoto meteor radar observation at 90 km altitude.

amplitude of the zonal component of diurnal tides. The analysis was the same as semidiurnal tide shown by Figure 4.1 in the previous section; that is, a harmonic analysis was applied using 30 days of observations, and low pass filter with a cut off period of 18 days was used. The amplitude of the eastward component of diurnal tide became large above 90 km in May to August of summer in 1983 and in 1985, while smaller than the other two years in 1984. There was also enhancement of the amplitude in February–April 1985, late September to early October in 1983 and 1985.

Figure 4.18 shows amplitudes and phases of the diurnal tides at 90 km altitude for eastward and northward components. The phase of eastward component did not vary largely from April to August in these three years, being about 1200 LT in 1983 and about 1600 LT in 1984 and 1985. On the other hand, in the other months the phase was not confined to a small range but rather varying gradually. It seems that in December for three years, the phase came to 0400 LT, and that the phase was decreasing slowly from September to December. In January, the phase decreased rapidly from 0400 LT to about 0800 LT in 1984 and 1985. These phases in winter jumped up to 1600 LT in March for 1984, while it changed slowly to 1600 LT from February to April in 1985.

Thus, the phase of diurnal tide did not show clear bi-modal feature as is seen

for semidiurnal tide. The phase was stable for summer in April–August, and then it became earlier until December. The spring transition showed large variability in the two years. The northward component shows similar stability of the phase in summer from April to August, although the phase sometimes became unstable such as from May to June in 1985. The phase was about 0800 LT for these three years in summer months. From December 1983 to February 1984 and from late October 1984 to January 1985, the phases were comparatively stable at around 2000 LT, which were more stable than those of eastward component. This is probably because the amplitude of northward component in winter was larger than that of eastward component in the same figure. The phase transition from September to December was again a very slow decrease and that in spring was abrupt in April 1984 and gradual in February–April in 1985. The phase difference between eastward and northward component was 4–9 hours generally showing quadrature phase relation, while in winter sometimes irregular; e.g. December 1983 to March 1984.

Thus, seasonal variation of the phase of the diurnal tide at Shigaraki did not show a clear two-state variation like that of semidiurnal tide, but rather showed gradual variation and large year to year variation.

4.3.2 Hemispheric Asymmetry between Shigaraki and Adelaide

In this subsection, we describe the comparison of the diurnal tides between Shigaraki and Adelaide, which are located in the geographic conjugate to each other, and discuss the hemispheric symmetry. The comparisons have been reported by Vincent et al. [1988], so we here summarize their results.

Figures 4.19 and 4.20 show the monthly mean amplitudes and phases of the diurnal tide at 90 km altitude at Adelaide and Shigaraki [Vincent et al., 1988].

The amplitudes of diurnal tide at Adelaide became maxima in autumn (March–April) and spring (September–October) for both northward and eastward component. The maximum amplitude was observed to be 20–40 m/s with a significant interannual variation. Secondary peak of the amplitude can be seen in late winter (August) in 1984 and 1985. In December–January and May–June, amplitudes became as small as 10–15 m/s.

At Shigaraki the amplitudes of eastward component had a maximum in late summer as shown in Figure 4.17, and those of northward component became maxima in spring (February–April) and late summer/early autumn (July–September), respectively, with the maximum amplitudes of 10–20 m/s for both northward and eastward components as shown in the previous subsection.

The amplitude of diurnal tide shows the significant interannual variability

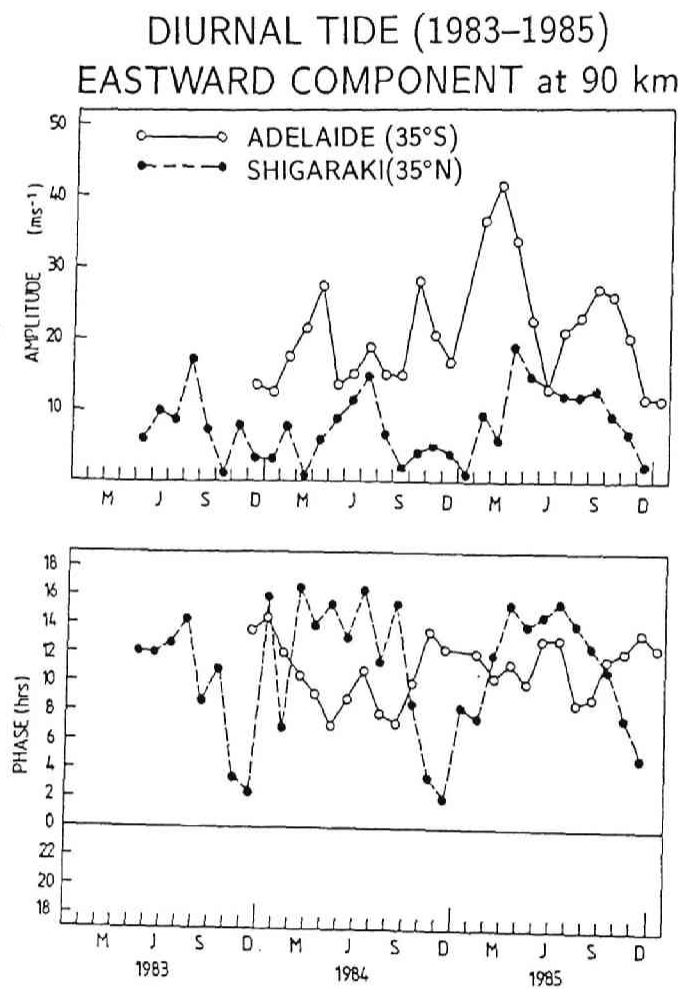


Figure 4.19: Amplitude and phases of the zonal component of the diurnal tide at 90 km (monthly mean). The phase means the local time of eastward wind maximum [Vincent et al., 1988].

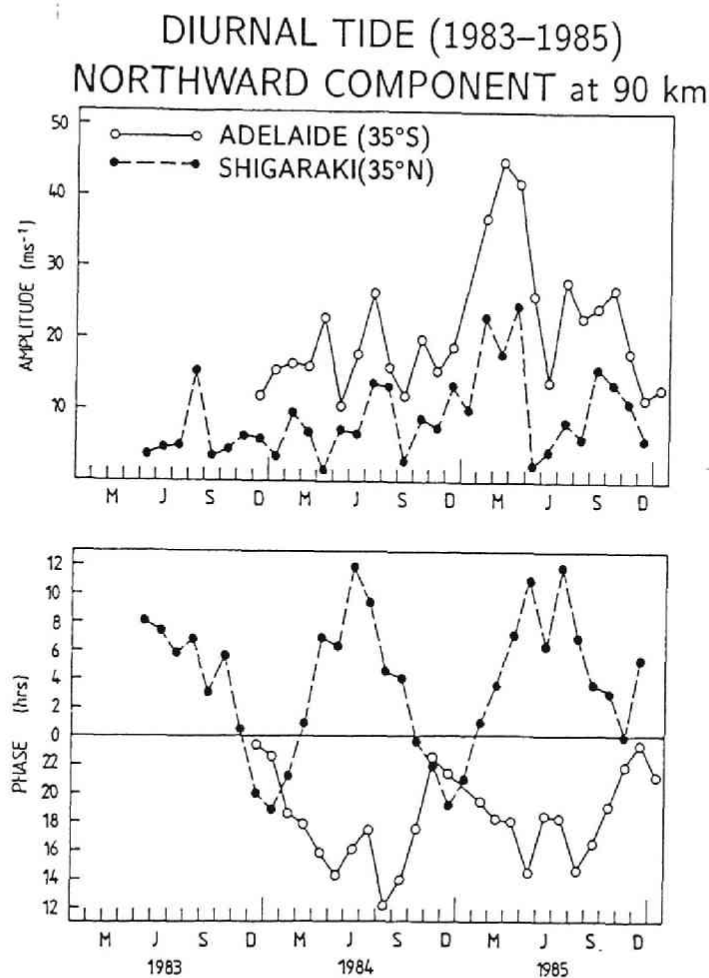


Figure 4.20: The same as Figure. 4.19, except for the meridional component. the phase is the local time of northward wind maximum [Vincent et al., 1988].

	December		June		Equinox	
	EW	NS	EW	NS	EW	NS
$\phi_A(LT)$	13	23	11	18	9	16
$A_A \text{ (ms}^{-1}\text{)}$	15	15	15	18	30–40	30–40
$\phi_K(LT)$	2	19	15	10	13	3
$A_K \text{ (ms}^{-1}\text{)}$	5	5–10	10	5	5–15	10
$\phi_K - \phi_A \text{ (h)}$	-11	-4	4	-8	4	-11

Table 4.1: Arithmetic mean amplitudes A and phases ϕ of the zonal and meridional diurnal wind components observed at 90 km at Shigaraki and Adelaide. The negative sign for the phase differences ($\phi_K - \phi_A$) means that the phase at Shigaraki leads that at Adelaide [Vincent et al., 1988].

at both location. In 1985 annual mean amplitudes were 40–60 % larger than those observed in 1984. The enhancement was conspicuous in February to April for both northward and eastward components. The variations of the amplitudes of northward components at the two locations had a better correlation than those of the eastward component. Furthermore, the most notable feature is that the amplitude of diurnal tide at Adelaide was considerably larger than that at Shigaraki by a factor of about 2, which supports the earlier findings by Aso and Vincent [1982].

The phase of the northward components at both Adelaide and Shigaraki show a clear annual variation, which were 6 months out of phase between the hemisphere. A similar variation can be seen for the amplitude of eastward component, but less strong.

It is found that the phase differences between the two locations changed with season. At the June solstice, the phase at Adelaide was about 6 hours earlier and 4–6 hours later than that at Shigaraki for eastward and northward components, respectively. On the contrary, at the December solstice, the phases at Shigaraki were earlier by 8–10 hours and about 2 hours for eastward and northward components, respectively. The duration of the phase of June solstice lasted longer than that of December solstice, which is more evident in the eastward component, suggesting the asymmetry between the northern and southern hemispheres [Vincent et al., 1988].

Table 4.1 shows a summary of the seasonal mean phase and amplitude of diurnal tides observed at 90 km. The quantities were obtained by arithmetically averaging the relevant amplitudes and phases.

Vincent et al. [1988] further examined the vertical profiles of the amplitudes

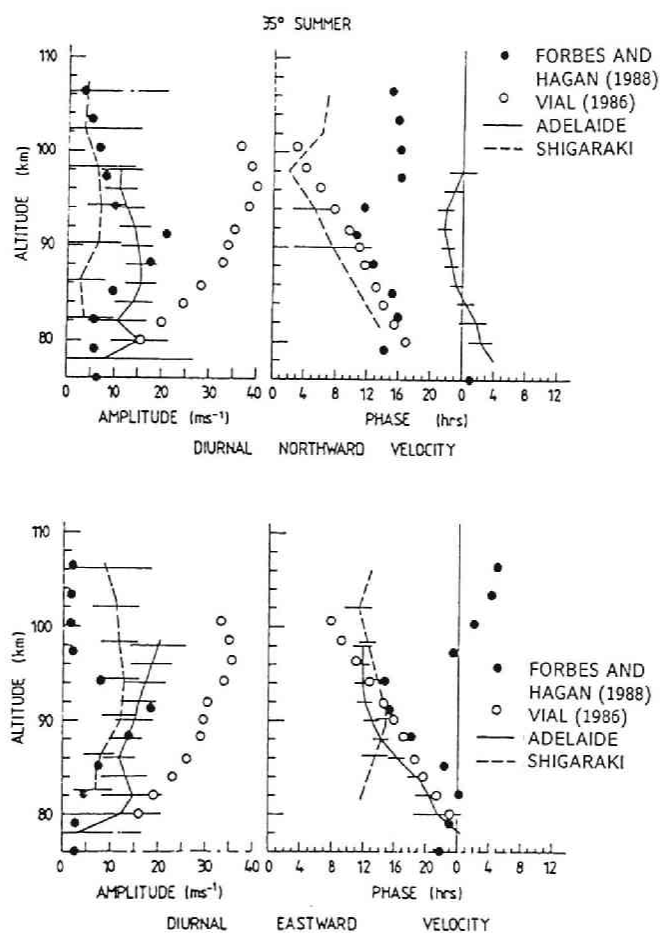


Figure 4.21: Height profiles of the amplitudes and phases of diurnal tide in summer at Shigaraki and Adelaide [Vincent et al., 1988].

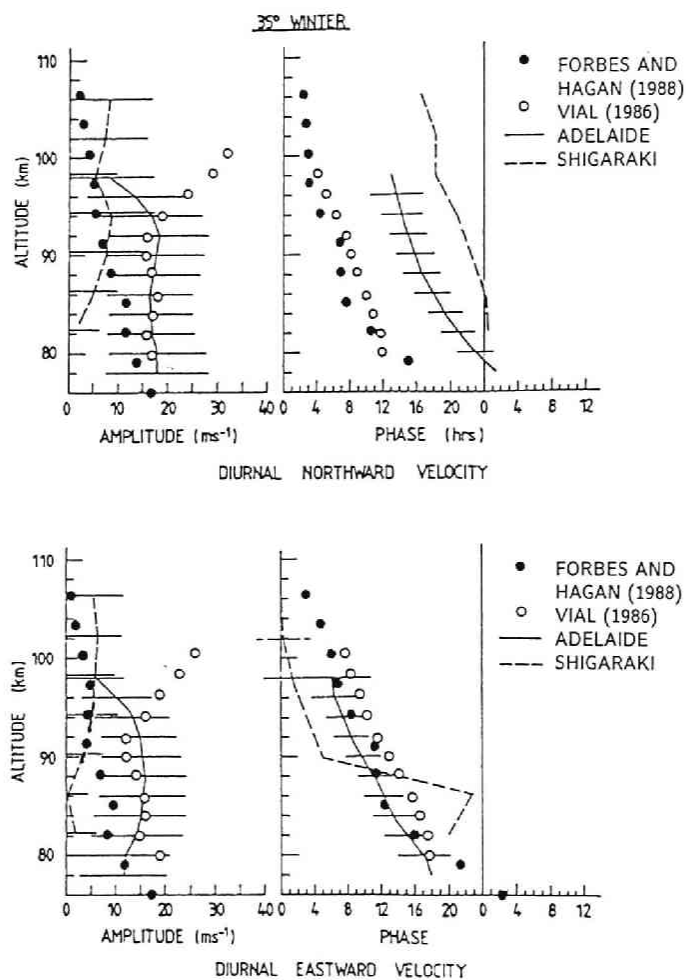


Figure 4.22: The same as Figure. 4.21, except for winter [Vincent et al., 1988].

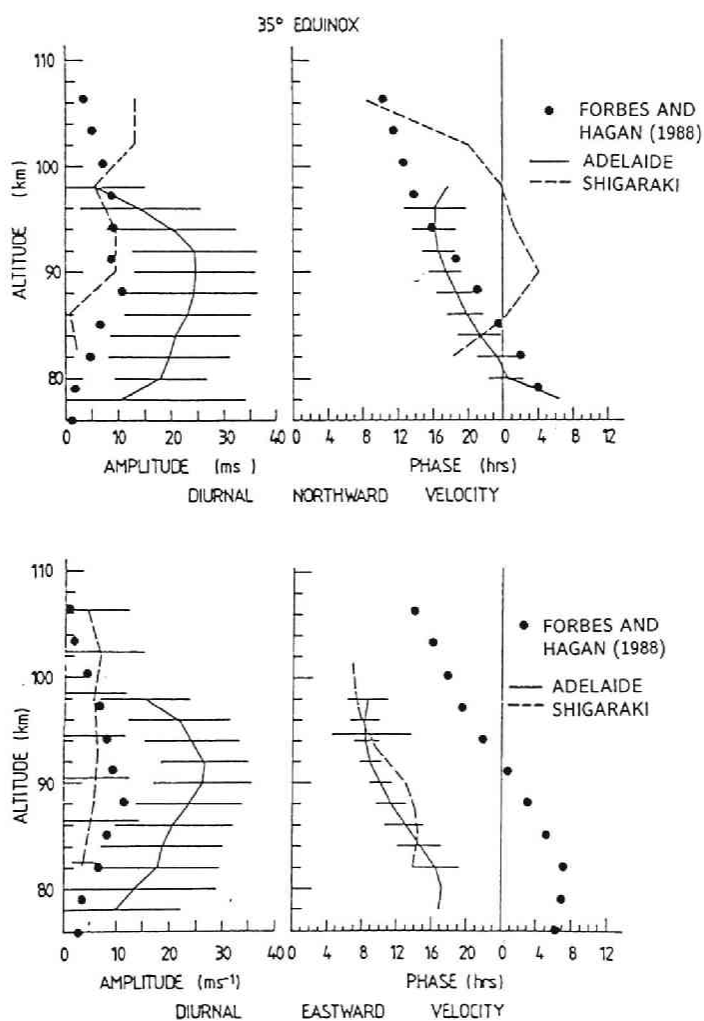


Figure 4.23: The same as Figure. 4.21, except for the equinoxes [Vincent et al., 1988].

	Summer		Winter		Equinox	
	EW	NS	EW	NS	EW	NS
Adelaide						
height < 90km	40km	85km	40km	40km	35km	35km
height > 90km	>100km	<i>E</i>	60km	80km	85km	80km
Shigaraki	<i>E</i>	35km	...	60km	40km	...

Table 4.2: Seasonal mean vertical wavelength λ_z for the diurnal tide. *E* signifies evanescent, and three dots signify indeterminate values [Vincent et al., 1988].

and phases of diurnal tides in summer, winter and equinoctial months in order to clarify the representativity of the climatology at 90 km. Figures 4.21, 4.22, and 4.23 show the averaged profiles for summer, winter, and the equinoxes, respectively, which correspond to the duration of May–August, November–February and the other equinoctial months. The amplitudes and phases were deduced by a vector average of the appropriate monthly mean values from 1983 to 1985, so that approximately 8-9 months of data were averaged for each profile. The bars show the degree of variability, although they were omitted from the phase profiles where the standard deviations were large. Note that a vector average is likely to cause the underestimation of the amplitude, but the amplitudes shown in the diagrams are not significantly affected, which can be recognized by the comparison of Figures 4.21 to 4.23 with the arithmetic mean amplitudes given in Table 4.1.

Results of models of Vial [1986] and Forbes and Hagan [1988] are also plotted for 35° latitude in Figures 4.21, 4.22, and 4.23. Note that the model of Vial provides the tidal parameters for June solstice, and the model of Forbes and Hagan [1988] describes December solstice. So that the phases of the northward component were shifted by 12 hours in order to show a reference for the northern hemisphere, although this assumes the tidal symmetry between northern and southern hemisphere [Vincent et al., 1988]. However, the phases of northward components observed at Adelaide were not added by 12 hours, so that the 12 hours discrepancy between the model and Adelaide observation means a good agreement between the model and observation for northward component.

It is evident that the amplitudes at Adelaide were always larger than those observed at Shigaraki at all heights. The phase profile seems smoother for the Adelaide data than those of Shigaraki, probably because of the small amplitude for the latter data. The amplitudes at both locations agreed better with

the Forbes and Hagan model than the Vial's one in summer, while in winter the amplitude at two locations are largely different and better agreement was observed with Vial's model below 94 km for the amplitudes at Adelaide. The phases at Adelaide agreed very well with both of the two models, as a comparison of absolute values. The observed phases were slightly earlier by about 2 hours. The vertical phase variation shows generally downward propagation of phase, especially below 90 km. If we assume a single tidal mode, the inferred vertical wavelength was about 35 km suggesting a strong contribution of the (1, 1) mode below 90 km, but above 90 km, the phase gradient became smaller indicating a feature of an evanescent mode. This characteristics at Adelaide was also observed at Shigaraki in summer and is also shown in the Forbes and Hagan model. However, for the phases of eastward component agreement between the observation at Shigaraki and the model is less clear than that for the northward component. The phases of zonal component indicated basically evanescent in summer, while in winter, phase gradient greatly changed at 86 km suggesting interference of several modes. The phase of the northward component agreed better with the results at Adelaide and with the model values, although the observed phases were earlier than the models by 4–8 hours.

At the equinoxes shown in Figure 4.23, the amplitudes at Shigaraki agreed well with the model of Forbes and Hagan, but the amplitudes at Adelaide were significantly larger than the model. The phases of eastward components observed at both locations show very similar variations, while the model of Forbes and Hagan shows a quite similar vertical gradient but the absolute phase values are different by 12 hours. The phases of the meridional components did not agree as well as the zonal component between two locations, while the phases at Adelaide agreed well with those by the model. It is noteworthy that below 90 km in the northward component, the amplitude at Shigaraki were as small as 2 m/s, indicating the phase value below 90 km in the northward component were not meaningful. The comparison of the phase of northward component above 90 km shows that the phase gradient at Shigaraki was similar to those at Adelaide and by model, while the absolute phase value was larger by about 10 hours. Mean values of λ_z are given in Table. 4.2.

Vial [1986] and Forbes and Hagan [1988] have shown that diffusive effects, that is the effect of the dissipation by the large diffusion of the atmosphere, are also important to determine the structure of tide as well as background mean wind distort the horizontal structure of the (1, 1) mode and produce changes in the local vertical wavelength. The effects of dissipation are more significant for the (1, 1) mode than in the lower-order semidiurnal modes because of the relatively short vertical wavelength (~ 25 km) where the dissipation does not occurred. Figure 4.24 shows the latitudinal structure of the amplitude and phase

CHAPTER 4. ATMOSPHERIC TIDES AND TWO-DAY WAVE

DIURNAL EASTWARD VELOCITY $Z = 90\text{ km}$

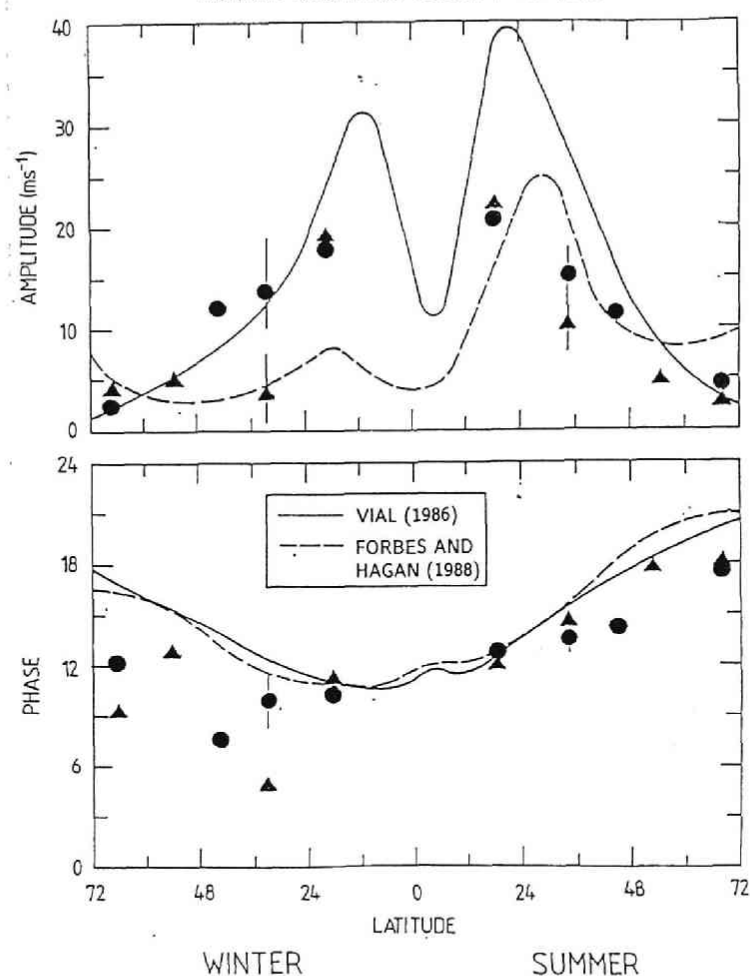


Figure 4.24: Latitudinal structure of the diurnal amplitudes and phases at 90 km from the models of Vial [1986] (solid line) and Forbes and Hagan [1988] (dashed line) compared with observed values at Adelaide (circles at 35° latitude) and Shigaraki (triangles at 35° latitude), together with the other MF and meteor radars. Circles and triangles correspond to southern and northern hemispheres [Vincent et al., 1989].

of eastward wind component of the diurnal tide at the solstices by Vial [1986] and Forbes and Hagan [1988] together with the results observed at Adelaide and Shigaraki at 90 km. The results of other MF and meteor radars are plotted as well; Poker Flat (65°N), Saskatoon (52°N), Puerto Rico (18°N), Townsville (19°S), Christchurch (44°S) and Mawson (68°S) [Vincent et al., 1989]. The models express larger amplitude in the summer hemisphere, which used to be symmetrical in the earlier models [e.g., Forbes, 1982a]. These new models also showed the phase difference in summer and winter, indicating that the phase of the eastward maximum in winter is earlier than that in summer. During the propagation through the middle atmosphere, the vertical wavelength became shorter and longer due to the Doppler shift by westward and eastward mean winds, respectively [Forbes and Hagan, 1988; Forbes and Vincent, 1989]. The seasonal variations of the phases observed at Adelaide and Shigaraki are similar to the prediction by the models of Vial [1986] and Forbes and Hagan [1988], and the observational results of the other latitudes also indicates the same sense as the models; early phase in winter and late in summer.

The difference of diurnal tide observed between Adelaide and Shigaraki is probably caused not only by hemispheric differences of mean wind and latitudinal temperature gradient but also by hemispheric differences in dissipation. Vincent et al. [1988, 1989] pointed out that the difference of dissipation between the northern and southern hemisphere is more important to explain the hemispheric difference. The diurnal tide at Adelaide showed the larger amplitudes and shorter vertical wavelengths than that observed at Shigaraki, which suggests that the diurnal (1, 1) mode at Adelaide is less significantly damped than at Shigaraki. The conspicuously large amplitudes observed in spring and autumn at Adelaide (Figure. 4.23) imply small dissipation in these seasons. Since the most of mesospheric turbulence is generated by breaking gravity waves [e.g., Lindzen, 1981; Fritts, 1984], seasonal and hemispheric variations of gravity wave activity could be expected to cause similar variations in eddy dissipation which are generated by turbulence. Vincent et al. [1988] further discussed that the large hemispheric asymmetry in land and sea distributions could be expected to generate a corresponding asymmetry in wave-induced turbulence in the mesosphere, at least in winter when low-phase speed waves are able to penetrate through the middle atmosphere winds easily.

The gradual change in the seasonal variation of the phase at 90 km altitude at both Adelaide and Shigaraki shown in Figures 4.19 and 4.20, which is different from the persistence of two phase values of the summer-like and winter-like months observed for semidiurnal tides, suggests that the modulation of single diurnal (1,1) mode throughout a year as discussed above is probable than the exchange of multiple modes.

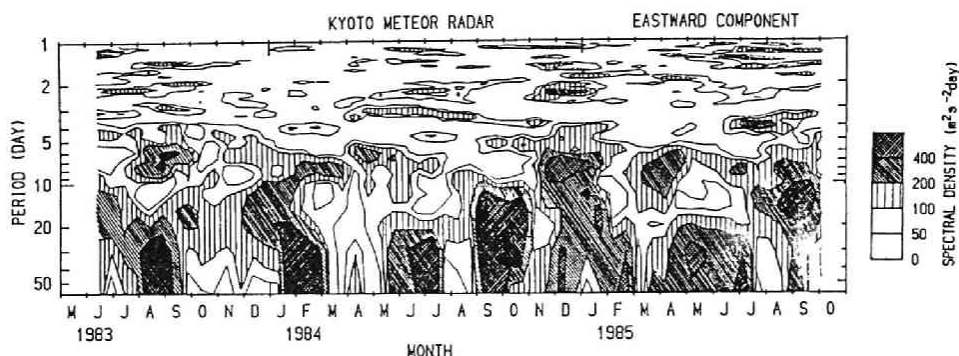


Figure 4.25: A moving power spectrum of the eastward wind component observed in 1983–1985 by the Kyoto meteor radar. Each spectrum is determined by calculating an autocorrelation function with a maximum time lag of 30 days [Tsuda et al., 1988b].

To summarize, there are a large difference between the northern and southern hemispheres in diurnal tides observed at Shigaraki and Adelaide even in the same season of the year, as well as the asymmetry between the summer and winter hemispheres at the same time. The difference between summer and winter is explained fairly well by the recent models by Vial [1986] and Forbes and Hagan [1988]. However, the hemispheric difference between northern and southern hemispheres in the same seasons is not included by those models and hence not expressed. The difference of the dissipations due to eddy viscosity owing to turbulence, which is mainly generated by breaking gravity waves in the mesosphere, is one possible candidate for the difference of the northern and southern hemispheres of diurnal tides. Therefore, the observational study of gravity wave activity and the comparisons between these observations at different locations especially between the northern and southern hemispheres, are quite important. These studies will be presented in Chapter 5 and 6.

4.4 Two-day Wave

In this section, we first present the observation of two-day wave at Shigaraki and Adelaide according to the results of Tsuda et al. [1988b], and then describe further analysis of the observation in 1983–1986 at both locations.

4.4.1 Spectral Analysis

Figure 4.25 shows a dynamic spectrum of eastward wind oscillations with periods longer than 1 day observed with Kyoto meteor radar observations in 1983–1985, which was calculated by shifting periods of the analyses of frequency spectra [Tsuda et al., 1988b]. Each frequency spectrum was calculated from the data of 60 days observations where the autocorrelation function with a maximum lag of 30 days were calculated and further processed to determine frequency spectra.

The two-day wave is found in summer months with a fairly large interannual variation in both amplitudes and wave periods. The active periods of this wave is found between early summer and autumn in 1983 and 1984. In 1985, the wave intensity was small in comparison with the other two years. Furthermore, the wave period of this year varied largely from about 2.1 days in spring to 2.5 days in July, and then decreased again in late summer and in autumn. The wave with periods of 2.3–2.4 days was also enhanced from December 1984 to January 1985. The two-day wave shown here are discussed in more detail in the following subsections.

Wind fluctuations with a period of 1.4 days emerged in summer months such as in July–September 1983, June–July 1984 and June–August 1985. Some other oscillations with periods shorter than 1.4 days can also be seen in Figure 4.25, although the behavior of these components is difficult to discern.

An oscillation with period of 3 day was enhanced in August–September 1983, and the period increased with time from 2.9 days in August to 3.0 days in September. This oscillation was again observed in April–August 1984, with much larger amplitude than the previous year. The wave period increased from 3.0 days in April to 3.4 days in August. The increase of the wave period seems to correlate with the decrease of the periods for the two day in both 1983 and 1984. If both of them are caused by a Doppler shift effect due to the mean wind which varies during these periods as shown in Chapter 3, these variations mean the different direction of horizontal phase velocities, although detailed discussions need the observation of horizontal structure. In 1985 an oscillation with a 3.4 day period is enhanced in August–September, corresponding to the waves with the same wave period in the same months in the previous year.

In summary, the waves with periods of about 3 days was detected in three successive years, and was enhanced in late summer in 1983 and 1985 and in the entire summer months in 1984. The waves with the period of 3 days were also observed in the meridional wind component in the same season of the year in 1978–1983 observation at Shigaraki [Ito et al. 1984]. However, it is not clear whether those waves with periods of near 3 days are identical to 4-day oscillation described below.

The 4-day oscillation was evident in June–August 1985. In June–September 1983, enhancement of the wave with period of 4.3–4.0 day occurred repeatedly.

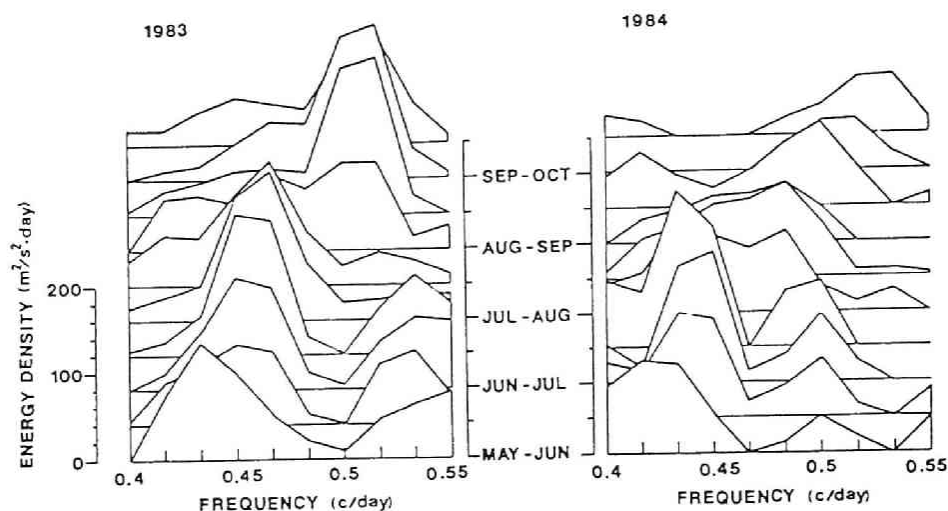


Figure 4.26: Power spectra of the eastward component of the two-day wave in 1983 and 1984. A spectrum is calculated every 15 days by using 60 days of data [Tsuda et al., 1988b].

In June–July 1984, an oscillation with 4.7 day period was found. Oscillations with periods of 4.0 days and 4.3 days were also detected in winter; in January–February 1984 and December 1984–January 1985, respectively. Thus, the oscillation with periods of 4 days are often observed in the summer months, and also in winter months.

It is more difficult to determine the wave component with periods longer than 5 days because of the frequency resolution corresponding to an observation duration. However, several oscillations are found as followings. In August–September 1983, oscillations with wave periods of 6–8 days were enhanced. A weaker enhancement with a period of 8 days was detected in August–September 1985, but no such waves can be seen in the corresponding months in 1984. Oscillations with periods of 7–8 days emerged from February to March in 1984, and were also enhanced in winter 1984 centered in December.

Oscillations with periods of 10–20 days were enhanced in August–September 1985, and in September–October 1984, respectively. The components with the periods longer than about 20 days seem to represent transience of mean zonal winds as a seasonal variation.

In the stratosphere various periods of planetary waves were observed by the satellite observations. Among these, several waves are identified as normal mode Rossby waves such as 5 days wave of symmetric (1,1) mode, 4 day waves

of symmetric (2,1) mode, 10 day waves of antisymmetric (1,2) mode and 16 day waves of symmetric (1,3) mode [Madden, 1978; Hirota and Hirooka, 1984; Hirooka and Hirota, 1985]. However, the identification of the global structure of the waves observed in the mesosphere and lower thermosphere mentioned above in this section needs a comparison of the number of similar observations at various location, which is beyond the scope of this study.

