Tropospheric Mid-Level Detrainment Flow Obtained from High-Resolution Non-Hydrostatic Atmospheric Model Experiments

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Abstract

Time integration using the Regional Atmospheric Modeling System (RAMS), a non-hydrostatic cloud-resolving model, was performed for 12 days over a low-latitude band (45°S–45°N) circling an aqua planet with 5-km horizontal grid-point intervals. Tropical and subtropical regions with active precipitation and clear sky, respectively, were clearly divided at 10° latitudes. The numerical experiments derived obvious tropospheric mid-level detrainment (TMD) flows near the 0°C level ($z \sim 5$ km) out of the tropics into the subtropics. The TMD flows became largest near the border (10° latitude). In this paper, the time-longitudinal mean field was spotlighted and the atmospheric structure accompanying the TMD flow was investigated.

When averaged over time and longitude, the subtropical mid-level troposphere, into which the TMD flows move, is approximately in a state of local thermodynamic equilibrium sustained mainly by the balance between the net radiative cooling and adiabatic heating due to mean subsidence flow. Considering the heat balance, a thermodynamic diagnosis of the mean subsidence flow field suggests the following mechanisms for the mean TMD flow: (1) The mean atmosphere near the melting level has stronger radiative cooling and a larger temperature lapse rate than the atmosphere above it. (2) Free subsidence in the mean subtropical mid-level troposphere, which is consistent with the vertical variation of thermal structure and suffers from no direct dynamic forcings, such as buoyancy, involves a vertically mass-divergent layer just above the melting level. (3) The steady poleward mean TMD flow out of the convective tropic atmosphere exists so as to compensate for the vertical mass divergence in the subtropical atmosphere. Because net meridional transports of sensible heat and water vapor in the middle troposphere are influenced by the mean TMD flow, the existence and the maintaining mechanisms of the mean TMD flow could be important elements of the climate system.

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1. Introduction

Recent observational studies have reported poleward flows from tropical regions into subtropical regions around the melting level of cloud ice with 0°C temperature (e.g., Johnson et al. 1999; Kemball-Cook and

Weare 2001; Kikuchi and Takayabu 2004; Yasunaga et al. 2004; Takayabu et al. 2006; Zhang et al. 2008). Because such flows are considered to be caused by horizontal mass divergence from cumulus convection, i.e., detrainment, they are referred to as tropospheric mid-level detrainment (hereafter, TMD) flows. These studies have investigated dynamic aspects of TMD flows and suggested that the flows originate in detrainments associated with mesoscale disturbances. Namely, a stable layer at a level in the middle troposphere could discourage ascending motion of air parcels in the cumulus convection, thereby generating vertical mass convergence and hence horizontal mass divergence (detrainment) near that level. With the equator being the symmetric boundary between the hemispheres, the ensemble effect of the TMDs of individual cumulus convections in the tropics appears as poleward flows at the global scale. In most studies, detrainment flow occurring near the melting level (near 550 mb) is more prominent than other detrainment flows in the lower to middle troposphere. These findings suggest that latent heat release from melting ice is one candidate for generating the stable layer in the middle troposphere. However, the dynamics of TMD flow does not appear to be so simple. For example, from analysis of Tropical Ocean Global Atmospheres/Coupled Ocean Atmosphere Response Experiment (TOGA COARE) data, Yasunaga et al. (2004) reported that TMD flow generation may depend on the existence of a stable layer, vertical shear of horizontal wind, and melting processes. Recent studies have also suggested the effect of thin stratiform cloud formed just below the melting level (Katsumata et al. 2006; Yasunaga et al. 2006; Zuidema et al. 2006; Yasunaga et al. 2008).

Air parcels rising in cumulus updrafts dry due to condensation and precipitation. Because TMD flows draw air through the cumulus flank at a level much lower than the tropopause, these flows can convey more water vapor from the moist tropical region to the dry subtropical region than can other detrainment flows near the tropopause. The effect of TMD flows in net (i.e., average over time and space) meridional transports of mass, heat, and water vapor thus constitutes an important issue when considering the role of cumulus convection in the climate system. However, in previous studies of observations and models covering limited areas and times, TMD flows have mainly been treated as local and short-lived phenomena, and the net transports have been discussed little.

There is another approach to understand outflow of a convective area into a non-convective area. Sherwood (1996), Salathé and Hartmann (1997), Pierrehum-

bert (1998), and Dessler and Sherwood (2000), using general circulation model (GCM) integrations, demonstrated that outflows in the middle to upper troposphere in non-convective areas occur so as to maintain local thermodynamic equilibria in the atmosphere (in this paper, the words “local thermodynamic equilibrium” are used, not for the meaning referred to in the statistic physics, but for a heat balance maintained by diabatic and adiabatic processes of a moving air parcel). Using simplified radiative-convective equilibrium models (including a non-hydrostatic dynamic model) in a 2-D domain, Iwasa et al. (2002, 2004) investigated net transports of mass and water vapor out of cumulus convection that develops intermittently at a fixed horizontal location into the surrounding atmosphere. The net transports were explained by a field of mean subsidence, in which excess adiabatic heating was locally balanced with the net radiative cooling, and its compensating horizontal flow. Although being different from the traditional approach to explaining local and short-term atmospheric motions from detailed dynamic features of cumulus convection, this approach, considering the local thermodynamic equilibrium in the atmosphere outside cumulus convection, can be effective for discussions of net transport out of a convective area and into a non-convective area.

High-resolution non-hydrostatic numerical models provide less ambiguous properties of the low latitudinal atmosphere compared with coarse-resolution parameterized GCMs. This advantage is due not only to the higher resolution of such models but also to their direct treatment of cumulus convection. The recent availability of huge computers, such as the Earth Simulator, has enabled us to conduct large-scale numerical experiments using high-resolution non-hydrostatic numerical models (e.g., Miura et al. 2005; Nasuno et al. 2006; Satoh et al. 2008). In this paper, we present TMD flows obtained from a very large domain numerical experiment using the Regional Atmospheric Modeling System (RAMS), a high-resolution non-hydrostatic numerical model. Because a very simplified boundary condition of an aqua planet was used in the experiment, our aim was not to simulate the real atmosphere or provide detailed analysis of local and short-term phenomena. Instead, we diagnose the time- and zonal-mean element of the TMD flows based on the local heat balance in the subtropical atmosphere. We then show the contribution of the mean flow element to the net meridional transports of sensible heat and water vapor by the TMD flows.

The numerical model used in this study is briefly described in Section 2. In Section 3, after showing general features of the model results, structures of TMD flows

Fig. 1. Snapshots at 10-days integration for latitude-height distributions of the zonal mean values of (a) eastward wind velocity $\bar{u}$, (b) northward wind velocity $\bar{v}$, (c) temperature $\bar{T}$, and (d) water vapor mixing ratio $\bar{q}_v$.

and atmospheric stabilities are demonstrated and then the result of a supplementary experiment of suppressed latent heat exchanges in water freezing and melting is presented, leading to a novel approach to the mean element of the TMD flows. In Section 4, by considering minimal physical processes maintaining the local thermodynamic equilibrium in the mean middle-tropospheric atmosphere in the subtropical region averaged over time and longitude, a mechanism for the mean element of the TMD flows is diagnosed, and its contribution to net meridional transport is evaluated. Finally, concluding remarks are provided in Section 5.

