A Total Flow Perspective of Atmospheric Mass and Angular Momentum Circulations: Boreal Winter Mean State

Ming Cai and Chul-Su Shin

Department of Earth, Ocean and Atmospheric Science, The Florida State University, Tallahassee, Florida

(Manuscript received 14 June 2013, in final form 1 February 2014)

ABSTRACT

This paper reports a comprehensive diagnostic analysis of mass and angular momentum (AM) circulations and their budgets in boreal winter using the 32-yr daily NCEP–Department of Energy (DOE) reanalysis (1979–2010). The diagnosis is performed using instantaneous total flows before taking time and zonal average without decomposition of time mean and transient flows and separation of zonal mean and wavy flows. The analysis reveals that embedded in a broad hemispheric thermally direct meridional mass circulation in each hemisphere are three distinct but interconnected thermally direct meridional cells. They are the tropical Hadley cell, the stratospheric cell, and the extratropical zonally asymmetric Hadley cell. The tropical Hadley cell corresponds to the Hadley cell of the classic three-cell model whereas the extratropical Hadley cell and the stratospheric cell correspond to the eddy-driven extratropical residual circulation. The joint consideration of meridional mass and AM circulations helps to substantiate Hadley’s original view that the hemispheric-wide thermally direct meridional circulation can have broad surface easterly in the tropics and westerly in the extratropics. Because the mass circulation cannot have a net divergence anywhere in long time mean and the earth’s AM decreases toward the poles, the companion AM transport in the equatorward cold air branch inevitably has to be divergent. The downward transfer of westerly AM to the cold air branch by the pressure torque associated with westward tilted baroclinic waves dominates such divergence in the extratropics, explaining the prevailing surface westerly there. In the tropics and polar region where the meridional circulation is nearly zonally symmetric, the dominance of this divergence results in a surface easterly there.

1. Introduction

One of the main goals of the theory for the atmospheric general circulation is to explain the meridional variation of the zonal mean surface wind pattern, as was first attempted by Hadley (1735). The topic of the climatological mean of the zonal mean zonal flow is the core component of the atmospheric general circulation theory (e.g., Lorenz 1967; Schneider and Lindzen 1977; Held and Hou 1980; Johnson 1989; and see Lindzen 1990 for a succinct review on this topic). Studies about the atmospheric general circulation have traditionally been focused on the time and zonal mean atmospheric flow in pressure coordinates (e.g., Lorenz 1967; Oort and Peixoto 1983; Peixoto and Oort 1992). The mass streamfunction in pressure coordinates shows a classic three-cell pattern from the equator to the pole—namely, the Hadley cell in the tropics, thermally indirect Ferrel cell in midlatitudes, and weak polar cell. The three-cell model is useful to explain the zonal mean surface zonal wind pattern by the Coriolis deflection of the zonal mean meridional wind. However, such mass streamfunction is derived from the time mean and zonal mean velocity field without considering the zonal variation of the atmospheric mass and, therefore, it represents the time mean velocity circulation rather than mass circulation.

The eddy-driven meridional circulation in the extratropics can be obtained through the transformed Eulerian mean (TEM) formulation that was first proposed by Andrews and McIntyre (1976). In the framework of TEM formulation, the convergence of Eliassen–Palm (EP) fluxes associated with baroclinic waves acts as not only a “drag” that causes deceleration of the westerly jet but also (adiabatic) forcing that drives a meridional residual circulation through the “downward control principle” (Haynes et al. 1991; Holton et al. 1995; Haynes 2005). Although in the TEM theory diabatic heating/cooling

Denotes Open Access content.

Corresponding author address: Dr. Ming Cai, Department of Earth, Ocean and Atmospheric Science, The Florida State University, 1017 Academic Way, Tallahassee, FL 32306.

E-mail: mcai@fsu.edu

DOI: 10.1175/JAS-D-13-0175.1

© 2014 American Meteorological Society
also plays a role in the residual circulation, the majority of studies place emphasis on the role of eddies in giving rise to the residual circulation (e.g., Edmon et al. 1980; Andrews et al. 1987; Haynes et al. 1991; Holton et al. 1995; Tanaka et al. 2004; Haynes 2005 and references therein). It is the adiabatic warming–cooling in the sinking–rising branch of the residual circulation that leads to the temperature above–below its radiative equilibrium in high–low latitudes. This in turn causes a reduction in the meridional temperature gradient and acts to maintain the thermal wind balance with the change in the westerly jet. However, the same eddies also drive a thermally indirect circulation, besides the thermally direct residual circulation. The sum of them is the net meridional (velocity) circulation driven by eddies. The middle latitude portion of the net meridional circulation driven by eddies corresponds to the Ferrel cell, which is weak and thermally indirect. In addition, despite its apparent success in explaining the climate variability in the zonal mean westerly jets in the upper troposphere and polar stratosphere, the theory of the residual circulation does not explain the prevailing surface westerly wind in the extratropics as the three-cell circulation theory.

Johnson and his collaborators (e.g., Gallimore and Johnson 1981; Townsend and Johnson 1985; Johnson 1989) have diagnosed the time and zonal mean mass circulation in isentropic coordinates. By explicitly taking the zonal variation of atmospheric mass into consideration, they showed a single thermally direct circulation cell in each hemisphere that links the heat source in the tropics to the heat sink in high latitudes. They have further linked the angular momentum (AM) circulation that is driven by both diabatic heating and force to the mass circulation between the equator and the poles. Since then, the mass circulation in isentropic coordinates has been analyzed in many studies (e.g., Juckes et al. 1994; Held and Schneider 1999; Schneider 2004; Schneider et al. 2006). More recently, Pauluis et al. (2008, 2010) presented the time and zonal mean moist isentropic mass circulation, highlighting water vapor transport in mid-latitudes. Pauluis et al. (2011) proposed a statistical TEM (STEM) formulation to easily analyze the meridional mass circulation.

We here follow the theoretical framework of Johnson and his collaborators and expand their diagnostic analysis that was mainly based on a few years of data. We have diagnosed diabatic heating–cooling and its associated cross-isentropic mass and AM fluxes, meridional mass and AM fluxes, cross-isentropic downward transfer of westerly AM by the pressure torque, the surface frictional torque, and the mountain torque using the daily National Centers for Environmental Prediction–Department of Energy (NCEP–DOE) Reanalysis II dataset (Kanamitsu et al. 2002). All of these fields are independently and explicitly calculated from the instantaneous total flow without the decomposition of time mean and transient flows and without separation of zonal mean and wavy flows. By doing this, their averages (time mean, and/or zonal mean, and/or vertical mean) reflect the total transport in the intervals of the averaging.

The primary objective of this study is to delineate the roles of instantaneous and continuous couplings among diabatic heating–cooling, meridional mass and AM transport, the pressure torque, the surface frictional torque, and the mountain torque in giving rise to the atmospheric meridional mass and AM circulations. The total flow approach helps to gain a generic conceptual and principle-based understanding on the nature of the atmospheric general circulation and the climatological meridional profile of the zonal mean surface zonal wind in a single framework. The physical principles that are explicitly applied in our conceptual explanation include the conservation principles of mass and AM, hydrostatic and geostrophic balance, and the baroclinic instability theory.

The rest of this paper is organized as follows: Section 2 presents the diagnostics equations in isentropic coordinates. Data and analysis procedure are summarized in section 3. Section 4 depicts salient features of the mean meridional mass and AM circulations in boreal winter [December–February (DJF)]. Mass and AM transport along the poleward and equatorward branches of the DJF-mean meridional mass circulation are examined in section 5. Section 6 is devoted to explain the origins of the surface westerly wind in the extratropics and easterly wind in the tropics by considering the mass and AM circulations jointly. The main findings and the total flow view of the atmospheric general circulation are summarized in section 7.