Among these waves, the two-day wave was dominant in summer in 1983 and 1984 as shown in Figure 4.25. The activity in 1985 summer was smaller than in the preceding two years. Periods of the two day wave were 2.1–2.5 days in summer and become as short as 2 days in autumn, although there seems to be a relatively large year-by-year variation. Figure. 4.26 shows frequency spectra determined every 15 days using 60 days of data obtained in 1983 and 1984. The frequency resolution is 0.017 day^{-1} , which corresponds to 1.6 hours resolution in periods at around the periods of two days. The wave period was about 52 hours in June–August in 1983, and then abruptly decreased to 47 hours in September. In June–August in 1984, the wave period was 53–55 hours, which was slightly larger than that in 1983. The wave period again jumped to about 48 hours, and became shorter than 47 hours in October. Although the amplitude was small, an oscillation with a period of 2.5 days in June–August was in 1985 as shown in Figure. 4.25. Again the wave periods became shorter in autumn, in September–October.

The changes of wave periods of two-day wave coincided with the rapid decrease of eastward mean wind at the altitude of 85–105 km as shown in Chapter 3. Tsuda et al. [1988b] pointed out the possibility of the correlation of these wave period variations of two-day waves and zonal mean wind intensity in the lower altitudes, which can partly affect the mean wind at meteor heights, since the eigen frequency of the normal mode depends on the background atmosphere conditions [Salby, 1981a]. They also suggested that the wave period could vary by the structure of the stratospheric jet if it is related to the baroclinic instability [Plumb, 1983]. The Doppler shift of the mean wind at the observed altitudes can also affect the observed wave periods, as well. The eastward wind in summer causes the longer observed period for the westward propagating waves, so that the normal mode Rossby gravity waves were in this sense. Thus, the change of the wave period of two day wave between summer and autumn could be explained by assuming either model of Salby [1981a] and Plumb [1983].

4.4.2 Antisymmetry between Shigaraki and Adelaide

Figure. 4.27 shows typical height profiles of amplitudes and phases of the two-day wave observed in June 1983 by the Kyoto meteor radar [Tsuda et al., 1988b].

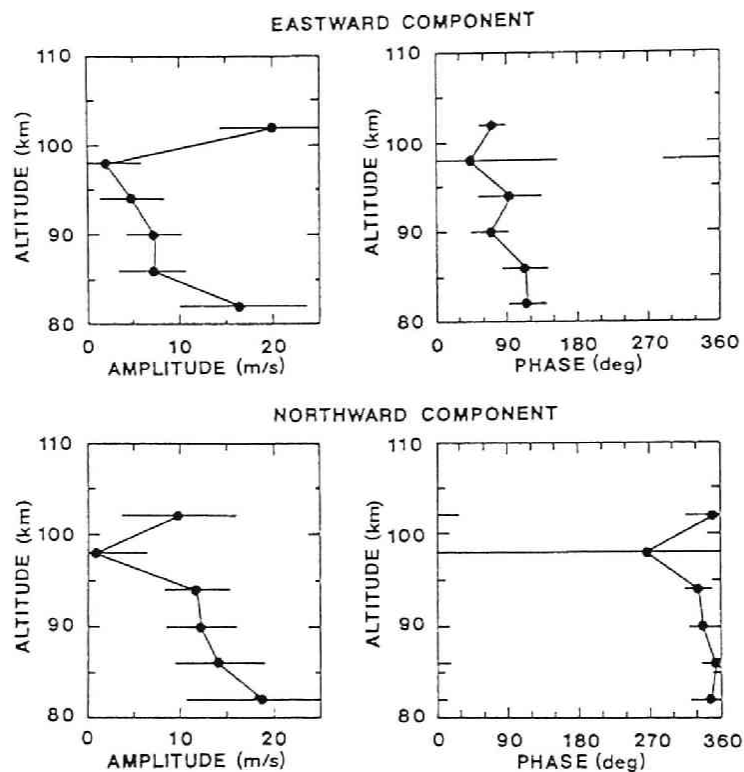


Figure 4.27: Amplitude and phase profiles of eastward (top) and northward (bottom) components of the two-day wave observed as 53 hrs by fitting a sinusoidal wave at each height. Phase correspond to relative local time of northward maxima, and their absolute values are not meaningful [Tsuda et al., 1988b].

The periodogram method is used to determine the wave period which was 53h in Figure. 4.27. The amplitude of the northward component was roughly twice as large as the eastward one and generally decreases with height from 20 m/s at 82km to 10 m/s at 102km. The phase progression suggests that the vertical wavelength was longer than 100km. The phase of the eastward maximum was delayed by about 120° relative to that for the northward maximum, indicating clockwise rotation of the wind vector and upward energy propagation. These features of the two-day wave agree well with the earlier determinations in August 1979 and July 1980 [Ito, et al., 1984], and are consistent with the general characteristics of the two-day wave [Vincent, 1984b].

Since the two-day waves become large in amplitude in summer in each hemisphere, the simultaneous observation between the hemispheres is not reported frequently. Craig et al. [1983] showed that the two-day wave observed at Sheffield (53°N) in August, 1980 could be simultaneously observed at Townsville (19°S), although the enhancement of the wave was not observed at Adelaide (35°S). They suggested that strong two-day wave excited in the northern hemisphere leaked into the southern hemisphere. Rodgers and Prata [1981] also showed from the upper stratosphere and lower mesosphere observations by satellite that this wave is essentially confined to the summer hemisphere. The difference of the wave periods between the northern and southern hemispheres were also reported by Vincent [1984b].

A conspicuously strong enhancement of two-day wave was observed in the southern hemisphere in January 1984 [Plumb et al., 1987]. During the same period, Kyoto meteor radar simultaneously observed the two-day wave. Figure. 4.28 shows vertical profiles of amplitudes and phases of the wave determined at Shigaraki and Adelaide on 18–31 January 1984 by using the sinusoidal curve fitting with a wave period of 48 hours [Tsuda et al, 1988b]. Note that the phases 0° and 180° in this figure correspond to 0000 LT on even and odd days in January 1984, respectively.

Amplitudes of the northward component at Adelaide were as large as 30–35m/s at 82–98 km altitude, which were reported as more than 50 m/s for an instantaneous value [Plumb et al., 1987]. The amplitudes observed at Shigaraki were less than 10 m/s and hence about $1/4$ of those at Adelaide. The amplitude of the eastward component observed at Adelaide was about 20 m/s, which was roughly a half of the amplitude of northward component. However, the eastward component at Shigaraki showed similar amplitude with that of the northward component. The vertical profiles of phases at Shigaraki and Adelaide show very similar height progressions, indicating a vertical wavelength of about 100km. The eastward and northward components were out of phase and in phase, respectively, between the determinations at Shigaraki and Adelaide.

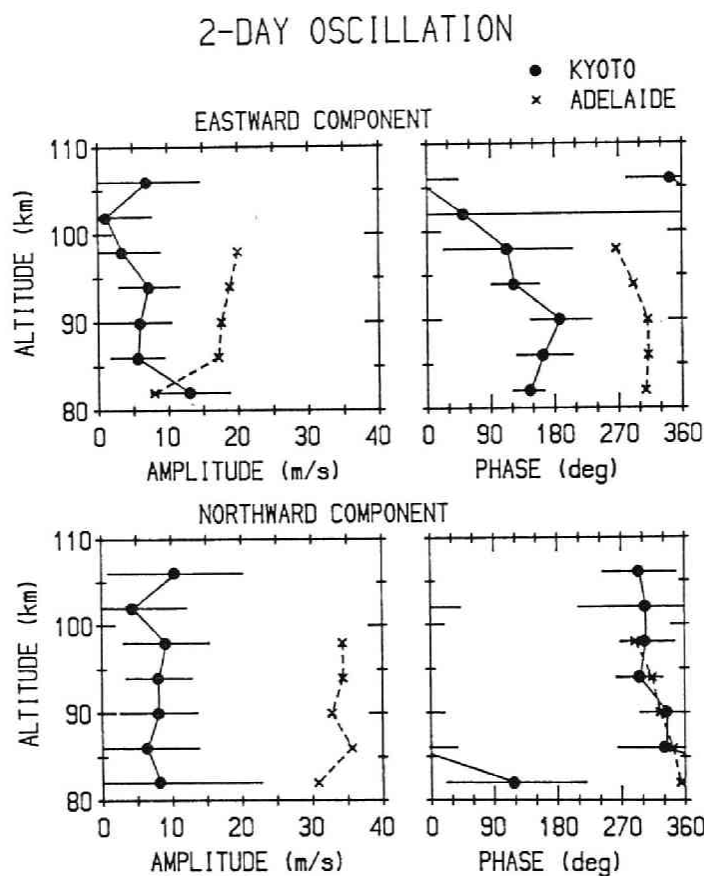


Figure 4.28: The same as Figure 4.27, but for the results from simultaneous observations between Shigaraki (full circles with solid lines) and Adelaide (cross symbols with broken lines) on 18–31 January 1984. The wave period is assumed as 48 h. Phase values of 0° and 180° corresponds to 00LT on even and odd days in January, respectively [Tsuda et al., 1988b].

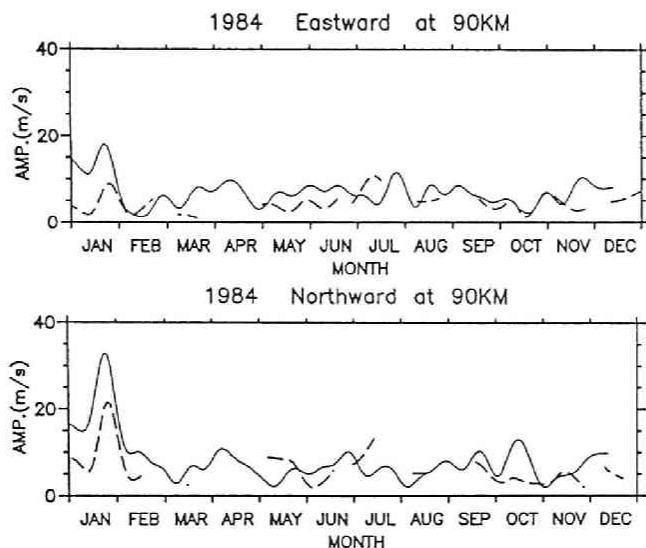


Figure 4.29: Amplitudes of two-day waves ($48 \text{ hours} \pm 3 \text{ hours}$) in 1984 observed at Adelaide (solid) and Shigaraki (dashed). The amplitudes are determined for 10 days of data, and further smoothed by a low pass filter with a cut-off of 20 days.

Tsuda et al. [1988b] showed that these phase relations suggest a manifestation of an antisymmetric normal mode oscillation, but that the difference of the amplitudes between hemispheres did not agree with a pure antisymmetric oscillation. However, this cooperative observation showed a clear evidence of the fact that the two-day wave is not entirely confined to the summer hemisphere, and that it is observed as an antisymmetric phase feature and asymmetric wave amplitude between the hemisphere. Salby [1981b] showed that the normal mode Rossby gravity wave of (3,0) have a larger amplitude at mesosphere in the summer hemisphere at solstice than the winter hemisphere due to the hemispheric difference of mean wind fields using a numerical model. Thus, a distortion of pure normal modes can be the reason of the hemispheric difference of the wave amplitude of two-day wave.

4.4.3 Interannual Variation of Two-day Wave between Shigaraki and Adelaide

So far we presented a case study of an event of two-day wave and its amplitude and phase profiles which suggests the contribution of antisymmetric normal mode according to the results of Tsuda et al. [1988b]. Next we examine the

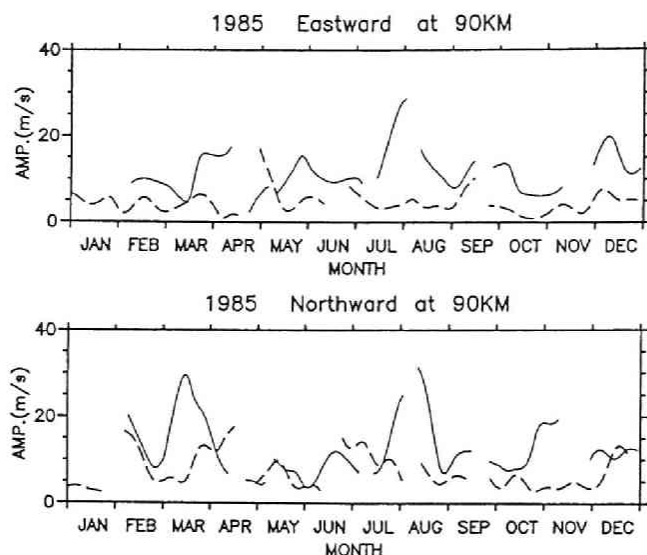


Figure 4.30: The same as Figure 4.29 except for the data in 1985.

behavior of two-day wave as a seasonal variation. Figure 4.29 shows the variation of the amplitude of two-day wave at 90 km altitude observed at Shigaraki and Adelaide in 1984. Two-day oscillations are determined from 10 days data by simultaneous fitting of diurnal, semidiurnal and 48 hours component. The determinations are made by shifting the observation periods by two days and the amplitudes are further smoothed by a low pass filter with a cut-off period of 20 days. Note that because the observation periods are comparatively short, the pass band of the fitting is wide enough to cover the wave periods of 45–51 hours.

The enhancement of the two-day wave in the latter half of January 1984 shown in Figure 4.28 is found to be the most significant event in 1984 for both Adelaide and Shigaraki in Figure 4.29. The determination was not successful when data acquisition was not sufficiently done or the observations were stopped by malfunction of radar in the months of March, April and July. Nevertheless, the enhancement were recognized in early July at Shigaraki with the amplitude around 10 m/s for both northward and eastward components, corresponding to the summer activity of the two-day waves, while the maximum of the amplitude was not clear because of the data gap in July. The corresponding enhancement at Adelaide was not manifest around July, but rather weak amplitude in comparison with the summer (January) peak were observed constantly.

The determination in 1985 were better and worse at Shigaraki and Adelaide than in 1984, respectively, as shown in Figure 4.30. Because of data gap in Jan-

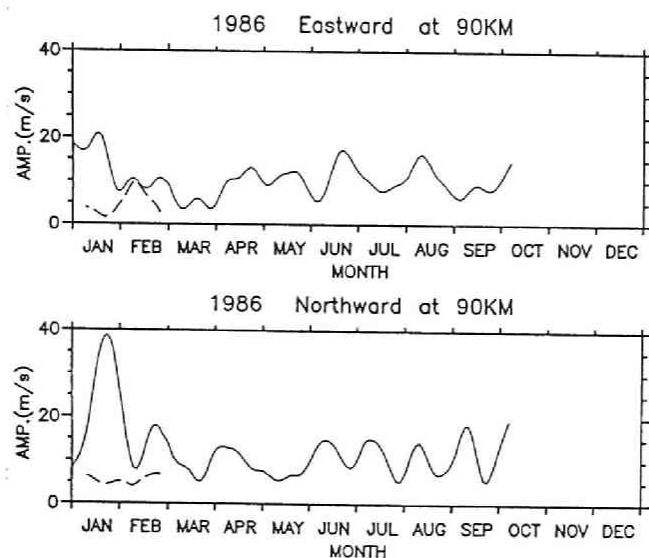


Figure 4.31: The same as Figure 4.29 except for the data in 1986.

uary 1985, the winter peak was not detected at Adelaide in this year. However, the enhancement of the amplitude at Shigaraki was not recognized in January this year. On the contrary, an enhancement with the amplitude of 15–20 m/s were detected in the northward component in February 1985 at both Adelaide and Shigaraki, while the peak was not clear because of the data gap in the beginning of February 1985. In this case the amplitudes of northward component were similar between Adelaide and Shigaraki. In June/July 1985, the amplitude of northward component at Shigaraki became as large as 15 m/s, showing summer enhancement, but eastward component was not so strong in summer months. This is probably the reason why the two-day wave was observed weaker than in 1983 and 1984 in the moving spectra of eastward component shown in Figure 4.25. The amplitude at Adelaide did not show an enhancement at the same time, but became as large as 30 m/s in the end of July and the beginning of August. Besides the large amplitude at solstice, there were some enhancements in March for Adelaide and in April for Shigaraki in this year, respectively. The simultaneous enhancement in September at both locations were also recognized for both northward and eastward component, while the amplitude was larger at Adelaide by a factor of two.

Figure 4.31 shows the similar amplitude variation in 1986, while the data at Shigaraki was limited to January and February. The amplitude of Adelaide again became as large as 40 m/s and 20 m/s for the northward and eastward components, respectively, in the latter half of January, while the corresponding

	$\overline{v'^2}$ (m ² /s ²)	$\overline{u'^2}$ (m ² /s ²)	$\overline{u'^2 + v'^2}$ (m ² /s ²)
Adelaide(35°S)	45.2	24.0	69.2
Shigaraki(35°N)	13.2	7.7	20.9
Ratio(A/S)	3.4	3.1	3.3

Table 4.3: Comparison of annual mean variance of two-day wave observed at Adelaide and Shigaraki in January 1984 – December 1985.

amplitude at Shigaraki was very small in this year.

Müller and Nelson [1978] reported from the observations of several meteor radars in the northern hemisphere that the amplitudes of two-day waves had maxima in summer and a secondary peak in winter with amplitude of 20 m/s and 6–7 m/s, respectively. Rodgers and Prata [1981] showed a similar variation in temperature of the upper stratosphere and the lower mesosphere at both southern and northern hemispheres at the low latitudes by satellite observations. Our observation at Shigaraki and Adelaide also showed the same tendency, but the summer peak was more conspicuous at Adelaide. Also there was a large interannual variability in both amplitude in solstice and other seasons. The peak of winter was not so large, and sometimes smaller than the peaks at spring and autumn, however, the peak at winter could become strong as shown in January 1984 at Shigaraki.

These comparisons between Adelaide and Shigaraki suggest that the two-day wave amplitude have a large interannual variability especially in winter, so that the enhancement of two-day waves are sometimes global and sometimes confined in the northern or southern hemispheres.

Annual mean wave variance of the two-day waves are compared in Table 4.3 between Adelaide and Shigaraki. The variances of northward and eastward wind component are deduced by the frequency spectra obtained in 1984–1985. At both location the variance of northward component was larger than that of eastward component. The total variance $\overline{u'^2 + v'^2}$ at Adelaide was 3.3 times larger than that at Shigaraki. Thus, the amplitude of two-day wave is 1.8 times larger in southern hemisphere than in northern hemisphere. The difference of the wave amplitude between summer and winter hemisphere can be explained by the difference of background mean wind between the hemispheres as shown in Salby [1981b]. However, to explain such a large difference of annual mean wave activity, large asymmetry of background state between the hemisphere is required throughout a year. The difference of the eddy viscosity as discussed in the previous section on diurnal tide can also be a reason of the difference of the

4.4. TWO-DAY WAVE

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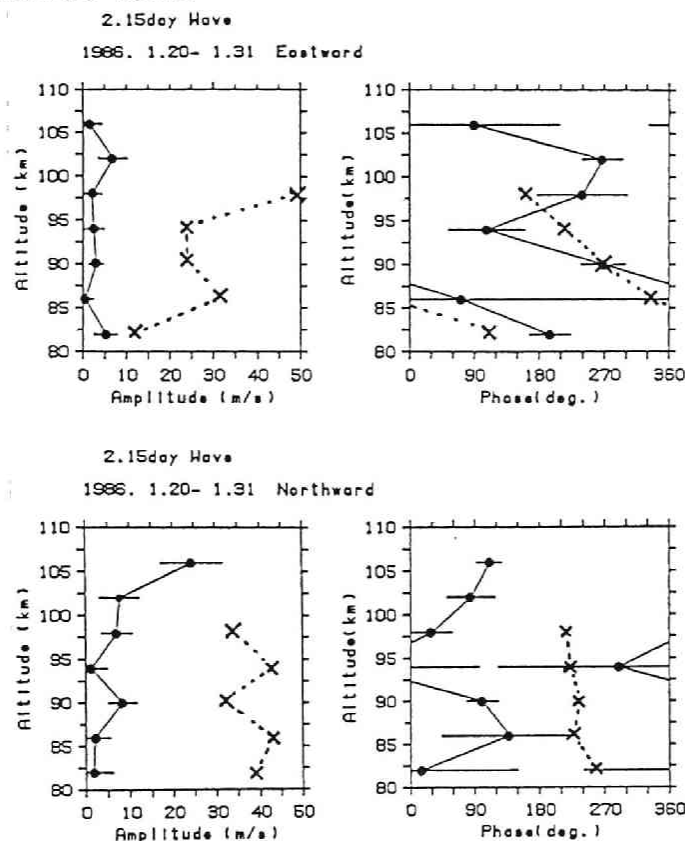


Figure 4.32: Amplitudes and phases of 2.15 days component observed in January 20-31, 1986 at Shigaraki (closed circles) and Adelaide (crosses). The phase of 0 degree means the maximum wind velocity at 0 LT on January 20.

two-day wave activity between northern and southern hemisphere. Thus, if the hemispheric difference is due to the difference of dissipation, that of northern hemisphere should be larger than that of southern hemisphere, which are the same sense as the amplitude of diurnal tides discussed previously.

Since the amplitude of two-day wave are not always enhanced simultaneously between the hemispheres as shown in Figures 4.29 to 4.31, it is sometimes difficult to compare the phase of two-day waves between the hemispheres because of the difficulty of determining the phase of the waves with small amplitude. Figure 4.32 shows a comparison of the vertical profiles of amplitudes and phases of two-day waves between Adelaide and Saskatoon. These profiles show a more complicated phase structure than those presented in the previous section as the

data of strong enhancement of two-day wave at both hemispheres in January 1984. Thus, it is considered that the clear antisymmetric phase structure can only be seen when the two-day waves are propagated up to upper mesosphere and lower thermosphere without dumping due to the background wind or dissipations.

4.5 Concluding Remarks

In this Chapter we have summarized the climatology of semidiurnal and diurnal tides and two-day waves observed by Kyoto meteor radar at Shigaraki, and compared them with the results at Adelaide and Saskatoon by using the results of Tsuda et al. [1988a,b] and Vincent et al. [1988, 1989] and some new analyses.

The semidiurnal tide observed at Shigaraki showed the maximum amplitude in September with the amplitude about 15–20 m/s. The phase profiles can be divided into two different states in summer-like and winter-like months. In summer-like months (April–October) fairly long (>100 km) vertical wavelength was observed, while in winter-like months (November–February) short vertical wavelength ranging 15–90 km was detected. The phases at 90 km altitude were almost constant during the summer-like and winter-like months, respectively, and differed by 4–6 months between the two states, for both northward and eastward component.

The amplitude of semidiurnal tide observed at Saskatoon were similar (~ 20 m/s) but the duration of large amplitude was longer than at Shigaraki: the differences between the two locations become conspicuous in winter months. The phase profiles agreed well with those of Shigaraki, showing summer-like (April–October) and winter-like (November–March) months. The phase at 90 km altitude agreed well not only in variation but also the absolute values. The vertical wavelength in summer-like and winter-like months was long (>100 km) and short (~ 45 km), respectively, although intermediate wavelengths existed in equinoctial months at Saskatoon. The phase variation at 90 km at Shigaraki and Saskatoon agreed well with the recent model by Forbes and Vial [1989], except that the phase in winter at Shigaraki was later by 4–5 hours than the model.

The transitions between summer-like months and winter-like months were different by up to one month: the autumnal transition was earlier at Shigaraki (early October) by about a month than at Saskatoon (early November). The phase of winter-like months continued until March at Saskatoon, while the transition started in February at Shigaraki. If we assume that the fundamental and higher-order modes of classical tidal theory [e.g., Kato, 1980] are dominant in summer-like and winter-like months, respectively, the seasonal variation could

be explained by the alternative appearance of these modes. For example, in spring the higher-order mode attenuates, and the fundamental mode becomes dominant. However, we showed that by assuming the dominance of another mode at equinoxes besides the two modes, the transition of the phase at 90 km and the seasonal variation of vertical wavelength are better explained. This equinoctial enhancement of another mode is also suggested in Forbes and Vial's model.

The semidiurnal tide at Adelaide also had the maximum amplitude at late summer with the values of about 20 m/s, while the annual average amplitude was larger than that at Shigaraki. The phase profiles were also divided into the summer-like and winter-like months with long winter-like months and short summer-like months. In each hemisphere, semidiurnal tides with long vertical wavelengths become dominant in summer-like months, with a tendency for short vertical wavelengths in winter-like months. Vincent et al. [1988] suggested that a classical antisymmetric tidal mode, with a single vertical wavelength in both northern and southern hemispheres, cannot explain the global structure of semidiurnal tides. The autumnal and vernal transition between winter-like and summer-like months coincided well between Shigaraki and Adelaide, i.e. between the hemispheres of the same latitude. These observational results further support the consideration by Vincent et al. [1989], that is an modification of the classical modes such that they give hemispheric asymmetry in their vertical structure.

The diurnal tides at Shigaraki had the amplitude of 10–20 m/s, with the maximum amplitude in summer. The phase at 90 km is comparatively stable during the summer months (April–August), while it did not show a clear two state phase as shown in semidiurnal tide in summer-like and winter-like months. The diurnal tide was usually larger in amplitude at Adelaide by a factor of about 2. The height structure of the phase indicated a larger contribution from the (1, 1) mode at Adelaide. Comparisons with recent theoretical models suggest that some of these differences can be due to hemispheric differences in forcing and to interactions with the background mean winds between the summer and winter hemispheres in the middle atmosphere. However, these models are not successful in expressing the hemispheric difference between the northern and southern hemispheres. The hemispheric difference of the eddy viscosity which are caused by the turbulence mainly due to the breaking of gravity wave can be a potential candidate of the hemispheric difference observed for the diurnal tides. The observation of the gravity waves between the hemisphere is therefore very important for studying the hemispheric difference observed above.

The spectral analysis of the wind velocity at Shigaraki showed the activities of the oscillations with periods of 1.4 and 3–15 days as well as two-day waves.

Periods of the two-day wave ranged from 2.1 to 2.5 days in summer and decrease to shorter than 2 days in September and October, which seems to correlate with the behavior of the mean winds.

The enhancement of the two-day wave on 18–31 January 1984 has been simultaneously observed between Shigaraki and Adelaide. The wave amplitudes at Adelaide were 30–35m/s and 15–20m/s for the meridional and zonal components, respectively, and they were larger than those at Shigaraki by a factor more than 2. Phase profiles clearly showed antisymmetric relations for both northward and eastward components, and a vertical wavelength of about 100km. This comparison suggests that the two-day wave is not confined to the summer hemisphere, but has a global structure extended over both hemispheres. The further comparisons of the amplitude of two-day waves have shown the summer maxima and winter subpeak of the amplitude at both Adelaide and Shigaraki, while the winter enhancement had a large year-by-year variations, suggesting strong interaction with background condition such as mean winds. The annual mean amplitude at Adelaide was about 2 times larger than that of Shigaraki, again indicating strong hemispheric asymmetries. The difference of the eddy viscosity between hemispheres again can be the reason of these differences.

Chapter 5

Gravity Waves Observed with the MU radar

5.1 Observation Techniques

In this chapter we present the characteristics of gravity waves observed with the MU radar.

Theoretical studies have proposed an important role of upward propagating gravity waves in the middle atmosphere dynamics, that is, gravity waves excited in the lower atmosphere transport energy and momentum to the mesosphere, where the breaking of gravity waves causes deceleration of the mean flow [Houghton, 1978; Lindzen, 1981; Matsuno, 1982; Holton, 1982, 1983; Fritts and Rastogi, 1985]. Basic mechanisms of these processes have been clarified by observations by using various remote-sensing and *in-situ* measurement techniques [e.g., Fritts, 1984]. However, a climatological study on the characteristics of the gravity waves are still under progress, especially on mesospheric gravity waves, which are expected to be useful for the study of excitation and propagation of gravity waves.

Among the various observations, Doppler radar measurement has become a very useful technique in determining the upward flux of the horizontal momentum [Vincent and Reid, 1983], which enabled us to measure the momentum deposition from gravity waves to the background mean flow. Fritts and Vincent [1987] found that about 70 % of the momentum flux is attributed to gravity waves with periods less than 1 hour. Reid and Vincent [1987] found that amplitudes of momentum flux showed a semiannual variation with maxima occurring at solstices, and that zonal mean flow accelerations have typical values of 50–80 m/s/day near 85 km. Tsuda et al. [1990a] studied gravity waves with periods from 5 min to 120 min, and found that zonal momentum flux showed eastward

and westward maxima in July/August and in December and February, respectively, and further determined the zonal acceleration for some cases as 7–13 m/s/day and –8 to –11m/s/day at height of 66–79km in summer and winter, respectively.

Since one of the most important characteristics of gravity waves is the horizontal structure, a number of studies tried to detect horizontal phase velocity of gravity waves with a ground-based radar, where various measurements are designed for observations at a single location.

Vincent and Reid [1983] determined typical values of the zonal horizontal wavelength and zonal phase velocity at 80–94 km as about 70 km and 70 m/s, respectively, utilizing multiple Doppler beams observations with the Adelaide MF radar. Meek et al. [1985a] observed gravity waves at 60–110km altitude from a spatial correlation analysis among horizontally spaced radar systems, and found that gravity waves with intrinsic periods from 10 to 100 min had horizontal scales ranging from 44 km to 211 km and phase velocity from 20 to 93 m/s. From meteor radar observations at 90–100 km Yamamoto et al. [1986] determined that gravity waves with periods from 2 to 8 hrs had mean values of zonal phase velocity and wavelength as ~ 25 m/s and ~ 360 km, respectively, and they further found that the westward propagating waves were dominant in winter.

Distribution of horizontal propagation directions of gravity waves was also studied by many authors. Ebel et al. [1987] found that wind perturbation in the height range of 61 km to 105 km was biased towards north-south direction with a considerable seasonal variation. Manson and Meek [1988] found that horizontal propagation of gravity waves in a height range from 64 km to 97 km were mainly pointed toward south, while eastward and westward momentum fluxes were observed in summer and winter, respectively.

Difference in the distribution of propagation directions of gravity waves between summer and winter was also reported by using radar observations [Vincent and Fritts, 1987] and rocket observations [Eckermann and Vincent, 1989] in the southern hemisphere. Tsuda et al. [1990b] determined the propagation direction of dominant gravity wave component at 60 to 90 km altitude in October 1986, and found that the preferential direction was south-southeastward.

In this chapter, we present the characteristics of gravity waves in the mesosphere (60–90km) with the MU radar observations, especially the momentum flux and propagation direction are discussed.

The observation duration of the mesosphere with the MU radar used in this chapter is summarized in Table 5.1, which are composed of four campaign of October 1986, June 1987, July 1990 and January/February 1991, and the monthly observations of 4–5 day per a month in December 1985 to December

Month	Oct 1986	Jun 1987	Jul 1990	Jan/Feb 1991	Dec 1985–Dec 1988
Days	19 days	24 days	21 days	15 days	159 days
L.T.	8–16 LT	7–17 LT	6–18 LT	7–17 LT	8–16 LT or 7–17 LT
Time Resolution	60/145sec (14/5days)	60/145sec (10/14days)	185sec	130/185sec (5/10days)	145sec

Table 5.1: Observation periods

1988.

Basic observation technique and analysis procedures of the mesospheric observations with the MU radar were described by Tsuda et al. [1990b], which is here briefly summarized. The antenna beam was steered for each transmission pulse sequentially into the vertical and four oblique directions declined from the zenith by 10° into north, east, south and westward. Binary phase modulation by the 8-bit complementary code with $4 \mu\text{sec}$ sub-pulse width were used. Received signal was oversampled every $2 \mu\text{sec}$, although the range resolution was determined by the sub-pulse length to be 600 m. From each Doppler spectrum obtained by 1 minute observation, radial wind velocity and spectral width were determined by using a non-linear Gaussian fitting.

It is noteworthy that a part of Doppler spectra were contaminated by a radio interference due to other communications or broadcasting. For such a spectrum the ordinal Gaussian fitting was sometimes unable to be successfully applied. In June 1987 and July 1990, significant interference was often observed, and a part of the data in January/February 1991 were also contaminated by weak interference. By using a more complicated algorithm, parameters of Doppler spectra could appropriately be estimated. However, in this study we exclusively used data without the influence of interference.

Although the observation parameters were the same during the four observation periods, time intervals between successive spectra were different as summarized in Table 5.1, because other observations in the lower height ranges were interleaved between the successive mesospheric observations. As a result, the time resolution was better in two observation periods in October 1986 and June 1987.

Turbulent scattering in the mesosphere is normally detected during only daylight hours, so that, a duration of observation was 10 and 12 hrs in June 1987 and July 1990, respectively. Although the observation interval was also extended to 10 hrs in January/February 1991, echo power was not strong enough

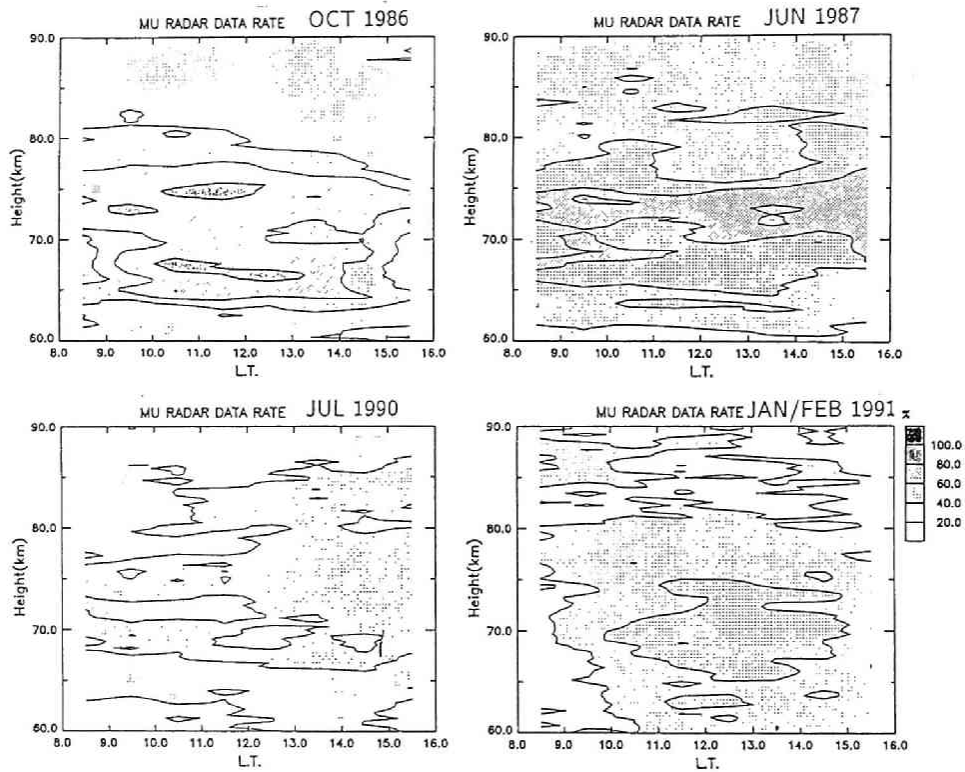


Figure 5.1: Time height cross section of hourly rate of the wind determination for vertical beam of the MU radar in October 1986, June 1987, July 1990 and January/February 1991. After data screening of the wind velocity.

in the early morning and late afternoon, causing shorter coverage of observation interval than in summer.

