The seasonal cycle of redistribution of atmospheric mass between continent and ocean in the Northern Hemisphere

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Using NCEP/NCAR and ERA-40 reanalyses, we studied the seasonal cycle of redistribution of air mass between continents and oceans over the Northern Hemisphere. Our results demonstrate that air mass in the Northern Hemisphere shifts clearly between continents and oceans when the season cycles. In July, the air mass reaches its lowest over Eurasia and its highest over the Pacific, and the opposite occurs in January. However, a different scenario is observed over the north Atlantic; the accumulated air mass reaches its maximum there in May. The maintenance of the accumulation or loss of air mass in a region is found to be related to the areal mean air mass flux divergence and the difference between precipitation and evaporation in an air column. The zonal-vertical circulations change with season, with the air ascent and decent reversed between land and sea. Besides, there also exists a noticeable difference of water vapor content of the air between continents and oceans, and this difference is season-dependent. Physically, the vapor content is able to significantly affect the atmosphere in absorbing solar short- and earth’s long-wave radiations, hence influencing atmospheric thermal conditions. The land-sea thermal contrasts inclusive of the diabatic heating rate changes their signs with season going on, resulting in the reversal of orientations of the temperature gradient. These thermal forcings not only facilitate the formation of the monsoons but also indirectly induce the seasonal cycle of the air mass exchanging over regions between continents and oceans.

atmospheric mass, seasonal cycle, redistribution between continents and oceans, land-sea thermal contrast


Atmospheric mass (AM) serves as an important quantity characterizing the changes in atmospheric general circulations. The AM distributions above the earth surface are believed to result from the changes of the atmospheric circulations in our climate system (Lorenz, 1951; Christy and Trenberth 1989). No doubt, to examine the variation of AM distributions and the related AM transport process on a hemispheric and even global scale is important for us to understand how and why the atmospheric circulation changes. Trenberth et al. (1981, 1985, 1994, 2005) investigated the variability and the conservation of the global dry air mass. For the annual cycle, global mean vapor mass is estimated about 0.29 hPa. The water vapor alteration in the atmosphere causes the total AM over an entire globe to change seasonally; the globally averaged surface air pressure reaches its maximum in August at a value of 985.64 hPa, and its minimum in January of 985.41 hPa. Lu et al. (2008) and Lu and Guan (2009) explored the variations in both AM and vapor content over a globe using NCEP/ NCAR reanalysis, showing the similar seasonal cycle of these two components. There are also some other studies (e.g., Chen et al., 1997; Hoinka, 1998; Carrera and Gyakum 2003; Zhao and

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Li 2006) in which the seasonal cycle of AM on global and hemispherical scale is discussed by means of different data sources.

The sun heats the earth, inducing the seasonal change in temperature of the earth surface and the atmosphere. The land and sea have different physical properties in terms of the thermal capacity. Due to the diabatic heating, the thermal contrast between land and sea is hence produced, leading to redistribution of AM over regions between the southern and northern hemispheres and also between continent and ocean in zonal. Striking exchange of AM occurs between the hemispheres, resulting in a close association with AM variation in them (Trenberth and Christy 1985; Christy and Trenberth 1985; Holl et al., 1988; Chen et al., 1997; Hoinka, 1998; Carrera and Gyakum 2003; Zhao and Li 2006; Lu et al., 2008). Guan and Yamagata (2001) examined the time series of surface pressures, discovering a fluctuation of AM distribution between the boreal and austral hemispheres that is denoted as interhemispheric oscillation (IHO). Lu et al. (2008) analyzed IHO seasonal variation, uncovering its pronounced seasonal cycle, with the variability in air and water vapor mass in an opposite manner. This IHO is driven mainly by the solar radiation. Evidently, the AM redistribution is able to drive the adjustment of atmospheric circulations, consequently showing its impact on monsoon variation is able to drive the adjustment of atmospheric circulations. The AM  redistribution between the boreal and austral hemispheres is considered to be responsible for temperature gradient that changes as a function of season (Tao and Chen 1957; Zhu et al., 1990; Zeng and Li 2002). Yang (1956) noted that land-sea thermal contrast causes the corresponding AM transport between land and sea. This transport reverses its direction between continent and ocean as season changes, in a close association with seasonal alteration of surface air pressures. It is found that in the Northern Hemisphere mid-latitudes the seasonal variations in AM flow are very striking (Van Den Dool and Saha 1993; Chen et al., 1997).