2. Diagnostics equations

The conservation equations for mass and AM in isentropic coordinates are

\[
\frac{\partial m}{\partial t} + \frac{\partial (mu)}{\partial l} + \frac{\partial (mv \cos \phi)}{\partial \phi} + \frac{\partial (m\theta)}{\partial \theta} = 0 \quad \text{and} \quad (1)
\]

\[
\frac{\partial (mA)}{\partial t} + \frac{\partial (mAu)}{\partial l} + \frac{\partial (mAv \cos \phi)}{\partial \phi} + \frac{\partial (mA\theta)}{\partial \theta} = -m \frac{\partial M}{\partial l} + mF_a \cos \phi, \quad (2)
\]

Where \(\lambda\) is longitude, \(\phi\) is latitude, \(\theta\) is potential temperature, \(a\) is the earth’s radius, and \(u\) (\(v\)) is zonal (meridional) wind. The quantity \(A = (\Omega a \cos \phi + u)a \cos \phi\) is
the (absolute) AM per unit mass with $\Omega = 2\pi \text{ day}^{-1}$ and $m = -(1/g)\partial p/\partial \theta$ is the air mass per unit volume ($g = 9.8 \text{ m s}^{-2}$ is the gravitational acceleration and $p$ is pressure in unit of Pascal). Also, $\theta$ denotes the diabatic heating rate, $M$ is the Montgomery potential, and $F_d = -(1/m)\partial \omega /\partial \theta$ with $\omega$ being the zonal component of frictional stress on isentropes. Note (1) is for dry atmospheric mass only because the source (evaporation) and sink (precipitation) terms of atmospheric moisture are not included in (1).

Let us consider a longitudinal tube placed in the isentropic layer between two adjacent isentropic surfaces ($\theta_1$, $\theta_2$) and between two latitudes ($\phi_1$, $\phi_2$). The zonal integral of an arbitrary variable $X$ over the longitudinal tube is defined as

$$[X] = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} X \cos \phi \, d\theta \, d\phi \, d\lambda,$$

(3)

where $[\cdot]$ is used to define quantities along the longitudinal tube. Also following Johnson and Downey (1975) and Johnson (1980), the mass-weighted zonal average of $X$ along the tube is defined as $X = [mX]/[m]$.

Using (3), the zonally integrated form of (1) over the tube is

$$\frac{\partial}{\partial t} [m] = (F_{\theta, \phi_1}^{\text{ad}} - F_{\theta, \phi_2}^{\text{ad}}) + (F_{\theta, \phi_1}^{\text{d}} - F_{\theta, \phi_2}^{\text{d}}),$$

(4)

where

$$F_{m, \phi}^{\text{ad}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} (mv) a \cos \phi \, d\theta \, d\lambda \quad \text{and}$$

$$F_{m, \phi}^{\text{d}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} (m\theta) a^2 \cos \phi \, d\theta \, d\lambda,$$

(5)

$F_{m, \phi}^{\text{ad}}$ represents adiabatic mass fluxes, corresponding to the total northward meridional mass transport along the isentropic layer crossing latitude $\phi$ between $\theta_1$ and $\theta_2$ while $F_{\theta, \phi}^{\text{d}}$ is diabatic mass fluxes, equaling the total upward vertical mass transport crossing isentropic surface $\theta$ between $\phi_1$ and $\phi_2$. Therefore, (4) states that temporal change of air mass in the longitudinal tube is equal to the sum of convergences of adiabatic and diabatic mass fluxes into the tube.

The zonal integration of (2) over the longitudinal tube yields

$$\frac{\partial}{\partial t} [mA] = (F_{\phi, \theta_1}^{\text{ad}} - F_{\phi, \theta_2}^{\text{ad}}) + (F_{\phi, \theta_1}^{\text{d}} - F_{\phi, \theta_2}^{\text{d}})$$

$$+ (T_{\phi, \theta_1}^{\text{pres}} - T_{\phi, \theta_2}^{\text{pres}}) + (T_{\phi, \theta_1}^{\text{fric}} - T_{\phi, \theta_2}^{\text{fric}}),$$

(6)

where

$$F_{\phi, \theta}^{\text{ad}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} (mA') a \cos \phi \, d\theta \, d\lambda;$$

$$F_{\phi, \theta}^{\text{d}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} (M a) a^2 \cos \phi \, d\theta \, d\lambda;$$

$$T_{\phi, \theta}^{\text{pres}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} \left( -p \frac{\partial z}{\partial \lambda} \right) a^2 \cos \phi \, d\theta \, d\lambda;$$

$$T_{\phi, \theta}^{\text{fric}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} \left( \tau a \cos \phi \right) a^2 \cos \phi \, d\theta \, d\lambda;$$

(7)

$F_{\phi, \theta}^{\text{ad}}$ represents the AM transport associated with the meridional mass transport crossing latitude $\phi$ between $\theta_1$ and $\theta_2$ while $F_{\phi, \theta}^{\text{d}}$ is the AM transport associated with the diabatic mass transport crossing isentropic surface $\theta$ between $\phi_1$ and $\phi_2$. The other terms in (6) correspond to the pressure torque ($T_{\phi, \theta}^{\text{pres}}$) and frictional torque $T_{\phi, \theta}^{\text{fric}}$ that cause AM exchange among adjacent isentropic layers or between the atmosphere and the ground surface without involving direct mass exchanges. In deriving (6), we have utilized (Tung 1982; Andrews 1983; Andrews et al. 1987; Egger et al. 2007; Egger and Hoinika 2008)

$$\left[ -m \frac{\partial M}{\partial \lambda} \right] = T_{\phi, \theta_1}^{\text{pres}} - T_{\phi, \theta_2}^{\text{pres}},$$

(8)

where $T_{\phi, \theta}^{\text{pres}}$ is given by the third equation of (7). Positive $T_{\phi, \theta}^{\text{pres}}$ corresponds to a gain of westerly AM in the layer above $\theta$ and a loss in the layer below $\theta$ and vice versa. In the interior atmosphere, the pressure torque is associated with vertically tilted waves (e.g., Johnson 1989). Under the hydrostatic balance, air in the layer above $\theta$ exerts an eastward torque to air below if the longitudinally alternating sloped isentropic interfaces tilt westward with height. Therefore, westward tilting waves act to transfer westerly AM downward ($T_{\phi, \theta}^{\text{pres}} < 0$) and vice versa. Note that the atmospheric pressure torque in isentropic coordinates is equivalent to the vertical EP flux in the TEM formulation (Andrews 1983; Andrews et al. 1987; Juckes et al. 1994; Tanaka et al. 2004). By adopting the Lorenz convention (1955), one can easily show that for the lowest layer above the ground, $T_{\phi, \theta_1}^{\text{pres}}$ in Eq. (8) corresponds to the mountain torque, $T_{\phi, \theta_1}^{\text{pres}}$.

$$T_{\phi, \theta_1}^{\text{pres}} = \int_0^{\theta_2} \int_{\phi_1}^{\phi_2} \left( -p s \frac{\partial z}{\partial \lambda} \right) a^2 \cos \phi \, d\theta \, d\lambda,$$

(9)

where $\theta_{\text{surface}}$, $p_s$, and $z_s$ are the surface potential temperature, pressure, and topography, respectively. Similarly, in the lowest layer, $T_{\phi, \theta_1}^{\text{fric}}$ in (6) represents the
levels running from the Montgomery potential, and temperature at the 15 isentropic levels (360, 385, 425, 500, 600, and 800 K) and pressure, Monte-
guy fluxes have the same units (Hadley; 1 Hadley = 1018 J) as their divergences and all torque terms.

By writing the mass continuity in the form of (4)–(5), both diabatic and adiabatic mass fluxes have the same units (kg s\(^{-1}\)) as their divergences. And by (6)–(10), AM fluxes have the same units (Hadley; 1 Hadley = 1018 J) as their divergences and all torque terms.