2. Model description

Numerical experiments using RAMS 3b (Walko et al. 1995) for atmosphere over a band between 45°S and 45°N encircling an aqua planet were performed on the Earth Simulator. The greatest advantage of using RAMS is that this non-hydrostatic model, which has been tested in various simulations, can be applied to the whole tropical and subtropical region without any modification. The model treats cumulus convection explicitly and considers seven types of condensed water species besides water vapor. Here, the original radiation code was replaced by a more sophisticated scheme of Nakajima et al. (1995). Horizontal resolution was $\Delta x = \Delta y = 5$ km, with 35 layers of varying thickness in

the vertical up to 26 km. Most of the free troposphere was resolved with a vertical grid interval of $\Delta z = 1$ km. The time increment was $\Delta t = 30$ s.

The initial model state was derived from the output of the Center for Climate System Research/National Institute for Environmental Studies (CCSR/NIES) atmospheric GCM (AGCM) (T42-L20 resolution) (Numaguti et al. 1995; K-1 model developers 2004), an atmospheric model that was integrated from a stationary isothermal atmosphere for one model year. Sea surface temperature was fixed as

$$T_{\text{SST}} = \begin{cases} 27.0 \times \left[1.0 - \sin^2 \left(\frac{\pi}{120} \phi \right) \right] ^\circ\text{C} & (-60^\circ < \phi < 60^\circ) \\ 0.0^\circ\text{C} & (\phi \leq -60^\circ \text{ or } 60^\circ \leq \phi), \end{cases} \quad (1)$$

where ϕ is northern latitude in degrees. Given that the model had high SST and no land surface, the model atmosphere may be regarded as that over the warm pool region of the Western Pacific. Annual mean insolation with diurnal change was maintained at the top of the atmosphere. The model values at the northern, southern, and top boundaries were nudged using AGCM values recorded every 12 hours.

Fig. 2. Snapshots at 10-days integration for (a) the longitude-latitude distribution of precipitation rate and (b) the meridional profile of the zonal-mean precipitation rate.

3. Results of the numerical experiment

3.1 General features

In this paper, the value of a physical quantity ϕ averaged over longitude is denoted using an overline as $\overline{\phi}$, whereas that averaged over both time and longitude

is denoted using brackets as $\langle \phi \rangle$. After 10-days integration, little change was observed in zonal mean values for wind fields and atmospheric conditions, such as temperature and humidity distributions (Fig. 1). The model atmosphere seemed to have approached a quasi-

Fig. 3. Snapshots at 10-days integration for (a) the vertical profiles of log-scaled mixing ratios of water species averaged over the tropics (between 10°S and 10°N) and (b) those averaged over the typical northern subtropics (between 13°N and 17°N). The horizontal broken line in (a) shows the constant height $z = 5$ km near the melting level in the tropics.

steady state, with cumulus convection activated in the tropical region. The atmosphere had a double intertropical convergence zone (ITCZ) structure in the tropics (Fig. 2a). The zonal mean precipitation rate (Fig. 2b) was large between 10°S and 10°N, corresponding to the activated cumulus convection. The meridional profile roughly resembled previous observations (e.g., Johnson et al. 1999) in the locations and the values of the double

peaks. These results suggested that, despite the simplified boundary condition of an aqua planet, the low-latitude atmospheric state obtained from the RAMS calculation was not extremely different from that of the real atmosphere. Figure 3 shows vertical profiles of log-scaled mixing ratios for water species horizontally averaged in the tropical region ((a), between 10°S and 10°N) and in the typical northern subtropical region

Fig. 4. Latitude-height distributions of the zonal mean heating rates in the middle troposphere between the equator and 25°N averaged over 2 days after 10-days integration for (a) shortwave, (b) longwave, and (c) the net radiation.

((b), between 13°N and 17°N). In the tropics (Fig. 3a), the level $z = 5 \text{ km}$ has an estimated 0°C temperature, corresponding to the melting level of cloud ice. Water vapor predominates over the other water species in amount at every level. Abundant water vapor fills the atmosphere below the melting level, whereas water vapor

content decreases quickly above the melting level (e.g., $5.80 \times 10^{-3} \text{ kg kg}^{-1}$ at $z = 5 \text{ km}$ and $3.51 \times 10^{-3} \text{ kg kg}^{-1}$ at $z = 6 \text{ km}$). The contents of cloud and rain water also decrease quickly above the melting level. Most species of solid-phase water exist above $z = 5 \text{ km}$. Only hail and graupel, having large terminal velocities, fall

Fig. 5. Snapshot at 10-days integration for the vertical profile of upward mass flux averaged over the tropics (between 10°S and 10°N).

across the melting level, although the contents are small compared to those of the liquid water. In the subtropics (Fig. 3b), the melting level decreases in height with latitude as $z \sim 5$ km at 10°S and N and $z \sim 4$ km at 20°S and N. Except for the descent of the melting level, the general features in the figure are common for both the tropics and subtropics, although the latter always has smaller mixing ratios than the former (e.g., water vapor content in the subtropics is 3.75×10^{-3} kg kg⁻¹ at $z = 5$ km and 1.75×10^{-3} kg kg⁻¹ at $z = 6$ km). Most of the total water amount exists below the melting level. For example, water vapor amounts vertically integrated below the melting level occupy about 93% and 95% of the total water vapor amount in the tropics and subtropics, respectively. The difference of water amounts across the melting level is attributed to the difference in water phases. The solid phase has larger efficiency in water removal from the air than the liquid phase: The solid phase has lower saturation water vapor pressure and larger efficiency in collecting condensed water particles compared to the liquid phase (Pruppacher and Klett 1997).

Figure 4 shows the latitude-height distributions of time- and zonal-mean heating rates in the middle troposphere for (a) shortwave, (b) longwave, and (c) net radiation (summation of (a) and (b)). Shortwave and long-

wave processes work as heating and cooling, respectively. Both effects are stronger near the melting level than in the adjacent domains above and below. The tropical region has the radiatively active layer near a constant height of $z \sim 5$ km, whereas the subtropical region is characterized with decrease of the height with latitude as $z \sim 5$ km at 10°N and $z \sim 4$ km at 20°N, corresponding to the respective melting level heights. In general, the subtropical region is radiatively more active, reaching larger heating and cooling values (e.g., 1.24×10^{-5} K s⁻¹ and -3.81×10^{-5} K s⁻¹ at $y = 17^\circ$ N and $z = 4$ km for shortwave heating and longwave cooling, respectively) than the tropical region (e.g., 9.28×10^{-6} K s⁻¹ and -2.84×10^{-5} K s⁻¹ at $y = 2^\circ$ N and $z = 5$ km for shortwave heating and longwave cooling, respectively). In the middle troposphere, the cooling rates due to longwave radiation are larger than the heating rates due to shortwave radiation. Consequently, the net radiation effect is essentially cooling, mainly determined by the longwave radiation (e.g., -2.57×10^{-5} K s⁻¹ and -1.91×10^{-5} K s⁻¹ at $y = 17^\circ$ N and $z = 4$ km, $y = 2^\circ$ N and $z = 5$ km, respectively).

3.2 Tropospheric mid-level detrainment flows

Figure 5 is the vertical profile of mean vertical mass flux in the tropics between 10°S and 10°N in snapshot at 10-days integration. Two local maximum peaks occur at $z = 10$ km and $z = 4$ km. The decrease in the mean upward mass flux near the tropopause indicates the top of deep convection. The other decrease near the melting level was also reported by Liu et al. (2001) from TOGA COARE simulations by non-hydrostatic cloud resolving models (CRMs) with 2-km horizontal resolution. Synchronous with this mean upward mass flux decrease, TMD is generated near the melting level in the model atmosphere. Figure 6 shows the detrainment on the global scale as a zonal mean poleward flow of air from the tropics to the subtropics. Observation studies have reported similar poleward flow near the melting level (e.g., Zuidema 1998). The TMD flow is strongest at 10°S and 10°N, just on the edges of the convective range in the tropics.