Figure 5.1 shows the hourly rate of the successful wind determination by Doppler wind spectra as a time height cross section in the four campaign periods. Note that the data with significant interference were removed before the calculation of the data rate, so that in June 1987 and July 1990, probability of determinations was smaller than that shown in Figure 5.1 due to the interference. The wind velocities were determined most frequently in October 1986, and less frequently in January/February 1991. The height range of the data rate larger than 40 % were 63–80 km, 65–78 km, 67–80 km and 66–74 km in October 1986, June 1987 July 1990 and January/February 1991, respectively. In July 1990, wind data at 80–85 km were also taken at the rate of 40 % in the afternoon, probably because of high electron density due to high solar activity which reached the maximum in 1989–1990. On the contrary, in January/February 1991, the duration of the wind data determination of 40 % was limited between 10–15 LT, in the small height range around 70 km. Therefore, we mainly utilized the data obtained between 68–80 km for the analysis of time series such as frequency spectra of the wind velocities and momentum flux.

Because of the difference of the wind determination rate, time resolution of the observation and the occurrence of the radio interference, the data quality is better for the former two periods; October 1986 and June 1987. So that, data are more closely analyzed in subsection 5.2.2 and 5.2.3 for those observation durations.

5.2 Momentum Flux Profiles

5.2.1 Frequency Spectra of Wind Velocity and Momentum Flux

We first analyze in this section frequency spectra of vertical and radial wind velocity, then further derive frequency spectra of momentum flux.

We analyzed a frequency spectrum by adopting the procedure described by Tsuda et al. [1990b] except that the maximum lag of an auto-correlation function (ACF) was set equal to 150 min for results in June 1987, July 1990 and October 1986, while it was 80 min in January/February 1991 because of the data lack in early morning and late afternoon. We restricted the height range for the spectral analysis to 68–80 km, since the time series were most continuously obtained at these altitudes.

Figure 5.2 shows frequency spectra of vertical wind velocity as well as those for radial wind velocity obtained in the four oblique directions during the four

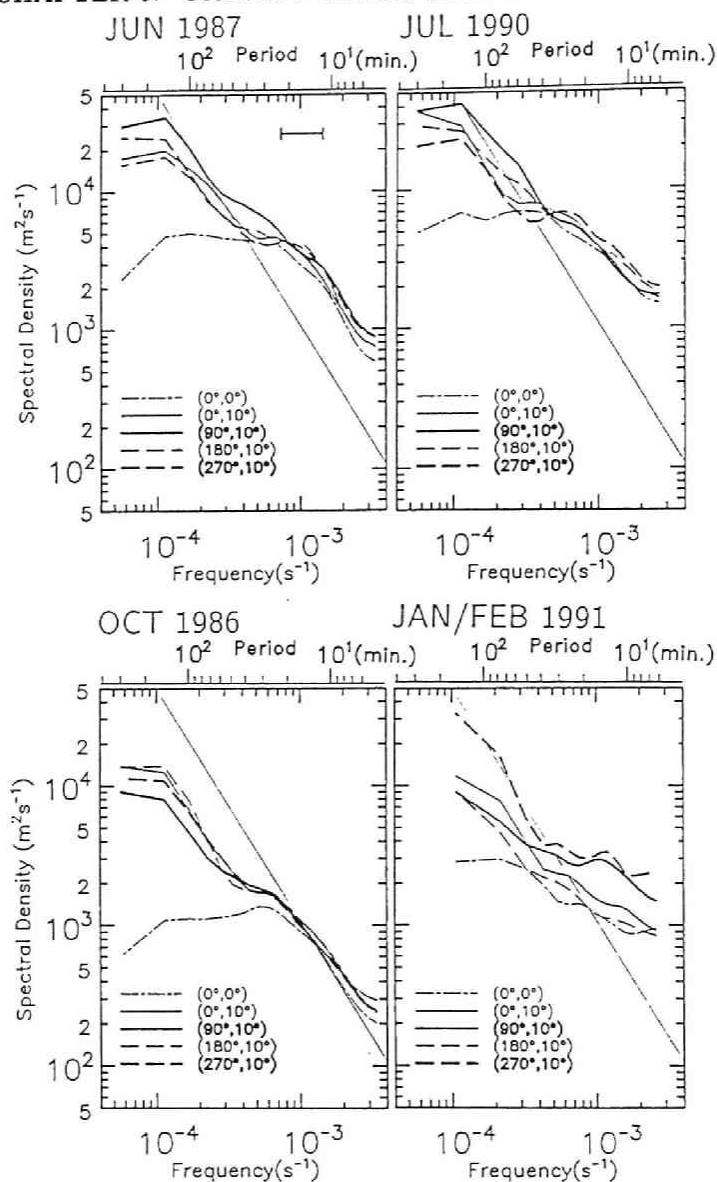


Figure 5.2: Frequency spectra of radial wind velocities observed on June 6–29, 1987 (a), July 2–21, 1990 (b), October 13–31, 1986 (c), and January 21 to February 8, 1991 (d). Thin chained lines are of vertical beams, and other lines are of oblique beams; thin solid lines (northward), thin dashed lines (southward), thick solid lines (eastward), and thick dashed lines (westward), respectively. Thin lines indicate the spectral slope of $f^{-5/3}$.

observation periods. Note that the spectral density at the lowest frequency point shown in Figure 5.2 tends to be smaller than the real value because of the removal of linear trend in the time series and the Hanning window multiplied to the ACF of the time series [Tsuda et al., 1989]. The spectra of high frequency component are smoothed by a spectral window of a raised cosine of the logarithmic frequency, where the resulting spectral resolution became approximately $\pm 35\%$ of any specified frequency as indicated in the figures. Thus, the reliability of the spectral density were much improved but frequency resolution became worse than the original spectrum. The 95 % confidence intervals of the spectral stability at each frequency are generally 14 % and 4 % for the frequency of $1 \times 10^{-4}\text{s}^{-1}$ and $1.7 \times 10^{-3}\text{s}^{-1}$, respectively. However, the confidence intervals are about twice larger for the data in January/February 1991 because of the low data acquisition rate.

General structure of the oblique spectra was similar among determinations in four observation periods. Logarithmic slopes of the oblique spectra were approximately $-5/3$ in a frequency range from 10^{-4}s^{-1} to $3 \times 10^{-3}\text{s}^{-1}$, suggesting that the oblique spectra were mainly contributed from the projection of horizontal wind components. While, the spectra for frequency component larger than $4-5 \times 10^{-4}\text{s}^{-1}$ were enhanced, which seems to be caused by the effects of vertical winds. However, in the frequency range from $7-8 \times 10^{-4}\text{s}^{-1}$ the spectra in June 1987, July 1990 and October 1986 again showed a slope of about $-5/3$, which could be interpreted by the Doppler shift effects on the spectrum of vertical winds [Fritts and VanZandt, 1987; Tsuda et al., 1990b]. In January/February 1991, the spectral slope was fairly gradual in the entire frequency range larger than $4-5 \times 10^{-4}\text{s}^{-1}$, and it did not show a bent at $7-8 \times 10^{-4}\text{s}^{-1}$ detected in other three periods. The determination rate of the Doppler velocity in January/February 1991 was lower than those in the other observation periods because of the low signal to noise ratio (SNR) in winter and the large interval between successive Doppler spectra due to the interleave of other observations. Such a condition might cause the noisy spectra in January/February in 1991 especially at the high frequency components.

Discrepancy of the spectral amplitudes among four oblique directions was recognized in the results which was most conspicuously seen for frequency components smaller than about $5 \times 10^{-4}\text{s}^{-1}$ where the horizontal component had more dominant contribution to the oblique spectra. It is noteworthy that in January/February the spectral amplitudes for zonal components were larger than the meridional ones, which can be recognized in a frequency range larger than $5 \times 10^{-4}\text{s}^{-1}$. It is appreciable that in October 1986 spectral amplitudes for the eastward component were smaller than those in the other three directions for frequency smaller than $3 \times 10^{-4}\text{s}^{-1}$. In June 1987 spectrum for the eastward

component showed largest values in the frequency range smaller than $8 \times 10^{-4} \text{s}^{-1}$. Spectra in July 1990 also showed larger values for the eastward winds in a low frequency range. The azimuthal anisotropy of the oblique spectra can be interpreted in terms of the anisotropy in the distribution of gravity wave propagation [VanZandt et al., 1990], which can further be related to the dominant direction of the wave-induced momentum flux.

For the vertical wind component the spectral slope was nearly zero but showed some variation with months for frequency range of $1 \times 10^{-4} \text{s}^{-1}$ to $4 - 5 \times 10^{-4} \text{s}^{-1}$, where the slope can be estimated as -0.05 , 0.05 , 0.15 and -0.4 in June 1987, July 1990, October 1986 and January/February 1991, respectively.

The spectral amplitudes for both vertical and oblique components were generally larger in summer months. That is, the density of oblique spectra in June 1987 and July 1990 were 3 and 3.5 times larger than in October 1986, respectively, in the frequency range between 1×10^{-4} and $2 \times 10^{-3} \text{s}^{-1}$. The spectral amplitudes in January/February 1991 were smaller than the summer results, but were larger than those in October by 1.5 to 2.5 times in the frequency range from 2×10^{-4} to $1 \times 10^{-3} \text{s}^{-1}$.

These comparisons suggest that the gravity wave energy in the mesosphere has a semiannual variation with the the smallest values at equinoxes, which is identical to the seasonal variation of the activity of mesospheric gravity waves with periods of 2 hr to 5 min at 60–90 km in 1985–1988 reported by Tsuda et al. [1990a], and is consistent with other radar observations [Vincent and Fritts, 1987; Meek et al., 1985b]

Next, we discuss frequency spectra of momentum flux. We subtracted spectral densities between a pair of oblique frequency spectra collected by using antenna beams pointed in the opposite azimuth angles. Then, we derived a frequency spectrum of momentum flux by dividing the resultant spectral density by $2 \sin 2\theta$ (θ : zenith angle of the oblique beam). That is,

$$\begin{aligned} F_{u'w'}(\omega) &= \frac{F_{V(90^\circ)}(\omega) - F_{V(270^\circ)}(\omega)}{2 \sin 2\theta} \\ F_{v'w'}(\omega) &= \frac{F_{V(0^\circ)}(\omega) - F_{V(180^\circ)}(\omega)}{2 \sin 2\theta} \end{aligned} \quad (5.1)$$

where, $F_{V(0)}(\omega)$, $F_{V(90)}(\omega)$, $F_{V(180)}(\omega)$, $F_{V(270)}(\omega)$ are spectra of the radial wind velocity at 0, 90, 180 and 270 degrees in azimuth angle, and θ represents the zenith angle of the oblique antenna beam direction, i.e., 10° in this study.

Figure 5.3 shows the observed momentum flux spectra, $F_{u'w'}(\omega)$ and $F_{v'w'}(\omega)$, for the northward and eastward component, respectively. Note that momentum flux spectra are plotted in an area preserving form, so that the area surrounded by the curve and horizontal axis is proportional to the integrated momentum

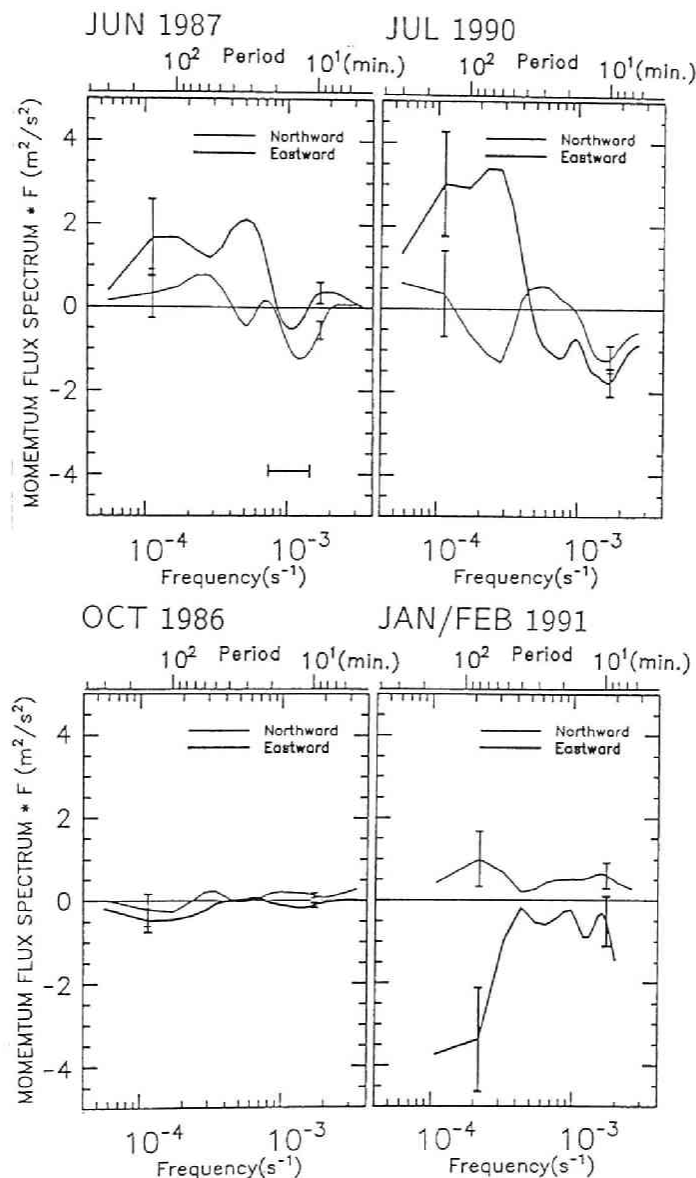


Figure 5.3: Frequency spectra of zonal (dashed) and meridional (solid) momentum flux observed on June 6–29, 1987 (a), July 2–21, 1990 (b), October 13–31, 1986 (c), and January 21 to February 8, 1991 (d). Note that curves are plotted by an area preserving form. The error bars indicate the confidence intervals of 95 %.

flux in a corresponding frequency range. The error bars indicate the confidence intervals of 95% of the spectral stability.

There is a clear seasonal variation that the zonal momentum flux was eastward in summer and westward in winter, which was already reported by Tsuda et al. [1990a]. We here discuss the frequency distribution of the momentum flux. In June 1987, the zonal momentum flux was largely enhanced in the eastward direction in a frequency range smaller than $8 \times 10^{-4} \text{s}^{-1}$, corresponding to the wave period of about 20 min. On the other hand, the meridional momentum flux was generally northward and southward in the lower and higher frequency ranges than $4 \times 10^{-4} \text{s}^{-1}$, respectively, but it did not show a clear enhancement in the observed frequency region. The zonal momentum flux was fairly small in a frequency range larger than $8 \times 10^{-4} \text{s}^{-1}$.

In July 1990, the large eastward momentum flux was detected in a frequency range smaller than $4.0 \times 10^{-4} \text{s}^{-1}$, that is, about 40 min in wave period. For a frequency range higher than $4.0 \times 10^{-4} \text{s}^{-1}$ the momentum flux was westward, being opposite to that for the lower frequency components. The meridional flux was fluctuating near zero, and it did not show a clear tendency of dominant directions.

On the other hand, the zonal momentum flux in January/February 1991 showed large values in the westward direction in a frequency range smaller than about $4 \times 10^{-4} \text{s}^{-1}$, or 40 min in wave period. For the higher frequency component the zonal momentum flux was weakly westward. The meridional flux was generally northward in the entire frequency region, although their amplitudes were fairly small. These results support a theoretical prediction that the zonal momentum flux should be reversed between summer and winter months [e.g. Lindzen, 1981].

In October 1986, amplitudes of both the zonal and meridional momentum were fairly small in the entire observed frequency range. However, the zonal momentum flux was generally westward in a frequency range smaller than $3.5 \times 10^{-4} \text{s}^{-1}$ (about 50 min in wave period), which is consistent with results in winter months, again suggesting that the condition in October 1986 was rather winter-like than summer-like.

Meek et al. [1985b] reported using Saskatoon MF radar observations that gravity waves with periods shorter than 1 hour mainly contribute to the acceleration of mean wind. Fritts and Vincent [1987] and Reid and Vincent [1987] pointed out using Adelaide MF Doppler radar observations that about 70 % of the momentum flux was attributed to the components with period less than 1 hrs. Our results, however, show that amplitudes of the zonal momentum flux were considerably large for gravity waves with the periods between 1 and 2.7–5 hrs. It can be suggested that the gravity waves with periods longer than 2.7–5

hours can also be important in carrying momentum flux, which, however, was not clarified yet by this study due to a limit in minimum frequency of the spectral analysis.

5.2.2 Vertical Profiles of Momentum Flux

In this section we present vertical profiles of momentum flux. Observations in June 1987, July 1990 and October 1986 were analyzed because the wind velocity data in these three months were collected well enough to analyze the vertical profile. The data were analyzed in the same way as in Tsuda et al. [1990b] except for the pass band characteristics of a filter. For each range gate and each day a linear trend was subtracted from the ten-hour (or eight-hour) time series of radial wind velocities. Then the fluctuating radial wind components were separated into two frequency ranges by using low pass and band pass filters with a cut off period of 30 min and a pass band of 30 – 5 min, respectively. We hereafter address the two bands as 8–0.5 hrs and 30–5 min components, respectively. By subtracting the variances of radial wind velocity collected in each range gate by using a pair of oblique beams we obtained profiles of both zonal and meridional momentum flux.

The values of the momentum flux in each range bin were averaged over a height range of 4.5 km, then the profiles of $\overline{\rho u'w'}$ and $\overline{\rho v'w'}$ were determined as shown in Figures 5.4– 5.6, where ρ was taken from the CIRA 1986 model. Note that for the data in July 1990, 30–5 min component was not deduced because the Nyquist period determined by the interval of the Doppler spectra was larger than 5 minutes and the band pass filter cannot be composed.

All the profiles showed that the momentum flux decreased with height, except for that of the 30–5 min component of zonal flux in June 1987 and 8–0.5 hrs component of meridional flux in July 1990. It is remarkable that the eastward momentum flux of the 8–0.5 hrs component in June 1987 and July 1990 was conspicuously larger in comparison with other profiles at the altitudes of 71–77km. At 71 km the eastward flux for the 8–0.5 hrs component was $2 \times 10^{-4} \text{kgm}^{-1}\text{s}^{-2}$ and $3 \times 10^{-4} \text{kgm}^{-1}\text{s}^{-2}$ in June 1987 and July 1990, respectively, while other components were about $0.5 \times 10^{-4} \text{kgm}^{-1}\text{s}^{-2}$.

The eastward flux of 8–0.5 hrs component in June 1987 and in July 1990 decreased rapidly with height indicating strong eastward drag force, while the 30–5 min component of June 1987 was small in all the height range. Other determinations from the MU radar observations by Tsuda et al. [1990a] were plotted in Figure 5.4, which were deduced for the wave periods of 2 hours to 30 minutes. They showed smaller momentum flux in comparison with the results of June 1987 and July 1990 by a factor of 2/3 to 1/2, although the general

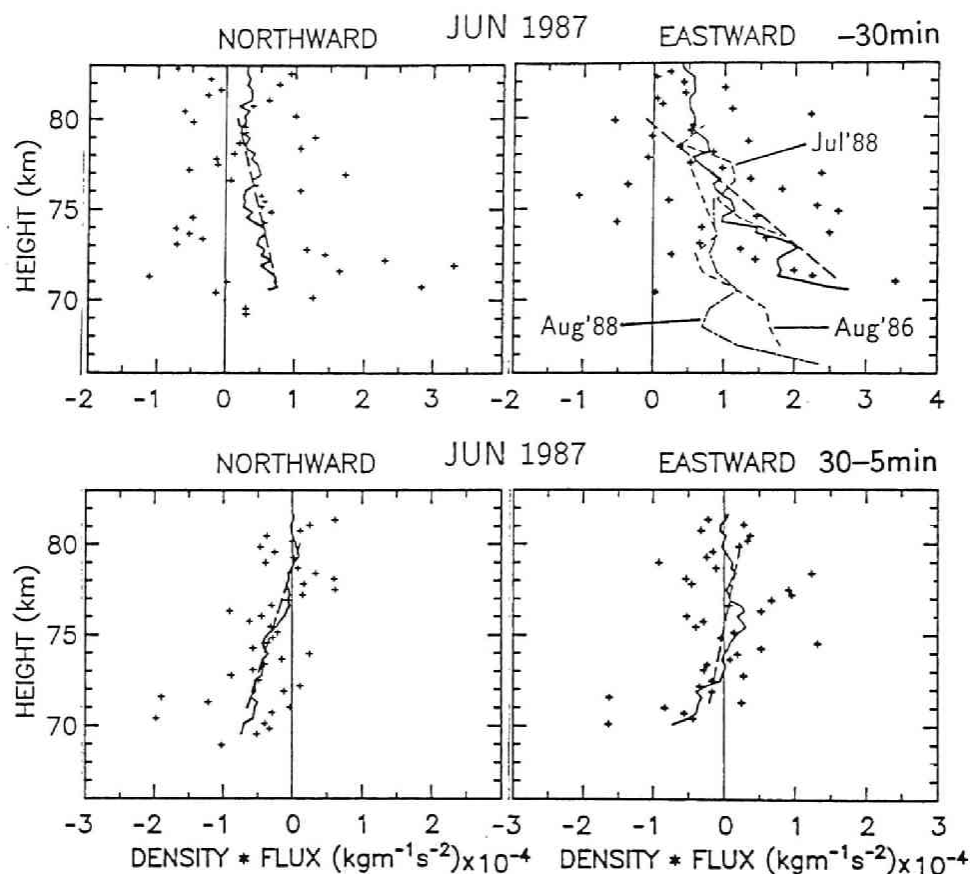


Figure 5.4: Vertical profiles of northward (left) and eastward (right) components of momentum flux in June 1987. The top and the bottom panels are corresponding to the components with periods of larger than 30 min and 30-5 min, respectively. Solid line shows 4.5km average of the momentum flux. Dashed lines are best fit lines between 71 and 80 km. In the right and top panel, the momentum fluxes observed in August 1986, August 1988 and July 1988 for the waves with periods of 2 hours to 30 minutes are also plotted by thin lines as a reference [Tsuda et al., 1990a].

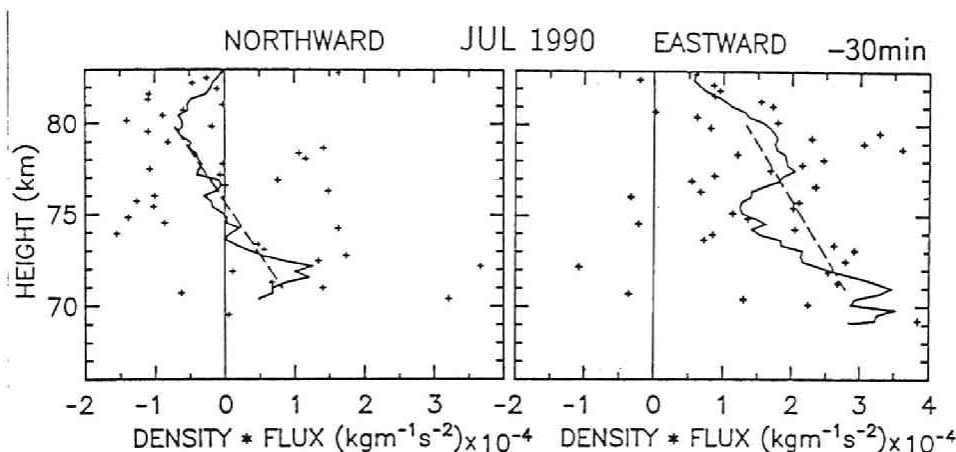


Figure 5.5: The same as Figure 5.4 except for the data in July 1990. Only the component with the periods larger than 30 minutes are deduced.

behavior was quite similar. The discrepancy seems to be attributed to the additional momentum flux of the wave component with period of 2 hours to 8 hours. Meridional component in June 1987 shows northward and southward flux in 8–0.5 hrs and 30–5 min ranges, respectively, which is also recognized in the frequency spectrum of momentum flux in Figure 5.3. The meridional component in July 1990 shows the northward and southward momentum flux below and above 75 km, resulting a total northward drag at around 75 km.

In October 1986 as shown in Figure 5.6, the zonal component was westward and eastward in 8–0.5 hrs and 30–5 min ranges, respectively. The amplitude of the zonal flux due to 8–0.5 hrs components was about $5 \times 10^{-5} \text{ kgm}^{-1} \text{ s}^{-2}$ at 70 km, which was larger than that for the 30–5 min components by a factor of about 2. Therefore, the resultant total flux seemed to be westward. The earlier results in December 1985 and February 1986 by Tsuda et al. [1990a] plotted in Figure 5.6 showed similar amplitude of westward momentum flux at 68–78 km altitude. However, Tsuda et al. [1990a] picked up the wave component with periods of 2 hours to 30 minutes, so that if they are compared as the momentum flux of 8 hours to 30 minutes components, the amplitude in October would be significantly smaller than those in December and February. The meridional component in October 1986 also shows the different directions between 8–0.5 hrs and 30–5 min components; northward and southward flux for 8–0.5 hrs and

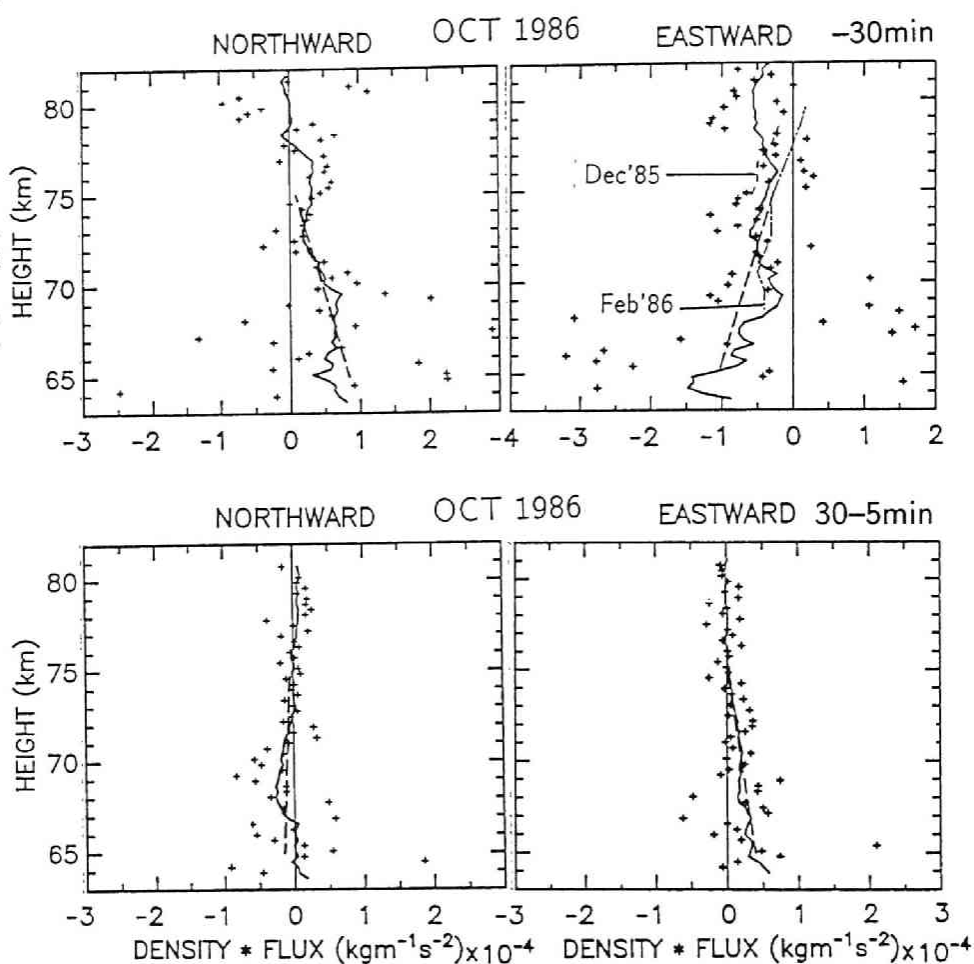


Figure 5.6: The same as Figure 5.4 except for the data in October 1986. Dashed lines are best fit line between 65 and 75 km. The momentum fluxes observed in February 1986 and December 1985 for the waves with periods of 2 hours to 30 minutes are also plotted by thin lines as a reference [Tsuda et al., 1990a].

Month	Direction	Drag (8 h–30min)	Drag (30min–5min)	Total	Remarks
Jun 1987	Zonal	+61 m/s/day	–10	+51	at 75 km
	Meridional	+13	–17	–4	
July 1990	Zonal	+36			at 75 km
	Meridional	+33			
Oct 1986	Zonal	–8.7	+4.3	–4.4	at 70 km
	Meridional	+8.4	–1.0	+7.4	
Aug 1986	Zonal	+7.2 (2h–30min)			at 70–77 km *)
July 1988	Zonal	+12.8 (2h–30min)			at 73–79km *)
Aug 1988	Zonal	+9.4 (2h–30min)			at 66–75km *)
Feb 1986	Zonal	–7.6 (2h–30min)			at 70–78km *)
Dec 1985	Zonal	–11.0 (2h–30min)			at 73–77km *)

*) Tsuda et al. [1990a]

Table 5.2: Mean flow drag due to the gravity wave observed in June 1987 and October 1986.

30–5 min components, respectively, although amplitude of the latter was small.

The most reliable profiles of momentum flux were estimated at 71–80 km in June 1987 and July 1990, and at 65–75 km in October 1986, respectively. Therefore, that part of the momentum flux profiles was fitted by a linear curve as indicated in Figures 5.4 – 5.6. Then, we estimated mean flow accelerations, or drag force, F_E and F_N , for zonal and meridional components, respectively, from the vertical derivative of the momentum flux profiles as

$$F_E = -\frac{1}{\rho} \frac{\partial}{\partial z} (\overline{\rho u' w'}) \quad (5.2)$$

$$F_N = -\frac{1}{\rho} \frac{\partial}{\partial z} (\overline{\rho v' w'})$$

The values of F_E and F_N can be regarded to be taken at 75 km in June 1987 and July 1990, and at 70 km in October 1986, respectively. The resultant values of F_E and F_N are summarized in Table 5.2.

The zonal mean winds plotted in Chapter 3 showed decrease with height above 74 km, 75 km and 69 km in June 1987, July 1990 and October 1986, respectively, thus the results of F_E and F_N shown in Table 5.2 were determined at these altitude regions.

In June 1987 the eastward mean flow drag due to 8–0.5 hrs waves was as

large as 61 m/s/day, while 30–5 min component showed westward drag of only 10 m/s/day, so that in total the eastward drag of 51 m/s/day resulted in. Meridional acceleration in June 1987 was northward and southward for 8–0.5 hrs and 30–5 min ranges, respectively. Their amplitudes were about 15 m/s/day for both cases. As a result, the drag force due to 8–0.5 hrs and 30–5 min components canceled with each other. Thus, the total acceleration became as small as -4 m/s/day. The zonal drag in July 1990 of 8–0.5 hrs component also showed eastward drag as large as 33 m/s/day. The northward drag was also up to 36 m/s/day, although 30–5 min component, which were not determined in the present study, might cancel the strong meridional flux as is the case of June 1987. In October 1986 westward and northward acceleration for the 8–0.5 hrs component was 8.7 m/s/day and 8.4 m/s/day, respectively, which was partly canceled by the high frequency component, causing a small acceleration.

To summarize, contribution of the low frequency components (8–0.5 hrs) was large not only to the momentum flux itself, but to their vertical gradient, i.e., mean flow drag. And, the high frequency component (30 min–5 min) had relatively smaller effects except for the meridional component in June 1987.

The zonal drag by the gravity wave with the periods from 2 hrs to 30 min was also reported by Tsuda et al. [1990a], which are also shown in Table 5.2. They observed that the zonal momentum flux in summer was 7–13 m/s/day at about 70–75 km altitudes. Our result here shows the eastward drag larger than them by a factor of 3–7 times, which is probably because our results were calculated for the waves with periods extending to 8–10 hours and also the height range was located higher from 71 km to 80 km. For the winter months, Tsuda et al. observed two cases in February 1986 and December 1985 which showed the westward drag of 7.6 and 11.0 m/s/day at 70–78 km for the wave periods of 2 hours – 30 minutes. These values are comparable with the westward flux observed for the periods of 8 hours– 30 minutes in October 1986. However, if the direction of the drag do not change between the wave component with periods larger and smaller than 2 hours as suggested from Figure 5.3, the drag by the wave periods of 8 hours – 30 minutes, i.e. the major component of the zonal drag can be estimated to be larger in winter than October.

The momentum flux and mean flow drag in October 1986 was also already reported by Tsuda et al. [1990b], who showed the momentum flux due to the wind fluctuation with period longer than 10 minutes. They calculated the westward and northward drag as 5.1 and 4.0 m/s/day, respectively, indicating the difference with the present results of 0.7 and 3.4 m/s/day, respectively. This difference is considered to be caused by small westward and comparatively large northward momentum flux with periods of the 10–5 min component, which can be seen in Figure 5.3.

The consistency of the current observations with the previous studies with the MU radar [Tsuda et al., 1990a,b] suggests that the momentum fluxes obtained in the present study are typical values in the seasonal variation.

Reid and Vincent [1987] reported that gravity waves with periods ranging from 8 min to 8 hrs induced a zonal mean wind acceleration at 80–98 km between 50–80 m/s/day as a typical value, and sometimes became almost 200 m/s/day. They further showed a good agreement of the observed acceleration with Coriolis torque due to mean meridional wind, while they showed disagreement when planetary waves were enhanced. Meek and Manson [1989] also investigated $\overline{u'w'}$ at 60–110 km and found that zonal acceleration by gravity waves with periods of 10–100 min was 10–50 m/s/day in summer and winter, which was also consistent with Coriolis torque. Although our results are limited to only during three campaigns and at the low altitude (75 km in June 1986 and July 1990, and 70 km in October 1986), they do not contradict to other observations. Moreover, the corresponding Coriolis torque from the meridional wind presented in Chapter 3 was -50 m/s/day at 75 km in June 1987, -55 m/s/day at 75 km in July 1990 and $+12$ m/s/day at 70 km in October 1986, which agree well with the zonal drag presented in this section except for the larger value than the observed drag in July 1990 where the drag due to the waves with periods of 30–5 min were not determined as described above.

5.2.3 Day to Day Variation of Gravity Wave Activity

Next, we describe day to day variations of gravity wave activity during the two observation campaigns; June 1987 and October 1986. As we reported in this section, frequency spectra suggested that gravity waves with periods longer than 20 – 50 min produced a major part of zonal momentum flux. Furthermore, in this section the momentum flux due to gravity waves with periods longer than 30 min was found to be dominant in accelerating mean wind. Therefore, we study the variance and momentum flux induced by gravity waves with periods from 30 min to 4 hours.