**3. Data and analysis procedures**

The data used in this study are derived from the NCEP–DOE reanalysis II, which provides daily spectral analyses on sigma levels and surface potential temperature, pressure, and wind stress from 1 January 1979 to 31 December 2010 (Kanamitsu et al. 2002). The sigma level data are interpolated onto 2.5° x 2.5° grids with zonal and meridional winds at the 16 full isentropic levels (θ = 235, 245, 255, 265, 275, 285, 295, 305, 320, 340, 360, 385, 425, 500, 600, and 800 K) and pressure, Montgomery potential, and temperature at the 15 isentropic levels running from θ = 240 to 650 K, which are roughly at middle points between two adjacent full isentropic levels. Because the layer mass and its meridional mass and AM fluxes are defined at the full isentropic levels, we refer to the other 15 isentropic levels as the “(interior) interface levels.” The diabatic heating rate (θ, the total heating rate equaling the sum of radiative, latent, and subgrid mixing heating rates) is diagnosed using the residual method (Nigam 1994; Chan and Nigam 2009) from the sigma level analysis and is interpolated into the 15 interface levels.

For the sake of brevity, we refer readers to Shin (2012) for the otherwise straightforward finite difference methods but with some tedious details used in evaluating each term in (4)–(10). It suffices just to highlight the following three special procedures that have been applied on the daily instantaneous fields:

(i) Isentropic surfaces intersect with the earth’s surface when their potential temperatures are lower than \(\theta_{\text{surface}}\). As a result, longitudinal tubes bordered by these isentropic surfaces are interrupted by the surface. To apply the zonal integral (3) to an isentropic longitudinal tube that intersects with the ground, we follow the Lorenz convention (Lorenz 1955) as in the literature (e.g., Andrews 1983; Johnson 1989; Juckes et al. 1994; Held and Schneider 1999) by allowing the isentropic surface that intersects with the ground to continue below the ground. Because the underground portion of an isentropic surface is assigned with the surface pressure and its elevation, the underground portion of a longitudinal isentropic tube, by definition, has no air mass. As a result, the underground portion of the longitudinal isentropic tube has no contribution to the zonal integral of each term in (4) and (6) along the isentropic layer that intersects with the ground.

(ii) The instantaneous daily fields of mass, \(F_{m,\theta}\), and \(F_{A,\theta}\), are defined at full isentropic levels whereas \(F_{m,\theta}^d\), \(F_{A,\theta}^d\), \(T_{\text{pres}}\), and \(T_{\text{fric}}\) are defined at interface levels. In addition to the 15 interface levels, there are two boundary levels in the vertical. The top boundary level is symbolically denoted \(\theta_{\text{top}}\), representing the surface where air pressure is zero (i.e., \(\theta_{\text{top}} > 800 \text{ K}\)). The top boundary condition is that

\[
F_{m,\theta_{\text{top}}}^d = F_{A,\theta_{\text{top}}}^d = T_{\text{pres}}|_{\theta_{\text{top}}} = T_{\text{fric}}|_{\theta_{\text{top}}} = 0.
\]

The bottom boundary level is symbolically denoted as \(\theta_0\). By the Lorenz convention, \(\theta_0\) is an imaginary isentropic surface that is always under the ground, or \(\theta_0 < \min[\theta_{\text{surface}}(\lambda, \phi, \ell)]\), but whose pressure, elevation, and frictional torque are always equal to their values at the surface. When applying Eq. (8) to the lowest isentropic layer whose bottom interface level is \(\theta_0\), one can easily show that the pressure torque on \(\theta_0\) surface is indeed equal to the mountain torque defined in (9). Therefore, the bottom boundary condition is that

\[
F_{m,\theta_0}^d = F_{A,\theta_0}^d = T_{\text{pres}}|_{\theta_0} = T_{\text{fric}}|_{\theta_{\text{surface}}} = 0.
\]

(iii) All terms on the right-hand side of (4) and (6) are evaluated from instantaneous fields and therefore they contribute to instantaneous local changes of mass and AM in longitudinal tubes. The local tendency terms on the left-hand side of (4) and (6), however, are evaluated using forward finite difference scheme with data on two successive days, representing daily-mean changes of mass and AM in longitudinal tubes. As a result, neither (4) nor (6) are in balance on daily basis. In addition, errors in diagnosing these terms on the right-hand side of (4) and (6) also contribute to the imbalance in daily budgets. To obtain self-consistent budget analyses,
we apply the method of Lagrange multiplier to minimize the imbalance in daily budgets. The additional constraints via the Lagrange multipliers include (i) local balance of mass and AM, (ii) no net accumulation of mass and AM over each isentropic layer globally due to adiabatic processes, (iii) no net accumulation of mass in the vertical due to diabatic processes, and (iv) requiring that the net AM accumulation in the vertical due to diabatic AM fluxes and frictional torque is balanced with the surface frictional torque. It is confirmed that the self-consistent adjusted fluxes of mass and AM are highly correlated with their original values and the residual is at least three order of magnitude smaller than the smallest term in (4) and (6) on daily basis.

From the instantaneous daily meridional mass and AM fluxes defined at full isentropic levels \( \theta_k \), we can calculate the vertically integrated total mass and AM fluxes across latitude \( \phi \) at time \( t \) in all layers above each of the interface levels,

\[
\Psi_m(\phi, \theta_k, t) = \sum_{\theta_{top}}^{\theta_k} F_{m,\phi}(\phi, \theta', t) \quad \text{and} \quad \Psi_A(\phi, \theta_k, t) = \sum_{\theta_{top}}^{\theta_k} F_{A,\phi}(\phi, \theta', t),
\]

where \( \theta_k \) is one of the 15 interface levels or at \( \theta_0 \) with \( \theta_{top} > \theta_k > \theta_0 \). In the literature, \( \Psi_m(\phi, \theta_0, t) \) is also referred to as the mass streamfunction because its meridional derivative would be exactly equal to the vertical mass flux after a long time average. However, for instantaneous flows, the meridional derivative of \( \Psi_m(\phi, \theta_0, t) \) at each level, representing the vertically integrated convergence of the meridional mass transport in all layers above, is not equal to the vertical mass flux crossing that level. Because of the existence of sink–source terms of AM at the earth’s surface that are not along a fixed isentropic surface, the meridional derivative of \( \Psi_A(\phi, \theta_0, t) \), even after a long time average, is not equal to the sum of the vertical AM flux and the vertical transfer of AM via the pressure/mountain and frictional torques in the layers near the boundary. Therefore, the meridional derivative of \( \Psi_A(\phi, \theta_0, t) \) at each level cannot be used to infer the sum of vertical AM fluxes and vertical AM transfer by the torque terms even after a long time average, although it still represents the vertically integrated convergence of the meridional AM transport in all layers above. For this reason, we cannot call \( \Psi_A(\phi, \theta_0, t) \) an AM streamfunction. Although the vertically integrated meridional AM flux—in height or sigma–pressure coordinates at the surface level where surface sinks–sources of AM are defined is always at a fixed coordinate surface (i.e., \( z = 0 \) or \( \sigma = 1 \)—can be called an AM streamfunction (e.g., Oort and Peixoto 1983; Egger and Hoinka 2011; Mak 2011).