As time goes on, the zonal land-sea thermal gradient is altered in the Northern Hemisphere, thereby changing AM flows and the related source/sink distributions between continent and ocean, whereby new atmospheric active centers are generated (Wang 1962; She et al., 2001). Despite a lot of work devoted to the research into such centers in the Northern Hemisphere, little has been done of AM redistribution between continent and ocean on a hemispheric scale. The current work is devoted to examining the seasonal variations of AM redistribution as well as their corresponding mass flows between land and sea in the Northern Hemisphere by dint of the reanalysis data.

1 Data and methods

Data employed consist mainly of NCEP/NCAR reanalysis (Kalnay et al., 1996), with the variables consisting of monthly mean surface pressure ($p_s$), surface and upper-air winds ($u$, $v$), geopotential height ($h$), vertical velocity ($\omega$), precipitable water ($W$), specific humidity ($q$), and temperature ($T$). The horizontal resolution is 2.5°×2.5° in the grid mesh, with 17 levels from 1000 up to 10 hPa and time spanning January 1979 through December 2011. For comparative purposes ECMWF reanalysis (Uppala et al., 2005) with the resolution being 2.5°×2.5°, and covering January 1979 to August 2002, is also employed. For a particular physical quantity the multi-year monthly mean from the ECMWF (reanalysis spanning January 1979 to August 2002) was compared to that of 33-year mean equivalent from NCEP/NCAR reanalysis over 1979–2011 and that from ERA-40-year project for 1979–2001 (23-year average) except the January–August mean covering 24 years.

Surface air pressure $p_s$ is an important parameter describing the variations in the climate system as well as AM distribution. The AM $m$ can be expressed as $m = \frac{a^2 f_0}{g} \int_\varphi \frac{p_s \cos \lambda d\lambda d\varphi}{\int_\psi},$ in which $f_0$ is the earth’s deformation parameter, $a$ the earth’s radius, $g$ the gravitational acceleration, and $\varphi(\lambda)$ the latitude (longitude).

To reveal AM transporting process during seasonal shift, the monthly mean NCEP/NCAR reanalysis data were employed to evaluate air mass flow. Following Chen et al. (1997) and Carrera and Gyakum (2007), the AM flow from the earth surface up to the top of the atmosphere was integrated by the equation of continuity, approximatively yielding

$$\frac{\partial p_{dry}}{\partial t} = -g \nabla \cdot M_{dry},$$

$$M_{dry} = \frac{1}{g} \int_\psi p_{dry} (1-q) V dp = \frac{1}{g} \int_\psi \int_\lambda V_{dry} dp, \tag{2}$$

in which $p_{dry}$ denotes the surface pressure of dry air, $M_{dry}$ the flow of dry air mass integrated vertically from $p_s$ to 10 hPa, and $p_t(=10$ hPa) the top-level pressure, $V_{dry}$ the wind speed of dry air and

$$\nabla = \left( i \frac{\partial}{\partial x} + j \frac{\partial}{\partial y} \right)$$

the differential sign of a vector.
Vapor pressure is represented by \( p_v = g W \), where \( W \) is the vertically integrated precipitable water amount. Based on the moisture budget equation we arrive at

\[
\frac{\partial W}{\partial t} = -g \nabla \cdot \mathbf{M}_{\text{wet}} + g (E - P_{\text{recip}}),
\]

where \( \mathbf{M}_{\text{wet}} \) stands for the vertically integrated wet air mass flow. \( V_q \), for the related wind speed of wet air, \( E \) for evaporation rate, \( P_{\text{recip}} \) for rainfall rate. The NCEP/NCAR moisture data are limited to 300 hPa, which is thus taken as the upper limit for integration for wet air.