In terms of the maximum zonal mean poleward velocity $\bar{v}$ shown in Fig. 6a, the detrainment flow near the tropopause (e.g., 2.49 m s⁻¹ at 10°N) is larger than the TMD flow (7.02×10^{-1} m s⁻¹ at the same location). However, because the atmospheric density ρ decreases with height, the corresponding maximum poleward mass fluxes $\bar{\rho}\bar{v}$ shown in Fig. 6b become similar to each other between those near the tropopause (7.88×10^{-1} kg m⁻² s⁻¹) and near the melting level (5.30×10^{-1} kg m⁻² s⁻¹). Furthermore, because wa-

Fig. 6. Left figures: Latitude-height distributions of the zonal mean meridional (a) velocity $\bar{v}$, (b) mass flux $\bar{\rho v}$, and (c) water vapor flux $\bar{q}_v \bar{v}$. Right figures: Vertical profiles of the respective velocity and fluxes at 10°S (broken lines) and 10°N (solid lines). All the values are the snapshots at 10-days integration.

ter vapor content q_v in the middle troposphere is much larger than that near the tropopause, the maximum zonal mean water vapor flux $\bar{q}_v \bar{v}$ in the middle troposphere ($5.12 \times 10^{-3} \text{ m}^{-2} \text{ s}^{-1}$) predominates over that in the tropopause ($4.32 \times 10^{-4} \text{ m}^{-2} \text{ s}^{-1}$) as shown in Fig. 6c. Just below the levels of the two prominent detrainment flows, opposite zonal mean flows toward the equator (i.e., entrainment flows) occur in the 8~10-km and 3~4-km levels. Such entrainment flows have also been

reported in observation studies (e.g., Mapes 2001).

3.3 Structure of stability in the troposphere

The stability of air parcels can be examined by comparing the moist static energy (Bohren and Albrecht 1998) of the air parcels

$$E = gz + c_p T + L_{CE} q_v \quad (2)$$

Fig. 7. Snapshots at 10-days integration for the vertical distributions of grid frequency in the tropics (between 10°S and 10°N) from 0 to infinity for (a) moist static energy E and (b) saturation moist static energy E^* . The contour interval is 10%. The thick broken lines show vertical profiles of the respective energies averaged horizontally over the same region.

and the saturation moist static energy of the ambient atmosphere

$$E^* = gz + c_p T + L_{CE} q_v^*, \quad (3)$$

where $g = 9.8 \text{ m s}^{-2}$, $c_p = 1.005 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$, and T are the gravitational acceleration, specific heat of air at constant pressure, and temperature (K), respectively. The latent heat constant L_{CE} of water condensation and evaporation is given by

$$L_{CE} = 2.501 \times 10^6 \text{ (J kg}^{-1}\text{)}, \quad (4)$$

q_v is mixing ratio of water vapor in the air parcel, and q_v^* is the saturation mixing ratio of water vapor in the ambient atmosphere calculated from temperature and pressure, respectively. Figures 7a and 7b for E and E^* , respectively, show the distributions of grid frequency between 10°S and 10°N in the plane of static energy and height (contours), together with the vertical profiles of the values averaged over the same region (thick broken lines). The vertical profiles resemble observed data (e.g., Mapes 2001). Air parcels ascending in the cu-

mulus convection are basically supplied from the mixing boundary layer ($0 \leq z \leq 1 \text{ km}$). In the boundary layer, the air parcels typically take E values between 3.33×10^5 and $3.54 \times 10^5 \text{ J kg}^{-1}$ (the mean value at the top of the boundary layer $z = 1 \text{ km}$ is $3.393 \times 10^5 \text{ J kg}^{-1}$), as shown in Fig. 7a. If the updraft in the cumulus convection is assumed to occur quickly, the parcels keep these E values during the upward motion. When the value of E in an air parcel is larger than the value of the saturation moist static energy E^* of the ambient atmosphere ($E > E^*$) at a specified level, then the air parcel is unstable during ascent. On the other hand, when E of the air parcel takes a value smaller than E^* of the ambient atmosphere ($E < E^*$), the parcel becomes stable. In contrast to the simple profile of mean E in Fig. 7a with a single minimum value $E = 3.349 \times 10^5 \text{ J kg}^{-1}$ at $z = 3.25 \text{ km}$, the profile of mean E^* in Fig. 7b has a more complicated structure. A level with a local minimum mean value $E^* = 3.442 \times 10^5 \text{ J kg}^{-1}$ at $z = 2.5 \text{ km}$ appears in the lower troposphere. Although the cause is not clear, this layer corresponds to the stable layer at 800 mb reported by Zuidema (1998). A local minimum value of $E^* = 3.429 \times 10^5 \text{ J kg}^{-1}$ also appears at $z = 9 \text{ km}$.

Most remarkably, however, the profile of mean saturation moist static energy attains a minimum value $E^* = 3.413 \times 10^5 \text{ J kg}^{-1}$ near the melting level ($z \sim 5 \text{ km}$). As mentioned in Subsection 3.1, strong net radiative cooling is generated near the melting level (Fig. 4c). This cooling is thought to keep working on the air near that level, making the equilibrium temperature slightly lower than the temperatures of adjacent domains above and below. Such a small decrease in temperature can produce a conspicuous decrease in the saturation moist static energy E^* because of the strong temperature dependence of the saturation mixing ratio of the water vapor term q_v^* included in Eq. (3). The typical values of E^* at $z = 5 \text{ km}$ are between $3.38 \times 10^5 \text{ J kg}^{-1}$ and $3.45 \times 10^5 \text{ J kg}^{-1}$, which are similar to the E values in the air parcels in the upper part of the mixing boundary layer. Because a stable layer $\partial E^* / \partial z > 0$ exists just above the level $z = 5 \text{ km}$, detrainment flows may possibly occur near that level. Interestingly, as shown in Fig. 8, the layer of minimum E^* values in the middle troposphere coincides roughly with the paths of the TMD flows (Fig. 6). This layer extends not only in the tropical region, but also over the subtropical regions. The layer of the minimum E^* values may be regarded as a duct layer along which the stable air parcels flow. Precisely comparing Figs. 4, 6, and 8, however, one can notice that the exact level of the detrainment flows appear slightly higher than the level of the maximum radiative

Fig. 8. Snapshot at 10-days integration for the latitude-height distribution of zonal mean saturation moist static energy $\overline{E^*}$.

Fig. 9. Same as Fig. 7 except for the temperature lapse rate Γ calculated from temperatures in adjacent pairs of vertical grid points.

cooling and the minimum saturation moist static energy.
Temperature lapse rate

$$\Gamma \equiv -\frac{\partial T}{\partial z} \quad (5)$$

is another measure for the atmospheric stability. Atmosphere with smaller Γ values is more stable. Figure 9 shows the grid frequency distribution of the temperature lapse rate and the mean vertical profile in the tropical region (between 10°S and 10°N). Stable layers are exhibited with the local minimum lapse rates at the top of the boundary layer ($\Gamma = 3.28 \text{ K km}^{-1}$ at $z = 1.25 \text{ km}$), in the lower troposphere ($\Gamma = 3.33 \text{ K km}^{-1}$ at $z = 2.88 \text{ km}$), and just above the melting level ($\Gamma = 3.82 \text{ K km}^{-1}$ at $z = 5.5 \text{ km}$). The stable layers agree with observational findings (e.g., Johnson et al. 1996; Zuidema 1998;

Johnson et al. 1999; Kemball-Cook and Weare 2001; Kikuchi and Takayabu 2004; Yasunaga et al. 2004).

As shown above, a stable layer exists just above the level of the TMD flows in terms of both the moist static energy and temperature lapse rate. This result means that the vertical structure of the atmospheric stability allows the TMD flows. However, it does not explain anything about the quantity of the TMD flows.