In the remaining sections, we will only present the DJF-mean fields. At each latitude, the time averages of the instantaneous fields in the layer bordered by the lowest interior interface level that has data and \( \theta_0 \) (referred to as the surface layer) and at the bottom level \( \theta_0 \) are done along the surface layer and along the \( \theta_0 \) surface, respectively. Because \( \theta_0 \) represents the surface via the Lorenz convention, we have data at \( \theta_0 \) and in the surface layer at all times. For other (full or interface) isentropic levels, the time averages of all instantaneous flux terms are done along isentropic surfaces with zero values assigned to these instantaneous fields at the times when their isentropic layer becomes the surface layer (for the full isentropic levels) or below the full isentropic level where the surface layer is defined (for the interface levels). We note that, for time averages of vertically integrated mass and AM fluxes, their instantaneous values at the lowest interface level are always assigned at the level below the surface layer where instantaneous adiabatic mass and AM fluxes are always zero. By such a proper time averaging procedure, which extends the Lorenz convention to the time domain, the commutative properties of the time mean and vertical differential–integral (difference–summation) are retained. For example, the time mean vertically integrated mass (or AM) fluxes can be obtained from the vertical integral of the time mean meridional mass (or AM) fluxes and the vertical differential can be used to obtain the time mean mass (or AM) fluxes from the time mean vertically integrated mass (or AM) fluxes. Moreover, such time mean average ensures not only that the time mean pressure torque at an interface level (including \( \theta_0 \)) is equal to the vertical summation of the time mean of the net pressure torque in each layer from the top layer to the layer above that interface level, but also that the time mean pressure torque along \( \theta_0 \) is the time mean mountain torque, just as an instantaneous field. The same can be said to the time mean of the diabatic AM fluxes and frictional torque.

---

\(^1\) Note that, in doing so, we have implicitly assumed that the pressure torque and mountain torque estimated using (8) and (9) are accurate. Also we have combined the terms \( F_{m,\theta} \) and \( T_{\text{fric}} \) into a single term \( (F_{m,\theta} + T_{\text{fric}}) \) because we have no way to directly estimate \( T_{\text{fric}} \) in the atmosphere. Since \( |T_{\text{fric}}| \ll |F_{m,\theta}| \) in the atmosphere, we still refer to this combined term as the diabatic AM flux. At the surface, it is the surface frictional torque.
4. DJF-mean meridional mass and angular momentum circulations

Because the June–August (JJA)-mean pattern is similar to the DJF-mean pattern except the reversal between the two hemispheres, we will use the phrases Northern Hemisphere (NH) and Southern Hemisphere (SH) with winter hemisphere (WH) and summer hemisphere (SH) interchangeably. To examine the association of the time mean zonal flow with the meridional mass and AM circulations, we also plot the DJF-mean mass-weighted zonal mean zonal wind as contours in Figs. 1–4. The zonal mean zonal wind exhibits three axes of maximum westerly wind in each hemisphere. One of them is tilted poleward upward from the subtropical jet to the polar stratosphere. The second axis slopes poleward downward from the subtropical jet level to a lower isentropic level at the Pole. The third axis, although less noticeable, is almost vertical below 300 K, connecting to the maximum surface westerly wind in midlatitudes.

In each hemisphere, poleward mass fluxes are dominant from the upper troposphere to the stratosphere, whereas equatorward mass fluxes are dominant mainly in the lowest tropospheric layers (Fig. 1a). Poleward mass fluxes reach local maxima on the equator side parallel to the axes of maximum westerly jet. The equatorward mass fluxes start from the pole in the lower troposphere of each hemisphere and the tropical portion of equatorward mass fluxes in the WH crosses the equator toward the SH and connects the poleward mass fluxes above the equatorward mass fluxes in the SH.

The tropical tower of upward mass fluxes is rooted at equatorial latitudes in the SH and splits into two distinct branches in the stratosphere toward the extratropics in each hemisphere (Fig. 1b). Outside the tropics, the diabatic mass fluxes are downward except in the lowest tropospheric layers where they are upward. It is the diabatic heating that lifts the air in the equatorward cold air branch from lower to higher potential temperature layers as it moves toward the tropics. In each layer, the maximum downward mass flux is found on the poleward side of the maximum westerly flow, showing a poleward-upward tilting from the subtropical jet to the polar stratosphere and a poleward-downward tilting from the subtropical jet level to the polar lower troposphere.

To depict the meridional mass circulation as a whole, streamlines of adiabatic and diabatic mass fluxes are plotted in Fig. 2, which has not been shown in the literature. Overall, there is a hemispheric wide cell of thermally direct mass circulation in each hemisphere. The poleward warm air branch of the meridional mass circulation connects the diabatic heating region in the tropics to the diabatic cooling region in the extratropics while the equatorward cold air branch brings air mass back to the diabatic heating region in the tropics from the cooling region in high latitudes. The broad hemispheric cell of meridional mass circulation consists of three distinct but interconnected cells—namely, the tropical Hadley cell (THC), the stratospheric cell (SC), and the extratropical Hadley cell (EHC). As to be shown shortly, both the EHC and SC are accompanied with strong pressure torque, and therefore they are zonally asymmetric meridional circulations. The weak pressure torque in the tropics indicates the THC is a zonally quasi-symmetric circulation. In the WH subtropics, the air in

![Fig. 1. Time and zonal mean isentropic mass circulation (shadings, $1 \times 10^9$ kg s$^{-1}$) and zonal wind (contours, m s$^{-1}$) in DJF for 32 yr from 1979 to 2010. (a) Adiabatic mass fluxes along the isentropic surfaces and (b) diabatic mass fluxes across the isentropic surfaces. The green colored shading, which is not included in the color bar, corresponds to zero values, indicating locations where mass fluxes equal zero exactly. Blank areas indicate the mass-less layers under the ground with the borderline between color shading and blank areas defined by the minimum DJF surface potential temperature in these 32 yr. Thick black lines roughly follow the boundary lines between the northward and southward air branches (see section 5 for details).]
the warm air branch has two different pathways. The downward pathway in the subtropical troposphere corresponds to the descend portion of the THC, which then is connected to the equatorward cold air branch of EHC. The poleward pathway corresponds to the warm air branch of the SC, which is part of the Brewer–Dobson circulation (BDC) in the winter hemisphere. The downward mass fluxes of the SC are connected to the warm air branch of the EHC. Over the polar latitudes, the downward mass fluxes are dominant over entire stratosphere and join the downward portion of the EHC, connecting to the cold air branch of the EHC. The cold air branch of the EHC joins its counterpart originated from the THC, forming the continuous equatorward mass fluxes from the pole to the equator in the lowest tropospheric layers.

We next discuss the companion meridional AM circulation associated with the meridional mass circulation. It is found that the meridional AM fluxes (Fig. 3a) are positively correlated with the meridional mass fluxes (Fig. 1a) at nearly all latitudes and all levels. The few exceptions are found outside the tropics and in the regions where mass fluxes are equatorward. Note that the poleward air mass fluxes carry larger AM and the equatorward air mass fluxes bring smaller AM. Hence, for the zonally asymmetric meridional circulation, the zonal integrated meridional AM fluxes can still be poleward even
when the zonal integrated mass fluxes are equatorward. The poleward transport of AM at each layer in the warm air branch is the strongest on the equator side of the maximum westerly wind and then decreases rapidly on the pole side of the maximum westerly wind. This indicates that the latitudinal positions of the westerly jets coincide with the convergence zones of the poleward AM transport in the warm air branch associated with the THC, EHC, and SC.

Shown in Fig. 3b (shadings) is the DJF-mean pattern of diabatic AM fluxes defined as the sum of the AM fluxes associated with diabatic mass transport and frictional AM transfer between two adjacent layers. Because the frictional AM transfer is much smaller except in the lowest tropospheric layer, diabatic AM fluxes tend to closely follow diabatic mass fluxes. Because of the accumulation of the westerly AM from the poleward air mass transport, the poleward moving warm air of the THC slows down and starts to move eastward in the subtropics. As it circulates along subtropical latitude circles, the warm air originated from the tropics cools via thermal radiation there, sinking downward crossing isentropic surfaces. The companion downward transport of AM in the subtropics acts to remove the accumulation of the westerly AM in the THC warm air branch, which in turns helps to maintain the poleward air mass transport of the THC. There exists another pair of upward and downward transport of AM from midlatitudes to high latitudes associated with the EHC. Similarly, because of the thermal radiative cooling, air mass sinks downward crossing the isentropic surfaces as it circulates around the polar cap following the stratospheric polar jet, giving rise to strong downward

**FIG. 3.** As in Fig. 1 but for (a) adiabatic and (b) diabatic angular momentum fluxes (Hadley). Note that the latter is the sum of angular momentum transport associated with diabatic mass fluxes and vertical angular momentum transfer by the frictional torque.