Given that \( p_v = p_{\text{dry}} + p_{\text{wet}} \), we combine eq. (1) with eq. (3), obtaining the tendency expressions of \( p_v \) as follows:

\[
\frac{\partial p_v}{\partial t} = \frac{\partial p_{\text{dry}}}{\partial t} + \frac{\partial p_{\text{wet}}}{\partial t},
\]

\[
\frac{\partial p_{\text{dry}}}{\partial t} = -g \nabla \cdot \mathbf{M}_{\text{rot}} + g (E - P_{\text{recip}}).
\]

where \( \mathbf{M}_{\text{rot}} \) denotes a rotating component, i.e., a non-divergent component, and \( \mathbf{M}_{\text{dry}} \) is a divergent component. And \( V_q \), for the related wind speed of wet air, \( E \) for evaporation rate, \( P_{\text{recip}} \) for rainfall rate. The NCEP/NCAR moisture data are limited to 300 hPa, which is thus taken as the upper limit for integration for wet air.

Following eq. (6), surface pressure variability rate depends mainly on the flux divergence and the difference between evaporation and precipitation. In fact, for monthly variations, atmospheric \(-g \nabla \cdot \mathbf{M} \sim 10^{-5} \text{ hPa s}^{-1}\) and \( g (E - P_{\text{recip}}) \) are larger than the pressure variation \( (\partial p_v/\partial t) \) by roughly 1–2 orders of magnitude. \( \mathbf{M} \) is the air mass flow integrated over \( p_v \), being a 2D vector, which can be separated into a flux stream function (\( \psi \)) and a flux potential function (\( \zeta \)), meaning that \( \mathbf{M} = k \times \nabla \psi + \nabla \zeta = \mathbf{M}_{\text{rot}} + \mathbf{M}_{\text{div}} \), of which \( \mathbf{M}_{\text{rot}} \) denotes a rotating component, i.e., a non-divergent component, and \( \mathbf{M}_{\text{div}} \) is a divergent component. And \( V_q \), for the related wind speed of wet air, \( E \) for evaporation rate, \( P_{\text{recip}} \) for rainfall rate. The NCEP/NCAR moisture data are limited to 300 hPa, which is thus taken as the upper limit for integration for wet air.

\[
\text{2 Seasonal cycle of AM redistribution in the Northern Hemisphere}
\]

Land-sea thermal contrast plays a crucial role in driving the monsoons (Tao and Chen, 1957; Zhu et al., 1990; Zeng and Li, 2002). The presence of thermal gradient is responsible for season-dependent reversal of winds, and in association with this, a noticeable fluctuation of AM redistribution occurs between land and sea. Due to the striking difference in thermal contrast between boreal land and sea in zonal dimension, the AM transfer of AM fluctuating between them is pronounced as the season goes on. Figure 1(a) shows that in high summer of July the stronger negative departure center of \( p_v \) resides over Eurasia, with a positive counterpart over the Pacific. Similarly, a negative departure core is seen over North America in contrast to a positive equivalent in the boreal Atlantic. In July the AM accumulation is mainly over the cold oceans as opposed to the deficit occurring over the warmer continents. When solar radiation alters its intensity as time goes on, the \( p_v \) pattern is regulated between land and sea. As shown on Figure 1(b), the \( p_v \) pattern shows its transition from negative in July to positive deviations over Eurasia and North America and reversal happens over the Pacific and Atlantic in October. In January (Figure 1(c)) \( p_v \) positive- (negative-) departure cores are observed in the continents (oceans), indicating an opposite pattern of AM to that of July, meaning AM piling (deficit) over the terrestrial (seas). In April \( p_v \) experiences readjustment between land and sea as opposed to the adjustment in October.