Upper panels in Figure 5.7 show daily mean variance of radial wind fluctuations in the vertical and four oblique directions with periods of 4 hours – 20 minutes. Data with significant gaps in daytime were excluded from the figure. Note that the vertical axis of the figures are adjusted to the magnitude of variance in each month, so that the scale is larger for June 1987 than for October 1986. Lower panels in Figure 5.7 show day to day variation of the momentum flux. In June 1987 the eastward momentum flux showed large day to day variation, that is, it quite often became as large as about $5 \text{ m}^2/\text{s}^2$, while it became close to zero on other days. There were no cases that showed westward flux

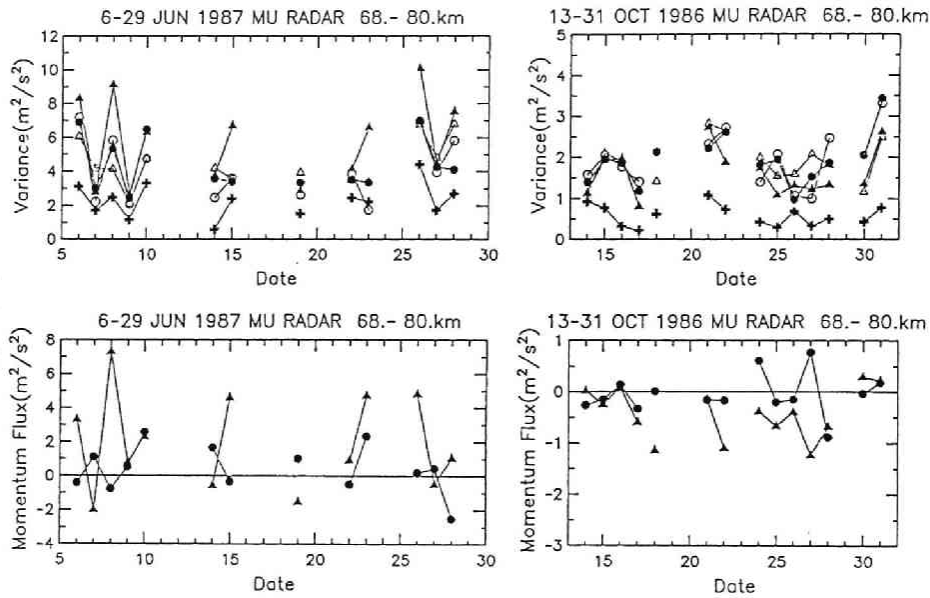


Figure 5.7: Day to day variations of the daily mean variance of the radial wind fluctuations (top panels) and momentum flux (bottom panels), respectively.

(Top panels)

Crosses, closed circles, open circles, closed triangles and open triangles show variance of vertical, northward, southward, eastward and westward beams, respectively. Left and right panels are for observation in June 1987 and in October 1986, respectively. Wind fluctuations with periods between 30 min and 4 hrs are extracted.

(Bottom panel)

Day to day variations of zonal (triangles) and meridional (circles) momentum flux with periods between 30 min to 4 hrs.

larger than $2 \text{ m}^2/\text{s}^2$.

Northward momentum flux also showed similar characteristics, although the magnitudes of northward momentum flux were smaller than the eastward components. It is noteworthy that there seems to have correlation between the enhancements of zonal and meridional components, but it did not always coincide.

In October 1986, the zonal momentum flux was generally westward, but it sometimes reversed to eastward with small amplitudes. On the other hand, the meridional flux was distributed in both northward and southward direction.

Reid and Vincent [1987] determined momentum flux and mean flow drag every 3–4 days, and found that the direction and magnitude of the flux and mean flow acceleration showed considerable variations between successive observation periods. Our result of the daily determination of momentum fluxes also showed a similar variability in amplitude and direction. These variations suggest that the propagation directions of gravity waves changed frequently.

Next, we examine a relation between the day to day variation of the momentum flux and the variability of the mean wind field. The top panel in Figure 5.8 shows a daily mean zonal wind velocity averaged over a height range between 70 and 85 km as well as the mean zonal wind shear in the same height region, which is calculated by using a linear fitting of daily mean wind velocity profile.

Amplitudes of both mean wind and wind shear showed large day to day variations. There seems to be a fairly good correlation between variations of momentum flux and wind shear except for the results on 26–28 June, 1987, while the variation of the mean wind amplitudes was somewhat different from those variations. A cross correlation between the zonal momentum flux and the zonal wind shear, and between the zonal momentum flux and the mean zonal wind are plotted as a scatter diagram in Figure 5.8. The correlation coefficient was 0.69 between the zonal wind shear and the zonal momentum flux, while it was -0.37 between the zonal wind amplitude and the momentum flux. Therefore, the former correlation, i.e. the correlation between the zonal wind shear and the zonal momentum flux, is relatively significant. The meridional momentum flux did not show a good correlation with either the mean meridional wind amplitude or the meridional wind shear.

Figure 5.9 shows similar comparison for observations in October 1986, where a height range for the calculation of the zonal wind and wind shear was 65–75 km. It again showed a good correlation between zonal momentum flux and zonal wind shear, while zonal mean wind amplitude did not have a good correlation with zonal momentum flux. The wind shear and the momentum flux showed large day to day variations, having large negative values between October 18 and 28, while the mean wind showed longer period variations with a minimum

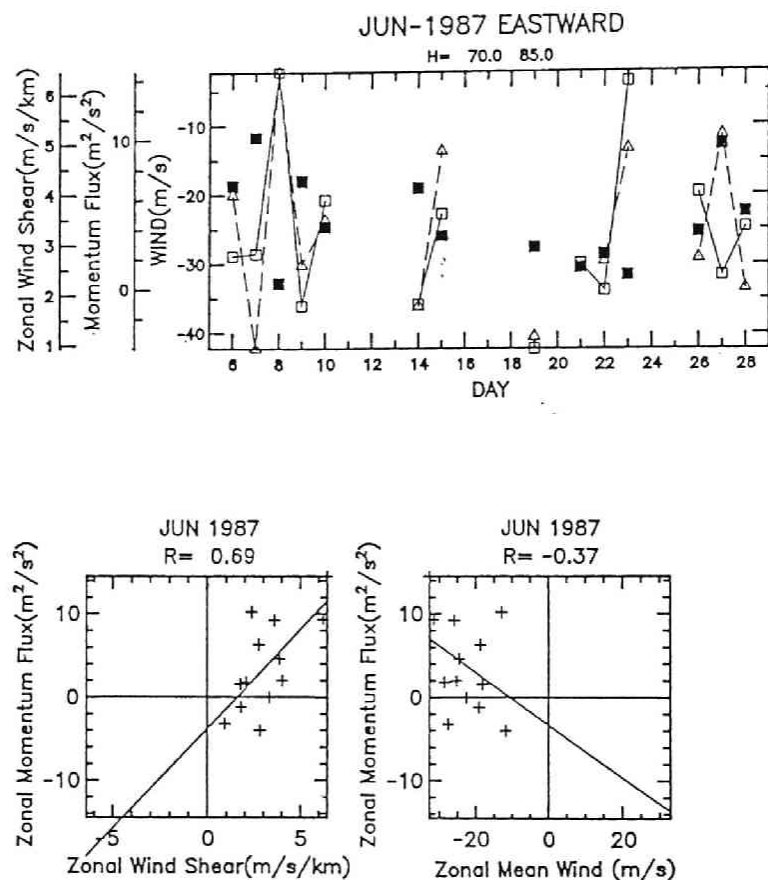


Figure 5.8: Day to day variation of the daily mean zonal wind (closed squares) and mean shear of the mean wind (open squares) at 70–85 km in June 1987 (top panel). Zonal momentum fluxes at 68–80 km are also plotted by triangles. The scatter diagram of mean shear and zonal momentum flux, and mean wind and zonal momentum flux are shown in the left and the right panels, respectively.

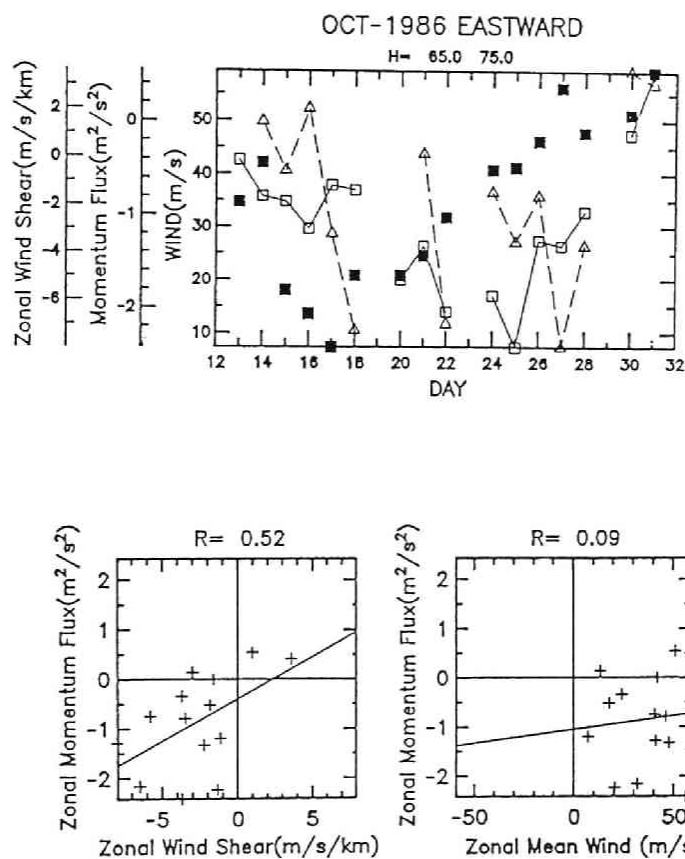


Figure 5.9: The same as Figure 5.8 except for the data in October 1986. Zonal mean wind and shear is calculated for 65 to 75 km.

eastward wind at October 17 and then almost monotonously increased up to 60 m/s until October 31.

The correlation coefficients between the zonal wind shear and the zonal momentum flux, and the zonal mean wind and the zonal momentum flux were 0.52 and 0.09, respectively. The meridional flux again did not show good correlation with meridional mean wind or wind shear.

The correlation between mean zonal wind shear and zonal momentum flux can be explained as the strong eastward momentum flux accelerates the mean zonal wind towards eastward by losing a part of momentum flux, i.e., these day to day variations in wind shear are caused by the day to day variation of momentum flux brought by gravity waves.

The day to day variations of the momentum flux can also be generated by the variation of the mean wind intensity in the mesosphere which modulates the effect of the gravity wave saturation due to convective instability. That is the gravity waves propagating in the same direction with the mean wind are attenuated by the increase of mean wind amplitude. This can also cause the fluctuation of momentum flux. However, the mean wind variation in that situation has to show a better correlation with the momentum flux variation than the shear variation, because the absolute amplitude of mean wind is concerned with the interaction with gravity waves. In the present result, the variation of zonal mean wind showed less correlation with momentum flux than the variation of zonal shear, so that this mechanism is not appropriate to this case.

Thus, the day to day variation of the zonal shear of mean wind in the present study is explained that they were caused by the day to day variation of zonal momentum flux carried by gravity waves, but also the variation of wind shear can cause these day to day variation of gravity waves for the case of June 1987.

5.3 Characteristics of Wave Propagation

5.3.1 Data Analysis

In this section we present the characteristics of the dominant gravity wave component which is observed in the vertical profile of the wind velocities, and especially study on the propagation directions of these waves.

Data analysis is described elsewhere [Tsuda et al., 1990b], while we here briefly summarize it again. Wind velocity profiles determined every 60–185 sec were averaged for 2 hours. These profiles are often dominated by sinusoidal fluctuations with vertical scales and amplitudes ranging from 5 to 15 km and up to a few tens of meters per second, respectively, which are thought to be interpreted as the gravity wave activity [Yamamoto et al., 1987, 1988; Muraoka

et al., 1987]. These fluctuations are not purely sinusoidal, but instead they appear to consist of a superposition of gravity waves with various vertical scales.

Hodograph analysis has been applied to some cases of these dominant gravity waves in order to study their characteristics and correlations with echo intensity [Yamamoto et al., 1987, Muraoka et al., 1988]. Tsuda et al.[1990b] investigated characteristics of those gravity waves more statistically by means of the hodograph analysis for the MU radar observations in October 1986. Here we adopted the same analysis procedure for data obtained in the other three observation periods in June 1987, July 1990 and January/February 1991 (See Table. 5.1).

Vertical and horizontal components of the wind velocity were smoothed out by a band pass filter with cutoffs at 3km and 18km, which was determined by considering vertical scales of the dominant components seen in the vertical profile. A hodograph was plotted from the filtered horizontal wind velocity profiles, then the wind velocities at the successive 10 to 20 heights were fitted by an ellipse. According to a linear gravity wave theory, horizontal propagation direction of gravity wave is aligned to a semi-major axis of the ellipse, and the propagation direction can finally be determined by considering the variations of the vertical wind velocity. The vertical propagation direction was determined by the rotation direction of the hodograph, that is, clockwise rotation with height in the northern hemisphere corresponds to the upward energy propagation, i.e., positive vertical group velocity. From the hodograph analysis we can determine other parameters of gravity waves such as intrinsic period, group and phase velocity in vertical and horizontal plane by using a dispersion relation of gravity waves.

We applied the hodograph analysis to bi-hourly profiles of wind velocity. On each analyzed day we normally picked up only one dominant gravity wave. However, two different waves were sometimes extracted, for example, one from the profiles in the morning and the other in the afternoon. We analyzed in total 59 cases, which were distributed as 14, 17, 19 and 9 in October 1986, June 1987, July 1990 and January/February 1991, respectively.

5.3.2 Vertical Propagation Characteristics

Figure 5.10 shows a distribution of vertical phase velocity of dominant components of gravity waves versus intrinsic wave period. The observed intrinsic periods are mainly distributed from 5 to 15 hours, with an average value of 9.9 hours. Since the analysis was done by using bihourly profiles, we were unable to detect waves with apparent wave periods shorter than 4 hrs due to the limit by a Nyquist sample rate. However, gravity waves with intrinsic periods of less than 4 hrs can also be observed in the bihourly profiles because of Doppler shift

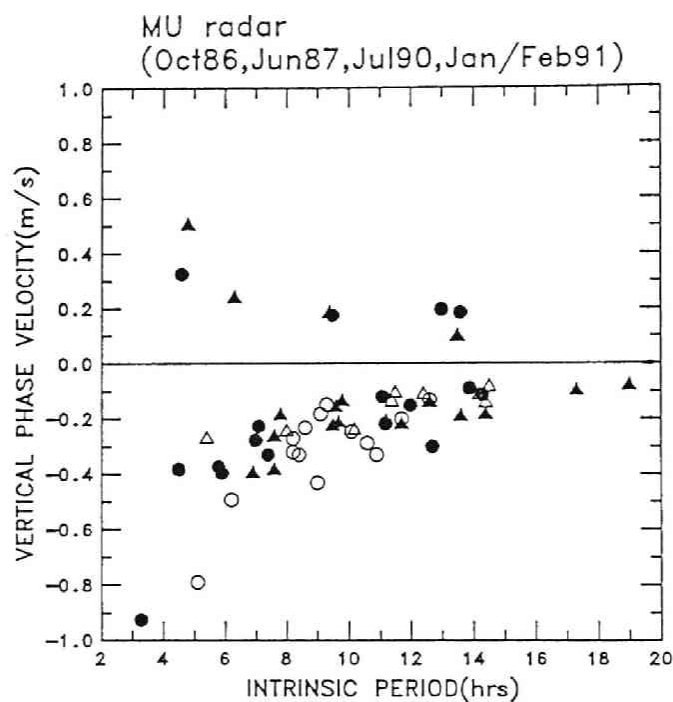


Figure 5.10: Vertical phase velocities vs intrinsic periods of the dominant gravity waves in the vertical profile at around 70 km, in June 6–29, 1987 (closed circles), July 2–21, 1990 (closed triangles), October 13–31, 1986 (open circles), and January 21 to February 8, 1991 (open triangles).

Month	sample	λ_z (km)	$\lambda_h(10^3\text{km})$	T^*	c_{gh} (m/s)	c_{ph} (m/s)
Jan/Feb'91	9	6.1 ($\sigma=1.4$)	1.02 (0.40)	11.3 (2.9)	20.1(4.7)	24.5 (5.8)
Oct '86	14	9.5 (2.9)	1.15 (0.39)	9.0 (1.9)	31.6 (9.4)	35.3 (10.0)
Jun '87	17	7.6 (2.3)	1.05 (0.65)	9.2 (3.6)	26.2 (8.0)	30.2 (10.2)
Jul '90	19	7.2 (1.7)	1.27 (0.74)	10.8 (3.6)	24.5 (5.9)	30.5 (8.6)
TOTAL	59	7.7 (2.4)	1.13 (0.60)	9.9 (3.3)	26.1 (8.3)	30.7 (9.7)

Table 5.3: Averaged parameter of dominant gravity waves for four campaign (mean (standard deviation))

effects caused by mean winds with the opposite direction to the phase velocity of the gravity wave. Nevertheless, Figure 5.10 does not show many cases of the wave with periods much smaller than 4 hrs.

Most of the observed waves showed negative phase velocity, demonstrating that gravity waves with upward energy transportation were dominant. However, 8 profiles showed positive phase velocity; 4 out of 8 occurred in June 1987 and the other 4 cases in July 1990. Therefore, in summer there seems to exist significant number of gravity waves with downward energy propagation. These gravity waves might be generated in the thermosphere, or the upward propagating gravity wave from lower atmosphere might be reflected at the height above mesosphere.

Vertical phase velocities were distributed in a range of 0.1 to 1 m/s, where most of the values were determined from 0.1 to 0.4 m/s. The corresponding group velocities were calculated as 0.1 to 0.4 m/s using a dispersion relation. It can generally be recognized that vertical phase velocity was larger when intrinsic wave period was shorter.

5.3.3 Vertical and Horizontal Wavelength

Figure 5.11 shows distribution of vertical and horizontal wavelength. Vertical wavelength was distributed between 4.5 and 15 km, while horizontal wavelength spread from 300 km to 2700 km. Monthly mean and standard deviations of these parameters are summarized in Table 5.3. A mean vertical wavelength was 6.1km and 9.5 km in January/February 1991 and October 1986, respectively. The difference in the mean vertical wavelength could be, at least partly, affected by the difference in a height coverage of the MU radar observations among four campaign periods, because those height range limitations work as a kind of high pass filter which eliminate the long vertical wavelength components.

The height range where the data rate was larger than 40 % at 1200 LT

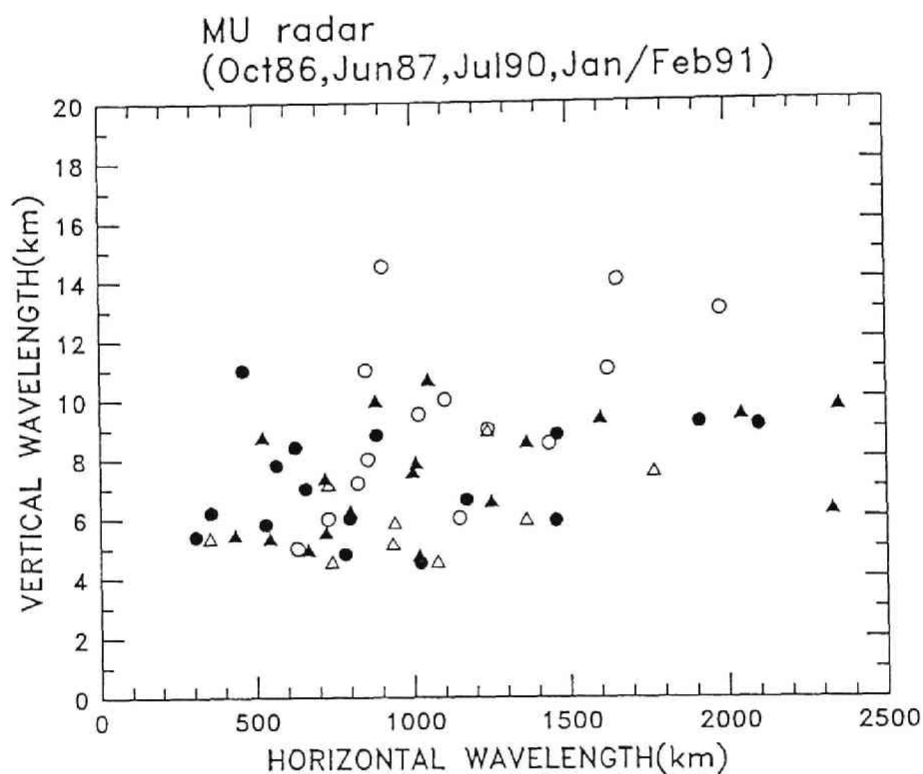


Figure 5.11: The same as Figure 5.10 except for horizontal wavelength versus vertical wavelength.

was 63–79 km, 65–76km, 67–77km and 65–75 km in October 1986, June 1987, July 1990 and January/February 1991, respectively, as shown in the previous section. Furthermore, in January/February the observed height range became remarkably smaller in the morning and afternoon than that during the mid day. As a result, the observed height range was large in October 1986, and small in January/February 1991, which probably caused the larger and shorter observed vertical wavelength of the gravity waves than those in the other two months.

Monthly mean horizontal wavelength and intrinsic period were 1.02×10^3 to 1.27×10^3 km and 9.0 to 11.3 hours, respectively. However the difference among the campaigns were small in comparison with their standard deviation. As a summary, we did not find any clear seasonal variation among the observation campaigns in λ_z , λ_h and T^* , except for the short λ_z in January/February 1991 and large λ_z in October 1986 which can be explained by the difference of observed height range.

It is noteworthy that the semidiurnal tides and the waves induced by non-linear interaction between tides and planetary waves have the wave periods in the range of the intrinsic periods of dominant gravity waves observed above.

As we described in Chapter 4, the semidiurnal tide observed by Kyoto meteor radar (35°N, 136°E) showed the wavelength larger than 60 km, while occasionally became 15–30 km in winter and spring. Therefore, in summer the basic mode of the semidiurnal tide, having long vertical wavelengths could be easily removed by the band pass filtering applied to the vertical profiles. While, contamination of the semidiurnal tides could be considered as a part of wind velocity fluctuations in winter and spring.

Manson and Meek [1991] pointed out from Saskatoon radar observation (52°N) that a nonlinear interaction between 2 day wave and diurnal or semidiurnal tides can produce a wave with period of 16 hrs, especially in summer, having the vertical wavelength of 10–15km, which could give contaminations to the analysis of the dominant components of gravity waves. However, this effect in the present study is rather small because of the following reasons. First, amplitudes of tides at 70 to 75 km altitudes seem to be as small as 5 m/s, being inferred from the Kyoto meteor radar observations as shown in Chapter 4. Second, the 2 day wave is normally enhanced only in mid-summer [Ito et al., 1984], while the semidiurnal tides in Kyoto are generally smaller than those in Saskatoon and become strong in September as described in Chapter 4, therefore, the occurrence of the 16 hrs waves will be limited, even if it exists.

Thus, we treated all the waves picked up by the hodograph analysis here as gravity waves, because the contamination by those nonlinear interaction waves and semidiurnal tides are expected to be small.

Figures 5.12 and 5.13 show distributions of vertical and horizontal wave-

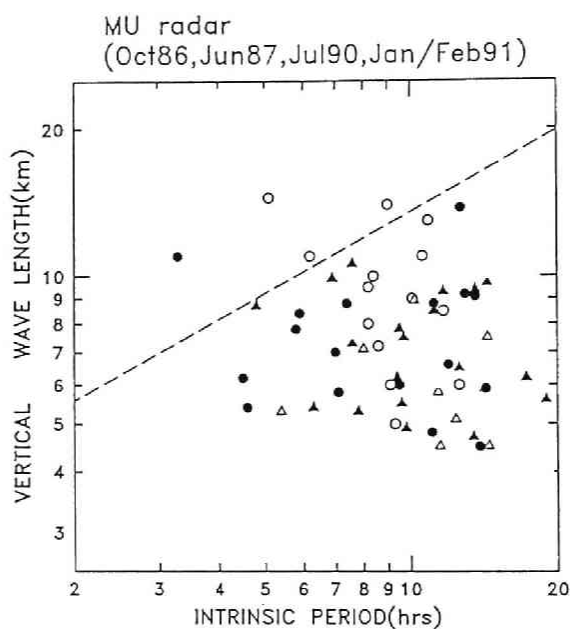


Figure 5.12: The same as Figure 5.10 except for intrinsic period versus vertical wavelength. Dashed line indicates the best fit of the results by sodium lidar observation, [Gardner and Voelz, 1987].

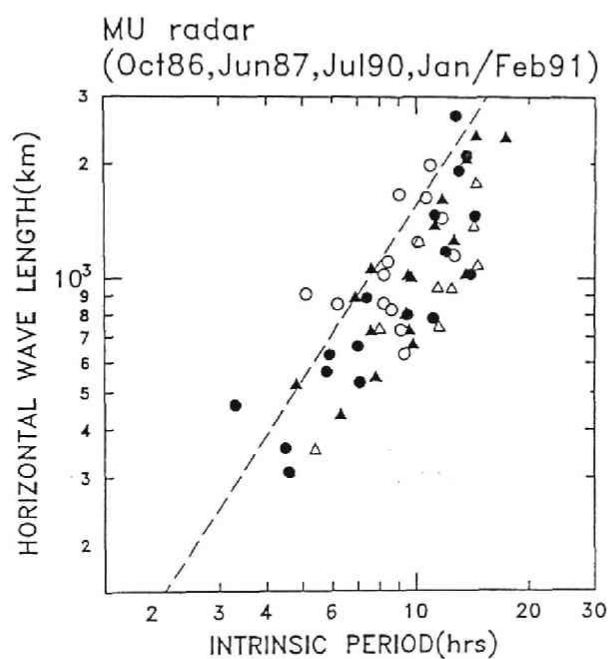


Figure 5.13: The same as Figure 5.10 except for intrinsic period versus horizontal wavelength. Dashed line shows the result of Gardner and Voelz [1987].

lengths as a function of intrinsic period, respectively. The intrinsic periods ranged from about 4 hrs to about 20 hrs, where the latter nearly corresponds to a Coriolis period at 35°N (20.9hrs). Note again that the maximum and minimum values of the vertical wavelength are limited by the observed height range and the pass band of a filter applied to the vertical profile. Figure 5.12 indicates that there is not a clear correlation between the vertical wavelength and intrinsic period. However, data points appeared in an area where intrinsic periods were longer and vertical wavelengths were shorter.

Figure 5.13 shows a clear positive correlation between the horizontal wavelength and the intrinsic period, which is partly affected by the fact that we picked up waves in a comparatively small range of the vertical wavelength. Gardner and Voelz [1987] observed gravity waves with vertical wavelengths less than 15 km at about 80–105km altitude by the Sodium layer observation with a lidar, and showed that monochromatic waves were more easily dissipated than saturated since these waves are damped by the eddy diffusion whose values range 36–300 m²/s at 80–100 km. They also showed that strong correlation existed between λ_z (in km) and T (observed period, in minute) and between λ_h (in km) and T (in minutes) as

$$\lambda_z = 0.40T^{0.55} \quad (r = 0.85) \quad (5.3)$$

$$\lambda_h = 0.093T^{1.52} \quad (r = 0.98) \quad (5.4)$$

where r denotes the correlation coefficient.

Manson [1991] compared these lidar data with the MF radar observations, and found that gravity wave, observed with MF radars, had wider distribution of parameters, and most of them were not easily damped by the eddy viscosity, however he explained that these differences occurred mainly because the two observational techniques have the different limit in observational vertical wavelength and the different time resolutions. Manson [1991] also showed that gravity waves observed by MF radars had a negative correlation between T and λ_z , while those by lidar observation show a positive correlation in the period range between 30 min and 14 hours. The present study showed no clear correlation between T^* (intrinsic period) and λ_z , and that the values of λ_z were rather widely distributed than those of lidar result.

The horizontal wave length λ_h observed with the MU radar /showed a relationship with intrinsic period, T^* , as

$$\lambda_h = 0.76T^{*1.13} \quad (5.5)$$

where the correlation coefficient $r=0.80$.

For a reference the best fit curve reported by Gardner and Voelz [1987] as in Eq. (5.4) is plotted in Figure 5.13 by the dashed line. The results observed with the MU radar show quite similar distribution with those by Gardner and Voelz. However, our result again showed wider distribution than the their results.

It is noteworthy that the dominant gravity waves in the vertical profile observed with our radar showed the characteristics similar to those observed with a lidar, although our radar observation use hodograph analysis of vertical profile of horizontal and vertical wind velocities at a selected local time while the lidar observation utilizes height-time cross section profile of the density and utilizes the apparent vertical phase velocity and vertical wavelength. The gravity waves observed with the MF radars [Manson, 1991] are different from the two observation mainly because of the comparatively rough height resolution (3–4 km) of the MF radars to observe the waves with vertical wave length of about 10 km.

Both the MU radar and lidar observations show that most of the waves were in a dumping range of eddy diffusion of $36\text{m}^2/\text{s}$ to $300\text{m}^2/\text{s}$ as shown in Gardner and Voelz [1987]. However, our results were slightly shifted towards longer periods and shorter wavelengths than the curve by Gardner and Voelz [1987]; that is, the distribution observed with the MU radar was shifted towards short wavelength by a factor of $2/3$, which corresponds to the eddy dissipation of $1/2.5$. This difference seems to be caused by the height range difference between the two observations, that is, the MU radar observed a region centered at 70–75 km which was lower than that of the sodium lidar observations by 10–20km, so that the gravity waves with shorter vertical wavelength and larger intrinsic periods are not dissipated yet at the height range of the MU radar observation.

5.3.4 Horizontal Propagation

Directions and amplitudes of horizontal phase velocities are shown in Figure 5.14 for four observation periods. In June 1987 almost all of the dominant gravity waves had northward component of horizontal phase velocity, being mainly concentrated in the northeastward direction, while only one wave showed small amplitude of southward component. Amplitudes of phase velocity ranged between 20 and 40 m/s with a mean value of 30.2 m/s. Group velocity, which can also be derived using the linear dispersion relation, showed the same horizontal direction aligned to the phase velocity, having slightly smaller amplitudes than the phase velocity as shown in Table 5.3.

In October 1986, the directions of the phase velocity was concentrated near south-southeastward, indicating no waves with the northward component. The amplitudes were between 30 and 50 m/s with mean value of 35.3 m/s, which was slightly larger than in June 1987.

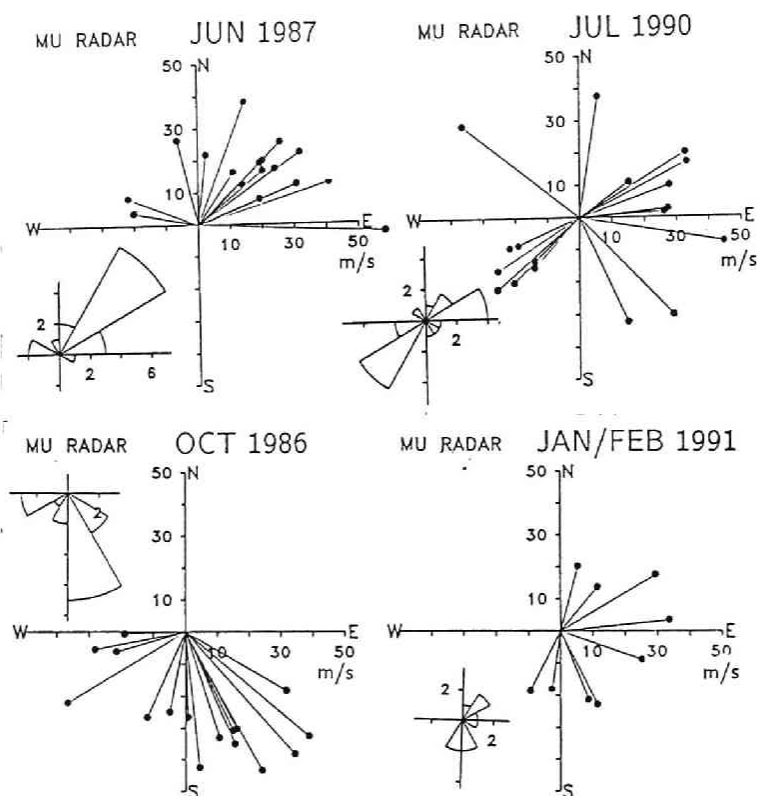


Figure 5.14: The distribution of the horizontal phase velocities of the gravity waves together with the histograms of the propagation directions, observed in June 6–29, 1987, July 1990, October 13–31, 1986, and January 21 to February 8, 1991.

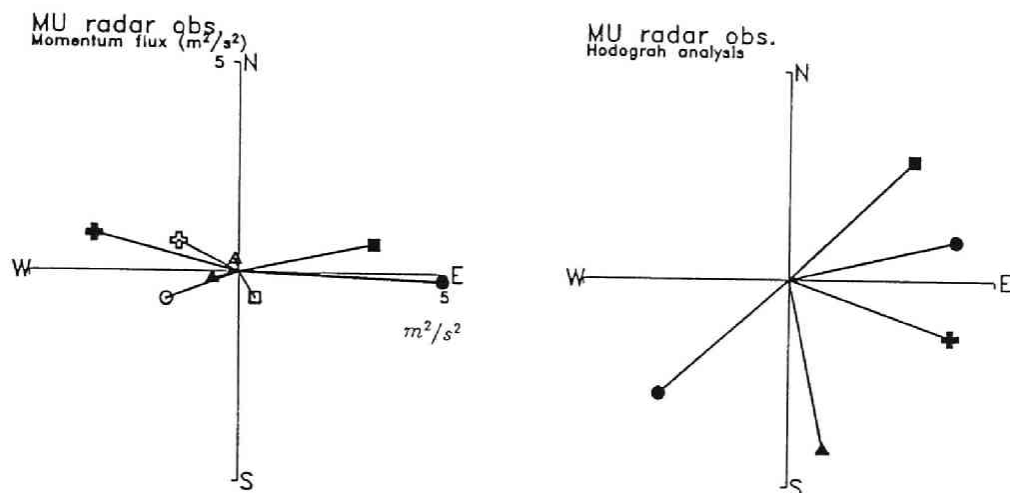


Figure 5.15: Averaged momentum flux of the gravity waves (left) and median directions of the dominant gravity waves in Figure 5.14 (right). The squares, circles, triangles and crosses correspond to June 1987, July 1990, October 1986 and January/February 1991. In the left panel closed and open symbols correspond to the component with periods of ~ 30 min and $30\text{--}5$ min, respectively.

In January/February 1991, 7 cases out of 9 had eastward component of phase velocity. The amplitude of phase velocity seems to be smaller than the previous two months, having the average value of 24.5 m/s.

In July 1990 the dominant propagation direction was not clear. Seven waves showed the directions between northward and eastward, while other seven profiles were between southward and westward. Gravity waves with the propagation direction between westward and northward, and between eastward and southward were three and one, respectively. Thus, the preferred direction of the gravity wave propagation in July 1990 was north-eastward and south-westward. Direction of horizontal propagation largely changed among four observation periods shown above.