**FIG. 4.** (a) DJF-mean zonal-mean pressure torque (shadings, Hadley). Negative shadings represent the downward transfer of westerly angular momentum, and vice versa. (b) As in Fig. 2 but for angular momentum streamlines (Hadley) and vertically integrated angular-momentum fluxes (shadings, Hadley). Note that at the surface, the vertically integrated angular-momentum flux is not zero even for the climatological annual mean condition, corresponding to the total meridional angular-momentum flux in the entire atmospheric column at each latitude. Contours are the DJF-mean zonal-mean zonal wind.
transport of westerly AM on the poleward side of the stratospheric polar jet in the WH.

Shown in the lowest level of Fig. 3b is the DJF-mean surface frictional torque. It is seen that the surface frictional torque tends to be negative in the extratropics where lies westerly at the surface, acting to transfer westerly AM from the stratosphere to the earth. In the tropics-subtropics and polar region where lies easterly surface wind, the positive surface frictional torque acts to add westerly AM to the atmosphere.

Figure 4a shows the vertical transfer of AM in the atmosphere by the pressure torque in DJF, which is in a good agreement with a recent study by Egger and Hoinka (2008) in terms of both spatial pattern and intensity even though the two results are derived from different datasets and for the different periods. Inside 30° of latitude, the pressure torque is relatively weak. The small amplitude of pressure torque indicates a lack of wave activities in the tropics. As a result, the vertical transfer of AM in the THC is mainly through diabatic processes. In the extratropics of both hemispheres, the pressure torque is negative from the stratosphere to the lowest tropospheric layers, indicating the downward transfer of westerly AM associated with westward tilted waves. In the extratropical troposphere, the maximum axis of negative pressure torque lies nearly along the boundary that separates the warm air branch from the cold air branch of the EHC. The downward transfer of westerly AM by the pressure torque not only alleviates the accumulation of westerly AM in upper layers associated with the net poleward mass transport but also supplies the needed westerly AM to the equatorward moving air mass that is along the upgradient direction of the earth’s AM. As the downward transport of westerly AM associated with the downward diabatic mass fluxes, the downward transfer of westerly AM by the pressure torque also helps to speed up the poleward advancement of warm air mass in upper levels but to slow down the poleward advancement of warm air below owing to the additional accumulation of westerly AM in comparison with air mass above. This explains why the axes of maximum westerly flow, maximum meridional mass, and AM fluxes are all tilted poleward and upward in the warm air branch.

In the WH tropics-subtropics and the both polar areas, the mountain torque acts to supply westerly AM to the cold air branch of the meridional mass circulation that is in the upgradient direction of the earth’s AM. In summer tropics and subtropics, the mountain torque acts to remove westerly AM from the cold air branch. Despite being along the upgradient direction of the earth’s AM, the midlatitude portion of the cold air branch in both hemispheres has excessive westerly AM owing to the downward AM transfer via the pressure torque and downward diabatic AM transport from the warm air branch—a fraction of which is removed by the mountain torque.

Color shadings in Fig. 4b represent \( \Psi_A(\phi, \theta_0, t) \), the DJF mean of the vertically integrated AM flux defined in the second equation of (11). Because of the positive correlation of diabatic mass and AM fluxes above the boundary layer, the spatial pattern in Fig. 4b is similar to that in Fig. 2. Although \( \Psi_A(\phi, \theta_0, t) \) cannot be regarded as the AM streamfunction as we explained in section 3, one should still be able to compare \( \Psi_A(\phi, \theta_0, t) \) in layers above the ground with the time mean AM streamfunction defined in height or sigma-pressure coordinate. The meridional AM circulation in the WH is much stronger and broader than that in the SH. Its maximum is ~420 Hadley at 315 K in the northern tropics while the minimum center is located in the southern subtropics with the value of ~160 Hadley, which are comparable to those in Oort and Peixoto (1983) and Mak (2011).

The meridional AM circulation can be succinctly and completely depicted from AM streamlines (Fig. 4b). Unlike the classical three cells of AM (and mass) streamfunction in height or sigma-pressure coordinates (e.g., Oort and Peixoto 1983; Mak 2011), the AM streamlines exhibits a thermally direct cell of AM circulation, as the isentropic mass circulation. Above the ground, the meridional component AM streamlines is the meridional AM flux in each layer and the vertical component is the sum of the AM flux companied with the diabatic mass flux, frictional torque, and pressure torque crossing each isentropic surface. At the surface, the meridional component by definition is zero and the vertical component is the sum of the mountain torque and surface frictional torque. The three distinct but interconnected thermally direct cells within the hemispheric wide thermally direct cell are also vividly identifiable from AM streamlines. In the extratropical upper troposphere, the streamlines of AM circulation are more downward than those of mass circulation mainly because of intense downward transfer of westerly AM by the pressure torque. The most striking difference between the meridional mass and AM circulations is found in the lowest layer
(representing the ground surface). The meridional mass circulation is closed in the atmosphere without direct mass transport from the earth (if we ignore the evaporation and precipitation). However, owing to the surface frictional torque and mountain torque, there is a net westerly AM transfer from the atmosphere to the earth in midlatitudes and from the earth to the atmosphere in the tropics, subtropics, and polar regions.

5. Mass and angular momentum circulations in warm and cold air branches

In this section, we wish to quantitatively describe the meridional mass circulation and its companion AM circulation in the warm and cold air branches in both hemispheres as a whole. The boundaries of the warm and cold air branches are defined at the levels that roughly follow the border lines separating northward mass fluxes from southward ones (heavy black lines in Fig. 1). In each hemisphere except in the tropics–midlatitudes of the SH, we can loosely divide the atmosphere into two layers: one is a warm air branch and the other is a cold air branch. The mass circulation of the WH warm air branch is toward the winter pole (WP) and that in the SH is toward the summer pole (SP) whereas the cold air branch in both hemispheres is toward the equator. The atmosphere over the tropics/midlatitudes of the SH can be loosely divided into three layers. The top layer is the beginning portion of the WH warm air branch, referred to as the SH portion of the WH warm air branch. The bottom layer is the continuation of the equatorward cold air branch of the SH. The mass circulation in the middle layer connects the WH equatorward cold air branch to the SH poleward warm air branch.

a. Meridional mass circulation

We begin with the winter hemisphere, which exhibits stronger and broader meridional circulations than the summer hemisphere. As shown in green bars of Fig. 5a, the WH warm air branch effectively starts from the SH tropics (in DJF, it is ~20°S), although its beginning can be further traced back to the beginning point of the SH BDC. The WH-bound warm air transport is driven by diabatic heating that regulates air mass (red curve) from the SH portion of the WH cold air branch below (blue bars). The continuous supply of air mass to the warm air branch (positive portion of the red curve) by heating air mass in the cold branch accelerates the WH-bound warm air transport. Soon after air mass in the warm air branch crosses the equator into the WH, it begins to suffer diabatic cooling. The diabatic cooling results in a loss of warm air mass to the cold air branch, causing the slowdown of the poleward mass transport in the warm air branch in the WH subtropics. In DJF, the warm air branch loses air mass to the cold air mass branch via diabatic cooling over the WH subtropics (5°–30°N) at a rate of ~9.3 × 10^{10} kg s^{-1}, which is 54.7% of that it acquires from the cold air branch in the SH tropics (~17 × 10^{10} kg s^{-1}). As the remainder part of air mass gained in the SH tropics in the warm air branch continues its journey toward the WP, the diabatic heating associated with the EHC pumps additional air mass from the cold air branch over the latitude span of (30°–40°N) at a rate of ~3.2 × 10^{10} kg s^{-1}. This leads to the second maximum of poleward mass transport at 40°N (~10.2 × 10^{10} kg s^{-1}). The two maxima of poleward mass fluxes correspond to the two maximum centers of vertically integrated mass fluxes shown in Fig. 2. The total amount of air mass that is brought down diabatically back to the cold air branch over the latitude span covering the poleward side of the maximum of the WH poleward warm air branch is ~1.14 × 10^{11} kg s^{-1}. At any given latitude, the equatorward mass flux in the cold air branch nearly equals the poleward mass flux in the warm air branch, which also nearly equals the total upward mass flux from the cold air branch measured from the beginning of the warm air branch.