In boreal summer (winter) the atmosphere over the land (sea) is warmer in comparison to that over the ocean (land), thereby leading to the lower (higher) summer (winter) surface pressure on the continents compared to that over the oceans. As evidenced in Figure 1, Eurasia (North America) is delimited in \((10^\circ–62.5^\circ\text{N}, 30^\circ–140^\circ\text{E})\) \((10^\circ–75^\circ\text{N}, 60^\circ–125^\circ\text{W})\) and the boreal Atlantic (Pacific) in \((45^\circ–85^\circ\text{N}, 0^\circ–60^\circ\text{W})\) \((27.5^\circ–62.5^\circ\text{N}, 150^\circ\text{E}–125^\circ\text{W})\). The departures of monthly area-averaged \( p_v \) (with its year mean deducted) for the four study regions are estimated, separately, using NCEP/NCAR and ERA-40 reanalysis data (Figure 2). As shown in these regions, these NCEP/NCAR and ERA-40 data evaluated area-averaged \( p_v \) differ little in between and the \( p_v \) seasonal variation takes the main form of a single wave but \( p_v \) varies over the continents in an opposite phase to those over the oceans. In these regions but North America \( p_v \) displays a roughly identical annual variation by 10 hPa. It can be computed that 10 hPa variation denotes the change in column AM on a unit area by 102.04 kg.

Zonally, \( p_v \) exhibits a seesaw-like pattern, with the amplitude differing by 20 hPa, AM over Eurasia reaches its minimum (higher values) in July (January). Over the Pacific and Atlantic, AM is minimal in January whereas it reaches the maximum in July (May) over the Pacific (Atlantic). Temporally, the out-of-phase variation in AM redistribution between land and sea is likely to be linked to the difference in spatial scale between Eurasia and North America and between Pacific and Atlantic basins. The discrepancy in thermal forcing is caused by varying-degree thermal inertia in relation to the difference in space scale between the regions.

\[
\text{3 Air mass exchange between land and sea in the boreal hemisphere}
\]

Seasonal variation in atmospheric circulations bears an
intimate relation to $p_s$. Following eq. (6), air mass flow is examined to look into how and why the $p_s$ changes with the season going on.

3.1 Air mass flow

Noticeable exchange of AM occurs between land and sea in

Figure 1  Multi-year averaged monthly mean surface pressure (hPa) calculated from NCEP/NCAR reanalysis, with July (a), October (b), January (c), and April (d), where $p_s$ annual mean has been deducted before plotting, with positive departures denoted by shading.
the NH. For these continents and oceans the maintenance of AM excess or deficit is related to area-averaged AM flux divergence. As shown in Figure 3, the strongest divergence (convergence) of AM takes place in summer (winter) over Eurasia and North America. In January, over Eurasia the strong convergence happens, and the flux divergence alters from negative into positive values in April, arriving at higher values in July and they change from positive to negative values in October. The seasonal variation in AM flux divergence is opposite between land and sea, with the strongest convergence (divergence) in summer (winter) over the oceans. In early-middle March, the oceanic divergence alters from positive into negative values and the Atlantic divergence changes from negative to positive values in September. The seasonal cycle over these four regions exhibits a predominant single-wave form in pattern.

In the period from April to September, the divergence of AM flux over land is positive, resulting in the AM deficit there, as shown in Figure 2, whereas the oceanic flux divergence is negative, implying its support of AM accumulation there. From November to December and from January to March, the terrestrial flux divergence is negative, indicative of convergence, by which to maintain AM accumulated there in sharp contrast to the oceanic situation for its deficit.

It is noted here that similar results as shown in Figure 3 can also be derived from ERA-40 data (not shown) though some differences in magnitudes rather than the signs in values are found. These differences are likely to be induced by errors in winds during data assimilation in different reanalyses.