3.4 Effect of latent heat exchanges in water freezing and melting

Some previous studies (e.g., Scala et al. 1990; Johnson et al. 1996; and Yasunaga et al. 2004) suggested that the melting process may play an important role in enhancing horizontal flow at the 0°C level. Yasunaga et al. (2004) reported that when they suppressed the latent heat exchange in the melting process, outflow at

Fig. 10. Snapshot at 10-days integration for the vertical profiles of zonal mean meridional mass flux $\bar{\rho}v$ at 10°S (broken line) and 10°N (solid line) obtained from the experiment of suppressed latent heat exchanges in water freezing and melting processes.

Fig. 11. Snapshot at 10-days integration for the longitude-height distribution of poleward mass flux ρv at 10°N.

the mid-troposphere became weak in a cloud-resolving two-dimensional model (Advanced Regional Prediction System: ARPS) simulating TOGA COARE. The effect of latent heat release may be strong in a local area where large solid water content undergoes melting while little liquid water freezes. Such a situation would be realized

locally in a severe squall line or a strong convection in the mature stage with a large amount of precipitation, like situations reported in the above studies.

To investigate the effect due to latent heat exchange in freezing and melting processes within the model, we attempted a warm rain experiment in which no solid

Fig. 12. Snapshots at 10-days integration for the longitudinal profiles of poleward mass flux ρv at 10°N at (a) $z = 13\text{ km}$ and (b) $z = 5\text{ km}$ levels in Fig. 11. Horizontal lines show the mean values ($7.88 \times 10^{-1} \text{ kg m}^{-2} \text{ s}^{-1}$ and $5.30 \times 10^{-1} \text{ kg m}^{-2} \text{ s}^{-1}$, respectively). Histograms of the longitudinal distributions normalized to have a unit standard deviation around the zero mean value, together with the curves of the ideal normal distribution at (c) $z = 13\text{ km}$ and (d) $z = 5\text{ km}$ levels. Each histogram has nine classes.

water existed. However, the experiment failed, producing an unrealistic atmospheric state in which the whole free troposphere (except for the mixing boundary layer near the surface) filled with liquid water cloud. This result is attributed to inefficient water removal resulting from the lack of the solid water phase. In place of the warm rain experiment, an alternative experiment, similar to that of Yasunaga et al. (2004), was performed. The microphysical processes were calculated normally, but latent heat exchanges in freezing and melting were suppressed in the model integration. This was accomplished by changing the value of the latent heat constant for freezing and melting

$$L_{\text{FM}} = 3.34 \times 10^5 \text{ (J kg}^{-1}\text{)} \quad (6)$$

to be zero. The value of the latent heat constant L_{CE}

for condensation and evaporation given by Eq. (4) was used normally. The initial state was the same as that used for the main experiment described in Section 2. The integration time was 10 days.

The resulting zonal mean TMD flow in the case of the zero L_{FM} value, given in Fig. 10, shows no clear differences from that in case of the normal L_{FM} value presented in Fig. 6. It means that the latent heat exchanges of freezing and melting are not primarily responsible for the TMD flows in the model. This may be because the latent heat exchanged in freezing and melting is too small compared with that in condensation and evaporation to affect the mean flow. There is also a possibility that the TMD flows observed in previous studies may have differed from those obtained in this study, and hence, the mechanisms between the flows may also

Table 1. Statistics of the longitudinal distributions of the poleward mass fluxes ρv at 10°N shown in Figs. 12a and 12b.

	Near the tropopause $z = 13 \text{ km}$	Near the melting level $z = 5 \text{ km}$
Zonal mean value $\overline{\rho v}$ ($\text{kg m}^{-2} \text{ s}^{-1}$)	7.88×10^{-1}	5.30×10^{-1}
Standard deviation σ ($\text{kg m}^{-2} \text{ s}^{-1}$)	1.92	2.32
χ^2 test statistic	0.399	0.348

differ: the mechanism for the mean TMD flow could be independent of the latent heat exchange, whereas the mechanisms for local and short-term detrainment flows may depend on that process.

3.5 Longitudinal distribution of the TMD flows

To obtain clues regarding the quantity of the TMD flows, we examine here the longitudinal distribution of the TMD flows in the model atmosphere. Figure 11 shows the longitude-height distribution of poleward mass flux ρv at 10°N in snapshot at 10-days integration. The longitudinal profiles at $z = 13 \text{ km}$ and $z = 5 \text{ km}$ levels are shown in Figs. 12a and 12b, respectively. The maximum values of poleward detrainment flows are $5.37 \text{ kg m}^{-2} \text{ s}^{-1}$ at $x = 105^\circ$ and $6.75 \text{ kg m}^{-2} \text{ s}^{-1}$ at $x = 25^\circ$, at $z = 13 \text{ km}$ and $z = 5 \text{ km}$ levels, respectively. Between the domains of poleward detrainment flows, domains of opposite equatorwise entrainment flows appear. The maximum magnitudes for the entrainment flows, $3.76 \text{ kg m}^{-2} \text{ s}^{-1}$ at $x = 218^\circ$ and $7.03 \text{ kg m}^{-2} \text{ s}^{-1}$ at $x = 299^\circ$, at the respective levels, are similar to those for the detrainment flows.

A simple statistical treatment was used for the longitudinal distributions of the flows. As shown in Table 1, the mean values $\overline{\rho v}$ were $7.88 \times 10^{-1} \text{ kg m}^{-2} \text{ s}^{-1}$ and $5.30 \times 10^{-1} \text{ kg m}^{-2} \text{ s}^{-1}$, and the standard deviations σ were $1.92 \text{ kg m}^{-2} \text{ s}^{-1}$ and $2.32 \text{ kg m}^{-2} \text{ s}^{-1}$ for the poleward mass fluxes at the levels of $z = 13 \text{ km}$ and $z = 5 \text{ km}$, respectively. Using these statistical values, the distribution of the poleward mass flux ρv was normalized to that of a variable ζ with a zero mean value and a unit standard deviation by

$$\zeta = \frac{\rho v - \overline{\rho v}}{\sigma}. \quad (7)$$

The normalized distributions of ζ are shown in Figs. 12c and 12d for the levels of $z = 13 \text{ km}$ and $z = 5 \text{ km}$, respectively. The figure also displays the ideal normal distribution curve

$$f(\zeta) = \frac{1}{\sqrt{2\pi}} \exp\left(-\frac{\zeta^2}{2}\right). \quad (8)$$

Applying the χ^2 test to these distributions gives results of 0.399 and 0.348 for the respective levels. These results show that at each level, the distribution observed in the model and the corresponding ideal normal distribution are not significantly different. That is, the detrainment flows at each level can be regarded as composed of a common non-zero mean element and random disturbing elements around the mean value. These results suggest that, instead of investigating in detail individual TMD flow elements at local longitudes and specific times, clarifying the mechanism that drives the mean element of the TMD flows can be an approach for understanding the TMD flows.

4. Discussion of mean TMD flow

In this section, we try to present an explanation for the mean TMD flow. It is not a mechanism based on causality, such as in an initial-value time-development problem, but a mechanism that is involved in the field of local thermodynamic equilibrium of the mean atmosphere. First, Subsection 4.1 surveys the kinds of thermodynamic process that dominate around the TMD flows. Subsection 4.2, the main part of the discussion, shows that the mean TMD flow element can be reconstructed diagnostically by considering an extremely simplified heat balance in an steadily moving mean atmosphere with a few assumptions. Finally, in Subsection 4.3, the mean TMD flow element is shown to contribute significantly to the net poleward transports of sensible heat and water vapor by the whole TMD flows.