The SH meridional mass circulation (Fig. 5b for the DJF mean) consists of a poleward warm air branch above and equatorward cold air branch below, similar to the NH except that the strength is much weaker. The SH cold air branch ends at the equator, but its warm air branch is connected to the WH cold air branch. The total mass flux into the SH warm air branch from the NH cold air branch in DJF is ~1.57 × 10^{11} kg s^{-1}. Most of the mass entering the SH from the WH is subject to intensive diabatic heating in the SH tropics. As a result, the total mass flux that goes back to the WH from the SH via the WP-bound WH warm air branch is nearly identical to that entering the SH. The convergence of the SH cold air mass fluxes in the tropics also feeds to the tropical portion of the SH warm air branch via diabatic heating. Between 20°S and 0°, the total air mass fluxes from the SH cold air branch into its warm air branch are ~7.4 × 10^{10} kg s^{-1}, which is nearly identical to the mass fluxes toward the SP via the SH portion of the THC. Therefore, the strength of the THC mass circulation is ~23.1 × 10^{10} kg s^{-1}, 68% of which is associated with the WH circulation and only 32% with the SH circulation. In the SH subtropics, the SH warm branch of the THC loses mass to its cold air branch diabatically at a rate of 2.3 × 10^{10} kg s^{-1} (over 20°–40°S). The relative weak cooling over the SH subtropics compared to that over WH subtropics also means that a large percentage of the mass flux of the SH THC (65%) continues poleward into the SH extratropics than that in the WH, which is
FIG. 5. DJF-mean mass transport in both the warm and cold air branches of the meridional mass circulation for the period as in Fig. 1 in the (a) winter hemisphere and (b) summer hemisphere (see text for details). Green (blue) bars represent poleward (equatorward) mass fluxes in the warm (cold) air branch in each hemisphere. Red curves are for diabatic mass fluxes from the warm to cold air branches. Red bars represent accumulated diabatic mass transport into the warm air branch starting from the beginning of warm air branch in each hemisphere.
45.3\% for the WH THC, although the absolute value over the SH (4.8 \times 10^{10} \text{ kg s}^{-1}) is only 62\% of that over the WH (7.7 \times 10^{10} \text{ kg s}^{-1}). The mass transport associated with the SH EHC is also less intense. For example, the upward mass fluxes to the warm air branch in mid-latitudes are \sim 2.7 \times 10^{10} \text{ kg s}^{-1} in the SH but \sim 3.2 \times 10^{10} \text{ kg s}^{-1} in the WH. The total downward diabatic mass transport to the cold air branch in the SH high latitudes are \sim 5.7 \times 10^{10} \text{ kg s}^{-1}, which is a half of that in the WH. The equatorward mass transport in the cold air branch of the SH EHC is less than half of that in the WH.

There is a small but coherent imbalance of meridional mass fluxes between the warm and cold air branches in each hemisphere in DJF. As indicated in Fig. 6, air mass in the cold air branch tends to increase at all latitudes in the WH at the expense of air mass loss in the warm air branch except in the tropics where it loses air mass to the warm air branch. In the SH, however, air mass in the warm branch has a positive tendency in all latitudes at the expense of loss of air mass in the cold air branch except in the summer polar region where air mass in the cold air branch increases. Over the WH high latitudes, the loss of air mass in the warm air branch is slightly less than that the gain of air mass in the cold air branch below, owing to the relatively faster convergence of the warm air branch than the divergence of the cold air branch, giving rise to the positive tendency of surface pressure there. In the subtropics and midlatitudes of the WH (20^\circ–47.5^\circ N), the surface pressure falls because the convergence of the warm air branch is relatively weaker than the divergence of the cold air branch. The same situation is found in the SH except the pattern of negative surface pressure tendency in low latitudes and positive surface pressure tendency is shifted equatorward compared to the NH.

b. Meridional angular momentum circulation

The upward diabatic mass transport in the heating place and downward mass transport in the cooling place drives an opposite meridional mass transport between the warm and cold air branches. Such meridional mass transport has to cross constant surfaces of the earth’s
AM. To overcome the AM inertia for continuous meridional mass circulation, the air mass that moves poleward needs to get rid of its excessive westerly AM whereas the equatorward moving air mass needs to gain AM. Therefore, the strength of the meridional mass transport is also dictated by both the removal rate of the westerly AM from the poleward mass transport and the addition of the westerly AM to the equatorward mass transport.

According to Fig. 7, the air mass in the SH portion of the WH warm air branch gains westerly AM from the air mass below via diabatic heating. In DJF, the maximum total poleward fluxes of westerly AM by the WH warm air branch is found the latitude of 58N (420 Hadley), where the WH warm air branch begins to lose westerly AM to the poleward mass transport and the addition of the westerly AM to the equatorward mass transport.

As the warm air mass moves poleward, it continues to lose westerly AM to the cold air mass branch diabatically until it merges with the warm air branch of the WH EHC (30N, in DJF). The total amount of the westerly AM that is removed from the tropical portion of the WH warm air branch per unit time is ~247 Hadley. Because of the lack of large-scale baroclinic waves, the downward transfer of westerly AM by the pressure torque is only ~8 Hadley in the tropics. In the extratropical portion of the WH warm air branch where the poleward mass transport is done mainly by westward tilted baroclinic waves, the pressure torque plays a major role in removing westerly AM from the warm air branch. Between the subtropics and WP, the downward transfer of westerly AM by the pressure torque from the warm to cold air branch is ~188.3 Hadley, which is nearly six times greater than the downward diabatic AM transport there (~31 Hadley).

6. Latitudinal variation of the time mean zonal mean surface zonal wind

In the literature, the time mean zonal mean surface zonal wind has been explained from the vertically integrated AM budget (e.g., Lorenz 1967; Oort and Peixoto 1983; Egger et al. 2007). The vertical integral of meridional AM fluxes from the top of the atmosphere to the surface equals the net meridional transport of AM
crossing each latitude (black dashed line in Fig. 8a), which is qualitatively consistent with that in Fig. 11b of Oort and Peixoto (1983). It is seen that in each hemisphere, there is a net poleward transport of westerly AM by the atmosphere over all latitudes except the polar latitudes where the net transport of AM by the atmosphere, although very small, is equatorward. This feature is in contrast with the mass circulation in which the poleward mass transport is nearly identical to the equatorward mass transport at each latitude (and become exactly identical for a long time average over all seasons). In the extratropics, the convergence of the net AM transport by the atmospheric circulation (blue bars in Fig. 8a) is nearly balanced with the transfer of AM to the earth by the surface frictional torque (orange bars) and mountain torque (dark brown bars). On the equator side of maximum poleward transport of AM, the divergence of the net AM by the atmospheric circulation is compensated by the supply of AM from the earth to the atmosphere via the surface frictional torque with minor contribution from the mountain torque. Over the polar region where the vertical integrated AM is not only equatorward but also divergent, the surface friction torque supplies the needed AM there. From the meridional pattern of the surface frictional torque, we immediately conclude that the zonal mean surface zonal wind would have to be easterly in the tropics and westerly over the vast area of the extratropics except over the polar region where it would be weakly easterly. However, such explanation for the meridional pattern of surface zonal wind is still not completely causal because now one needs to answer a new question: Why is the vertically integrated meridional AM circulation divergent in the tropics and convergent in the extratropics? Or why is the net atmospheric AM transport dominated by the poleward transport in each hemisphere?