The interesting spatiotemporal variations in AM exchange between land and sea can be observed. Figure 4 depicts a pattern of rotational and divergent components of air mass flow, with the annual mean deducted from the winds, for the middle-lower troposphere (integrated from surface to
Figure 4  The rotational (stream line) and divergent components (vector) of air mass flow with yearly mean subtracted. 1000–500 hPa integrals shown in (a) and (c), and 500–10 hPa equivalents in (b) and (d). There, (a) and (b) are for July, and (c) and (d) for January. Shaded contours are for air mass flux divergence. All the quantities are derived from NCEP/NCAR reanalysis.
500 hPa) and above (integrated from 500 to 10 hPa). In July (Figure 4(a)) the air mass flows converge below 500 hPa over Eurasia, centered on the coastal belt of eastern China, with a divergent core in the eastern Pacific, and at the same time convergence (divergence) occurs over North America (the Atlantic). Zonally, a pattern of alternate convergence and divergence is observed. In relation to the divergence zones, a cyclonic (anticyclonic) circulation shows up to the northwest of the Asian convergent (Pacific divergent) region. Since a non-divergent component also indicates air mass transport, the westward transport from the Pacific is related to the AM inflow into Eurasia. Likewise, the same is true for the Atlantic and North America with regards to air mass shift. As evidenced in Figure 4(b), the reversal happens in air mass flow pattern in middle-upper layer above 500 hPa, with divergent (convergent) centers over the continents (oceans) in a pattern of alternating divergence and convergence in zonal direction, indicating AM flows from the large terrestrials into oceans.

The distribution of divergent centers and rotational components in January is contrary to that in July as shown in Figure 4(c), with a pattern of alternate divergence and convergence in zonal direction, meaning air mass flowing from land to sea in layers below 500 hPa. In contrast, Figure 4(d) presents a pattern above 500 hPa isobaric level opposite to that below it, with a distribution of alternating convergence and divergence in zonal direction, and divergent cores are located on the left side of the exit region of “air mass jet flow”, the air mass flowing from the oceans into the continents.

3.2 Zonal vertical circulation

Monthly mean vertical velocity from the earth surface up to 100 hPa is provided by NCEP/NCAR in its reanalysis and agrees with that from ERA-40 project. However, we use ERA-40 reanalysis for the vertical velocity available for levels from surface upward to 1 hPa to construct and investigate the multi-year mean cross sections of vertical circulations (composed of the divergent wind \( u_f \) and vertical velocity, wherefrom the annual mean has been deducted) and disturbed geopotential height \( H' \) (with annual and zonal means subtracted, deducting the latter makes zonal difference more evident) for July and January.

Besides the horizontal circulations, the vertical circulations also link atmospheric motions over continents to those over oceans. In July, there occur four vertical cells (Figure 5(a)). Between western Eurasia and the eastern Atlantic (eastern Eurasia and the eastern Pacific) there is a counterclockwise (clockwise) circulation cell. Between the eastern Pacific and western North America a weaker counterclockwise cell is present and a clockwise cell exists between northern Europe and western Atlantic. Deep updraft is seen over Eurasia, with an intense downdraft over the eastern Pacific. Enfeeble rising (sinking) air happens over North America (the northern Atlantic). Extending from 500 hPa into the stratosphere, the geopotential height shows its positive (negative) departure centers over Eurasia (the central-eastern Pacific, and Atlantic). However, in the layer below 500 hPa the distribution of height departures is opposite to the pattern above. Over the study continents and oceans, the sign differences in geopotential height departures between layers above and below 500 hPa are consistent physically with the direction of vertical motion.

In January, there exist four vertical cells opposite directionally to those in July (Figure 5(b)). A clockwise (counterclockwise) cell is seen between western Eurasia and the eastern Atlantic (eastern Eurasia and the eastern Pacific), and a clockwise (counterclockwise) cell resides between the eastern Pacific and western North America (between eastern North America and the western Atlantic). In that case, subsidence (updraft) happens over the two continents (the eastern Pacific and Atlantic). The distribution of geopotential heights in January is in anti-phase to the pattern in July over these continents and oceans.