4.1 Estimating each term in the thermodynamic equation

Before beginning the main discussion, let us estimate and compare the effects of each thermodynamic process in the atmosphere into which the TMD flows move. The thermodynamic equation for the temperature $\bar{T}(t, y, z)$ of zonal mean atmosphere is

Fig. 13. Snapshots at 10-days integration for the latitude-height distributions of the zonal mean values (those except for the right-bottom figure) and the vertical profiles of the horizontal mean values (the right-bottom figure) of the terms in the thermodynamic equation (9) between 10°N and 25°N. V_2 : advective heating due to mean horizontal velocity, V_3 : advective heating due to disturbing horizontal velocity, V_4 : advective and adiabatic heating due to mean vertical velocity, V_5 : advective heating due to disturbing vertical velocity, and V_6 : net radiative heating. The contour interval $1.5 \times 10^{-5} \text{K s}^{-1}$ for V_4 is larger than that $3.5 \times 10^{-6} \text{K s}^{-1}$ for the others.

$$\begin{aligned}
 \frac{\partial \bar{T}}{\partial t} &= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{w} \frac{\partial \bar{T}}{\partial z} + \sum_i \frac{\partial \bar{T}}{\partial t} \Big|_i \\
 &= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{v}'^x \frac{\partial \bar{T}'^x}{\partial y} - \bar{w} \frac{\partial \bar{T}}{\partial z} - \bar{w}'^x \frac{\partial \bar{T}'^x}{\partial z} + \frac{\partial \bar{T}}{\partial t} \Big|_A + \frac{\partial \bar{T}}{\partial t} \Big|_R + \frac{\partial \bar{T}}{\partial t} \Big|_L + \frac{\partial \bar{T}}{\partial t} \Big|_D \quad (9) \\
 &= -\bar{v} \frac{\partial \bar{T}}{\partial y} - \bar{v}'^x \frac{\partial \bar{T}'^x}{\partial y} - \left(\frac{\partial \bar{T}}{\partial z} + \frac{g}{c_p} \right) \bar{w} - \bar{w}'^x \frac{\partial \bar{T}'^x}{\partial z} + \frac{\partial \bar{T}}{\partial t} \Big|_R + \frac{\partial \bar{T}}{\partial t} \Big|_L + \frac{\partial \bar{T}}{\partial t} \Big|_D,
 \end{aligned}$$

V_1 V_2 V_3 V_4 V_5 V_6 V_7 V_8

where the overlines and primes with x represent zonal mean values and disturbing elements from the zonal mean values, respectively. The terms $\partial \bar{T} / \partial t|_i$ are the local heating rates due to physical processes indicated

by the subscript i standing for adiabatic **A**, radiative **R**, latent heat **L**, and diffusive **D** processes, respectively. The adiabatic heating term is given as

$$\left. \frac{\partial T}{\partial t} \right|_{\mathbf{A}} = -\frac{g}{c_p} \bar{w}, \quad (10)$$

where $g/c_p = 9.75 \text{ K km}^{-1}$ is the dry adiabatic lapse rate. In the final form of Eq. (9), two terms proportional to the mean vertical flow velocity $\bar{w}$, those for mean vertical advection heating and adiabatic heating, were unified.

The symbol V_i ($i = 1, 2, \dots, 8$) below each term in Eq. (9) indicates the magnitude of the value of the term. Regrettably, data for estimating the values V_1 , V_7 , and V_8 were not available at the time of our analysis. However, the temperature distribution of the zonal mean atmosphere integrated for 10 days does not change over time significantly, as mentioned at the beginning of Subsection 3.1. Additionally, it seems unlikely that water phase change or wind shear would make a large contribution to the heat balance in the atmosphere of the subtropical mid-troposphere, where subsidence controls and strong motions, such as cumulus convection, are inactive. Therefore, the contributions of V_7 and V_8 as well as of V_1 may be dropped from the comparison in Eq. (9). For a rough estimate of the contribution of each term, the values of V_2 , V_3 , V_4 , V_5 , and V_6 in the zonal mean atmosphere snapshot at 10-days integration are compared in Fig. 13. The distributions are shown between 10°N and 25°N . The vertical profiles of the values, given in the right-bottom part of Fig. 13, were those averaged horizontally in the same region. Note that the values shown here are snapshots provided as rough estimates and for comparison. Thus, they are not balanced in an exact way, and detailed comparison between them is meaningless.

Heating rates due to mean horizontal velocity V_2 are negligibly small. The small values of this advection term $\bar{v}\partial\bar{T}/\partial y$, despite the finite values of the mean horizontal velocity $\bar{v}$, are attributable to the very small horizontal gradients in temperature $\partial\bar{T}/\partial y$ (Mapes 2001). This also means that buoyancy is ineffective in this domain. Although some values for heating rates due to disturbing elements of horizontal V_3 and vertical V_5 velocities are shown, especially in the higher half of the latitudes, these values are not large and seem to have opposing signs that cancel each other out. Compared to the small values of these terms, heating rates due to mean vertical velocity V_4 and cooling rates due to net radiation V_6 are large. The radiative cooling field V_6 has a smooth latitude-height distribution. In contrast, the heating field due to mean vertical velocity includes the fine structures of many pin-shaped cooling domains. The model has 8000 grid points in the longitudinal di-

rection. At most grid points, subsidence on a typical order of 1 mm s^{-1} occurs. Cumulus convection with updrafts, typically on the order of 1 m s^{-1} , develops occasionally. Therefore, it is not surprising to see the pin-shaped disturbances in figure V_4 , provided that a latitudinal point is not rare, at which cumulus updrafts occur on a few grid points over the whole longitude. When the values V_4 are horizontally averaged over the region, the result is the smooth primary heating effect roughly balancing with the primary cooling effect of net radiation V_6 (Fig. 13 right bottom).

4.2 Thermodynamic diagnosis for the mean TMD flow

To explain the mechanism that produces the mean element of TMD flows, we apply a diagnostic approach similar to that of Iwasa et al. (2002, 2004) to a radiatively and convectively equilibrated atmosphere with simplified physical processes. The analysis is performed with the zonal mean atmosphere averaged over 2 days after 10-days integration, when the model is considered to be in thermodynamic equilibrium. Note that the subtropical region into which the TMD flows move is investigated, not the tropical region out of which they move.

Imagine a mean atmosphere $\mathbf{A}$, averaged over time t and longitude x , spreading over a two-dimensional latitude-height plane (y, z) in the subsidence region. Our objective is to obtain the steady flow field (V, W) (y, z) in this mean atmosphere. To do so, let us trace a mean air parcel $\mathbf{P}$ initially located at an arbitrary point (y, z) and having the properties of air at (y, z) in the mean atmosphere $\mathbf{A}$. First, we accept a fact from the model result:

fact-I:

The mean atmosphere $\mathbf{A}$ is settled in a state of local heat balance at every point, and this state will not change over time.

Next, we make two assumptions:

assumption-i:

The mean air parcel $\mathbf{P}$ moves only in a dynamically neutral manner. That is, it does not suffer from buoyancy and is not directly affected by convection in the tropical region or by baroclinic waves in the mid-latitude region.

assumption-ii:

All the thermodynamic effects acting on the moving mean air parcel $\mathbf{P}$ are adiabatic effects due to its vertical velocity W and effects due to net radiation. Thermodynamic effects due to other processes can be neglected.

Fig. 14. Latitude-height distributions in the mid-troposphere of (a) time- and zonal-mean net radiative heating rate $\langle \partial T / \partial t \rangle_{\mathbf{R}}$, (b) mean temperature lapse rate $\langle \Gamma \rangle$, (c) vertical velocity W diagnosed using Eq. (15) from (a) and (b), (d) time- and zonal-mean density $\langle \rho \rangle$, (e) vertical mass flux $\langle \rho \rangle W$ calculated from (c) and (d), and (f) vertical mass flux $\langle \rho w \rangle$ observed in the model. The values for (a) and (d) are averaged over 2-days after 10-days integration. The values for (b) are calculated from the approximated mean temperatures. See text.