We next attempt to demonstrate that the meridional pattern of the time mean zonal mean surface zonal wind can be qualitatively determined by considering the mass and AM circulations jointly. Because the climatological mean of the vertical integrated meridional mass circulation has to be zero by the mass conservation principle, we have to consider the mass circulation in the warm and cold air branches, respectively; and the subscript W_to_C stands for the flux from warm air branch to cold air branch when the term inside the bracket is positive.

The equation for the long time mean of the AM budget in the cold air branch is

$$-rac{1}{a \cos \phi} \frac{\partial (mv \cos \phi)}{\partial \phi} \bigg|_{C} - \frac{1}{a \cos \phi} \frac{\partial (mv \cos \phi)}{\partial \phi} \bigg|_{W} - \frac{(-m \theta)}{W_{\text{to}, C}} = 0,$$

where the overbar stands for the long time average; the bracket [ ] was defined in (3); the subscripts W and C stand for the vertical integral over all layers in the warm and cold air branches, respectively; and the subscript W_to_C stands for the flux from warm air branch to cold air branch.

As indicated in (12), the diabatic mass transport between the warm and cold air branches can balance the divergence–convergence of meridional mass transport in each branch in a long time mean. However, its companion diabatic AM transport cannot do so for the depletion–accumulation of westerly AM associated with the meridional mass transport. The poleward decreasing profile of the earth’s AM alone always implies a net divergence of the equatorward AM transport in the cold air branch, as indicated by the second term in (13). This extra divergence in the AM transport by the nondivergent
FIG. 8. (a) DJF-mean angular-momentum budget for the whole atmosphere for 32 yr from 1979 to 2010. Dashed line represents net northward transport of angular momentum (scale on right); and positive (negative) values indicate mountain and surface frictional torques corresponding to gain (loss) of westerly angular momentum in the atmosphere. (b) DJF-mean angular momentum budget in the cold air branch associated with five different processes. Positive (negative) values indicate accumulation (removal) of westerly angular momentum in the cold air branch.
meridional circulation in the cold branch alone causes a depletion of westerly AM there.

For a zonally quasi-symmetric circulation such as the THC, (13) can be approximated as

$$T_{\text{fric}} \big|_S \approx [mu]_C \frac{1}{a} \frac{\partial A}{\partial \phi} C > 0.$$  \hfill (14)

In deriving (14), we have invoked the approximation that for a zonally quasi-symmetric circulation, $[mA] \approx [A][m\theta]$ and

$$\left[ \frac{A}{a \cos \phi} \frac{\partial (mu \cos \phi)}{\partial \phi} \right] \approx [A] \left[ \frac{1}{a \cos \phi} \frac{\partial (mu \cos \phi)}{\partial \phi} \right],$$

and then made use of (12). We also have utilized the facts that the downward transfer of westerly AM by the pressure torque is zero for a zonally symmetric meridional mass circulation and that the friction in the surface boundary layer is always much greater than that in the interior atmosphere; namely, $\|(-T_{\text{fric}})_S \| \gg \|(-T_{\text{fric}})_{W_{\text{to},C}} \|$. The right-hand side of (14) is the net divergence of the equatorward AM transport in the cold air branch of a zonally symmetric nondivergent meridional mass circulation. It follows that the surface wind in the tropics would have to be easterly so that the surface frictional torque can balance out the extra divergence in the equatorward AM transport associated with the zonally quasi-symmetric THC. Note that the right-hand side of (14) approximately equals the Coriolis torque on the meridional mass flux—that is, $-\left[ mu/a \times \frac{\partial A}{\partial \phi} \right] \approx a \cos \phi (2\Omega \sin \phi [mu])$—when the relative AM is ignored. Therefore, the alternative way to interpret (14) is that for a zonally symmetric circulation, there has to be a positive surface frictional torque to balance the negative Coriolis torque on the equatorward mass transport in the cold air branch (e.g., Johnson 1989; Held and Schneider 1999; Schneider 2006).

Over the equatorial region where the meridional gradient of the earth’s AM is close to zero, the net AM divergence associated with mass circulation is also close to zero. Then the mountain torque, although it is small, together with a small residual in the net AM divergence, determines the sign of the frictional torque over the equatorial region, explaining why in an aquaplanet GCM model, the time mean surface wind over the equatorial region could be a weak westerly instead of a weak easterly (see Fig. 9).

Equation (14) is also applicable over the polar region where the EHC becomes more zonally symmetric circulation and the role of the pressure torque diminishes rapidly inside the polar circle (purple bars in Fig. 8b) as the circumference of latitude circles approaches zero toward the pole. This explains the existence of a weak easterly wind in polar latitudes, which yields necessary positive frictional torque to compensate the unavoidable depletion of westerly AM in the zonally quasi-symmetric equatorward mass transport over the polar latitudes.

For a zonally asymmetric meridional circulation, such as the EHC in the extratropics, we need to consider the contributions of waves to the angular momentum budget besides the extra AM divergence term in the equatorward cold air branch. Because of baroclinic instability, atmospheric wave motions in the extratropics are dominated by westward tilted baroclinic waves. Being in hydrostatic and geostrophic balance, such westward tilted baroclinic waves do all of these three things at the same time—namely, (i) a net poleward mass and AM transport in the warm air branch, (ii) a net equatorward mass and AM transport in the cold air branch, and (iii) downward transfer of AM via the pressure torque. As the zonally symmetric part, the equatorward mass transport can always be balanced by the diabatic mass transport from the warm to cold air branch air branch such that there is no net divergence of the equatorward mass transport in the cold branch. Therefore, the extra divergence term in the companion equatorward AM transport still exists although it is weaker compared to that under the THC because the zonally asymmetric meridional mass circulation itself is weaker. This contributes to a depletion of the westerly AM in the cold air branch despite of no net mass divergence/convergence there. By the hydrostatic balance, pressure on the west side of the zonally slopping isentropic surface is higher than that on the east side. This results in a downward transfer of westerly AM from the warm to cold air branch (purple bars in Fig. 8b) while AM
is transported around by the wave-driven EHC. Over most part of the EHC except in the polar latitudes where the meridional circulation is zonally quasi symmetric, the downward transfer of westerly AM by the pressure torque is greater than the loss of westerly AM due to the divergence of the equatorward AM transport in the cold air branch. The mountain torque (black bars in Fig. 8b) due to longitudinal variation of surface pressure over large-scale topography in the NH midlatitudes acts to remove only a fractional portion of the gain of the accumulation of westerly AM in the cold air branch by the pressure torque there. Such a net accumulation of westerly AM by nonfrictional processes in the extratropical portion of the cold air branch has to be removed by the surface frictional torque, which is possible only when the surface zonal wind is westerly.

7. Summary

In this study, we follow the theoretical framework of Johnson and his collaborators and extend their diagnostic analysis that was based on a few years of data (Gallimore and Johnson 1981; Townsend and Johnson 1985; Johnson 1989). Specifically, we have diagnosed diabatic heating/cooling and its associated diabatic mass and absolute angular momentum (AM) transport, meridional (adiabatic) mass and AM transport, downward transfer of westerly AM by the pressure torque, the surface frictional torque, and the mountain torque using daily NCEP–DOE Reanalysis II dataset for 32 years from 1 January 1979 to 31 December 2010. All of these fields are independently and explicitly calculated from the total flow using daily data. The explicit diagnosis of these fields related to the mass and AM circulations from instantaneous total flows without artificial decomposition of time mean and transient flows and without separation of zonal mean and wavy flows enables us to gain a better understanding on the atmospheric general circulation and its companion AM transport and budget in a single framework.