Although \( p_s \) over Eurasia shows the remarkable excess (deficit) of AM in winter (summer), this feature is in close relation mainly to vertical velocity over the eastern Atlantic and eastern Pacific. The seasonal cycle of Pacific (North American) \( p_s \) corresponds largely with vertical motion over eastern Eurasian and western North America (the eastern Pacific and western Atlantic), with the Atlantic \( p_s \) variation associated mainly with vertical motion over eastern North America and western Eurasia.

3.3 Vapor budget between land and sea in the boreal hemisphere

Variations in surface pressure are determined by both dry air and vapor pressures (eq. (5)). Vapor content greatly affects the atmospheric absorption of long- and short-wave radiation, thereby impinging on thermal condition of the atmosphere. The regional distribution of atmospheric vapor pressure \( p_w \) is indicative of vapor contents. Vapor contents over land and sea in NH are distinctively different as a function of season. In Figure 6(a) the rotational component of the water vapor transporting shows that in July there exist three anticyclonic centers of the tropospheric vapor, which are situated separately over the central Pacific, northern Atlantic, and Indian Oceans, of which the first and the second correspond to subtropical highs over Pacific and Atlantic respectively, and the last is related to the summer monsoon, Somali jets and the equatorial buffering zone over the Indian Ocean. The divergent components of vapor transport are able to show the source/sink of water vapor, with a convergent core in eastern Asia and the western Pacific and the divergent core at eastern-Pacific subtropics that provides rich vapor for eastern Asia. These results are in agreement with Zhou et al. (2005).

The distribution of water vapor pressure is associated...
with the convergence and divergence of the vapor transports in the troposphere. Figure 6(a) delineates that in July positive departures of vapor pressures are seen in Eurasia and the northern part of the northwestern Pacific, and also in North America where the $p_w$ departures are weaker. The departures of vapor content are larger over the land than over the sea. Because vapor acts as a predominant absorber of short- and long-wave radiation, diabatic heating intensity in the atmosphere will be stronger over Eurasia and North America than over the oceans, thereby causing the air mass exchange between land and sea.

The tropospheric vapor pattern in January is contrary to that in July as shown in Figure 6(b). The vapor pressure departures are negative over South Asia, Northeast Asia, and southeastern North America. In boreal winter, the vapor pressure departure is positive in the Southern Hemisphere. Water vapor pressure $p_w$ and surface air pressure $p_s$ vary in an anti-phase way, as seen in Figures 6 and 1(a), (c). The vapor (dry air) mass reaches its annually higher (lower) values in July over Eurasia and North America. The zonal difference in air density exists (Lu et al., 2008). This difference can be explained partly by the water vapor content changing in the air column. The moles of vapor (H$_2$O–18) are much smaller than those of O$_2$ and N$_2$ so that vapor piling over Eurasia reduces air density there, producing zonal pressure gradient that causes air mass exchange between land and sea.

4 Land-sea thermal contrast

Land-sea thermal contrast drives the air mass to redistribute over the continents and oceans (Liu et al., 1999; Wu et al., 2003; Liu et al., 2004; Wu et al., 2009; Wu et al., 2012). The contrast is season-dependent, influencing the direction and magnitude of land-sea temperature gradient, thereby driving the AM seasonal cycle between land and sea. Atmospheric circulation depends on both dynamic and thermal effects for its variation. However, only the thermal impacts are discussed in the following.
4.1 Zonal difference in air temperature