These assumptions are not yet validated, although they seem plausible as shown in the previous subsection. The validity of these assumptions can be checked by how well the diagnosed results explain the model results.

By assumption-i, the equation of motion for the mean air parcel $\mathbf{P}$ can be dropped from the main discussion here. When the mean air parcel $\mathbf{P}$ moves vertically with velocity $W(y, z)$ for a small time δt , the temperature increase in the ambient mean atmosphere $\mathbf{A}$ is given by

$$\delta \langle T \rangle_{\mathbf{A}} = -\langle \Gamma \rangle W \delta t, \quad (11)$$

where $\langle \Gamma \rangle(y, z) = -\partial / \partial z [\langle T \rangle(y, z)]$ is the temperature lapse rate in the mean atmosphere $\mathbf{A}$. From assumption-ii, on the other hand, the only thermodynamic effects that a moving air parcel $\mathbf{P}$ experiences are those of the adiabatic process due to subsidence and those of the net radiative process. Therefore, temperature increase in the mean air parcel $\mathbf{P}$ during the same motion as that for Eq. (11) is given as

$$\delta \langle T \rangle_{\mathbf{P}} = \left(-\frac{g}{c_p} W + \left\langle \frac{\partial T}{\partial t} \right|_{\mathbf{R}} \right) \delta t, \quad (12)$$

Fig. 15. Latitude-height distributions in the mid-troposphere of (a) divergence of the diagnosed vertical mass flux $\partial(\langle \rho \rangle W) / \partial z$ calculated from Fig. 14e, (b) horizontal mass flux $\langle \rho \rangle V$ diagnosed using Eq. (17) from (a), and (c) time- and zonal-mean horizontal mass flux $\langle \rho v \rangle$ observed in the model. In (b), the vertical broken line indicates the end point $y = y_E = 22^\circ \text{N}$ of the integration in Eq. (17), and the distribution in latitudes higher than y_E is not shown. The values for (c) are averaged over 2-days after 10-days integration.

where $\langle \partial T / \partial t |_{\mathbf{R}} \rangle (y, z)$ is the mean net radiative heating rate. When and only when the temperature increase in the ambient mean atmosphere $\mathbf{A}$ is the same as that in the mean air parcel $\mathbf{P}$ as

$$\delta \langle T \rangle_{\mathbf{A}} = \delta \langle T \rangle_{\mathbf{P}}, \quad (13)$$

the local heat balance is sustained in the mean atmosphere $\mathbf{A}$ as specified in fact-I. Otherwise, net heating or cooling occurs locally, changing the temperature of the ambient mean atmosphere $\mathbf{A}$ over time and breaking fact-I. In contrast, for the mean air parcel $\mathbf{P}$, assumption-i is realized when and only when Eq. (13) is satisfied.

Otherwise, the mean air parcel $\mathbf{P}$ is not dynamically neutral but will have either positive (when $\delta\langle T \rangle_{\mathbf{A}} < \delta\langle T \rangle_{\mathbf{P}}$) or negative buoyancy (when $\delta\langle T \rangle_{\mathbf{A}} > \delta\langle T \rangle_{\mathbf{P}}$), which would break assumption-i. In this framework of discussion, therefore, fact-I and assumption-i are not independent of each other, but are equivalent to each other.

Using Eqs. (11) and (12), Eq. (13) leads to

$$-\langle \Gamma \rangle W = -\frac{g}{c_p} W + \left\langle \frac{\partial T}{\partial t} \middle|_{\mathbf{R}} \right\rangle. \quad (14)$$

Equation (14) can be interpreted as a thermodynamic equation for the mean atmosphere $\mathbf{A}$ that drops terms other than those of vertical advection, adiabatic heating, and radiative heating. We can then solve this equation for W as

$$W = \left(\frac{g}{c_p} - \langle \Gamma \rangle \right)^{-1} \left\langle \frac{\partial T}{\partial t} \middle|_{\mathbf{R}} \right\rangle. \quad (15)$$

At an arbitrary point (y, z) in the mean atmosphere $\mathbf{A}$, once the temperature lapse rate $\langle \Gamma \rangle$ and the net radiation heating rate $\langle \partial T / \partial t |_{\mathbf{R}} \rangle$ are given, the corresponding vertical velocity W is obtained from Eq. (15). Note that Eq. (15) does not include any direct dynamic forcings. Thus, W can be called a steady vertical velocity generated in the field of thermodynamically equilibrated mean atmosphere.

Figure 14a shows the latitude-height distribution of the net radiative heating rate averaged over time and longitude, which is clipped between latitudes 10°N and 25°N from Fig. 4c. All the values in the figure are negative, demonstrating that longwave radiative cooling dominates over shortwave radiation heating. Near the melting level, which decrease in height with latitude, strong radiative cooling is maintained over the whole latitudinal region in the figure (e.g., $-2.42 \times 10^{-5} \text{ K s}^{-1}$ at $y = 15^\circ\text{N}$, $z = 4 \text{ km}$). A corresponding distribution of the temperature lapse rate $\langle \Gamma \rangle$ is shown in Fig. 14b. Because data for time-mean temperature were not available at the analysis time, we used time-mean data of potential temperature and pressure to calculate approximated $\langle \Gamma \rangle$ values. Large temperature lapse rates are also found along the melting level (e.g., 4.19 K km^{-1} at 15°N , $z = 4 \text{ km}$). Figure 14c presents the distribution of the vertical velocity W diagnosed using Eq. (15) from the distributions shown above. The motion of the air parcel becomes subsidence ($W < 0$), providing excess adiabatic heating ($-(g/c_p - \langle \Gamma \rangle)W > 0$) that balances the net radiative cooling ($\langle \partial T / \partial t |_{\mathbf{R}} \rangle < 0$ as shown in Fig. 14a) in the temperature field ($g/c_p - \langle \Gamma \rangle > 0$ as suggested from Fig. 14b). The large values of the net

radiative cooling rate and the large temperature lapse rate near the melting level make the subsidence velocity W large there (e.g., -4.36 mm s^{-1} at $y = 15^\circ\text{N}$, $z = 4 \text{ km}$).

Multiplying the vertical velocity W by the mean atmospheric density $\langle \rho \rangle(y, z)$, shown in Figs. 14c and 14d, respectively, provides the diagnosed vertical mass flux $\langle \rho \rangle W$ (y, z) shown in Fig. 14e. The vertical mass flux $\langle \rho \rangle W$ is likely to take negative values that are smaller above and larger below a layer just above the melting level over the whole latitudinal region in the figure (e.g., at $y = 15^\circ\text{N}$, $-2.56 \times 10^{-3} \text{ kg m}^{-2} \text{ s}^{-1}$ at $z = 6 \text{ km}$, while $-3.62 \times 10^{-3} \text{ kg m}^{-2} \text{ s}^{-1}$ at $z = 4 \text{ km}$). For comparison, Fig. 14f shows time- and zonal-mean vertical mass flux observed in the model $\langle \rho w \rangle(y, z)$. The latter has a similar feature near the melting level, especially between 13°N and 17°N , although it has fine structures due to small-scale disturbance in the region of higher latitudes. The increases in absolute value of the downward mass flux across the layer just above the melting level mean that the layer is a domain of mass divergence. Figure 15a shows vertical mass divergence $[\partial(\langle \rho \rangle W) / \partial z](y, z)$ calculated from the diagnosed subsidence mass flux $\langle \rho \rangle W$ in Fig. 14e. A layer of mass divergence appears just above the melting level (e.g., $5.27 \times 10^{-7} \text{ kg m}^{-3} \text{ s}^{-1}$ at $y = 15^\circ\text{N}$, $z = 5 \text{ km}$). The height of the layer decreases with latitude from $z \sim 6 \text{ km}$ near the tropical region to $z \sim 5 \text{ km}$ in latitudes higher than 13°N . Domains of weak mass convergence lie above and below the layer (e.g., $\partial(\langle \rho \rangle W) / \partial z = -6.08 \times 10^{-9} \text{ kg m}^{-3} \text{ s}^{-1}$ at $y = 15^\circ\text{N}$ and $z = 8 \text{ km}$, and $\partial(\langle \rho \rangle W) / \partial z = -2.31 \times 10^{-8} \text{ kg m}^{-3} \text{ s}^{-1}$ at $y = 14^\circ\text{N}$ and $z = 3.25 \text{ km}$).