Streamlines of the mean meridional mass circulation reveal that there are three distinct but interconnected thermally direct meridional circulation cells and they are embedded in one broad hemispheric thermally direct meridional mass circulation in each hemisphere. They are the tropical Hadley cell, the stratospheric cell (part of Brewer–Dobson circulation in the winter hemisphere), and the extratropical zonally asymmetric Hadley cell. The tropical Hadley cell corresponds to the Hadley cell of the classic three-cell model whereas the extratropical zonally asymmetric Hadley cell and the stratospheric cell correspond to the so-called eddy-driven residual circulation in the extratropics.

The global meridional mass and AM circulations can be succinctly summarized as a pair of the poleward warm air branch above and equatorward cold air branch below in each hemisphere. The warm air branch is connected to the cold air branch via diabatic heating in the equatorial region and in the subtropics–midlatitudes and the cold air branch is connected to the warm air branch via diabatic cooling in the subtropics and in the extratropics. The two hemispheric meridional overturning circulations are interconnected over the summer hemisphere tropics. The winter hemisphere equatorward cold air branch connects to the summer hemisphere warm air branch, although most of the air mass coming to the summer hemisphere from the winter hemisphere cold air branch is brought back to the winter hemisphere warm air branch via diabatic heating associated with the tropical Hadley cell centered in the summer hemisphere.

The joint consideration of the meridional mass and AM circulations of the total flow helps to gain a generic conceptual understanding on the nature of the atmospheric general circulation based on the first principles. In particular, the total flow approach enables us to substantiate Hadley’s original view that the hemispheric wide thermally direct meridional circulation between the equator and the poles can have broad surface easterly in the tropics and westerly in the extratropics. The physical principles that are explicitly applied in our conceptual explanation of the key features of the atmospheric general circulation include the conservation of mass and AM, hydrostatic and geostrophic balance, and the baroclinic instability theory.

The poleward decreasing solar forcing drives a hemispheric-wide meridional mass circulation that moves warm air mass aloft poleward and cold air mass below equatorward. The warm and cold air branches are connected by diabatic heating in the tropics and in subtropics–midlatitude and cooling over the vast area of the subtropics and extratropics, completing the hemispheric wide meridional mass circulation. Because the energy used to heat the cold air coming from high latitudes is carried away by the poleward warm air branch to high latitudes, there exists a net radiative energy flux surplus in the tropics and a deficit in high latitudes at the top of the atmosphere.

Such hemispheric-wide thermally direct meridional mass circulation is zonally quasi symmetric only in the tropics–subtropics. The diabatic heating associated with intensive convections in the tropics lifts air to the warm air branch. On its poleward journey, the conservation of its (total) AM makes air mass move more toward the east than toward the pole. As a result, the poleward moving warm air begins to converge in subtropical latitudes. The converged air recycles back to the cold air
branch via radiative cooling, causing no net mass accumulation there in long time mean. However, because the earth AM decreases toward the poles, the convergence of the poleward AM transport in the warm air branch is much larger than the mass convergence. As a result, only part of the accumulation of westerly AM in the subtropics can be removed by the downward air mass transport to the cold air branch below via radiative cooling, which is responsible for a net large accumulation of westerly AM in the subtropical portion of the warm air branch. Because of the small viscosity of air, it requires a much stronger zonal wind to transfer the remaining accumulation of westerly AM from the warm air branch to the cold air branch via frictional process. This explains the formation of the subtropical westerly jet at the upper level. The formation of the subtropical westerly jet by itself means that the zonally symmetric meridional mass circulation cannot prevail beyond the subtropics since warm air mass carried upward and poleward by tropical convections just keeps circulating the subtropical latitudes while waiting for being moved downward via radiative cooling.

The continuous radiative cooling in high latitudes due to much weaker solar forcing there and the warmness (in reference to the local solar energy) in the subtropics due to energy transported from the tropics act to strength the meridional temperature gradient, which powers baroclinically amplifying waves whose wind and pressure fields are nearly in geostrophic balance. According to baroclinic instability theory, baroclinically amplifying waves have to tilt westward with height. Being in hydrostatic and geostrophic balance, these westward tilted waves inevitably cause a net poleward transport of warm air mass and its AM in upper levels and a net equatorward transport of cold air mass and its AM in lower levels, giving rise to the zonally asymmetric meridional mass and AM circulations in the extratropics. The diabatic heating (latent heat) generated by slantwise convection associated with westward tilted baroclinic waves and radiative cooling connect the two branches of the zonally asymmetric meridional mass circulation. The extratropical zonally asymmetric meridional mass circulation is distinct but interconnected with the tropical zonally quasi-symmetric meridional because part of the warm air mass (20%–40%) that circles subtropical latitudes is pulled toward the extratropical portion of the warm air branch by these waves. Again, conservation of (total) AM of the poleward moving air makes it move more toward the east than toward the pole. This, together with the rapid poleward meridian convergence, slows down the poleward mass fluxes. Moreover, the rapid meridian convergence in high latitudes causes a convergence of the poleward moving warm air not only directly but also indirectly by limiting physical space for waves because of much shorter latitude circles in polar regions. The convergence of the poleward moving warm air is redirected to the equatorward cold air branch via diabatic cooling. Again, because the earth AM decreases toward the poles, the convergence of the poleward AM transport by the warm air branch is much larger than the mass convergence. As a result, only part of the accumulation of westerly AM in the extratropical portion of the warm air branch can be removed by the downward air mass transport to the cold air branch below via radiative cooling. Unlike the zonally symmetric circulation, the same westward tilted baroclinic waves that drive the extratropical mass circulation can transfer westerly AM downward via pressure torque on the zonally alternating sloping isentropic surfaces without any exchange of air mass in the vertical. This leads to a relatively smaller residual of the accumulation of westerly AM in the extratropical portion of the warm air branch in comparison with the subtropics, which is then transferred downward via frictional process. Given the same viscosity, a smaller accumulation of westerly AM in the extratropics means a slower westerly jet there in comparison with the subtropics.

The poleward decreasing of the earth’s AM alone means a net divergence of westerly AM transport in the equatorward cold branch even with no net mass accumulation anywhere in the cold air branch. In the tropics, the lack of zonal asymmetry means that both the pressure torque and mountain torque are very weak. As a result, the dominance of the divergence in the equatorward AM transport by the nondivergent meridional mass circulation due to the poleward decreasing the earth’s AM, or the Coriolis torque, is responsible for the prevailing surface westerly there, as originally postulated by Hadley (1735).

Over the equatorial region where the meridional gradient of Earth’s AM is close to zero, the net AM divergence associated with the equatorward mass circulation is also close to zero. Then the mountain torque, although it is small, together with a small residual in the net AM divergence, determines the sign of the frictional torque over the equatorial region, explaining why in an aquaplanet GCM model, the time mean surface wind over the equatorial region can be a weak westerly instead of a weak easterly.

The downward transfer of westerly AM throughout the extratropical atmosphere column by westward tilted baroclinic waves via the pressure torque exceeds the divergence of the equatorward AM transport due to the poleward decreasing the earth’s AM in the extratropical portion of the cold air branch. This results in a net accumulation of westerly AM by nonfrictional processes in the extratropics. Such accumulation of westerly AM in the extratropical cold air branch has to be removed
mostly by negative surface frictional torque because the mountain torque is much weaker, implying a surface westerly wind in the extratropics. By the geometric constraint of the much shorter latitude circle over polar region, the polar portion of the extratropical Hadley cell becomes zonally quasi symmetric again. As a result, the depletion of westerly AM in the polar portion of the equatorward cold air branch due to the meridional gradient of Earth’s AM has to be supplied by positive surface frictional torque, implying a weak surface easterly wind over polar latitudes.

Acknowledgments. This research was in part supported by research grants from the Chinese Ministry of Science and Technology (2010CB951600), the National Science Foundation (ATM-0833001), the NOAA CPO/CPPA program (NA10OAR4310168), and the DOE Office of Science Regional and Global Climate Modeling (RGC) program (DE-SC0004974). The authors are grateful to the two anonymous reviewers for their insightful and constructive comments.

REFERENCES