The zonal gradients of temperature are season-varying in the NH, directly impacting on the variations of atmospheric circulation and hence indirectly deriving air mass exchange between land and sea in terms of the thermal wind relation. We define $T'=T - [T]$, in which $T$ denotes a multi-year monthly mean temperature (with its annual mean subtracted from the monthly mean) and $[T]$ the zonal mean of temperature. The zonal mean temperature has been deducted from $T$ so that the zonal discrepancy becomes clearer (Figure 7). The temperature difference between land and sea at the same latitude is linked to diabatic heating of air under the impacts of solar radiation above the underlying surface. Note that the spatial patterns of temperatures $T'$ look similar in layers both above 500 hPa and below 500 hPa. In this work, the mean temperature of the air column from 500 to 200 hPa $\int_{p=200}^{p=500} T dp / (p_{500} - p_{200})$ is presented to depict the tropospheric thermal condition (Zhao et al., 2007; Zhao et al., 2008). It is found in Figure 7(a) that the mean temperature in July has a distinctively positive departure centered over Eurasia, stronger than over North America. In contrast, the negative temperature departures emerge over the Pacific and Atlantic, with a pattern of alternate high and low temperatures in zonal dimension. When positive departures are over Eurasia, the air column expands due to heating while it shrinks due to cooling over the oceans. As a response to the warmer temperature departures, low pressure systems appear often below 500 hPa isobaric level over the research continents, accompanied by convergence, with higher geopotential height and divergence generated above 500 hPa. Over the colder region, an opposite scenario appears. Such a distribution is responsible for driving air mass to flow from sea into land below 500 hPa, with air migration in a reversed direction at high levels.

Distribution of temperature departures in the atmosphere between land and sea for January is contrary to that for July, with alternating low and high values in zonal dimension (Figure 7(c)). This pattern is conducive to the genesis of highs (lows) below (above) 500 hPa accompanied by divergence (convergence) over the terrestrials, and, in a sharp contrast, the reversal happens over the oceans. This distribution favors AM to go from land to sea (from sea to land) at low (high) levels, resulting in an opposite AM shift in July.

4.2 Seasonal cycle of diabatic heating

Radiation and surface sensible heat flux are the dominant sources of atmospheric energy. Increased (decreased) radiation heating (cooling) and surface sensible heat transfer are responsible for the rise (fall) of air temperature, leading to expansion (contraction) of in-column air, and thus forcing...
the AM to exchange between land and sea. Here, the atmospheric apparent heat sources \(<Q_1>\) and apparent vapor sinks \(<Q_2>\) are analyzed. The difference between \(<Q_1>\) and \(<Q_2>\) (Luo and Yanai, 1984) denotes the diabatic heating produced by in-column net radiation heating and surface heat flux exchange. If the difference between \(<Q_1>\) and \(<Q_2>\) is positive, meaning that non-adiabatic heating is strong for the region and the negative difference suggests that radiative cooling happens, together with reduced thermal exchange between surface and air.

The large zonal differences in diabatic heating (\(<Q_1>\) minus \(<Q_2>\)) at mid latitudes are observed, which reverses as a function of season (Figure 8). In July, the differences of diabatic heating exhibit a pattern as zonally “positive-negative-positive-negative”, with positive centers over the continents. This implies that heating due to radiation and surface sensible heat transfer is strengthened, leading to convergence in the lower troposphere and divergence at high levels. A negative center of diabatic heating emerges over the Pacific and Atlantic, i.e., radiative cooling is intensified and the transport of heat flux is weakened from sea into air, with divergence (convergence) taking place at the lower (high) levels. This thermal contrast drives AM from the oceans to the continents at low-levels of the atmosphere, and in a reversed manner for high-levels. The AM transfer is season-dependent. In January, the difference (\(<Q_1>\) minus \(<Q_2>\)) has its pattern as ‘negative-positive-negative-positive’ in zonal direction, just opposite to that in July, thus causing the transfer of AM to be reversed between land and sea.

This reversal of heat transport between winter and summer is responsible partially for the monsoon generation in parts of Asia and Mexico. We notice that vigorous heating or cooling occurring in the northwest Pacific basin is related to the large zonal difference in heating over eastern Asia, conducive to the happening of AM shift and vertical circulation cell between land and sea (Figures 4 and 5).