Vincent and Fritts [1987] found from the MF radar observation that the propagation direction of gravity waves with periods of 8 to 24 hrs was different between summer and winter. Eckermann and Vincent [1989] also pointed out the difference in the preferential propagation directions in summer and winter from the falling sphere observation in the upper stratosphere.

Besides the hodograph analysis, we also estimated horizontal propagating direction of gravity waves from the direction of a momentum flux vector $(\overline{u'w'}, \overline{v'w'})$

as shown in the previous section. Figure 5.15 shows comparison of propagation directions of gravity waves between hodograph and momentum flux analyses at around 70 km altitude. Two sets of momentum flux vector are deduced in a period range from 8 – 10 hours to 30 min and between 30 min and 5 min, which correspond to the major and the minor parts of zonal momentum flux, respectively.

The right panel of Figure 5.15 shows the median direction of horizontal phase velocities, being determined by the hodograph analysis shown in Figure 5.14. In July 1990 two directions are indicated since the propagation directions distributed around northeastward and southwestward directions. Note that the components used in the hodograph analysis correspond to the low frequency gravity waves with the periods between 5 and 15 hrs. While, the momentum flux analysis was done for relatively higher frequency ranges with wave periods from 30 min to 8–10 hours and 5 min to 30 min (hereafter called as 8–0.5hrs and 30–5min momentum flux, respectively).

It is found that propagation directions by the 8–0.5hrs momentum flux had large seasonal variation; eastward in June and July and westward in October and January/February. The 30–5min momentum flux did not always coincide with the directions of the 8–0.5hrs momentum flux, however the directions of the two momentum flux vectors differed within 90° except for July 1990. On the other hand, the propagation directions obtained by the hodograph showed less significant seasonal variations, showing mainly eastward, but one direction determined in July 1990 was south-westward.

We here compare directions for these three different frequency ranges. In June 1987 these three directions agreed fairly well, especially the propagation directions determined by the hodograph and the 8–0.5hrs momentum flux were similarly directed toward north-eastward and east north-eastward, while the 30–5min momentum flux was slightly shifted toward south. In October 1986 and January/February 1991, these directions scattered widely, so the three determinations of directions did not always agree. Similar discrepancy in the propagation direction and momentum flux can be seen in the data in July 1990. Thus, the different frequency component of the gravity waves often showed different horizontal propagation direction.

Vincent and Fritts [1987] studied propagation direction of the gravity waves by the Stokes parameter analysis of SAD wind and momentum flux analysis of Doppler measurement with the Adelaide MF radar, finding that gravity waves in two wave period ranges, 1 to 8 hrs and 8 to 24 hours, propagated into a similar direction toward south-east in summer, while in winter the directions were almost opposite, that is, southward and northward for higher and lower period components, respectively. We also found by the MU radar observation

Month	sample	λ_z (km)	$\lambda_h(10^3\text{km})$	T^*	c_{gh} (m/s)	c_{ph} (m/s)
Nov-Dec	9	8.0 ($\sigma=0.9$)	1.18 (0.45)	10.2 (2.8)	26.5 (3.1)	31.3(4.7)
Jan-Feb	16	8.3 (1.5)	1.34 (0.47)	11.0 (2.4)	27.4(4.8)	33.0 (6.3)
Mar-Apr	22	8.0 (1.9)	1.16 (0.45)	10.3 (2.3)	26.5 (6.5)	31.1 (7.6)
May-Aug	41	9.0 (1.4)	1.42 (0.59)	10.7 (3.1)	30.3 (4.9)	36.3 (6.5)
Sep-Oct	16	8.5 (1.9)	1.11 (0.38)	9.5 (2.7)	28.3 (6.5)	32.3 (6.8)
TOTAL	104	8.5 (1.7)	1.29 (0.52)	10.4 (2.8)	28.4 (5.6)	33.6 (7.0)

Table 5.4: Averaged parameter of gravity waves for monthly mesospheric observations in 1985–1988 (mean (standard deviation)).

that different frequency component of gravity waves did not always have similar propagation directions.

5.3.5 Seasonal Variation in 1985–1988

So far we studied characteristics of gravity waves using data collected in the four campaign periods, and found that the gravity wave parameters such as intrinsic periods, horizontal wavelength, horizontal phase and group velocities did not vary with season, while the vertical wavelength showed slight variation which can be attributed to the seasonal changes in the height coverage of the MU radar observations. The horizontal phase velocity showed variations with seasons, while general preference seems to be eastward propagation. However, those variations are less clear as a seasonal variation because of the lack of spring data and a small sampling in winter when the background mean wind can vary largely by planetary waves and sudden warmings.

In order to confirm these variations more clearly, we further analyzed the monthly observation of the mesosphere with the MU radar from December 1985 to December 1988. Note that observations were not carried out in June 1986, December 1986, July 1987 and December 1987. Time resolution was 145 sec for all the observation period, and other observational parameters were the same as those used in the four campaign observations.

Bi-hourly wind velocity profiles were constructed on 159 days during the three year observation periods. Among them we clearly detected the dominant gravity waves in 104 cases. All the data were separated into five seasons; November to December, January to February, March to April, May to August and September to October. Note that the November to December and January to February are corresponding to the early winter and late winter, in which the background wind field is sometimes largely different because of the effect of sudden warming in the northern hemisphere as shown in Chapter 3.

Vertical Wavelength and Other Parameters

Characteristics of the dominant gravity waves such as vertical and horizontal wavelengths, intrinsic period, horizontal group and phase velocities are summarized in Table 5.4. Compared with the results obtained in the four campaigns shown in Table 5.3, λ_h were slightly larger, while other parameters agreed well within a discrepancy by up to 10 %.

We reported in 5.3.3 such that λ_z was shorter in January/February 1991 and longer in October 1986 as summarized in Table 5.3. The results in Table 5.4 showed that λ_z had the minimum value of 8.0 km in November–December and March–April, and the maximum value of 9.0 km in May–August, indicating the deviation of the mean vertical wavelength was smaller than that determined for the four campaign observations. Thus, the seasonal variations of the vertical wavelength are not clearly recognized from monthly observations. Average values of the horizontal wavelengths, intrinsic period, horizontal phase and group speed were 1.29×10^3 km, 10.4 hours 28.4 m/s and 33.6 m/s for the monthly observations, respectively, and again the differences of seasonal average values were significantly smaller than the standard deviation in each season. As a summary, the gravity wave parameters such as the vertical and horizontal wavelength, intrinsic wave period, and horizontal and vertical wavelength did not show clear seasonal variation.

Horizontal Propagation Direction

Distributions of horizontal phase velocities are plotted in Figure 5.16, where number of cases in every 30° in azimuth angle is also summarized as a histogram. In summer months propagation directions were concentrated into the north-eastward direction, which can more easily be recognized by the histogram, although there were several cases that waves propagated toward the opposite direction. In November and December the propagation directions were mainly distributed to the westward, while two cases showed the direction of east of northward. In contrast, in January and February the preferred direction again became north-eastward, and almost all the waves propagated into the direction within 90° from north-eastward. In September and October the preferential direction was centered near south south-eastward, while in March and April propagation directions were distributed fairly uniformly.

The propagation direction observed in the four campaign periods shown in Figure 5.14 showed good correlation with the results of the monthly observations shown above, suggesting these characteristics were repeated year by year. The remarkable feature is that the propagation direction was almost opposite in summer and early winter, north-eastward and westward for the former and latter

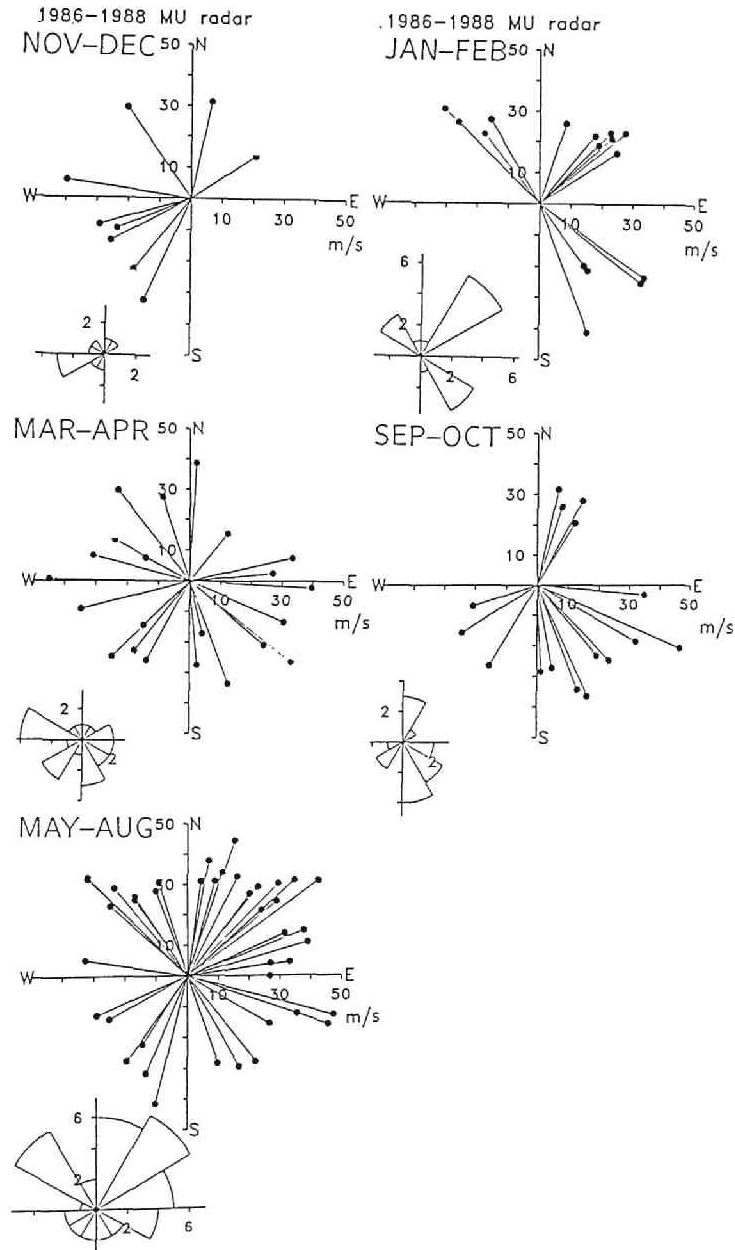


Figure 5.16: The distributions of the horizontal phase velocities of the gravity waves together with the histograms of the propagation directions, observed in 1985 December — 1988 December. Five plots correspond to the seasons; May-Aug, Sep-Oct, Nov-Dec, Jan-Feb and Mar-Apr.

months respectively, while the propagation direction became almost opposite between the early and late winter. Furthermore, the clear preference in the propagation direction in autumn but almost uniform distribution in spring were observed.

In order to estimate the average propagation direction more quantitatively, the momentum flux of each wave was deduced by using the amplitude of the horizontal wind component and a linear dispersion relation of the gravity wave as follows,

$$\overline{u'w'} = \frac{k}{|m|} \overline{u'^2} = \frac{1}{2} \frac{k}{|m|} U^2 \quad (5.6)$$

where u' is amplitude of horizontal wind fluctuation parallel to the azimuth of the wave propagation, and U is the amplitude of u' , i.e., the semi-major axis of the hodograph of the corresponding wave. Horizontal wavenumber k was inferred by using observed m and T^* by means of a dispersion relation.

The results are plotted in Figure 5.17, where the average momentum flux vector in each season is also indicated by closed circles with the bars indicating the standard deviations of the spread of the symbols. Note that confidence interval of the average momentum flux is roughly 1/4 of the standard deviation indicated here.

The mean direction of the momentum flux was north-eastward in summer and late winter, south-eastward to east south-eastward in autumn and westward in early winter which were similar with the phase velocity distribution shown in Figure 5.16. In spring, the average momentum flux shows east south-eastward direction, although the deviation of the momentum flux was large. However, if we consider the confidence interval of the mean momentum flux, the momentum flux carried by dominant gravity waves in spring is considered to be toward south of east. Out of the five seasons in a year shown in Figure 5.17, four seasons except for early winter in November and December, the momentum flux directions of the dominant gravity wave component were toward eastward rather than westward.

The reason of the non-uniform distributions of the propagation directions of gravity waves and their seasonal variations could be explained by the removal of waves at a critical level in the course of upward propagation of gravity waves [Ebel et al., 1987]. The non-uniform distribution of the propagation direction of the gravity waves could be also due to the distribution of the gravity wave source [e.g. Miyahara, 1986], while it is not yet revealed which caused the climatology of the propagation direction presented here.

We here compare these directions with the momentum flux with periods shorter than those dominant waves in order to clarify the difference of the direction of two components observed in the campaign periods. Figure 5.18 shows

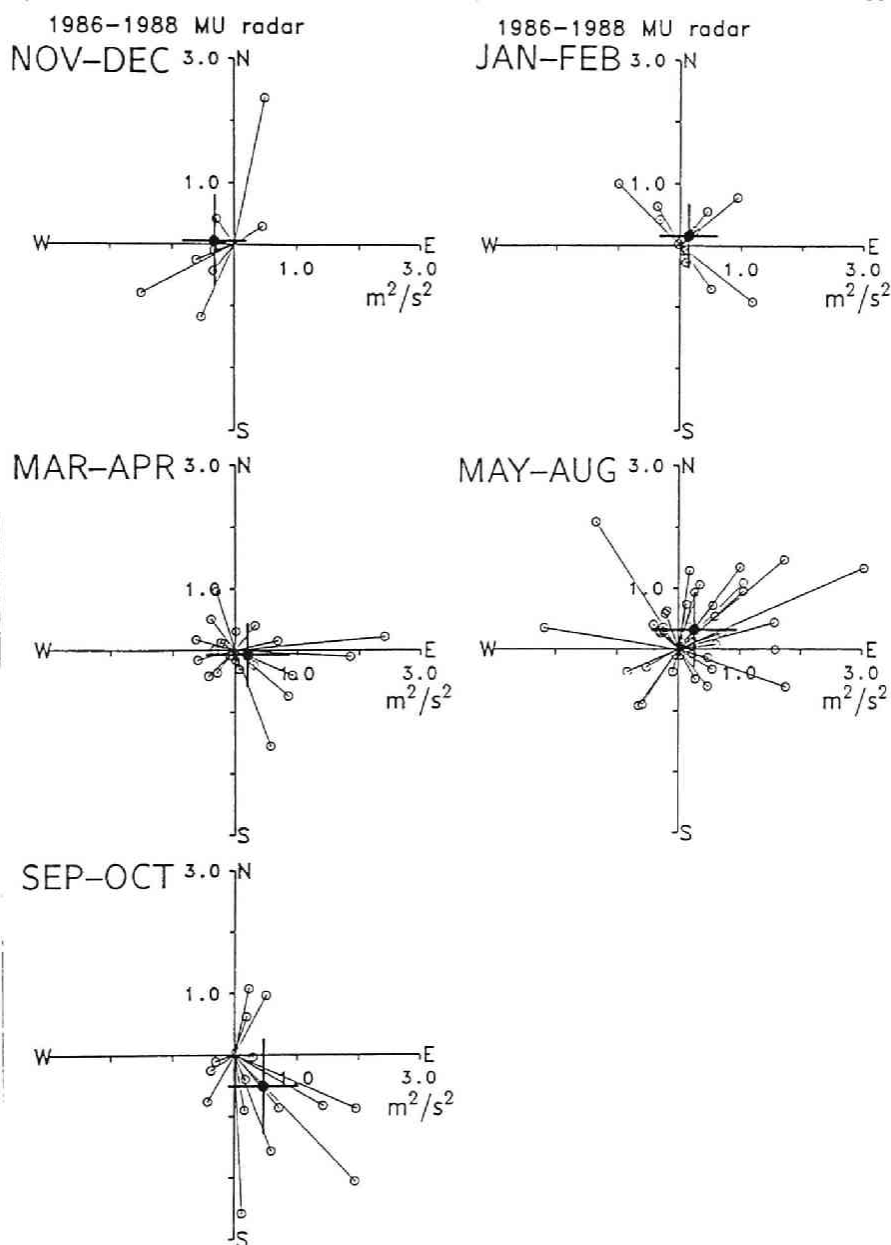


Figure 5.17: The same as Figure 5.16 except for the distributions of the momentum flux vector $(\overline{u'w'}, \overline{v'w'})$ of the each dominant gravity waves. Closed circles show the vector mean of the seasons and bars shows the standard deviation of the distribution.

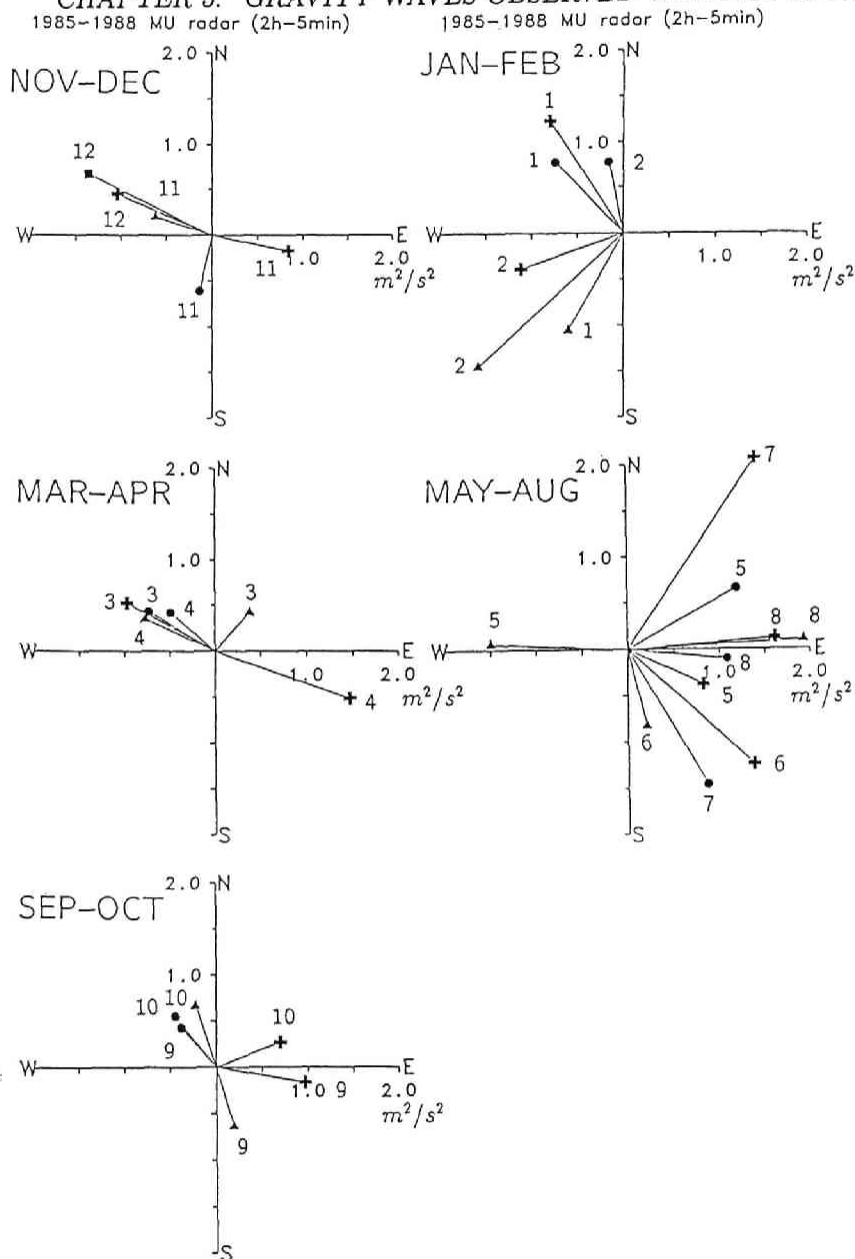


Figure 5.18: The monthly mean momentum flux in 1985 December — 1988 December after Tsuda et al., [1990a]. The squares, circles, triangles and crosses indicate the year 1985, 1986, 1987 and 1988. The numbers at the symbols shows the month of the year.

the distribution of the monthly momentum flux vectors for gravity waves with periods between 2 hours to 5 min observed by Tsuda et al. [1990a]. As shown in Figure 5.18, in summer the directions of momentum flux vectors were centered near eastward except for May 1987, while their directions were clearly reversed toward westward in winter [Tsuda et al, 1990a]. Furthermore, the westward flux were observed in both early and late winter with only one month showing the opposite direction in November 1988.

In spring and autumn, the tendency was not clear and the magnitude of the momentum flux was relatively small, indicating the transition between winter and summer. The meridional component of the momentum flux did not have clear preferential direction, but it showed a rather spread distribution having both northward and southward components. Therefore, the meridional component of the integrated momentum flux became fairly small because of a cancellation among these vectors.

Results shown in Figures 5.16 – 5.18 for monthly mesospheric observations generally agreed well with those determined in the four campaign observations, showing that dominant waves in the vertical profiles (long period waves) were likely to propagate towards eastward for all seasons, except for early winter, that is November and December, while the gravity waves with period of 2–8 hours to 5 minutes propagated into eastward and westward directions in summer and in the whole period of winter, respectively. The clear change of the preferred propagation direction of dominant gravity wave component from westward to eastward in the transition of early to late winter was not accompanied with the directional change of momentum flux with shorter periods, which were westward in both early and late winters.

Momentum Flux Carried by the Dominant Gravity Waves

It is noteworthy that the magnitude of the momentum flux in Figure 5.17 often became as large as $1\text{--}2 \text{ m}^2/\text{s}^2$, which was comparable to the momentum flux deduced for the waves with the period of 8–10 hours to 30 minutes in the previous section and of 2 hours 5 minutes by Tsuda et al. [1990a]. This fact suggests that the dominant gravity waves observed in the vertical profile of the wind velocity with wave periods of 5 to 15 hours can also carry a significant part of momentum flux.

The amplitude of momentum flux are summarized in Table 5.5 together with the results of four campaign periods. Here in Table 5.5, the amplitude of momentum flux vector is calculated as, $(\sqrt{(u'w')^2 + (v'w')^2})$. The amplitudes of momentum flux for dominant wave deduced by the hodograph analysis observed by monthly observations had the maximum in autumn with $0.83 \text{ m}^2/\text{s}^2$ and the

Month	hodograph (3–15hrs)	8hr–30min	30–5min	2hr–5min
June 1987	1.08 (m^2/s^2)	3.3	0.73	
July 1990	0.55	4.9		
Oct 1986	1.54	0.65	0.29	
Jan/Feb 1991	0.61	3.57	1.6	
May–Aug, 1986–8	0.64			2.97
Sep–Oct, 1986–8	0.83			0.54
Nov–Dec, 1985–8	0.57			1.02
Jan–Feb, 1986–8	0.46			1.90
Mar–Apr, 1986–8	0.47			0.97

Table 5.5: Magnitude of momentum flux ($|\overline{u'w'}, \overline{v'w'}|$) of different frequency components

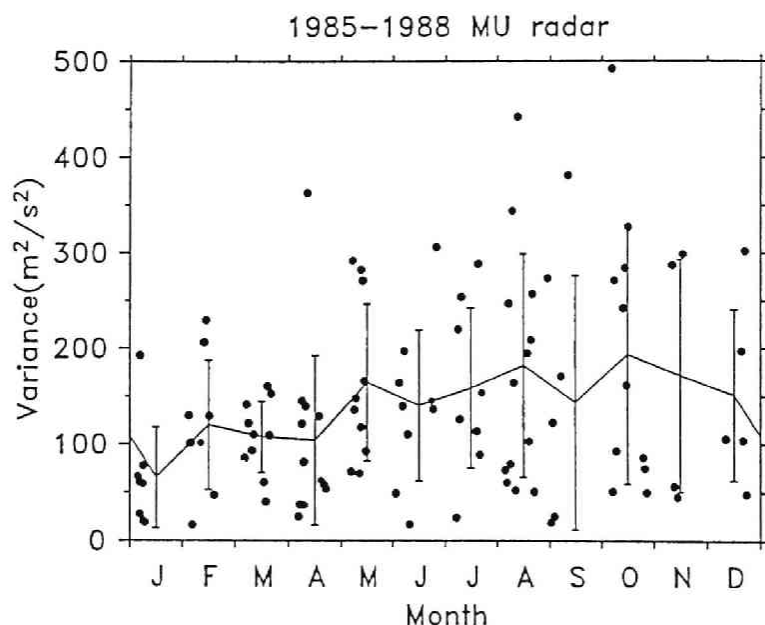


Figure 5.19: Seasonal variation of variance of the horizontal wind velocity ($\overline{u'^2 + v'^2}$) induced by the dominant gravity wave in the vertical profile in 1985 December — 1988 December. The line and error bars indicate the monthly mean and standard deviation of variances.

minimum in late winter and spring with 0.46 and $0.47 \text{ m}^2/\text{s}^2$, respectively. The variance of the horizontal wind velocities ($\overline{u^2 + v^2}$) induced by each dominant gravity wave also showed an annual variation with maximum in late summer to autumn and minimum in winter and spring, as shown in Figure 5.19.

The momentum flux of the dominant gravity waves observed in the four campaigns shown in Table 5.5 generally agreed with these monthly observations, showing large and small momentum flux with $1.54 \text{ m}^2/\text{s}^2$ and $0.61 \text{ m}^2/\text{s}^2$ in October 1986 and January/February 1991, while in July 1990 the momentum flux was as small as $0.55 \text{ m}^2/\text{s}^2$ suggesting a large year to year variation. On the other hand, the momentum flux due to the waves with periods 2 hours to 5 minutes in monthly observations showed maxima in summer and late winter with the values of 2.57 and $1.90 \text{ m}^2/\text{s}^2$, and small flux in spring and autumn as 0.97 and $0.54 \text{ m}^2/\text{s}^2$, respectively. The momentum flux of the wave component with the periods of 8 hours to 30 minutes, and 30 minutes to 5 minutes showed similar maxima in summer and winter and equinoctial minima, although the values were generally larger by a factor of 2 for the former component.

Therefore, contribution of the dominant gravity waves to the momentum flux was roughly about $1/6$ and $1/3$ of those due to the component with period of 8 hours to 30 minutes and 2 hours to 5 minutes, respectively. However, in September and October amplitude of momentum flux due to the dominant component became large but that of the shorter period component was small, thus the former became about 1.5–2.5 times as large as the latter. The momentum flux due to the dominant gravity wave component in the mesosphere is usually a minor component but it becomes dominant in autumn.

These features suggest that the activity of dominant component of the gravity waves in the vertical profile is not sensitive to the variation of vertical structure of the mean wind, showing westward and eastward in summer and winter in the upper stratosphere and the mesosphere, while that of the gravity waves with periods of 2 hours to 5 minutes does. It is possible to interpret the difference of the seasonal variation of dominant component of gravity waves, i.e. inertia gravity waves, from that of the short period gravity waves (e.g. 2 hours – 5min) occurs because the former is generated in the lower atmosphere and propagates upward with fairly small interactions with the mean winds while the latter is more closely concerned with the interactions with the mean zonal winds in the middle atmosphere. One possibility for the latter case is that the inertia gravity waves generated in the lower atmospheres propagated up to the upper stratosphere induce the short period gravity waves with the momentum flux direction similar to the original waves, when the inertia gravity waves break in the interaction with mean wind.

In order to clarify the difference between the seasonal variation of magnitudes

and directions of momentum flux transported by dominant gravity waves and short period gravity waves (8 hours – 5 minutes), the more extensive observations of the upper stratosphere and lower mesosphere are strongly desired.

5.4 Concluding Remarks

We studied characteristics of gravity waves in the mesosphere at 60–90 km observed by means of the MU radar in four periods; June 1987, July 1990, October 1986 and January/February 1991.

Frequency spectra of the radial wind velocity determined at 68–80 km during the four observation periods had a slope of about $-5/3$ for frequency smaller than about $4 \times 10^{-4} \text{s}^{-1}$. While, vertical spectra showed a slope of $+0.15$ to -0.4 . The spectral densities at the frequency of $2 \times 10^{-4} \text{s}^{-1}$ were the largest in June 1987 and July 1990 as $1.5\text{--}2 \times 10^4 \text{m}^2 \text{s}^{-1}$, became $6 \times 10^3\text{--}1.5 \times 10^4 \text{m}^2 \text{s}^{-1}$ in January/February 1991, and showed the smallest values of $4 \times 10^3 \text{m}^2 \text{s}^{-1}$ in October 1986.

Analysis of frequency spectra of momentum flux clarified that zonal momentum flux was predominantly contributed by gravity waves with period longer than 20 min in June 1987, 40 min in July 1990, 50 min in October 1986 and 40 min in January/February 1991, respectively. The meridional momentum flux did not show clear components that were dominantly detected in the momentum flux spectra, it rather fluctuated around zero.

Vertical profile of momentum flux, $\overline{u'w'}$ and $\overline{v'w'}$, in June 1987 and October 1986 showed that mean flow drag by low frequency waves (wave periods; 8 h–30 min) was larger than that by high frequency components (5 min–30 min) for both zonal and meridional components except that the meridional component in June 1987 was comparable to the zonal one. The directions of the drag of the high and low frequency components were opposite, causing cancellation between different frequency components. The total zonal and meridional acceleration observed in the daytime duration was $+51 \text{ m/s/day}$ and -4 m/s/day in June 1987 at 75 km, and -4 m/s/day and $+7 \text{ m/s/day}$ in October 1986 at 70 km, respectively. Since the momentum deposition by waves with different frequency components are often of different directions, it is important to observe the dominant frequency component of the momentum flux such as the waves with periods longer than 30 minutes. The zonal mean flow drag observed above showed good correspondence with Coriolis torque of the mean meridional wind observation.

A day to day variation of the gravity wave activity was larger in June 1987 than in October 1986. These variations in zonal momentum flux seem to have good correlation with the shear of zonal mean wind of each day, suggesting these

mean wind fields were accelerated and changed by the day to day variation of the gravity wave activity.

Dominant component of gravity waves in the mesosphere (at 70–75 km) detected in the bihourly vertical profiles were analyzed by using a hodograph. The analysis was done for data obtained in four campaigns June 1987, July 1990, October 1986 and January/February 1991 with the MU radar (35°N, 136°E), furthermore we included data collected by a monthly mesospheric observation from December 1985 to December 1988. Most of the dominant gravity waves showed upward energy propagation except for several cases in June 1987 and July 1990.

Mean values of vertical and horizontal wavelengths, and intrinsic period were 8.2 km, 1200 km and 10.2 hrs, respectively, which did not vary largely with season. Distribution of these parameters showed that the dominant gravity waves in the vertical profiles observed in this study would be dissipated by the eddy diffusion of 36–300 m²/s, which is similar to the results of Sodium lidar observation by Gardner and Voelz [1987]. However, the gravity waves observed with the MU radar can be more easily dissipated by the eddy viscosity with the magnitude of 1/2.5 of that the lidar observations showed, which corresponds to the low observation altitude for the MU radar by 10–20 km.

The amplitudes of horizontal group and phase velocities did not show clear seasonal variations, and their mean values were 28 m/s and 33 m/s, respectively. On the other hand, directions of the horizontal phase and group velocities, that is, the horizontal propagation directions, varied with season. The preferred directions were north-eastward in summer, westward in early winter, north eastward in late winter and south south-eastward in autumn. In spring the propagation direction was more uniformly distributed than the other seasons, while the mean momentum flux showed south of eastward direction. Thus, the eastward propagation was more prevailing than the westward one except for in early winter.

Besides the dominant waves (3–19 hrs in intrinsic period), propagation directions of 8–0.5 hrs and 30–5 min components of gravity waves were deduced by momentum flux analysis, which showed large seasonal variation; the major components of the momentum flux i.e. 8–0.5 hrs component shows eastward in summer and westward in both early and late winter. Then, in June these momentum fluxes and those of dominant waves showed similar direction, while other seasons differed widely, showing that the different frequency component often show different propagation direction and their momentum flux could cancel with each other.

The comparison of the momentum flux of dominant waves and momentum flux of the waves 8 hrs–5 min clarified that typical momentum fluxes are 0.5–1 m²/s² and 1.6–3 m²/s², respectively. The wind variance and the amplitude of

the momentum flux of those dominant component showed annual variation with maximum in autumn and minimum in late winter and spring, while those of the momentum flux with waves with 8hrs–5min periods varied semiannually with summer and winter maxima and equinoctial minima. The dominant components of gravity waves did not have a significant momentum flux in comparison with the waves with 8hrs–5min period, except for in autumn (September and October) when the momentum flux due to the dominant component became larger than those by waves with period less than 8 hours by a factor of 1.5–2.5.

The seasonal variations of the amplitude and direction of the momentum flux suggest that the dominant gravity waves are affected less significantly by the seasonal variation of the mean wind in the upper stratosphere and mesosphere than the short periods gravity waves.

Chapter 6

Latitudinal Comparisons of Gravity Wave Characteristics

In this chapter we compare the gravity waves observed by the MU radar (35°N , 136°E) with the results of Saskatoon MF radar (52°N , 107°W), and Adelaide MF radar (35°S , 139°E).

Spectral analysis of gravity waves in terms of frequency and vertical wavenumber were studied by several radar observations at various locations [Carter and Balsley, 1982; Vincent 1984a; Frezal et al., 1981; Meek et al., 1985b; Tsuda et al., 1989]. Vincent [1984a] found that the annual mean of frequency spectra for horizontal wind velocity fluctuations were similar between Adelaide (35°S) and Townsville (19°S) at 86 km altitude, and that these spectra were also similar to that observed at Poker Flat (65°N) in summer at 85 km. Yamamoto et al. [1986] showed that the gravity waves with periods longer than two hours observed with the Kyoto meteor radar (35°N) were smaller than the results obtained by the three radars mentioned above. Meek et al. [1985b] showed that the results at Saskatoon (52°N) agreed well with the spectra analyzed for the three stations shown by Vincent [1984a].

Meek et al. [1985b] further discussed the seasonal variation of the gravity wave energy in various frequency ranges by presenting a time height section; Ebel et al. [1987] extended this to other years (1979–84). Climatology of the gravity wave activity was also reported by Reid [1986], Vincent and Fritts [1987] and Tsuda et al. [1990b]. They found that the wind velocity variance due to gravity waves showed a semiannual variation with equinoctial minima, generally; however, the gravity wave with long period (>1 hours) at Saskatoon showed the annual variation with winter maximum below 75 km, with only a weak semiannual variation at 75–85 km [Reid, 1986].

In the stratosphere and the mesosphere meteorological rocket observations

revealed latitudinal and seasonal variations of gravity wave energy with a vertical scale of about 10 km [Hirota, 1984], showing semiannual and annual variations in the lower and higher latitudes, respectively. Rayleigh lidar observations also showed that gravity waves with the vertical wavelength of 15–0.6 km have annual and semiannual variations in the upper stratosphere and the mesosphere, respectively [Wilson et al., 1991].