The local variations in the vertical mass flux, whether divergent or convergent, must be balanced with opposite local variations in the horizontal mass flux following to the equation of continuity for the mean atmosphere

$$\frac{\partial}{\partial y}(\langle \rho \rangle V) = -\frac{\partial}{\partial z}(\langle \rho \rangle W). \quad (16)$$

Here we include an additional assumption:

assumption-iii:

All the compensating flows in such low latitudes of the subtropical region are supplied from and absorbed into the tropical region of active cumulus convection. They do not flow from or into the mid-latitudinal region.

Again, assumption-iii is not yet validated, but will be checked when the diagnosed results have explained the model results, as well as other assumptions. Now, from

assumption-iii, integrating Eq. (16) poleward from y to an end point y_E provides horizontal mass flux at an arbitrary point (y, z)

$$(\langle \rho \rangle V)(y, z) = - \int_y^{y_E} \frac{\partial}{\partial z} (\langle \rho \rangle W)(\eta, z) d\eta \quad (17)$$

or the horizontal flow velocity

$$V(y, z) = - \frac{1}{\langle \rho \rangle(y, z)} \int_y^{y_E} \frac{\partial}{\partial z} (\langle \rho \rangle W)(\eta, z) d\eta. \quad (18)$$

The end point of integration y_E should be taken at the virtual end of the extent of horizontal mass outflow from the convective region of the tropics:

$$\langle \rho \rangle V = 0 \text{ at } y = y_E. \quad (19)$$

The end point y_E is regarded as the center of the subsidence field, at which the thermodynamically consistent circulations from the tropics and the mid-latitudes run against each other with horizontal flows of the same magnitude but of opposite direction. Provided that these circulations are not extremely distorted near y_E , we may take y_E at the point at which the diagnosed subsidence flow becomes locally minimum or maximum in the meridional direction. With the vertical velocity W given from Eq. (15), the corresponding horizontal velocity V can be determined as compensating flow through the atmospheric continuity as Eq. (18). The flow field $(V, W)(y, z)$ represented by the equation pair (15) and (18) is the solution for the problem presented at the beginning of this subsection, namely, the steady flow field in the mean atmosphere **A**.

Figure 15 shows distributions of (b) diagnosed horizontal mass flux $\langle \rho \rangle V$ obtained from Eq. (17) and (c) mean horizontal mass flux $\langle \rho v \rangle$ obtained in the model, respectively. The end point in Eq. (17) was taken at $y_E = 22^\circ\text{N}$, at which the diagnosed subsidence mass flux takes a local maximum value $\langle \rho \rangle W = -4.16 \times 10^{-3} \text{ kg m}^{-2} \text{ s}^{-1}$ in the meridional direction at the $z = 4 \text{ km}$ level as shown in Fig. 14e. This is indicated with a vertical broken line in Fig. 15b. The diagnosed maximum outflow at 10°N is then $\langle \rho \rangle V = 8.08 \times 10^{-1} \text{ kg m}^{-2} \text{ s}^{-1}$, which is similar to the mean model value $\langle \rho v \rangle = 9.60 \times 10^{-1} \text{ kg m}^{-2} \text{ s}^{-1}$. The height $z = 5 \text{ km}$ of the maximum outflow at 10°N also shows agreement between the figures. This level is determined from the distribution of the divergent layer in the subtropics shown in Fig. 15a. Through horizontal integration in Eq. (17), the outflow $\langle \rho \rangle V$ is contributed more from

the larger integration interval ($13^\circ\text{N} \leq y \leq 22^\circ\text{N}$) with lower levels ($z \sim 5 \text{ km}$) of the mass divergent layer than from the smaller integration interval ($10^\circ\text{N} \leq y \leq 13^\circ\text{N}$) with higher levels ($z \sim 6 \text{ km}$) of the layer.

On the other hand, the diagnosed mean TMD flow $\langle \rho \rangle V$ keeps its strength and reaches higher latitudes than the observed mean TMD flow $\langle \rho v \rangle$. The reason can be inferred from the difference in the distribution of the respective corresponding mean vertical mass fluxes (Figs. 14e and 14f). In deducing the diagnosed flow field, no cumulus convection was assumed in the subtropical region (Fig. 14e). However, updrafts were found in the subtropical region in the model atmosphere (Fig. 14f) presumably associated with cumulus convections, especially at 18°N and higher latitudes. The high-latitude subtropical convections are thought to generate their own flow field, which is likely to weaken the TMD flows of poleward direction and to result in the weaker observed mean TMD flow in higher latitude. Weak mean flow toward the equator below the melting level (about $z \sim 3.25 \text{ km}$ level) is also reproduced and is due to vertical mass convergence in the subsidence over the subtropical region. Unrealistic large detrainment flow is diagnosed at a lower (below 3 km) level (Fig. 15b). This is because the diagnosed flow field was derived by taking only radiative cooling into account. Given that the flow field at such a lower level is strongly affected by the mixing layer physics, the mean flow would be better reproduced in a similar way if we considered heating due to turbulent mixing.

The general coincidence between Figs. 15b and 15c suggests that the mechanism of the mean TMD flow in the model atmosphere can be understood from the thermodynamic discussion on the mean subsidence field in the subtropical region. At the same time, all the assumptions made in the discussion are now justified: The mean air parcels in the subtropical atmosphere do not suffer from dynamic forcings, but instead are dynamically neutral (assumption-i). Net heating of the mean air parcel is determined between adiabatic heating due to subsidence and net radiative cooling (assumption-ii). The mean TMD flow is constrained by the atmospheric continuity between the tropics and subtropics, being separated from the atmosphere in the middle latitudinal region (assumption-iii).

The radiative process plays a primary role in the mechanism. As Eq. (15) shows, the vertical velocity W is driven in proportion to the net radiative heating rate $\langle \partial T / \partial t |_{\mathbf{R}} \rangle$. The radiative effect may also be involved in determining the temperature lapse rate $\langle \Gamma \rangle$ in the subtropical atmosphere. The lapse rate $\langle \Gamma \rangle$ regulates the vertical motion in the form of the coefficient

Fig. 16. Latitude-height distributions in the mid-troposphere for the diagnosed (a) sensible heat flux $c_p \langle T \rangle V$ and (b) water vapor flux $\langle q_v \rangle V$, and those of the snapshots at 10-days integration for zonal-mean (c) sensible heat flux $c_p \bar{T} \bar{v}$ and (d) water vapor flux $\bar{q}_v \bar{v}$.

$(g/c_p - \langle \Gamma \rangle)^{-1}$. In Subsection 3.4, similar TMD flows were shown in another experiment assuming the same conditions except ignoring latent heat exchanges in water freezing and melting. The similar findings further support the mechanism described in the discussion that is not significantly affected from the latent heat exchanges.