5 Discussions and conclusions

The seasonal cycle of the AM redistribution over areas of the continents and oceans in the NH has been investigated using NCEP/NCAR and ERA-40 reanalysis data. The results are summarized as follows:

As the season evolves, AM in the northern hemisphere experiences perceptible shift between land and sea. In July air mass over Eurasia (Pacific) reaches its minimum (maximum) for the year whereas the AM distribution is reversed in between in January. AM excess is maximized over the Atlantic (Pacific) in May (July).

As time goes on, surface air pressure departures (with yearly mean deducted) over the Tibetan Plateau vary in an opposite manner to those of the surroundings (Saha et al., 1994). In winter, the Eurasian temperatures are lower compared to
Figure 8 Differences between vertically integrated apparent heat sources $<Q_1>$ and apparent vapor sinks $<Q_2>$, from which yearly mean has been subtracted for July (a) and January (b) based on NCEP/NCAR data. Units are in W m$^{-2}$.

its surrounding oceans. And the lower temperature allows a rapid drop of pressure with altitude. For the higher-altitude Tibetan Plateau, its pressure reaches the minimum in winter for the year, with the summer pressure opposite to the winter arriving at its annual maximum. On the other hand, over the North American western coast (where the elevation is higher), the departures of surface pressures (with its annual average deducted) vary in an opposite way to the situation in its interior (Figure 1).

The vertically integrated air mass flow reveals a distinctive AM exchange at higher and lower levels between the continent and ocean. For the oceanic and neighboring terrestrial regions, the maintenance of AM excess and deficit is associated with its area-averaged flux divergence. In July, the air mass flow below 500 hPa exhibits a pattern of alternate convergence and divergence from west to east, with an opposite pattern above 500 hPa. In January, the pattern of air mass flows at higher- and lower-levels between land and sea is contrary to that in July. The zonal vertical circulation varies with season and its sinking and rising branches experience variation in direction between land and sea, thereby linking together air motions between land and sea.

Vapor contents differ greatly between land and sea, undergoing a seasonal cycle. It is known that vapor contents can impact greatly on the absorption of shortwave radiation from the sun and longwave radiation from the earth surface, thus affecting the air thermal condition. In July, the vapor departure reaches its higher value over the continents and the northern NW Pacific. Plentiful vapor in this season is conducive to the air absorption of more shortwave radiation over the land and in contrast, vapor deficit arrives at its greater value in January over southern Asia, NE Asia, and SE North America.

The boreal thermal difference between land and sea varies as a function of season, thus affecting the direction and value of temperature gradient between them, which drives the seasonal cycle of AM over them. In July, the mean temperature shows a pattern of alternate high and low value in the west-east direction and the reversal happens in January. The diabatic heating rate ($<Q_1>$ minus $<Q_2>$, with vapor latent heating excluded) shows a great discrepancy in zonal direction while the heating rate changes as a function of season. In July, such discrepancy exhibits a pattern as "positive-negative-positive-negative" and the reversal occurs in January. The reversal in diabatic heating between land and sea favors the genesis of monsoons and induces indirectly the seasonal cycle of AM exchange between land and sea.

In May the accumulated AM over the Atlantic reaches its maximum, possibly in relation to its smaller oceanic basin compared to the Pacific. This happens because the response time of air over the Atlantic to the heat forcing is probably
shorter, consequently causing accumulated AM there to be the largest in May. On the other hand, the variation is smaller in AM over North America from month to month. The Pacific AM has its largest accumulation in July, with the excessive mass compensated by the AM from Eurasia and the Atlantic where the deficit occurs correspondingly (Figure 3). Atmospheric mass is a basic parameter describing the atmosphere and its distributions, and its shift between land and sea is under the control of seasons, properties of land and sea, vapor transfer, and internal dynamics of the air-sea system. The annual course of AM over North America is very short, its cause being a problem of interest. Besides, Guan et al. (2010) analyzed IHO (inter-hemispheric oscillation) by means of model atmospheric data. But it is uncertain whether the shift and exchange of AM between boreal continent and ocean can be simulated with success by dint of different climate models that will give the same results. These problems deserve further researches.

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