Although qualitative comparison of the gravity wave energy has been done for several radar observations, detailed quantitative comparisons have not yet been carried out so far. It is noteworthy that gravity waves are normally observed as a superposition of waves with various frequencies and wavelengths. Each observation technique extracts only a part of spectra because of a limit in both time and height resolutions. Therefore, it is sometimes difficult to detect the difference in the gravity wave energy by using different observation techniques. In order to detect the difference of the gravity wave activity at different sites, first of all the data should be collected simultaneously in the same frequency range and the same height range with the same sampling volume. Furthermore, it is important to apply the same data analysis techniques to these data.

6.1 Data analysis

In the present study we compare the mesospheric gravity waves observed with the MU radar at Shigaraki (35°N, 136°E) and the MF radar at Saskatoon (52°N, 107°W) in January – December 1989, and further compare the gravity waves observed with the MU radar and Adelaide MF radar for two campaign periods of June 1987 and January/February 1991. We first discuss the data analysis in comparing the results of MU radar and Saskatoon MF radar closely. The comparison of the results of the Adelaide MF radar and the MF radar should be similar because the radar techniques of Adelaide and Saskatoon MF radars are basically same, as shown in Chapter 2.

Table 6.1 shows the comparison of the observation techniques and parameters between the MU radar and Saskatoon MF radar in that period. The Saskatoon MF radar was operated continuously throughout a year except for the periods of malfunction of the system such as in October 9–28, 1989. On the other hand, the MU radar carried out the mesospheric observation for 4–5 days per a month, because observations of lower atmosphere and ionosphere were allocated for the rest of the periods. Moreover, the radar echoes from mesosphere are very weak at night at the frequency of the MU radar (46.5MHz), so that the mesospheric observation is limited during the daytime. Thus, the Saskatoon MF radar provides better data in the sense of observation duration. In contrast, height and time resolutions are better for the MU radar (600m, 3.5 min.) than

	MU radar	Saskatoon MF radar
Location	Shigaraki, Japan (35°N, 136°E)	Saskatoon, Canada (52°N, 107°W)
Observation Period	4-5 days /month in 1989 (*1)	every day in 1989 (*2)
Observation time	daytime 6-8 LT to 16-18LT	24 hrs
Height range	60-90 km	58-109km
Time resolution	3.5 min	5 min
Height resolution	600 m	3 km
Wind measurement	DBS	SA (horizontal) Complex-correlation (vertical)
Beam Direction	(Az, El)=(0°, 0°), (0°, 10°), (90°, 0°), (180°, 10°), (270°, 0°)	(Az, El)=(0°, 0°)
Beam width	3.7°	26°

*1 no observation in March 1989

*2 no observation in October 9-28, 1989

Table 6.1: Comparison of the observation between the MU and Saskatoon

Saskatoon MF radar (3km, 5min.), which means the MU radar can observe the gravity waves with shorter vertical wavelength and shorter wave periods.

The observational height range also differs between two radars; the MU radar observes 60–90km altitudes, while the height range of good data acquisition throughout a year is between 68–80 km as shown in Chapter 5. On the other hand, Saskatoon MF radar can observe the height of 58–109 km, although in the lower several height the echo can be received in only daytime, and the observation of high altitudes is strongly affected and limited by the total reflection by the ionosphere.

The more important difference is that the observational technique of wind velocities. The MU radar is measuring the Doppler velocity using pencil beam with 3.7 degree in half power width. The beam directions are steered pulse by pulse in the vertical and four oblique directions. Thus, the radial wind for the five beam directions are determined. Saskatoon MF radar utilizes the spaced antenna (SA) method [Meek, 1980], which gives the horizontal wind velocities (u, v), and a simple complex-correlation method [Meek and Manson, 1989] provides the vertical velocity (w) as well, which is a kind of Doppler measurement. The antenna beam width of the transmission of Saskatoon radar is about 26° in calculation, but the effective beam width is estimated as smaller than 20° below 80 km by the experimental results in 1985. This wider beam causes a larger radar sampling volume in horizontal direction than that of the MU radar. It is noteworthy that vertical winds from MF radars have been used less often [Meek and Manson, 1987, 1989], but those interferometer studies reported that the complex correlation method used here provides useful estimates of the vertical velocity over 5 minutes intervals. This appears to be because the radar scatterers are rather uniformly distributed in azimuth, which minimizes the possible effect of oblique scatterers in the rather broad antenna beam. The agreement with VHF/MST radar measurements of w from Poker Flat (65°N) was shown by Coy et al. [1986] and Meek and Manson [1989].

Nevertheless, the differences in the observation methods for the MU and MF radars cause some difficulty in comparing the gravity wave activity such as wind variance, when the observed data are analyzed for each radar without paying any attention to the companion radar. For instance, the difference of observational local time may cause the difference of the magnitude of gravity waves by the local time dependency of gravity wave amplitudes [Fritts and Vincent, 1987].

The height difference also affects the amplitude of the gravity wave, because the amplitude varies with height. The time and height resolutions determine the frequency and vertical wavenumber range of the observation, so that high resolutions usually give larger intensity if simple variance of the fluctuation is calculated. The beam pattern of the transmitting and receiving antenna also leads

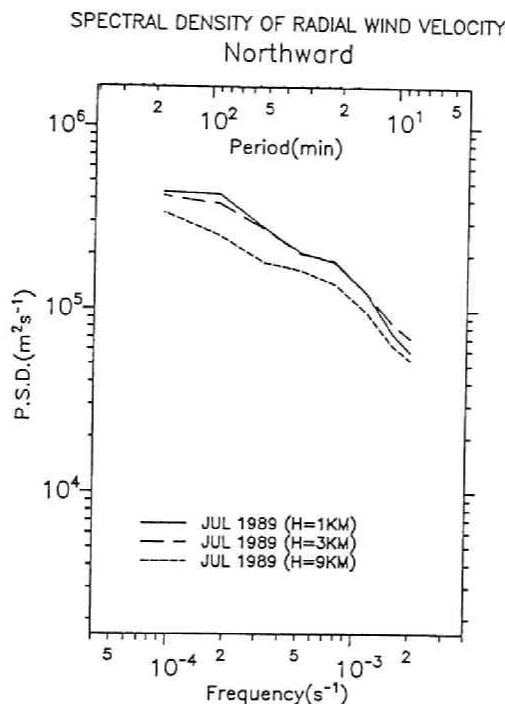


Figure 6.1: Effect of height averaging upon the power spectra of radial wind velocities observed with the MU radar on July 17–21, 1989. Average of the northward and southward beams is plotted. Three lines corresponds to the height averaging of 1 km, 3 km and 9km, respectively. Each averaging in height was done for 60 sec observations of the radial wind velocity.

to averaging of the wind fields within the radar sampling volume, mainly to the horizontal direction, which causes the elimination of the horizontal wavenumber component larger than the corresponding cutoff.

Thus, the comparison between the two radars needs careful treatment of data in order to minimize the difference of the observations. In the present study, only the data during the daylight hours are analyzed; from 06–08 LT to 16–18 LT for the MU radar and from 08 LT to 19 LT for Saskatoon radar. The height range to analyze is also adjusted to 68–80 km for MU radar and 67–79 km for Saskatoon data.

Next we consider the difference of the height resolution and antenna beam width more quantitatively. The low height resolution works as a low pass filter which removes the wave components with vertical wavelength shorter than the

twice of the height resolutions. The vertical wave number spectra of the wind velocity in the middle atmosphere is known as having the shape of m^{-3} in the range of $m > m^*$, where m^* is the characteristic wavenumber of the gravity waves [Smith et al., 1987; Tsuda et al., 1989]. Tsuda et al. [1989] observed that characteristic wavelength $1/m^*$ in the mesosphere was larger than 10 km and the spectra with wavelength shorter than that agree well with the shape of m^{-3} . Thus, the wave variances of the wind components with the vertical wavelengths smaller than 10 km are considered to be small.

Figure. 6.1 shows the effect of the height resolution on the frequency spectra of the observed radial wind fluctuations observed with the MU radar. The spectrum of each height resolution plotted here was obtained by averaging the wind fluctuation in the height bin of 1km, 3km and 9km. As is expected from the vertical wavenumber spectra discussed above, the effect of the averaging is small for the case of averaging with 3 km. On the other hand, averaging for 9 km loses 25 % of the energy in the range of the period of 10–100 min. The height resolutions of the MU radar and Saskatoon MF radar are 0.6 km and 3 km, respectively, and therefore, a significant difference between the two radars is not expected (Figure 6.1). The reduction of the frequency spectra due to averaging for 9 km in height is 40 % and 20 % at the frequencies of $2 \times 10^{-4}\text{s}^{-1}$ and $1 \times 10^{-3}\text{s}^{-1}$, showing large reduction in the low frequency component. This is probably because the vertical wavelengths of the observed gravity waves are generally smaller for the waves with longer periods, which is also shown in the result of GRAVNET system of the Saskatoon MF radar by Manson and Meek [1988].

Thus, we have shown that the averaging effect by the height resolution of 3 km is not significant for gravity waves with periods of 10–100 minutes. However, we average the data taken by the MU radar over 3 km in height in order to align the height resolution to Saskatoon MF radar, and to minimize data gaps caused by low S/N.

Yamamoto et al. [1986] studied the gravity waves spectra with the period longer than 2 hours observed by Kyoto meteor radar (35°N) and found that the spectral density was significantly smaller than the result of other MF and MST radars. They also suggested that the smaller spectra by Kyoto meteor radar can be due to its coarse height resolution (> 10 km). Figure 6.1 indicates that their suggestion seems to be true, and so an observation with better height resolution such as the MU radar is necessary in order to compare the variance of the wind velocity of gravity waves at Shigaraki and other locations.

The wind component observed by the MU radar is a Doppler velocity of the radial wind, so that the zonal and meridional component of the wind fluctuation are often substituted by the radial wind fluctuations obtained by the oblique

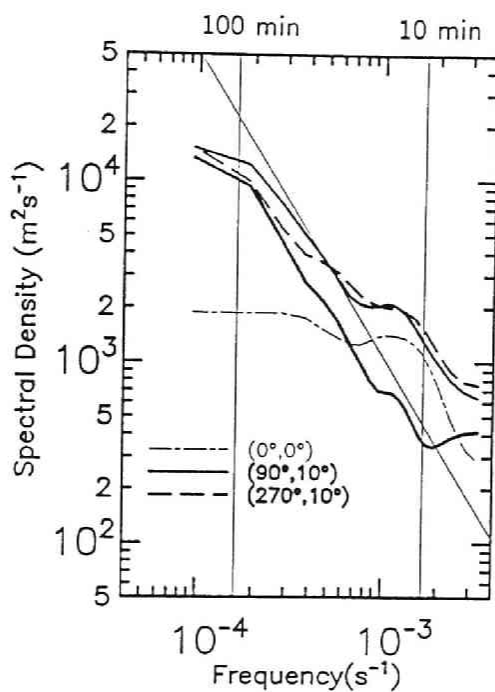


Figure 6.2: An example of the determination of the frequency spectrum of the zonal wind ($F_u(\omega)$: thick solid) from the frequency spectra of the radial wind velocities of the vertical ($F_{V_{RV}}(\omega)$: thin dash and dot), eastward ($F_{V_{RE}}(\omega)$: solid) and westward ($F_{V_{RW}}(\omega)$: dashed) beams. Data were taken in May 15–19, 1989.

beams [Tsuda et al., 1990a]. The radial wind V_R is written as,

$$V_R = \xi \sin \theta + w \cos \theta \quad (6.1)$$

where ξ , w and θ are horizontal and vertical wind velocities and zenith angle of the antenna beam, respectively. Note that horizontal wind component ξ indicates the wind component in the direction of the azimuth of the antenna beam, so that for example in the eastward beam ξ is denoted as $\xi = u$. Then, the power spectrum of radial wind $F_{V_R}(\omega)$ is written as

$$F_{V_R}(\omega) = F_\xi(\omega) \sin^2 \theta + 2F_{\xi,w}(\omega) \sin \theta \cos \theta + F_w(\omega) \cos^2 \theta \quad (6.2)$$

where $F_\xi(\omega)$, $F_w(\omega)$ are the power spectra of ξ and w , and $F_{\xi,w}(\omega)$ is the cross spectrum of ξ and w . By using Eq. (6.2), the spectra of zonal wind u and meridional wind v can be calculated as

$$F_u(\omega) = \frac{F_{V_{RE}}(\omega) + F_{V_{RW}}(\omega)}{2 \sin^2 \theta} - F_{V_{RV}}(\omega) \cot^2 \theta \quad (6.3)$$

$$F_v(\omega) = \frac{F_{V_{RN}}(\omega) + F_{V_{RS}}(\omega)}{2 \sin^2 \theta} - F_{V_{RV}}(\omega) \cot^2 \theta \quad (6.4)$$

where $F_{V_{RV}}(\omega)$, $F_{V_{RN}}(\omega)$, $F_{V_{RE}}(\omega)$, $F_{V_{RS}}(\omega)$ and $F_{V_{RW}}(\omega)$ are frequency spectra of the radial winds of vertical, northward, eastward, southward and westward beam, respectively.

Figure. 6.2 shows the example of the spectra of radial wind velocity for eastward beam $F_{V_{RE}}(\omega)$, westward beam $F_{V_{RW}}(\omega)$, vertical beam $F_{V_{RV}}(\omega)$ and the calculated eastward wind spectrum $F_u(\omega)$. The confidence interval of the spectral density of the horizontal wind $F_u(\omega)$ was 23% and 28% at the frequencies corresponding to the wave periods of 100 minutes and 10 minutes, respectively. Then, the confidence interval of the variance between the periods of 10–100 min deduced by integrating these spectra is about 10%. Therefore, differences of variance of the wind velocities between the MU and MF radars, which are larger than 10 % are probably significant in the following comparisons.

It is also noteworthy that horizontal wind velocities of SAD wind measurement also could include the contamination of vertical wind velocity component, that is the spatial difference of the vertical wind velocity in a horizontal extent of the scatterer are observed in the SAD wind [Royrvik, 1983]. This effect becomes significant at the short periods waves with periods around 10 minutes because of the small horizontal wavelength for those short period waves.

Another important point we have to consider is the data analysis method. For the MU radar observations, the power spectrum of the radial winds calculated for every 1 km height are integrated for a specified frequency range (e.g. 2 hours

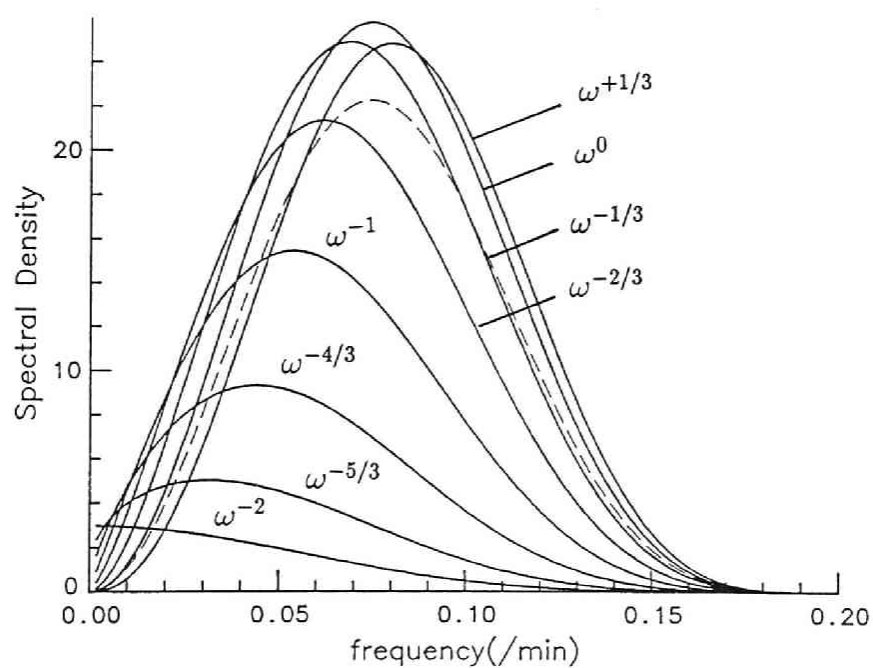


Figure 6.3: Output spectra of the difference filter for various spectral slope.

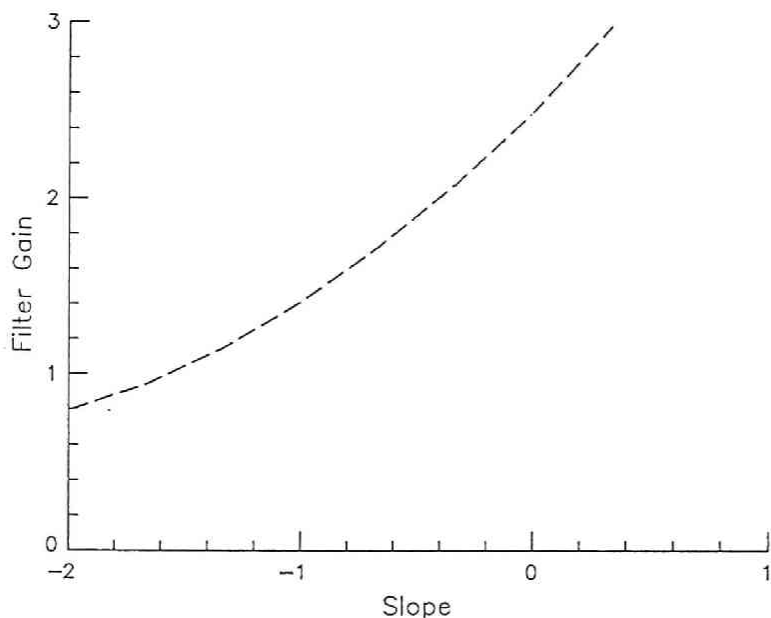


Figure 6.4: The gain of the difference filter (V_D/V_P) compared with the spectral window between 10 min to 100 min for a range of spectral slopes k of the spectrum $F(\omega) = a\omega^k$.

to 5 minutes) to investigate the gravity wave activity [Tsuda et al., 1990a]. On the contrary, for the Saskatoon MF radar, a difference filter method has been adopted to obtain the intensity of the gravity waves [Meek et al., 1985b]. They used the difference filter for the data taken every 5 minutes, and thereby extracted the wave component with periods of 1 hours to 10 minutes, although the longest period is slightly larger than 1 hour [Meek et al., 1985b]. At first, we examine the difference between the two analysis methods in detail.

Figure. 6.3 shows the frequency spectra of the output of the difference filter (DF) used by Meek et al. [1985b] for various input spectra; ω is the frequency and k is the spectral slope. It is found that the filter pass band largely depends on spectral slope k . The pass band of 90 % of the total energy is 140min–10min, 67min–8 min 40min–6.2min and 29 min–3.3min in periods for the spectral slope k of $-5/3$, -1 , $-1/3$ and $1/3$, respectively. Thus, a smaller slope (steeper in negative slope) in frequency spectra produces a lower frequency for the pass band.

The difference of the wind fluctuation variance between the integration of

the spectrum (PSD) and the differential filter (DF) can be estimated as follows. The variance V_P calculated by the integration of the spectrum is

$$V_P = \int_{\omega_1}^{\omega_2} F(\omega) d\omega \quad (6.5)$$

where $F(\omega) = a\omega^k$ is the spectrum of the wind fluctuation, and ω_1 and ω_2 are the lower and higher boundary frequency for integration. The variance V_D by the difference filter is calculated as

$$V_D = \int_0^\infty F(\omega) P(\omega) d\omega \quad (6.6)$$

where $P(\omega)$ is the frequency characteristics of the difference filter described as [Meek et al., 1985b],

$$P(\omega) = (2 - 2 \cos(2\pi\omega T_D)) \left(\frac{\sin(\pi\omega T_L)}{\pi\omega T_L} \right)^2 \quad (6.7)$$

where $T_D=5\text{min}$ and $T_L=4.5\text{min}$. The ratio of V_D and V_P is shown in Figure. 6.4, where ω_1 and ω_2 are set to $1/100\text{min}^{-1}$ and $1/10\text{min}^{-1}$. The ratio of V_D and V_P becomes large with increasing (more positive) spectral slope k ; 0.95 for $k = -5/3$ and 2.5 for $k = 0$. This means that the variance ratio is significantly influenced by the slope of the frequency spectra of the object.

Figure. 6.5 shows the ratios of V_D and V_P for the monthly mean variances of zonal, meridional and vertical winds ($\overline{u'^2}, \overline{v'^2}$ and $\overline{w'^2}$) for the Saskatoon MF radar in 1989. It is found that the ratio for $\overline{u'^2}$ and $\overline{v'^2}$ are almost constant for all season. The annual average and standard deviation of the ratios are 1.58 and 0.12 for $\overline{u'^2}$, and 1.75 and 0.09 for $\overline{v'^2}$, indicating the monthly ratios are almost constant. The average and standard deviation of $\overline{w'^2}$ are 2.07 and 0.29. The larger average ratio for $\overline{w'^2}$ than for $\overline{u'^2}, \overline{v'^2}$ corresponds to the more gradual slope of the frequency spectra of vertical wind w' as shown in Figure. 6.4, which will be discussed in the next section. The larger standard deviation of the ratio for $\overline{w'^2}$ suggests the spectral slope of w' has larger variability than u' and v' , which is also shown in the next section. The almost constant ratios of $\overline{u'^2}, \overline{v'^2}$ and $\overline{w'^2}$ throughout a year indicate that the difference filter [Meek et al., 1985b] can be applied reasonably to study the seasonal variation of the wind variance despite the simpleness of the calculation.

Thus, in conclusion we have found that variances of the wind fluctuation obtained by the two analysis techniques are significantly different and that their ratios depend on the spectral slope. Although the ratio of the two is rather constant over the year, it seems most appropriate to use the integration of the frequency spectra to determine the gravity wave intensity for the data of both the MU radar and the Saskatoon MF radar in order to avoid the difference of the analytical method.

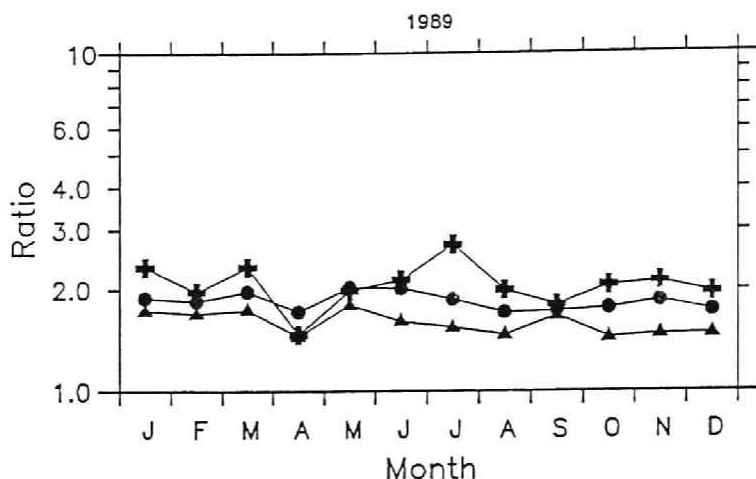


Figure 6.5: Ratios of the variances calculated by the difference filter and by integration of power spectral density for zonal (u' , triangles), meridional (v' , circles) and vertical (w' , crosses) components.

6.2 Latitudinal Difference

6.2.1 Frequency Spectra

In this subsection, we compare the frequency spectra of the wind velocities u , v , and w observed by the Saskatoon MF radar and the MU radar. The comparison is mainly focused on the spectral shape, and intensities of the spectra are discussed in the next subsection as a variance of the wind velocities. Since the spectral slope could affect largely to the band pass frequency of the difference filter, it is also important to examine the seasonal variation of the spectral slope for checking the gain of the difference filter. Also the spectral shape is important for checking the validity of the wind determinations.

Figures 6.6 to 6.8 shows the monthly mean spectra of zonal, meridional and vertical wind components. Note that the spectra of eastward wind in January was not able to be determined for the MU radar data because of the shortage of effective wind data. Figure 6.6 shows that the spectra at Saskatoon MF radar had the logarithmic slope of $-5/3$ for all months at the frequency range lower than $20\text{--}70\text{ day}^{-1}$ ($20\text{--}70\text{ min. in period}$); although the slope becomes gradual at the frequency of about 70 day^{-1} (20 min. in period) in January–February and September–December. On the other hand, the strong slope persists only until a frequency of $20\text{--}30\text{ day}^{-1}$ ($70\text{--}45\text{ min. in period}$) in May to August. Thus, the wind fluctuations with periods shorter than $70\text{--}45\text{ min}$ were enhanced in spring

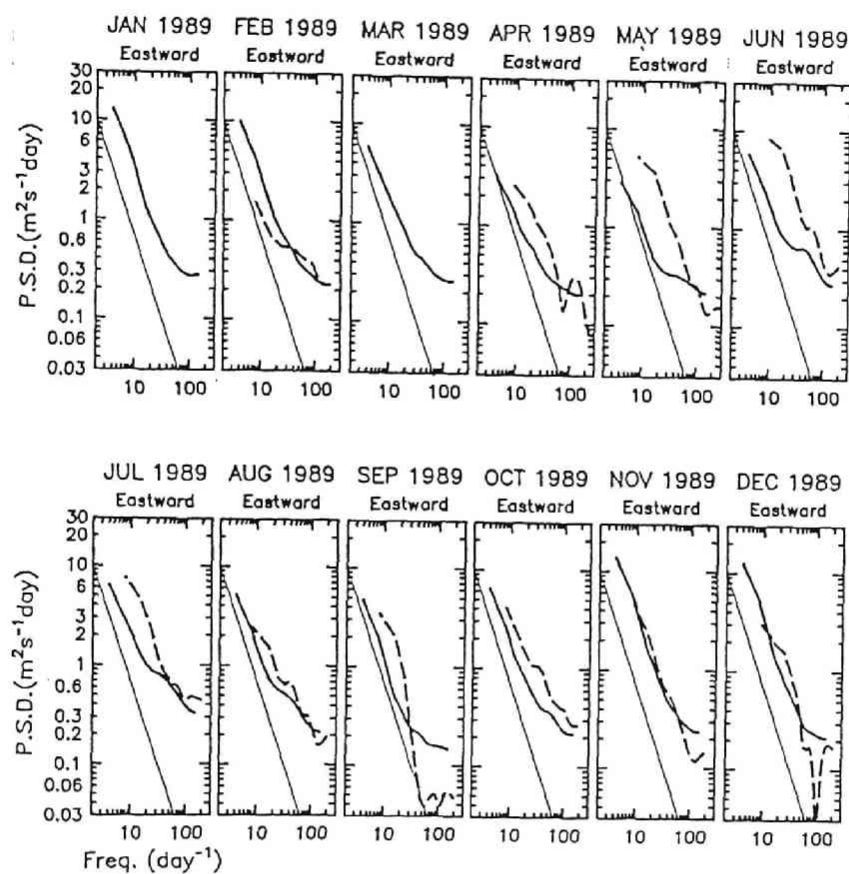


Figure 6.6: Monthly frequency spectra of zonal wind fluctuations u' by Saskatoon MF radar (solid lines) and the MU radar (dashed lines). Thin lines show the slope of $\omega^{-5/3}$.

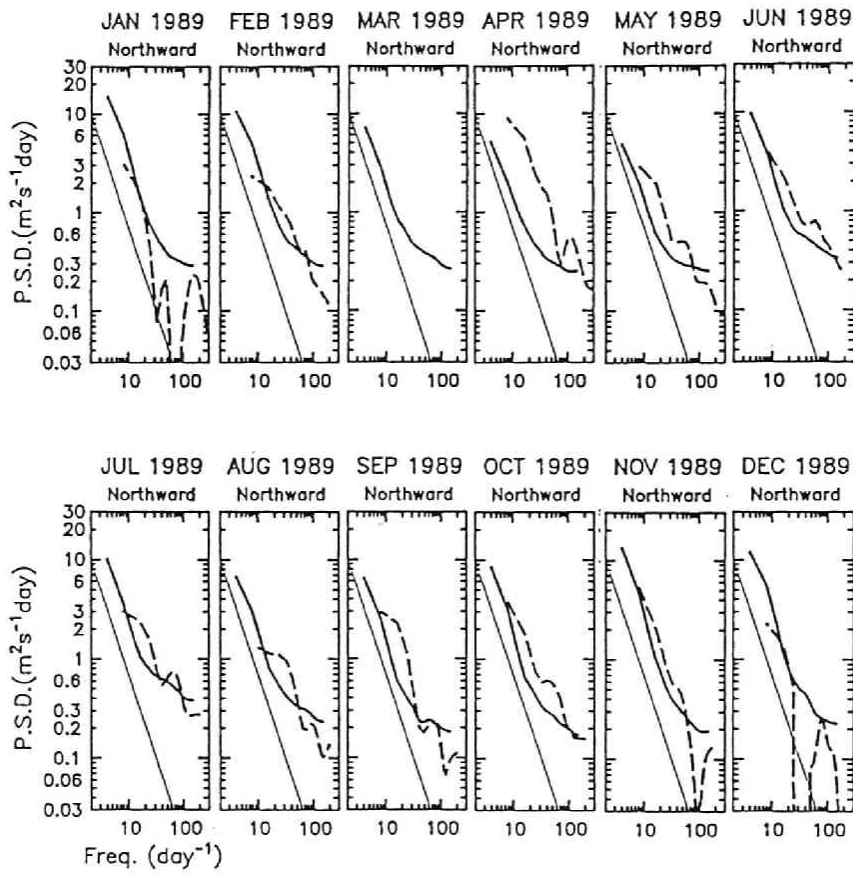


Figure 6.7: The same as Figure. 6.6, except for the meridional (v') component.

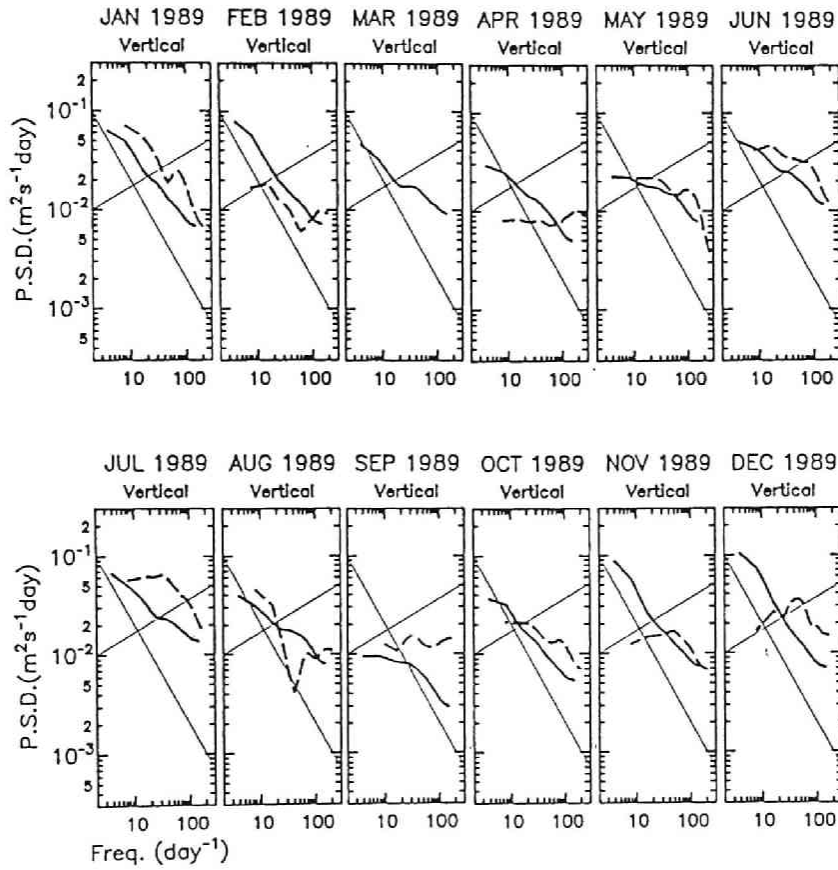


Figure 6.8: The same as Figure. 6.6, except for vertical (w') component. The thin lines shows the spectral slope of ω^{-1} and $\omega^{+1/3}$.

slightly and summer especially, i.e. from March to August.

The frequency spectra of the MU radar also showed that the spectral slope was basically $-5/3$ throughout a year. The enhancement at the high frequency component seen in Saskatoon data cannot be clearly found for the spectra obtained with the MU radar, except for July when the frequency component larger than 70 day^{-1} was seen to be enhanced. In winter such as in November and December, the spectra between the MU and Saskatoon became very similar in shape, showing good agreement with the spectral slope of $-5/3$.

We here discuss the error component in determining the the fluctuations of the wind velocity. When we take the frequency spectra of the wind fluctuation, the error of the spectral density is considered as two ways.

One is that the error in the wind measurement itself, that is, the error in the wind determination in one Doppler or SAD measurement. This is further divided into a random error and the contamination of other physical phenomena. The former is independent between the wind measurements, so that the observed spectra are added by noise floor corresponding to the error amplitude. The latter is concerned with some physical process such as the contamination of the differential vertical wind to the SAD wind [Royrvik, 1983], and could have a specified spectral shape.

The other is the error or stability of the spectra due to the small ensemble average of the spectra caused by a short observation periods in comparison with the frequency resolution of the spectra.

The enhancement of the frequency spectra of eastward wind in high frequency component observed in the Saskatoon data can be interpreted by the noise floor or the contamination of the vertical wind component, because the corresponding enhancement is not found in the MU radar data. Thus, the frequency spectra of the real eastward wind can be smaller than that shown in Figure 6.6.

On the other hand, the spectra obtained by the MU radar seems to be less stable than that by the Saskatoon MF radar especially in the frequency component larger than 70 day^{-1} (20 min. in period). This can be explain by two reasons. One is that the spectra of the MU radar were determined for the observation of 4–5 days per a month, while spectra were averaged for a full month for Saskatoon radar. The short average duration causes the confidence interval of spectral density at each frequency of the MU radar data to be larger by a factor of 2.5 than that of Saskatoon. Next, the frequency spectra of the horizontal wind velocity of the MU radar were deduced by the radial wind velocities obtained by the vertical beam and two oblique beams as shown by Eqs. (6.3) and (6.4). For the high frequency gravity waves, the amplitudes of the vertical wind fluctuation w' became large in comparison with the horizontal component u' and v' as derived from dispersion relation of the gravity waves. Therefore, the contribution

of the vertical winds in the radial wind velocities becomes dominant; so that the estimation of the spectra of the horizontal wind component by Eqs. (6.3) and (6.4) become to include larger uncertainty due to the subtraction of the spectral densities of a same order.