4.3 Transports by the mean TMD flow

A physical value ϕ (representing sensible heat $c_p T$ and water vapor mixing ratio q_v) and a meridional velocity v are respectively divided into time- and zonal-mean elements (indicated with brackets) and disturbing elements (indicated with a prime) as

$$\begin{cases} \phi(t, x, y, z) = \langle \phi \rangle(y, z) + \phi'(t, x, y, z) \\ v(t, x, y, z) = \langle v \rangle(y, z) + v'(t, x, y, z). \end{cases} \quad (20)$$

Then, the net meridional flux $\langle F_\phi \rangle(y, z)$ for the physical value ϕ is written as

$$\langle F_\phi \rangle = \langle \phi v \rangle = \langle \phi \rangle \langle v \rangle + \langle \phi' v' \rangle. \quad (21)$$

We now examine the extent to which the diagnosed mean element $V(y, z)$ of the TMD flows contributes to

the net meridional transports in the middle troposphere. This is investigated only with the first term on the right-hand side of Eq. (21), replacing $\langle v \rangle$ with V , as

$$F_{\phi \text{diag}}(y, z) = \langle \phi \rangle V. \quad (22)$$

The transports $F_{\phi \text{diag}}$ (Eq. 22) due to the mean TMD flow element should be compared to the exact net transports $\langle F_\phi \rangle$ (Eq. 21). Unfortunately, however, we did not calculate time-mean fluxes of sensible heat and water vapor. Thus, the zonal mean distribution

$$\overline{F_\phi}(t, y, z) = \overline{\phi v}, \quad (23)$$

where the overline indicates the zonal mean snapshot at $t = 10$ days, is shown for reference to create rough estimates of the net fluxes. Note that the latter fluxes are not averaged over time, and they are not exactly the net fluxes. Hence, detailed comparisons are meaningless between $F_{\phi \text{diag}}$ and $\overline{F_\phi}$.

Figure 16 shows the diagnosed fluxes of (a) sensible heat $F_{\text{Sdiag}} = c_p \langle T \rangle V$ and (b) water vapor $F_{\text{Vdiag}} = \langle q_v \rangle V$, together with the zonal-mean flux snapshots at 10-days integration of (c) $\overline{F_S} = c_p \bar{T} \bar{v}$ and (d) $\overline{F_v} = \bar{q}_v \bar{v}$. The maximum values in the diagnosed transports appear at $y =$

10°N and $z = 5 \text{ km}$, $F_{\text{Sdiag}} = 3.31 \times 10^5 \text{ W kg}^{-1} \text{ m}$ and $F_{\text{Vdiag}} = 5.19 \times 10^{-3} \text{ m}^{-2} \text{ s}^{-1}$, respectively. The maximum values in the snapshot zonal-mean transports are $\overline{F_S} = 3.04 \times 10^5 \text{ W kg}^{-1} \text{ m}$ and $\overline{F_V} = 6.41 \times 10^{-3} \text{ m}^{-2} \text{ s}^{-1}$, respectively. These values are from the same level of $z = 5 \text{ km}$ but from a higher latitude of $y = 13.5^\circ\text{N}$ compared with the diagnosed values. This is because TMD flows near 10°N at the snapshot time were weaker than those at other times. Although minor differences exist between $F_{\phi\text{diag}}$ and $\overline{F_\phi}$, the orders of the values coincide well between them. The disturbing elements of the TMD flows, which are one order larger than the mean element (as shown in Fig. 12b), might be expected to make large contributions to the net meridional transports. However, having both poleward and equatorward directions, their transporting effects seem to cancel each other when averaged. Finally, what remains in the net transports is the contribution of the mean TMD flow element.

5. Concluding remarks

By time integration of the low-latitude atmospheric over an aqua planet using the non-hydrostatic model RAMS, TMD flows were obtained and shown to flow from the tropics to the subtropics taking the maximum value near 10° latitude. This study focused on the time- and zonal-mean structure of the subtropical middle troposphere and investigated steady flow in the mean atmospheric field that is locally settled in a thermodynamic equilibrium. It appeared that it is possible to rebuild a consistent mean flow field including the mean TMD flow from free subsidence of zero-buoyancy mean air parcels suffering from radiative cooling in the locally equilibrated mean temperature field. The key mechanism is a vertically inhomogeneous thermal structure around the melting level. Locally, near the melting level, the strong radiative cooling and large temperature lapse rate result in the balancing subsidence velocity being larger than that in the above domain. The velocity difference generates a layer of vertical mass divergence just above the melting level, which induces compensating air flow horizontally from the tropical region. It is interesting that the moderate free subsidence of the subtropical atmosphere supplies the mechanism controlling the mean TMD flow. Forcing of direct circulation concerning tropical convection does not extend to the mean free atmosphere in the subtropics. In that view, cumulus convection in the tropical region (mainly its vertical transporting effect) is needed only for indirect effects such as for convective adjustment and to retain mass, heat, and water vapor in the tropical region.

The previous studies cited in the introduction, both

observations and model research, have treated TMD flows in limited time intervals and areas. Their objects were TMD flows as temporarily and spatially local components in direct circulation of convection. Certainly, the traditional dynamics-based approach is effective for phenomena such as disturbances occurring in the tropics. In this study, on the other hand, the object was the mean TMD flow averaged over time and space. At such large scales, fine structure convective circulation is averaged out, and instead, regulation of thermodynamic balance works predominantly. The difference of this study from the previous studies suggests that we may need to alter our perspective on the effects of cumulus convection depending on the temporal and spatial scales of interest. When we discuss net outflow from convective atmosphere into non-convective atmosphere, we should emphasize the flow-inducing mechanism of the thermodynamic balance outside the convection as presented by Iwasa et al. (2002, 2004) and in this study. Because the mean flow accompanies net transports of sensible heat and water vapor, the mean flow and its mechanism should be indispensable for discussing the feedback effects of cumulus convection on environmental changes such as global warming.

Although the mechanism was derived from the time- and zonal-mean field of the model atmosphere, it is also expected to work in the real atmosphere under the following conditions: there is a wide enough subsidence-predominant region adjacent to another region of active convection, and thermodynamic effects similar to those in the model atmosphere work in the former region.

This study did not investigate the causes of the strong radiative cooling or of the large temperature lapse rates near the melting level. One of the possible candidates for the strong radiative cooling is the water distribution in the atmosphere (Liou 2002), especially the distribution of water vapor and liquid cloud shown in Subsection 3.1. The amount of water was small in the model atmosphere above the melting level, but large below that level (Fig. 3). The difference in water distribution quite presumably cause a jump in optical depth, and hence, the strong radiative cooling near the melting level. Furthermore, the strong radiative cooling may be involved with generation of the large temperature lapse rate change. This situation is harmonious with the thin stratiform cloud just below the TMD flows that has been reported in recent studies (Yasunaga et al. 2006; Zuidema et al. 2006; Yasunaga et al. 2008), if we assume that strong radiative cooling occurs near the thin cloud layer. Detailed investigation is needed to clarify the cause of the vertical thermal structure near the melting level.

There is no guarantee that a state similar to that of the model atmosphere will occur in the real atmosphere. The aqua planet boundary condition is assumed to have supplied more water vapor into the model atmosphere, making it more humid than the real atmosphere. As the water content above the melting level is very small, the existence of more water in the model atmosphere would result in a more definite contrast in the vertical profile of water content, and hence, stronger radiative cooling near the melting level. On the other hand, even if the current real atmosphere does not exhibit the mean TMD flow, it is possible that future warmed environments will increase atmospheric water amounts and produce the mean TMD flow by the mechanism presented in this study. When the mean TMD flow appears, it can effectively transport water vapor, a greenhouse gas, into the dry subtropics and could contribute to accelerating global warming. Future studies should investigate these issues.

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