The 95 % confidence interval of the spectra shown in Figures. 6.6 and 6.8 is 26 % at the frequency of 10 day^{-1} (144 min) and 7 % at 144 day^{-1} (10 min) for the vertical wind w' and 23% at the frequency of 10 day^{-1} and 28 % at 144 day^{-1} for the horizontal wind u' and v' for the data of May 15–19, 1989; These figures are 9% and 2.4% for Saskatoon data. It is noteworthy that the reliability of the high frequency regions of horizontal wind spectra observed with the MU radar may change according to the spectral density of three radial wind velocities because the horizontal wind spectra are calculated by the subtraction of the spectra of three antenna beams. Although the relative confidence interval of spectral density was not good for the high frequency range for the MU radar data, the magnitude of spectral density itself was small in the high frequency range according to the slope of $\omega^{-5/3}$ approximately. Thus, the absolute confidence interval itself was not so large even in the high frequency regions. For example, the variance calculated by the integration of the spectra of zonal wind velocities for the period range of 10–100 minutes has the confidence interval of about 5 % as a standard deviation for the spectra of May 15–19, 1989. Thus, the variance comparison in the following sections is considerably reliable despite the large variation of the high frequency range.

The day to day variation of the gravity wave activity can also be the reason of the less stability of the frequency spectra, because the average of the spectra for short duration cannot smooth out the variability. This effect is also significant for the MU radar data, because a large day to day variation of the gravity wave variance was observed as shown in Chapter 5.

The spectra of the northward wind component shown in Figure. 6.7 are very similar in above features with eastward component. It shows basically the frequency slope of $-5/3$ for both the MU radar and Saskatoon MF radar, and the enhancement of high frequency waves relative to low frequency component in the frequency range larger than $20\text{--}30 \text{ day}^{-1}$ (50–70 min) in spring and especially summer at Saskatoon. The enhancement at the MU radar was not clear except for July 1989.

However, the vertical wind spectra showed discrepancy in shape between the Saskatoon MF radar and the MU radar. The vertical wind spectra at Saskatoon, shown in Figure. 6.8, had the spectral slope of $-2/3$ to -1 from November to February, and became $-1/3$ to $-1/2$ in March to October. In May and September, at the frequency range lower than $50\text{--}80 \text{ days}^{-1}$ the spectral slopes became almost flat with the slope of -0.1 to -0.15 . The spectra by the MU radar show

the slope of -0.6 in January and agreed well with that of Saskatoon; while in November and December, the slope in the high frequency range larger than 70 day^{-1} (20 min) shows about -0.6 and agreed with that of Saskatoon, but in the lower frequency there is a transition to slopes which were positive with the value of $+0.15$ – $+0.3$. In February, the transient frequency was smaller than that in November and December. In summer, the spectral slopes with the MU radar were also flatter in the low frequency region lower than 30 – 50 day^{-1} (30–50 min) (i.e. the same trend as at Saskatoon), but became steep with the range of -0.6 to -1 in the higher frequency region from May to August. The spectra in April and September showed very flat slopes. Thus, in general the spectral slope of the Saskatoon data were more likely to have steeper negative value than that of the MU radar, which showed both negative and positive slope according to the month.

The difference of the spectral shape between the Saskatoon MF radar and the MU radar can also be interpreted as due to the wide antenna beam of Saskatoon MF radar. Meek and Manson [1987] showed that the scattering point observed by the interferometry using Saskatoon MF radar sometimes drifts in the illuminating area of the antenna. In that case, the radial wind velocity contains the projection of horizontal wind as well, when the scatterer is not situated at the zenith. Because the difference of the spectra between the MU radar and Saskatoon MF radar are not well described, it is rather difficult to determine the source of the difference. However, we had better keep this difference in interpreting the result shown later.

Thus, we found that the slope of the frequency spectra were similar between the MU and Saskatoon for the horizontal wind component, except for enhancement of high frequency component at Saskatoon and variability at the MU radar. The spectra of vertical wind component showed some discrepancy between the Saskatoon MF and the MU radar data, indicating larger negative spectral slope for Saskatoon data.

Since the seasonal variation of the spectral slope could cause a variation of the pass band of the difference filter which has been used for Saskatoon data; it is rather better not to use the difference filter because it can provide the change of filter pass band according to the variability of spectral shape as described before especially for vertical component. Furthermore, the difference of absolute gain between the difference filter (DF) and integration of power spectra (PSD) is significant and have slight seasonal variations as shown in the previous section. So that in the following section, the integration of power spectra is adopted for the comparison of the data between Saskatoon and the MU radars.

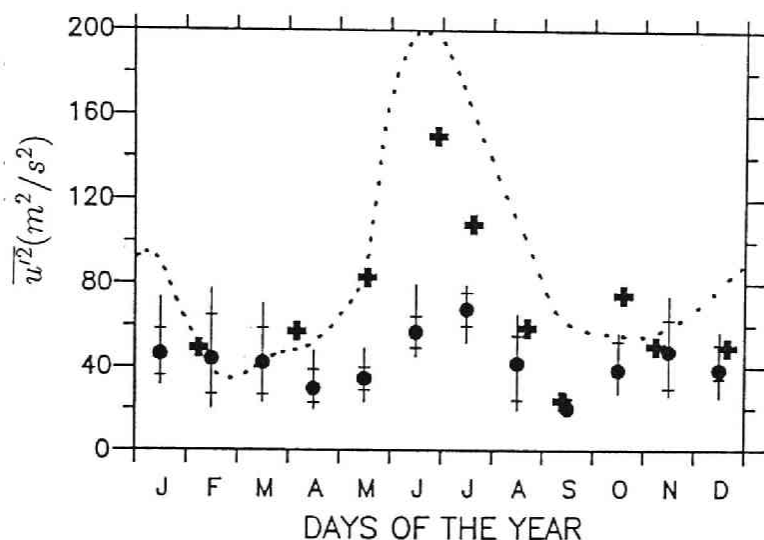


Figure 6.9: Monthly mean variance $\overline{u'^2}$ (zonal) for the MU radar (crosses) and Saskatoon MF radar (circles). The error bars and vertical lines indicate the representativity of the 5 days mean variance (see text). The dotted line shows the average profile of $\overline{u'^2}$ for the wave periods of 120–5 minutes observed in 1985–1988 with the MU radar [Tsuda et al., 1990a].

6.2.2 Seasonal Variation of the Wind Variance

Next we compare the monthly mean intensity of the gravity waves between the MU radar and Saskatoon MF radar. Figures. 6.9 to 6.11 show the monthly mean value of $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ for the frequency component of 10–100 min which were calculated by integrating the power spectral densities shown in Figures. 6.6 to 6.8. The detailed time-height cross section of the gravity waves observed with the Saskatoon MF radar is discussed elsewhere [Manson and Meek, 1991]. The MU radar data (cross symbols) are not plotted when the spectra of the wind were not reasonably determined (e.g. u' of January 1989), as well as in March when no observations were carried out.

The values of Saskatoon MF radar are plotted by the circles with lines and error bars in order to show the representativity of the 5 day average of such as the MU radar observation. Among the observation for a month, possible successive five days' data were chosen and integrated spectra were averaged for the five days. The top and bottom of lines indicate the maxima and the minima among the five day averaged variances. The error bar shows the standard deviation

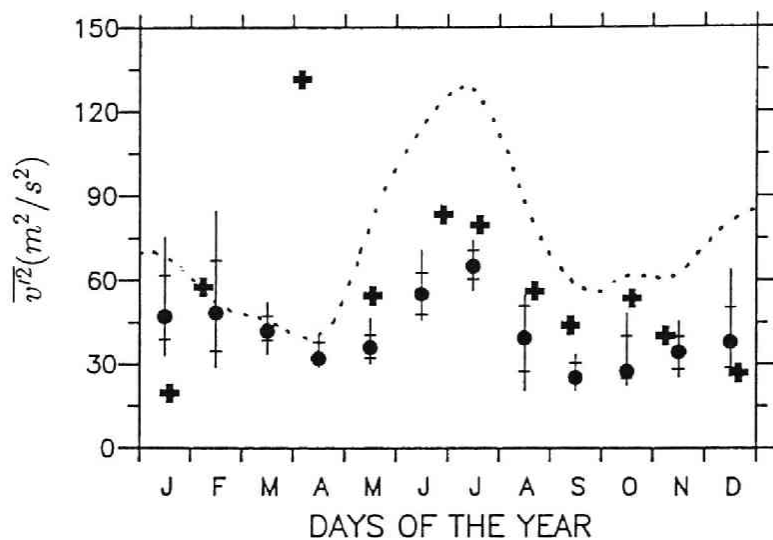


Figure 6.10: The same as Figure. 6.9 except for $\overline{v'^2}$ (meridional).

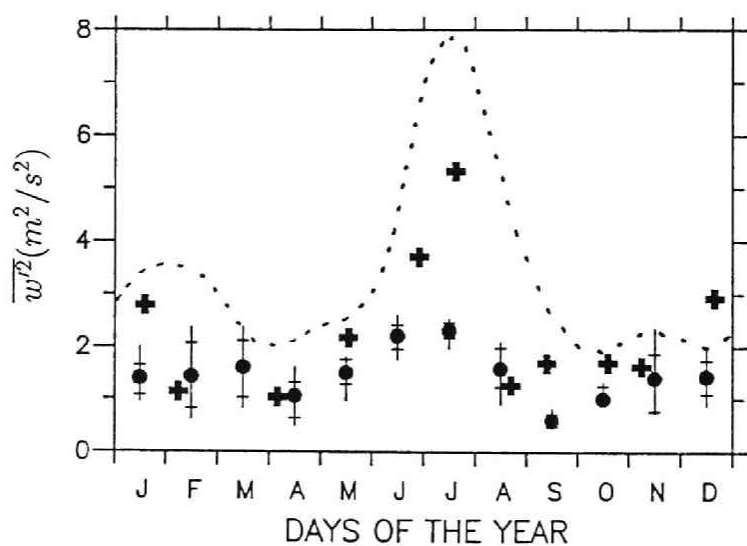


Figure 6.11: The same as Figure. 6.9 except for $\overline{w'^2}$ (vertical).

(s. d.) of the five day mean variance from the monthly average of the five days mean.

The monthly mean of $\overline{u^2}$ at Saskatoon in Figure. 6.9 shows the semiannual variation with the maxima in summer and winter and minima in April and September, same as reported by the previous observations at Saskatoon [Meek et al, 1985b; Ebel et al., 1987]. The error bar and vertical lines became large in November, January, February and March, while small in summer and some equinoctial months e.g. April–July, September, and October. This means the variability of the gravity wave intensity with period longer than 5 days is large in winter, which may include the variability associate with the winter sudden warmings and planetary waves. A large variation is also found in August, which probably is because of the rapid change from the summer maximum to the equinoctial minimum. In contrast, the comparatively small variation in May could reflect a more gradual variation from spring to summer than the end of summer.

The monthly variance $\overline{u^2}$ observed by the MU radar shows more complicated variations than that by Saskatoon data, probably because the MU radar observation is limited 4–5 days in a month as discussed in the previous section and the monthly variations are not well averaged. The variations of the wind variance observed by Tsuda et al. [1990a] with the MU radar in 1985–1988 are supplementally plotted in Figure 6.9 by a dotted line in order to clarify the seasonal variation from the variability due to the short duration average. Note that the wind variance by Tsuda et al. [1990a] was given as the component of the waves with periods from 2 hours to 5 minutes and the variance of radial wind of the vertical and oblique beam. Therefore, we calculate the variance of the eastward wind component by the relation of Eq. (6.3).

The variance $\overline{u^2}$ observed by the MU radar also shows the semiannual variation with summer and winter peaks, while the summer peak is much larger than the winter one. As a result, the variance in summer were significantly larger for the MU radar data than that of Saskatoon MF radar, while in winter the difference became small. The minima of the variance in September show good agreement for both the MU and Saskatoon MF radar, suggesting the autumnal minima occur globally.

It is noteworthy that Tsuda et al. [1990a] showed that the winter peak of the gravity wave intensity is less systematic than that in summer showing variability of intensity with months and years. The error bars of the Saskatoon variances also show that the five-day mean variances in winter can vary significantly in a month, supporting the suggestion by Tsuda et al. [1990a]; i.e. the winter variability of the variance by the MU radar was due to the variation in that month.

Component	mean ratio(MU/Saskatoon)	Standard dev.	Confidence int.
$\overline{u'^2}$	1.6	0.5	0.17
$\overline{v'^2}$	1.5	0.8	0.26
$\overline{w'^2}$	1.6	0.6	0.19

Table 6.2: Yearly average of the ratio of monthly mean variance of the wind fluctuation between the MU radar and Saskatoon MF radar.

The variation of $\overline{v'^2}$ shown in Figure. 6.10 were generally similar with that of $\overline{u'^2}$. The variance by the Saskatoon MF radar shows clear semiannual variation with summer and winter peaks and minima in April and September. The variance by the MU radar also shows a peak in summer but less strong than the eastward component, which is also found for the result of Tsuda et al. [1990a]. The variance decreased in September, while the winter subpeak was not found for the data in 1989, although the results in 1985–1988 showed a subpeak in December/January. This difference in winter is probably due to the year to year variability or the day to day variability. The data in April obtained by the MU radar was much larger than those in other months, which is not considered as usual seasonal value by the comparison with the data of Tsuda et al. [1990a].

The variance $\overline{w'^2}$ at Saskatoon shown in Figure. 6.11 again has the summer and winter peaks and minima in April and September, indicating similar variations with $\overline{u'^2}$ and $\overline{v'^2}$. The variation of $\overline{w'^2}$ at the MU radar had the maximum in summer and secondary peak in winter, and minima in April and August, showing clear semiannual variation, although the autumnal minimum is earlier by one month than the Saskatoon data.

Thus, the variance of $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ of the waves with period 10–100 min at 67–80 km altitude showed basically semiannual variation with summer maxima, winter secondary peak and equinoctial minima for both Saskatoon and Shigaraki, while the difference between the summer peak value and winter one was larger for the MU radar data than that of Saskatoon MF radar.

Next, we compare the difference of the wind velocity variance of the gravity wave between Saskatoon MF and the MU radars. In early summer, that is, in May, June and July, the values of $\overline{u'^2}$ for the MU radar were significantly larger than those for the Saskatoon MF radar results by a factor of about 2. The values of $\overline{v'^2}$ and $\overline{w'^2}$ also showed larger intensity for the MU radar during those months, but the ratios of the wind variance obtained by the MU radar to that by the Saskatoon MF radar were smaller in $\overline{v'^2}$. Also for $\overline{w'^2}$ in the winter, such as in December and January 1989, the variances by the MU radar were larger than those by the Saskatoon MF radar by a factor of 2. The variance $\overline{v'^2}$ by the MU radar in April was also larger by a factor of 3.5. In the other seasons,

the difference between the MU radar and Saskatoon MF radar is not always clear compared with the error bars in each month. In order to compare more quantitatively, including all the months and excluding the variability in a month for the MU radar data, we compare the annual average of the monthly variance ratio between the MU radar and the Saskatoon MF radar. The averaged ratio is summarized in Table. 6.2. As a result; the ratios of the wind velocity variance observed with the MU radar to that with Saskatoon MF radar were 1.6, 1.5 and 1.6 for $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$, which are significantly large compared with the confidence interval.

As we discussed in the previous section, the comparison of the wind variance shown above is affected by the larger horizontal averaging of the wind field for Saskatoon MF radar data than that for the MU radar data. This effect causes the variance of the Saskatoon observation to be smaller than the real variance of the wind fluctuations. We here discuss the loss of the variance by the spatial averaging more quantitatively. We assume that the measured wind velocity by the Saskatoon MF radar is the average of the space within 10 degree from the zenith, which is considered to be the effective antenna beam width. Then the wind field at 80 km altitude is averaged for 28km ($= 80 \times 2 \sin 10^\circ$), resulting in waves with horizontal wavelength smaller than 56 km (2×28) to be filtered out. Next we assume that the gravity wave spectra with respect to horizontal wave number k is proportional to $k^{-5/3}$; and further assume that the gravity waves with period of 10–100 minutes have the horizontal wavelength between 10–1000 km according to the observational results shown by Manson and Meek [1988] and Reid [1986].

Then, we can roughly estimate the loss of the variance η by horizontal averaging as

$$\eta = 1 - \frac{\int_{1/(1000km)}^{1/(56km)} k^{-5/3} dk}{\int_{1/(1000km)}^{1/(10km)} k^{-5/3} dk} = 0.106 \simeq 11\% \quad (6.8)$$

Thus, the variances observed at Saskatoon MF radar shown in this section are estimated to be reduced by 11 % of the real variances. On the contrary, no loss is expected for the MU radar data, because the width of the illuminating area of the MU radar antenna is about 5 km at 80 km altitude and smaller than the shortest scale of the gravity wave assumed here. Note that in our analysis of the MU radar data only the homogeneousness of the statistical characteristics of the wind fluctuation between the different antenna beams is assumed and the homogeneousness of the wind field is not necessary. Thus, the spatial effect on the obtained wind variance is limited to that of the averaging the wind field in a single pencil beam.

Consequently, the variances of the wind fluctuations of the periods 10–100 minutes at the MU radar are larger than those at the Saskatoon by the factor of at least about 1.3–1.4. The Figures 6.9–6.11 show the general similarity of the gravity wave intensities at the different locations of 35°N and 52°N, consistent with the comparison as reported by Vincent [1984] and Meek et al. [1985]; however, a consistently larger variance at the latitude 35°N than at 52°N has been revealed by a closer comparison in the present study. It is important to know the global distribution of gravity wave activity in order to clarify the wave sources and various interactions caused by the propagation of gravity waves. Especially, the latitudinal comparisons of observations in and above the mesosphere are very important because gravity waves are thought to interact strongly with mean winds of mesosphere and decelerate them. of that altitude so far. Further comparison among the radar observation network (e.g. MLTCS network) are desirable to reveal the latitudinal structure of gravity wave activity in the mesosphere and lower thermosphere.

6.2.3 Momentum Flux

We here compare the momentum flux by the MU radar and Saskatoon MF radar. Figures. 6.12 and 6.13 shows the zonal and meridional momentum flux $\overline{u'w'}$, $\overline{v'w'}$ for the waves with periods of 10–100 min. We first discuss the difference between the momentum flux deduced by the difference filter (open circles) and the integration of the spectrum (closed circle) with Saskatoon MF radar.

Note that the spectrum taken here is a cross spectrum between the vertical wind w' and horizontal wind u' or v' , and the real part of the spectrum is integrated for the frequency corresponding to the periods of 100 min to 10 min. The calculation of the cross spectrum was done by exactly same way as that used for auto spectra [Tsuda et al., 1990a] except that the CCF (cross correlation function) was taken instead of ACF in auto spectra. In this case, ratios between the momentum flux by means of the two method, such as we discussed for the wind variance, cannot be compared because the momentum flux is a value with sign. Hence the simple ratio of the two values and its average over a year does not have a meaning. Therefore, we will discuss the difference of the values for the two methods.

The zonal momentum flux by the difference filter and the cross spectra, shown in Figure 6.12 agreed with the difference of less than $0.5 \text{ } 0.3 \text{ m}^2/\text{s}^2$, while in June and October the difference become as large as $0.7\text{--}0.8 \text{ m}^2/\text{s}^2$. The annual variation, however, is clear for both methods showing that the eastward maximum (westward minimum) in summer and westward maximum in winter.

The meridional momentum flux shown in Figure. 6.13 shows good agreement

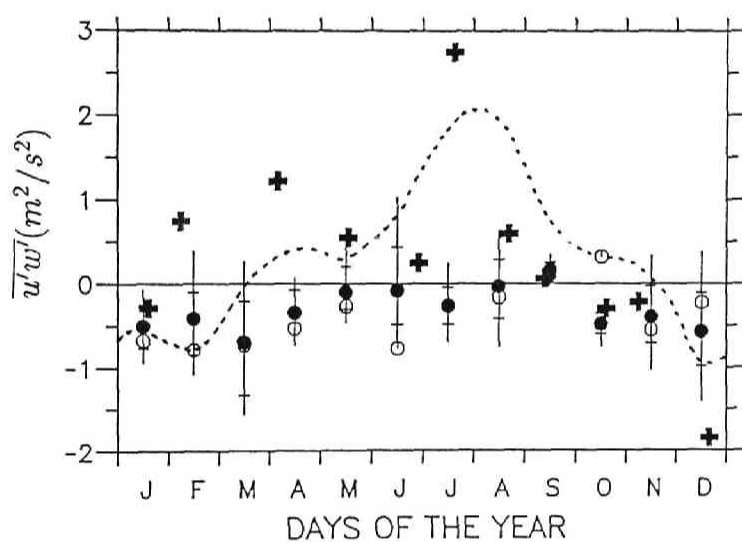
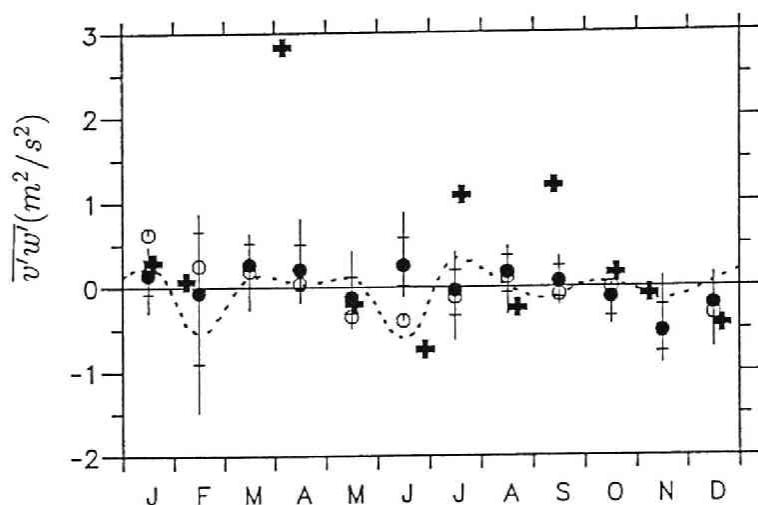


Figure 6.12: Monthly mean upward flux of zonal momentum $\overline{u'w'}$ for the MU radar (closed crosses) and Saskatoon MF radar (circles). Open and closed circles correspond to the flux calculated by DF and PSD methods, respectively. The error bars and vertical lines indicate the representativity of the 5 days mean value (see text). The dotted line shows the momentum flux observed with the MU radar in 1985-1988 by Tsuda et al. [1990a].

Figure 6.13: The same as Figure. 6.9 except for $\overline{v'w'}$.

between the two methods with the difference usually smaller than $0.2 \text{ m}^2/\text{s}^2$, except that in January and June the difference became as large as $0.5 \text{ m}^2/\text{s}^2$. The seasonal variation of the meridional momentum was not clear, but rather fluctuating around zero with smaller values than the zonal components.

Because the momentum flux $\overline{u'w'}$ and $\overline{v'w'}$ were calculated from the time series of the each wind component u' and w' or v' and w' , covariance $\overline{u'w'}$ and $\overline{v'w'}$ were largely affected by the filter characteristics of the measured wind velocities. Furthermore, the determination of u' or v' is done by SA method, while w' is determined by the complex correlation, that is a kind of Doppler measurement, with the Saskatoon MF radar. Both determination can be contaminated the other wind component as suggested by frequency spectra in 6.2.1. So that there is possibility that those contaminations produce some error or bias in determining the momentum flux.

Despite such difficulties, the zonal and meridional momentum fluxes shown in Figures 6.12 and 6.13 coincide well with the two determination by the difference filter and the cross spectral method. Thus, we compare these momentum fluxes with the results of the MU radar, although the uncertainty of the effect of the wind contaminations exists.

The momentum flux of the MU radar was determined by the difference of the variance of a pair of oblique beams [Vincent and Reid, 1983]. Since the

antenna beam width is small (3.7° half power width in one way), the effect of the wind contaminations such as discussed above are expected to be small. The zonal momentum flux of the MU radar in Figure. 6.12 shows eastward flux from February to August, and westward flux from October to January, and agreed well with the observations in 1985–1988 reported by Tsuda et al. [1990a], although that of February in 1989 was shown to have a westward phase. The summer and the winter peaks were found in July and December, with the values of $\sim +3\text{m}^2/\text{s}^2$ and $\sim -2\text{m}^2/\text{s}^2$, respectively. The typical eastward and westward fluxes are about $+0.5\text{--}1.0\text{m}^2/\text{s}^2$ for the eastward flux in February–August and about $-0.3\text{m}^2/\text{s}^2$ in October to January. These amplitudes of the momentum fluxes were similar to those of Tsuda et al. [1990a], although Tsuda et al. showed the momentum flux with the periods of 2 hours to 5 minutes which were slightly wider than our range. The meridional momentum flux by the MU radar did not show a clear seasonal variation. It sometimes becomes large for instance in April, while usually within $\pm 0.5\text{m}^2/\text{s}^2$ from zero, which again similar to the result in 1985–1988 by Tsuda et al. [1990a].

The seasonal variations of the zonal momentum flux in the MU radar and the Saskatoon MF radar show that momentum flux was eastward in summer and westward in winter, with the similar order of the amplitude between the two locations, which is the confirmation of Tsuda et al. [1990a] and Meek and Manson [1989], and support the theories of the gravity wave acceleration of the mean wind [e.g. Lindzen, 1981]. The amplitude of zonal momentum flux at Saskatoon seems to be smaller than that at Shigaraki especially in summer, while detailed comparison needs the discussions of the effect of the analysis at Saskatoon with the closer analysis.

6.3 Hemispheric Comparison between Shigaraki and Adelaide

In this section we compare the gravity waves observed by the simultaneous observation between the MU radar at Shigaraki (35°N) and Adelaide MF radar (35°S) carried out in June 1987 and January/February 1991. These observation periods are summarized in Table 6.3. The observations carried out with the MU radar were out of four campaign observations presented in Chapter 5. Corresponding observations with the Adelaide MF radar were compared with the MU radar data.

The Adelaide MF radar was operated continuously, and recording the mesospheric and lower thermospheric wind by means of spaced antenna technique in every 2 minutes. Besides that, Doppler wind measurements were carried out by

MU radar		
June 1987		
6-20	DBS, 60-90 km, 60 sec resolution	
21-29	150 sec resolution	
Jan/Feb 1991		
21-25 Jan	60-90 km, 210 sec resolution	
28 Jan-1 Feb	130 sec resolution	
4-8 Feb	185 sec resolution	
Adelaide MF radar		
June 1987		
5-14	SAD+Doppler, 70-90 km, 2 min resolution	
22-30	SAD+Doppler, 60-100 km, 2-4 min resolution	
other days	SAD, 60-100 km, hourly mean wind	
Jan/Feb 1991		
21 Jan -14 Feb	SAD 60-100 km, 2-4 min resolution	
28 Jan-14 Feb	+ Doppler, 2-4 min resolution	
14 Jan-20 Jan	SAD, 60-100km, hourly mean wind	

Table 6.3: Observation duration of June 1987 and January/February 1991 at the MU radar and Adelaide MF radar.

the two pencil beam of receiving antenna and time series of received data were recorded for the periods of 5–14 and 22–30 in June 1987. Furthermore, the data recording system was updated in January 1991 and the received signal of all the receiving channels were recorded continuously, from January 1991. Thus, the Doppler wind became available from January 28, 1991. These data bases provide the sample of the summer and winter data at the latitude of 35° in the northern and southern hemispheres, which enable us to discuss the symmetry of the wind fluctuation with periods between several minutes and several hours. In this section, frequency spectrum of the wind fluctuations and the propagation direction of the gravity waves at two locations are compared.

6.3.1 Frequency Spectra

Figure 6.14 shows frequency spectra of northward (left) and eastward (right) wind velocity components observed with the MU radar and Adelaide MF radar in June 1987 and January/February 1991. The frequency spectra of the wind velocity determined by spaced antenna technique were plotted for the Adelaide data, while the frequency spectra of the horizontal wind velocity deduced by the vertical and two oblique beams, as discussed in section 6.1. Besides the spectra of horizontal wind velocity, the horizontal projections of the average spectra observed by two oblique beams, for instance northward and southward beam, are also plotted for the MU radar observations, as well.

The frequency spectra of the horizontal wind observed with the MU radar had the spectral slope close to $\omega^{-5/3}$ in the frequency range from 10^{-4} s^{-1} to $2 \times 10^{-3} \text{ s}^{-1}$ for northward wind component and from 10^{-4} s^{-1} to $5 \times 10^{-4} \text{ s}^{-1}$ for eastward wind component, respectively. The projection of the radial wind spectra into the horizontal plane for the northward wind component showed the enhancement due to the vertical wind component in the frequency range higher than $4 \times 10^{-4} \text{ s}^{-1}$ in both June 1987 and January/February 1991, as shown in Chapter 5.

On the other hand, the spectra of the northward wind observed with Adelaide MF radar showed the spectral slope of about $-5/3$ from $1 \times 10^{-4} \text{ s}^{-1}$ to $3 \times 10^{-4} \text{ s}^{-1}$ and in the frequency range larger than those frequencies were significantly enhanced up to the highest frequency of $2 \times 10^{-3} \text{ s}^{-1}$. The spectral shape observed with Adelaide MF radar is almost identical to the horizontal projection of the spectra of radial winds of the MU radar, but different from that of the horizontal wind spectra by the MU radar.

The eastward component again shows the similarity between the frequency spectra of the wind velocity observed by Adelaide MF radar and the horizontal projection of the radial wind by the MU radar, both of which showed the en-

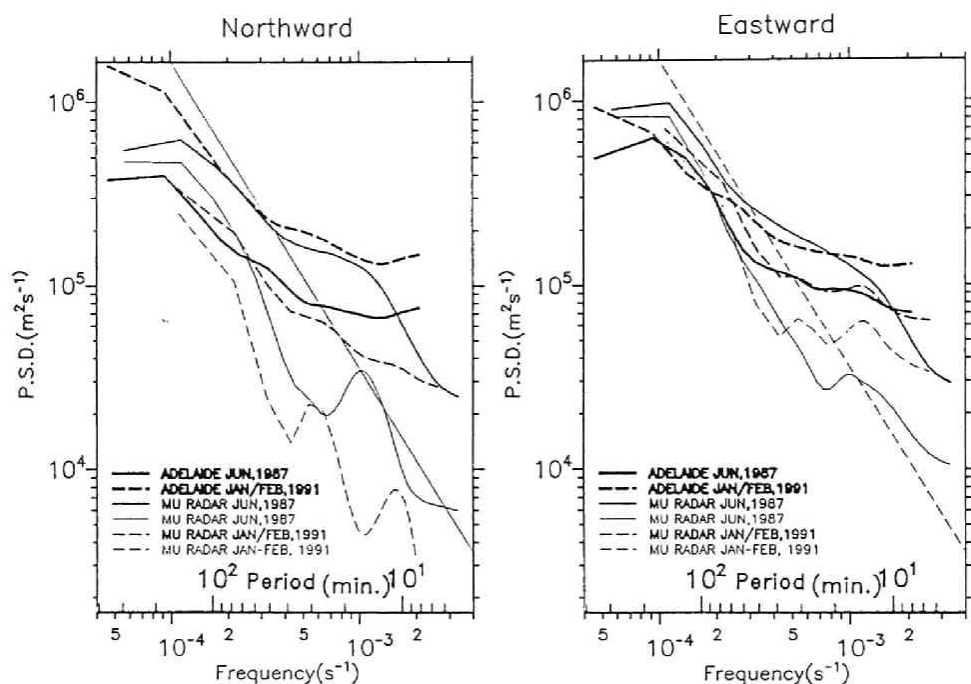


Figure 6.14: Frequency spectra of the northward (left) and eastward (right) wind components observed in June 1987 (solid lines) and in January/February 1991 (dashed lines), respectively. Thick lines are for Adelaide MF radar. Middle and thin lines are the projection of the horizontal plane of the radial wind velocity and horizontal wind component observed with MU radar, respectively. Straight lines indicate the spectral slope of $\omega^{-5/3}$. (see text)

hancement from the spectra with $\omega^{-5/3}$ shape at the frequency range higher than $2-3 \times 10^{-4} \text{s}^{-1}$, while the horizontal wind spectra by the MU radar had the spectral slope of $-5/3$ up to $4 \times 10^{-4} \text{s}^{-1}$ in January/February 1991 and $7 \times 10^{-4} \text{s}^{-1}$ in June 1987, respectively.

The enhancement of the frequency spectra of the horizontal wind velocity again can be considered to be the contamination of the horizontal spatial difference of the vertical wind in the radar sampling volume [Royrvik, 1983]. The antenna beam pattern for transmission at the Adelaide MF radar was about 40 degree in half width, and so was the pattern for reception. However, the effective antenna beam width is limited by the aspect sensitivity of the scatterer. Lesicar and Hocking [1991] recently made an estimation of the effective pattern as 10° in half width, and so the horizontal difference of the vertical wind in this sampling area can contaminate the horizontal wind determined by the spaced antenna method.

In Figure 6.14 the frequency spectra of the horizontal wind observed at Adelaide included the vertical wind component as well as the frequency spectra of the radial wind velocities obtained by the MU radar, although the magnitude of contamination in Adelaide data is now not well resolved. However, the similar shape of the frequency spectra between the Adelaide and the MU radar observations in the frequency range of 10^{-4} to 10^{-5}s^{-1} suggests the contribution of the vertical wind are similar at least in these frequency range. Therefore, we compare the frequency spectra obtained by the Adelaide MF radar and the horizontal projection of the radial wind velocity obtained by oblique beams with the MU radar. Note that the frequency spectra of the wind velocity determined by Saskatoon MF radar shown in the previous section also showed the enhancement from the spectral slope of $-5/3$ in the frequency range larger than $5-8 \times 10^{-4} \text{s}^{-1}$, which suggests the contribution of the vertical wind velocities. However, the enhancement was limited to high frequency component probably because of the difference of the characteristics of the scatterer.

Next, we compare the spectral density of the wind velocity more closely. The spectral density of the northward wind observed by the MU radar were $6 \times 10^5 \text{m}^2 \text{s}^{-1}$ at 10^{-4}s^{-1} and $1.5 \times 10^5 \text{m}^2 \text{s}^{-1}$ at 10^{-3}s^{-1} in June 1987, and $4 \times 10^5 \text{m}^2 \text{s}^{-1}$ at 10^{-4}s^{-1} and $4 \times 10^4 \text{m}^2 \text{s}^{-1}$ at 10^{-3}s^{-1} in January/February 1991, showing the larger values in summer than in winter as presented in Chapter 5. On the other hand, the frequency spectra of northward wind velocity at Adelaide were $4 \times 10^5 \text{m}^2 \text{s}^{-1}$ at 10^{-4}s^{-1} and $6 \times 10^4 \text{m}^2 \text{s}^{-1}$ at 10^{-3}s^{-1} in June 1987 and $10^6 \text{m}^2 \text{s}^{-1}$ at 10^{-4}s^{-1} and $1.5 \times 10^5 \text{m}^2 \text{s}^{-1}$ at 10^{-3}s^{-1} in January/February 1991, indicating the larger values in summer again than in winter. The spectral densities in June 1987 by the MU observation and that in January/February 1991 by the Adelaide MF radar agreed well within the discrepancy of 25 % at each frequency in the

		$\overline{v'^2}$	$\overline{u'^2}$	$\overline{v'^2 + u'^2}$
Summer	MU June 1987	105.2	121.6	226.8
	Adelaide Jan/Feb 1991	115.0	94.9	209.9
Winter	MU Jan/Feb 1991	51.4	88.5	139.9
	Adelaide Jun 1987	52.1	72.9	125.0

Table 6.4: Comparison of the wind variance of the waves with periods between 120 and 30 min.

range of $1.3 \times 10^{-4}\text{s}^{-1}$ to $1.3 \times 10^{-3}\text{s}^{-1}$. The spectral density in June 1987 by Adelaide MF radar and that in January/February 1991 also agreed within 35 % in the frequency range of 10^{-4}s^{-1} to $7 \times 10^{-4}\text{s}^{-1}$. Similar coincidence can be found in the eastward wind component within 50% discrepancy in the frequency range of $1.3 \times 10^{-4}\text{s}^{-1}$ to $1.3 \times 10^{-3}\text{s}^{-1}$. Thus, the frequency spectra observed by the MU radar and Adelaide MF radar were similar if those are compared between in the winter or summer of each hemisphere. The wind variance at both locations were estimated by integrating the frequency spectra in the the frequency range of $1.4 \times 10^{-4}\text{s}^{-1}$ to $5.6 \times 10^{-4}\text{s}^{-1}$ (120 min–30 min in period). The results are shown in Table 6.4.

The variance of the northward wind component of the MU radar and Adelaide MF radar observations were 105.2 and $115.0 \text{ m}^2\text{s}^{-2}$ in summer and 51.4 and $52.1 \text{ m}^2\text{s}^{-2}$ in winter, indicating the variance by Adelaide MF radar was slightly larger. The variance of the eastward component in the MU and Adelaide MF radars were 121.6 and $94.9 \text{ m}^2\text{s}^{-2}$ in summer, and 88.5 and $72.9 \text{ m}^2\text{s}^{-2}$ in winter, showing the larger variance for the MU radar. The difference of the variance between eastward and northward components is considered to be because the propagation direction of gravity waves are not horizontally isotropic, which is discussed more closely in the following subsection. The total variance ($\overline{v'^2} + \overline{u'^2}$) of the wind fluctuation with periods of 120–30 min in the MU and Adelaide MF radars were 226.8 and $209.9 \text{ m}^2\text{s}^{-2}$ in summer and 139.9 and $125.0 \text{ m}^2\text{s}^{-2}$ in winter, respectively. Thus, the variance observed with the MU radar was about 10 % larger than that with Adelaide MF radar, for both summer and winter. Because the Adelaide MF radar was sampling the vertical width of about 3.8 km and horizontal width of about 28 km for the beam width of 10° at 80 km altitude. Thus, the variance of the Adelaide can be estimated smaller by more

than 11 % by the same estimation in the present section as the Saskatoon MF radar. So that we can conclude that the variance of the wind fluctuation with periods of 120–30 min is almost the same as that observed by the MU radar at Shigaraki.

The mesospheric tidal wind has been observed and compared between Shigaraki (35°N) and Adelaide (35°S) [Aso and Vincent 1982; Vincent et al., 1988, 1989], and the amplitude of diurnal tides were found to be much larger in southern hemisphere than in the northern hemisphere at the latitude of 35°. Vincent et al. [1988] suggested this difference is due to the hemispheric difference in forcing, interactions between the tide and background winds or the difference of the dissipation caused by the turbulence due to gravity waves. However, we have shown that the gravity wave in the northern hemisphere at 35° latitude was almost the same amplitude as that observed in southern hemisphere at the same latitude. Thus, dissipations by the turbulence caused by the gravity waves are considered to be similar magnitude between the northern and southern hemispheres if our comparison shows the typical values of the conjugate latitude. This showed a different result from what is expected by the tidal observations shown in Chapter 4, that is the large dissipation in the northern hemispheres.

It is, however, noteworthy that there is also the possibility that our observation of gravity waves are just one sample of the latitude, and there can be large longitudinal variability in eddy dissipation. In that case, eddy viscosity which affects to the tidal amplitudes could be different from the viscosity at one longitude, and hence the large hemispheric difference of eddy viscosity might occur. So far we do not have any observations with the longitudinal variability of gravity waves and so those observations in future are desired.

6.3.2 Propagation Direction of Gravity Waves

Here we compare the propagation direction of the gravity waves observed at Shigaraki and Adelaide. In Chapter 5 we examined the propagation directions of the waves with periods of about 5–15 hours by means of hodograph analysis and those of the waves with periods from 8–10 hours to 5 minutes by means of momentum flux analysis. We here apply the two analysis to get the information on propagating directions for Adelaide observation. One is the hodograph analysis of the vertical wind profile of bi-hourly wind, which extracts the waves with long period. The other is the Stokes parameter analysis which has been applied to the Adelaide MF radar data to determine the propagation directions [Fritts and Vincent, 1987]. We applied these analyses to determine the gravity waves with periods of 6 hours–30 min.

Figure 6.15 shows the histogram of the propagation direction determined by

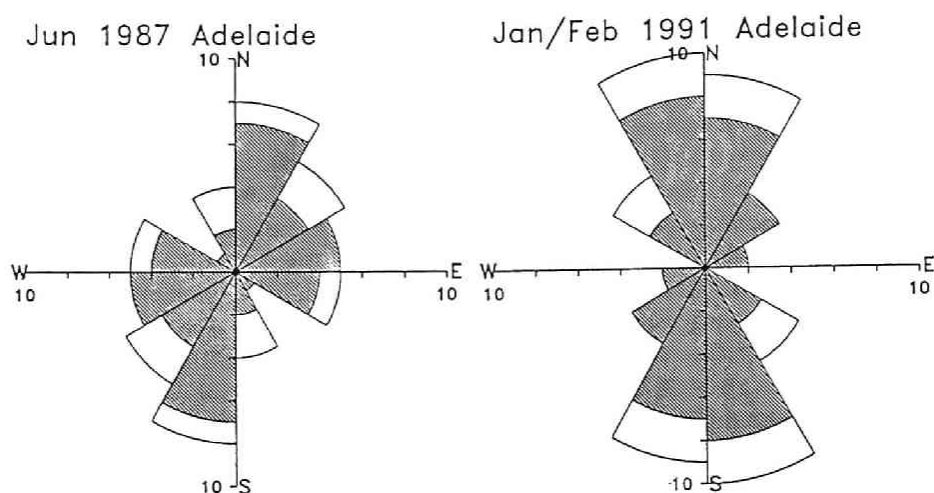


Figure 6.15: Histogram of the propagation direction of the gravity waves by the hodograph analysis of the vertical profiles of the bi-hourly wind. Shaded area shows the upward energy transporting waves. A sinusoidal component with periods of 24 or 12 hour are removed before the fitting of the ellipse.

	June 1987	Jan/Feb 1991
GW event	29	30
upward energy propagation	22	24
downward energy propagation	7	6
mean vertical wavelength	13.3km	13.2km
S.D.	3.2km	2.0km
mean intrinsic period	9.4 hrs	9.0 hrs
S.D.	4.2 hrs	3.0 hrs

Table 6.5: Extracted parameter of the hodograph analysis of the vertical profile of wind velocity observed with Adelaide MF radar.

the hodograph analysis in June 1987 and January/February 1991. The analysis is the same as that applied to the MU radar data described in Chapter 5, except that the fitting of an ellipse was applied after removing the tidal component of 12 and 24 hours period from the hourly mean wind at each height and filtering by a band pass filter with pass band of 8–20 km in vertical profiles. As a result, 29 and 30 wave events were extracted from the data in June 1987 and January/February 1991, respectively. The statistical parameters for the fitted wave are shown in Table 6.5. Note that the estimated parameters are vertical wavelength, intrinsic period and propagation direction, since the other parameters derived using the former two parameters are worse in accuracy because of the low height resolution of Adelaide MF radar than that of the MU radar.

The mean vertical wavelength and intrinsic period were 13.3 km and 9.4 hours in June 1987 and 13.7 km and 9.0 hours in January/February 1991, respectively. These vertical wave length were larger than that of the MU radar observation as shown in Chapter 5, which were 7.6 and 6.1 km in June 1987 and January/February 1991, respectively. This difference is because of the observational height range and height resolution between the two radar observations. The intrinsic periods observed by Adelaide MF radar did not show large difference from those by the MU radar, which were 9.2 hours and 11.3 hours in June 1987 and January/February 1991. The gravity waves with upward energy propagation were 22 out of 29 profiles and 24 out of 30 profiles in June 1987 and January/February 1991, respectively, indicating the dominance of the upward energy propagating waves.

The propagation directions determined here had the ambiguity of 180° because in the current analysis the vertical wind components were not deduced. In June 1987 the propagation directions were centered from north to east or south to west directions, with the center of the distribution of 32° east of north or west of south. On the other hand, the propagation directions in January/February 1991 was distributed around north-south directions with the central value of 3° east of north or west of south. Thus, the propagation directions in summer were aligned north-south direction and in winter they were shifted about 30 degree east of north, although ambiguity of 180° exists.

Eckermann and Vincent [1989] applied the Stokes parameter analysis to the falling sphere observation carried out near Adelaide, and found that the propagation of the gravity waves with vertical wavelength of 2–10 km at 30–60 km altitude were centered in the direction of south of east (or north of west) in summer and east of north in winter. The results of the hodograph analysis here and those of Eckermann and Vincent agree well in winter, showing the north-south direction and the direction close to that, while in summer they differs from each other, showing north-south for radar observation and south of eastward direc-

		$\overline{v'^2}$	$\overline{u'^2}$	$\overline{v'u'}$	Azimuth
Summer	28 Jan–14 Feb 1991	210.1	149.5	-6.4	-6° (+180°)
Winter	5–14 Jun 1987	101.2	144.7	7.0	81° (+180°) *)
	22–30 Jun 1987	59.9	88.8	0.2	90° (+180°) *)

Table 6.6: Propagation direction estimated using Stokes parameter.

*) wind velocity by Doppler beams shows eastward momentum flux

tion for the falling sphere observation. These results suggest that in summer, the gravity waves with the eastward or westward directions below 60 km strongly interact with the mean wind and other atmospheric waves during the upward propagation.

The other determination of propagation direction by Stokes parameter analysis [Fritts and Vincent, 1987] was applied to the wind variance for the waves with periods of 6 hours to 30 minutes. The results are summarized in Table 6.6. In January/February 1991 the propagation direction was aligned to north-south direction, while in June 1987, the direction became east-westward direction. The wind fluctuations by the Doppler velocity observed with eastward and westward beams showed the eastward momentum flux in June 1987 for the corresponding frequency component, but not determined in January/February 1991 because of low S/N from the height below 80 km.

Figure 6.16 shows the summary of the propagation direction of the gravity waves observed in June 1987 and January/February 1991 at the MU radar and Adelaide MF radar, together with the earlier works for the Adelaide data by Fritts and Vincent [1987] and Eckermann and Vincent [1989]. It is found that in both the MU radar and Adelaide MF radar, the directions of the hodograph analysis and the gravity waves with periods of 6–8 hours to 30 minutes were similar in summer, while different in winter, although the difference in winter was almost 180 degree for the MU radar and 50 (or + 180) degree for Adelaide. Since the direction determined for Adelaide had the ambiguity of 180 degree care must be paid for the comparison. However, the concentration of the direction of gravity wave propagation of different frequency component seems to be occur for both hemispheres.

When the directions by Fritts and Vincent [1987] for the MF radar observation at 80–90 km and by Eckermann and Vincent [1989] for the falling sphere observation at 30–60 km are included together, it is recognized that the propagation direction at Adelaide was generally biased to north-south directions, except

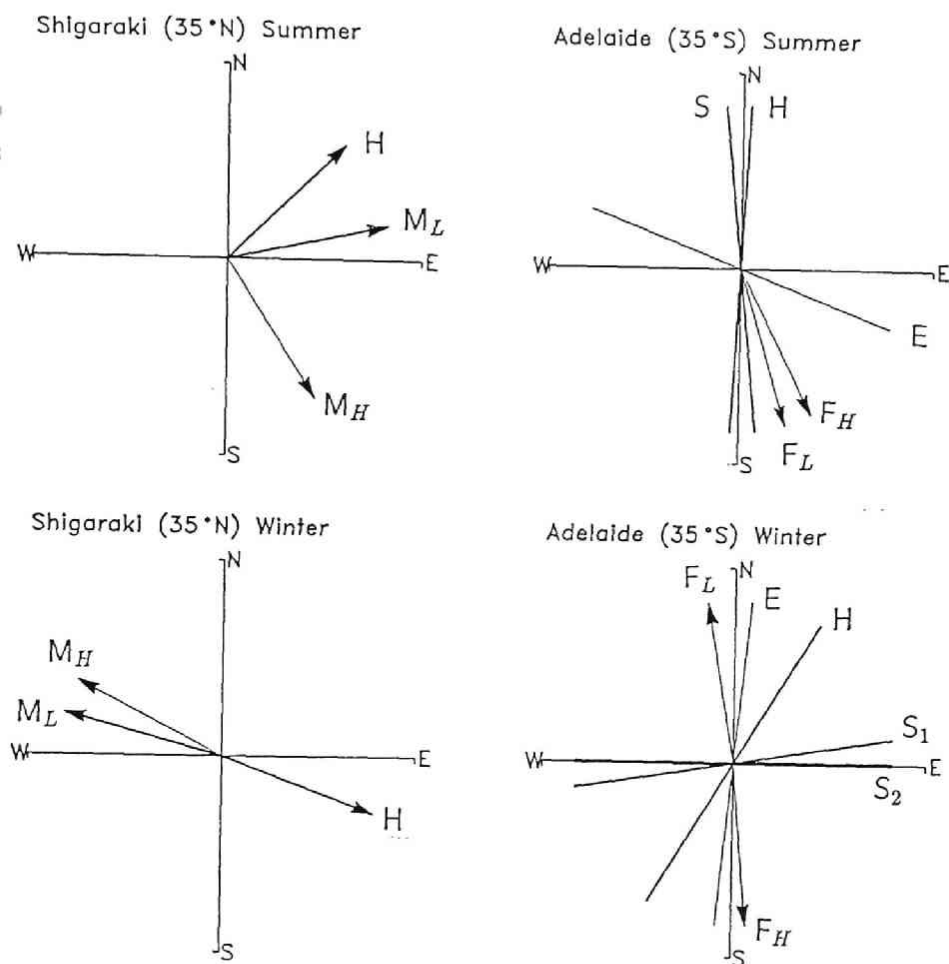


Figure 6.16: The summary of the propagation direction observed in June 1987 and January/February 1991 with the MU radar and Adelaide MF radar, together with the results of Fritts and Vincent [1987] of the Adelaide MF radar, and Eckermann and Vincent [1989] of the falling sphere.

H: Hodograph analysis (at 70–75 km and 80–88 km in Shigaraki and Adelaide, respectively)

M_L, M_H : momentum flux with periods of 8 h – 30 min and 30 min – 5 min at 68–80 km,

S, S_1, S_2 : Stokes parameter analysis at 70–80 km (1: June 5–14, 1987, 2: June 22–30, 1987),

F_L, F_H : 24–8 h and 8–1 h component by Fritts and Vincent [1987] at 85 km

E: falling sphere by Eckermann and Vincent [1989] at 30–60 km.

for the 6 hours–30 min component in winter and falling sphere observation in summer. On the other hand, the propagation directions of gravity waves observed by the MU radar show eastward–westward propagation for almost all the component except for the waves with periods less than 30 minutes in June 1987.

According to the height time cross section of the ratio $\overline{v'^2}/\overline{u'^2}$ at Adelaide MF radar [Fritts and Vincent, 1987], the variance $\overline{v'^2}$ of the waves of the periods of 1–8 hours were usually larger than $\overline{u'^2}$, while sometimes smaller below 80 km. This means that above 80 km the propagation direction of the waves are more aligned to north–south direction while below that height it sometimes becomes eastward–westward direction. The eastward–westward bias observed in summer for the falling sphere data and in winter for the MF radar Stokes parameter of the waves with the periods of 6 hours to 30 minutes can be due to the low observation altitudes.

The tendency of the eastward–westward direction of the propagation observed by the MU radar than that by the Adelaide MF radar data also be partly caused by the low height range for the MU radar observations as noted in the captions of Figure 6.16, because the eastward–westward propagation is more likely to be observed in the altitude below 80 km. However, the comparison at 68–70 km to 80 km altitude between the results of Stokes parameters observed at Adelaide and the momentum flux analysis done for the MU radar result also showed the preference of the eastward–westward propagation at Shigaraki and northward–southward at Adelaide. Same tendency can also be recognized in the difference of the spectral density between northward and eastward wind component in figure 6.14, that is the eastward wind component at Shigaraki was larger than the northward component in both summer and winter, while the Adelaide data shows that the northward component was larger in summer.

As a summary, the gravity waves observed at Shigaraki with the MU radar tend to propagate to eastward and westward directions while the northward and southward directions are preferred at Adelaide. As we stated in the last Chapter, the propagation direction of the gravity wave reflects the source of the waves or the interactions with the winds. However, it is rather difficult to attribute this hemispheric difference to the difference of the mean wind because the structure of the mean wind is not so different as to make the difference of the preference of zonal and meridional direction of the gravity waves. One possible reason for the difference of propagation direction of gravity waves between hemispheres observed here is the difference of the land–sea distribution around the radar stations, that is Shigaraki and Adelaide is located in the east and south of the continent, respectively. If the gravity wave source on or above the continent is larger than that in the ocean, the present result can be explained.

6.4 Concluding Remarks

In this chapter, we have compared the gravity wave intensities with periods of 10–100 minutes in the mesosphere at 67–80 km altitude between the MU radar, Shigaraki (35°N), and the MF radar at Saskatoon (52°N) in 1989. The duration of the analysis and the height resolution are adjusted to the daytime and 3 km, respectively, in order to minimize the difference caused by the difference of these parameters. The radial winds velocities obtained by the MU radar are used to calculate the frequency spectra of the horizontal wind fluctuation u' and v' and vertical wind w' . The spaced antenna method is used to provide horizontal winds from the MF radar, and a complex correlation method is used to provide vertical velocities.

The characteristics of the difference filter [Meek et al., 1985b] is shown and the filter pass band is found to be changed according to the spectral slope of the frequency spectra, so that the variance obtained by the difference filter can be affected by the spectral slope. However, comparison of the monthly mean variance by the difference filter and the integration of the spectra showed that the ratio between the two analytical methods is almost constant throughout a year, and that the difference filter can therefore be used to investigate the seasonal variation but with some adjustment to provide the absolute value.

The frequency spectra of the horizontal wind fluctuations u' and v' at both locations had the slope of $-5/3$, while the frequency component larger than 20–30 day⁻¹ (50–70 min in period) were significantly enhanced and consequently the spectral slope there became gradual in March to August. The spectra of vertical wind w' at Saskatoon showed that the spectral slope was $-2/3$ to -1 in November to February and $-1/3$ to $-1/2$ in March to October. The spectra observed by the MU radar also showed a less steep spectral slope than that by Saskatoon MF radar, and in November and December spectral slope was $+0.15$ – $+0.3$ in the frequency lower than 70 day⁻¹. Thus, the variation of the spectral shape between the two radars generally agreed well for u' and v' component, but disagreed for w' component.

The variance $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ showed semiannual variation with summer maximum, winter secondary peak and minima in September and April at both Saskatoon and Shigaraki, although the variance in summer was more conspicuous at Shigaraki than in Saskatoon. The variability of the 5 day mean variances of $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ were larger in winter than in summer, suggesting that the uncertainty in determining the gravity wave intensity by the 4–5 days observation a month was caused by this variability during a month rather than the month to month or year to year variation.

The annual averages of the ratio of the monthly mean variance between

the MU and Saskatoon MF radars were 1.6, 1.5 and 1.6 for $\overline{u^2}$, $\overline{v^2}$ and $\overline{w^2}$, respectively. The estimation including the difference of the spatial average due to the beam width of the antenna suggests that the variance of gravity waves at Shigaraki (35°N) is at least 30–40% larger than that at Saskatoon (52°N).

The zonal momentum flux showed an annual variation with the eastward peak in summer and westward peak in winter at both locations, although the momentum flux in summer at Saskatoon did not become completely eastward but rather close to zero, which might be biased by the contaminations of wind determination of horizontal and vertical winds. Nevertheless, the seasonal variation of zonal momentum flux at both locations were similar with transitions between summer and winter around April and October. The meridional momentum flux did not show clear seasonal variation, but the amplitudes were small for both locations.

The frequency spectra of the wind velocities and the propagation direction of the gravity waves are compared between the observations of Adelaide MF radar (35°S) and the MU radar (35°N) for the two campaign durations of June 1987 and January/February 1991, which were in summer and winter for both hemisphere. The frequency spectra of the wind velocities determined by the spaced antenna with Adelaide MF radar showed a similar enhancement of the spectra as the oblique spectra obtained with the MU radar at the frequency higher than $3 \times 10^{-4} \text{s}^{-1}$ suggesting the contamination of the vertical wind difference in the radar sampling volume.

Both the data of Adelaide MF and the MU radars shows larger spectral density in summer than in winter, while the spectral density of both local summers and winters between Adelaide and the MU radar were almost identical. The variances $\overline{u'^2 + v'^2}$ of the waves with periods of 2 hours – 30 minutes observed with the MU radar were about 10 % larger than that with the Adelaide MF radar, suggesting the gravity wave intensity at the 35° latitude in the northern hemisphere is almost the same as that at the same latitude in the southern hemisphere, considering the spatial resolution of the MF radar observation. This result further suggests that the dissipation due to the turbulence caused by the gravity wave are not much larger in northern hemisphere.

The propagation direction of the gravity waves at around 80–85km were determined by the hodograph analysis of the vertical wind profiles observed by the Adelaide MF radar for the waves with vertical wavelength and intrinsic periods of about 13km and 9 hours, respectively. They showed the alignment to north-south direction in January/February 1991 and about 30 degree east of north in June 1987. These directions agreed well in winter and differed in summer with the results of the falling sphere observation at 30–60 km, which shows north-south and east-west alignment in winter and summer, respectively.

The Stokes parameter analysis determined the propagation direction of the waves with periods between 6 hour and 30 minutes, which were aligned to west of north or east of south direction in January/February 1991 and east-west direction in June 1987, respectively.

These results are compared with the propagation directions of the gravity waves observed with the MU radar together with the earlier observation at Adelaide [Fritts and Vincent, 1987]. It is recognized that in summer propagation direction of the gravity waves with different frequency component tends to concentrate while spread in winter for both hemisphere. The results of the MU radar showed the directions with east-west alignment more than the results of Adelaide. This may represent the hemispheric difference of the propagation direction of gravity waves. One possibility for the origin of this difference is the difference of the topography between Shigaraki and Adelaide.

Chapter 7

Summary and Conclusions

This thesis is concerned with radar observation of the dynamics of the mesosphere and lower thermosphere. Two types of radars at Shigaraki (35°N); the MU radar and Kyoto meteor radar, were utilized for the observations. And their results were compared with two MF radars, Saskatoon MF radar (52°N) and Adelaide MF radar (35°S), in order to investigate the latitudinal and hemispheric difference.

The mean winds were deduced from Kyoto meteor radar observation (80–110 km) in 1983–1986 and the MU radar observation (60–90 km) in 1985–1988. The results showed that the summer eastward wind enhanced at around 95 km with peak value more than 30 m/s in July. In January and February eastward wind again enhanced below 95 km. Mean meridional wind was generally southward and enhanced up to 10 m/s in summer. The westward wind in summer below 80 km reached maximum in June at 65–75 km with the amplitude of about 40 m/s. In winter, the peak of the eastward wind was in the end of November at 70 km with the amplitude of 60 m/s. The eastward wind in winter showed large variability suggesting the effect of stratospheric warmings. The reversal height of the summer westward wind is then determined as 83 km altitude, while it seems to have interannual variation.

These mean wind profiles at Shigaraki were compared with the mean wind obtained with Adelaide MF radar in 1984–1986. The reversal height of zonal mean wind in summer does not show a significant interannual variation but rather constant at about 85 km at Adelaide, while it showed large interannual variation at Shigaraki. The mean wind characteristics were basically similar between the hemispheres except for the winter variability observed in northern hemisphere, and summer reversal height.

Three cases are reported that the stratospheric sudden warming was accompanied with the change of the eastward wind into westward at 70–90 km at

Shigaraki, suggesting close relation between sudden warmings and winter variability of the mean wind at Shigaraki.

The atmospheric tides and two day waves are studied using Kyoto meteor radar and Adelaide and Saskatoon MF radars. The results shown by Tsuda et al. [1988a,b] and Vincent et al. [1988, 1989] were reviewed and further analysis using these radars were presented.

Amplitudes of semidiurnal tides were generally larger at Saskatoon than at Shigaraki, and so were the amplitudes at Adelaide. At three locations, the seasonal variation was mainly composed of two fundamental states of summer-like and winter-like months which were characterized by long and relatively short vertical wavelengths, respectively. Seasonal variations of phases at 90 km agreed fairly well for both zonal and meridional components between Shigaraki and Saskatoon, while the phase of zonal and meridional component were out of phase and in-phase between Shigaraki and Adelaide, respectively, indicating that there are strong contributions from the antisymmetric (2, 3) mode. Although the agreement between recent modelings and our observations is relatively good, it is still important to study that whether the agreement become better if more realistic background atmosphere parameters are used in the models, or consideration of some new physical process is necessary.

Clear transitions between the two states observed in equinoxes (February–April and October–November) were studied more closely with 4- to 5-day observation intervals among Saskatoon, Adelaide and Shigaraki. The transitions were then found to be slightly delayed at Saskatoon in both vernal and autumnal equinoxes than at Shigaraki, while the transition at Adelaide and Shigaraki was in phase. Abrupt changes in phases of semidiurnal tides coincided with decrease and fluctuations of their amplitudes, suggesting the existence of interference between fundamental and higher modes which are dominant in summer-like and winter-like months, respectively, while the existence of the higher mode at the equinox could explain the whole transition better.

Comparisons of the diurnal tidal amplitude and phase parameters between Adelaide and Shigaraki also showed that there are distinct hemispheric asymmetries. The diurnal tide was usually much larger in amplitude at Adelaide, and it exhibited a vertical profile which indicates a relatively larger contribution from the propagating (1, 1) mode at this location. Comparisons with recent theoretical work suggest that while some of these differences may be due to hemispheric differences in forcing and/or to interactions between the tide and the background winds in the middle atmosphere, it is also possible that they indicate weaker turbulence induced damping of the (1, 1) mode in the southern hemisphere because of the stronger excitation of gravity waves in the northern

hemisphere.

Two-day waves were studied by simultaneous observations between Kyoto meteor radar and Adelaide MF radar. Wind oscillations with periods ranging from 1.4 to 20 days at 80-110km altitude were detected using Kyoto meteor radar observations made in 1983-1985. Among these oscillations, the two-day wave was repeatedly enhanced in summer and autumn. The period of the two-day wave ranges from 52 to 55h in summer, and became as short as 46 to 48h in autumn in 1983 and 1984. The change in the wave period seems to coincide with a decrease in the amplitude of the zonal mean wind. A two-day wave event was simultaneously observed in January 1984 at Shigaraki and Adelaide. Amplitudes at Adelaide were larger than those observed at Shigaraki by a factor exceeding 2. The meridional wind component was in phase and the zonal component was out of phase, respectively, implying antisymmetry of the two-day wave between the northern and southern hemispheres. The extensive comparison from 1984 to 1986 showed that the average amplitude of two-day wave at Adelaide was about two times as large as that at Shigaraki.

Thus, the comparison of the radar observation between Shigaraki and Adelaide showed that the mean wind in the mesosphere and lower thermosphere between the northern and southern hemisphere are comparatively symmetrical, while the wave amplitude such as tides and two day waves were much larger in the southern hemisphere than in the northern hemisphere. These difference could be caused by the difference of the eddy viscosity, that can be generated by the hemispheric difference of the gravity wave activity between the hemisphere, because the turbulences are considered to be mainly generated by gravity wave breaking in the mesosphere.

Therefore, the latter part of the thesis are devoted to the gravity wave studies, among the MU radar at Shigaraki, and the MF radars at Adelaide and Saskatoon. First, we studied the characteristics of the gravity waves at Shigaraki, which are necessary to study the gravity wave propagations and energy and momentum transport from the lower atmosphere to the mesosphere, and from the mesosphere to the thermosphere. Then, we compared the gravity waves observed at the three latitudes and discuss the global distribution of the gravity wave activity.

The characteristics of gravity waves (8-10 hours to 5 min periods) in the mesosphere at 60-90 km were observed by means of the MU radar in four periods; June 1987, July 1990, October 1986 and January/February 1991. The frequency spectra of radial wind velocities were basically similar among four observation periods, showing that oblique spectra had a logarithmic slope of about $-5/3$. The spectral density showed semiannual variation with large amplitudes in summer

and winter. Gravity waves with periods longer than 20–50 min to 5 hours (the lowest limit of the spectral analysis) were found to carry a large part of zonal momentum flux, while dominant frequency component was not detected for the meridional flux. This result is different from the previous work by the MF radars, which pointed out that the major part of momentum flux are transported by the gravity waves with periods less than 60 minutes. Zonal and meridional drag of mean winds induced by the gravity wave with periods from 5 min to 8–10 hours were +51 m/s/day and –4 m/s/day in June at 75km, and –4.0 m/s/day and +7.4 m/s/day in October at 70 km, respectively. The major part of drag force was also found in the wave component with periods larger than 30 minutes. Day to day variation of zonal momentum flux showed a good correlation with vertical shear of zonal mean wind, which suggests effects of the gravity wave activities on mean wind fields.

We further studied characteristics of dominant component of gravity waves in the vertical profile of the wind velocities observed in the mesosphere (at around 70–75 km) with the MU radar in the four campaign periods and the monthly mesosphere observations carried out for 4–5 days a month in December 1985–December 1988.

Most of the dominant waves propagated upward except that they sometimes propagated downward mainly in summer. A typical vertical wavelength and intrinsic period of the dominant gravity waves were 5–15 km and 4–15 hrs, respectively. Mean values of the vertical wavelength, horizontal wavelength and intrinsic periods were 8.2km, 1200km and 10.2hrs, respectively, which did not show significant seasonal variations. Most of the waves were found in the strong dissipation range of the eddy diffusion of 36–300 m²/s, which does not agree with MF radar observations, but rather consistent with Sodium lidar observation. Mean horizontal phase velocity was 33 m/s. Horizontal propagation direction was generally eastward throughout a year except for the westward propagation in early winter (November–December). This suggests the dominant gravity waves are less sensitive to the change of the mean zonal wind direction in the middle atmosphere than the short period gravity waves (2 hour to 5 minutes). Momentum flux carried by the dominant gravity waves was estimated as 0.5 ~ 1m²/s², which was generally smaller than the momentum flux of gravity waves with periods ranging from 8 hours to 5min. The amplitude of the momentum flux of the dominant component had the autumnal maximum, while the momentum flux with periods of 8 hours–5 minutes had equinoctial minima. Thus in September and October, the momentum flux due to the dominant component became significant.

The gravity waves observed at Shigaraki(35°N) and Saskatoon (52°N) were

compared for the data obtained in January to December 1989. The frequency spectra of horizontal winds u' and v' showed a spectral slope of $-5/3$ and in summer the slope became gradual at the frequency component larger than $20-30 \text{ days}^{-1}$ for both data taken by the MU radar and Saskatoon MF radar. The spectral slope of w' was basically $-2/3$ to -1 in winter and $-1/3$ to $-1/2$ in other seasons at Saskatoon, while the slope was smaller or sometimes positive at Shigaraki. The wind variance $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ showed semiannual variation, with maximum in summer, secondary peak in winter and minima in September and April for both locations. The annual average of the ratio of $\overline{u'^2}$, $\overline{v'^2}$ and $\overline{w'^2}$ between the MU and Saskatoon MF radars were 1.6, 1.5 and 1.6, suggesting the gravity wave variance at 35°N is stronger than that at 52°N by about 30–40% if the difference of the observations is considered. The zonal momentum flux showed an similar annual variation with the eastward and westward peaks in summer and winter, respectively, for both locations, although the zonal momentum flux at Saskatoon seems to be shifted towards westward even in summer, which might be due to some bias in the present analysis.

The comparison of the gravity waves observed with the MU radar and Adelaide MF radar were carried out in two campaigns in summer and winter, that is in June 1987 and January/February 1991. The frequency spectra were larger in summer than in winter at both locations, while the spectral density were almost identical if summer or winter in both hemispheres are compared. The variances $\overline{u'^2 + v'^2}$ of the waves with periods of 2 hours – 30 minutes observed by the MU radar were about 10 % larger than that with the Adelaide MF radar, indicating the gravity wave intensity between the hemispheres at the latitudes 35° is almost the same, if we compensate the difference of spatial resolution of the two radars. This suggests that the dissipation due to the eddy viscosity by the turbulence, that is caused by the gravity waves, is not much larger in the northern hemisphere, which is different from what the previous tidal observations suggested.

The propagation direction of the dominant gravity waves in the vertical wind profile at around 80–85km observed by the Adelaide MF radar showed the alignment to north-south direction in summer and about 30° east of north in winter. The Stokes parameter analysis determined the propagation direction of the waves with periods between 6 hour and 30 minutes showed that the propagation direction was aligned to west of north or east of south direction in summer and east-west direction in winter, respectively.

These results are compared with the propagation directions of the gravity waves observed with the MU radar together with the earlier observation at Adelaide [Fritts and Vincent, 1987; Eckermann and Vincent, 1989]. The propagation direction of the different frequency components seems to have similar direction in summer but the directions were rather widely spread in winter at both lo-

cations. The results of the MU radar showed the preferential directions of the gravity wave propagation were aligned to zonal direction more than the results of Adelaide. These results indicated the difference of the gravity wave propagation direction between the hemisphereis, or the difference at the two locations of Adelaide and Shigaraki, which might be caused by the distribution of the continent around these observational sites.

If we interpret this difference of the preferential propagation direction of the gravity waves as the influence of the continent distribution around the radar site, that means the gravity wave activity may have large zonal variability, and hence the comparison of gravity wave variance shown here does not always show a typical zonal mean viscosity that affects the tidal amplitudes. Further observations and comparisons will reveal the more detailed cause of the hemispheric difference of the tidal amplitudes.

In conclusion, we have examined the mesospheric and lower thermospheric dynamics such as mean wind, gravity waves, atmospheric tides and two day waves by radar observations carried out at Shigaraki (35°N , 136°E) by the MU radar and Kyoto meteor radar, and further compared the results with the two MF radar observations at Adelaide (35°S , 138°E) and Saskatoon (52°N , 107°W). These comparisons led us to some important findings on the global structure of the atmospheric wave characteristics, and thus the significance of the cooperative radar observations are confirmed. Further observational study using radar network such as MLTCS network will reveal the dynamic behavior of the middle atmosphere more closely.

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