Southeast Asian Pressure Surges and Significant Events of Atmospheric Mass Loss from the Northern Hemisphere, and a Case Study Analysis

MARCO L. CARRERA AND JOHN R. GYAKUM
Department of Atmospheric and Oceanic Sciences, McGill University, Montreal, Quebec, Canada

(Manuscript received 31 January 2006, in final form 31 October 2006)

ABSTRACT
A recent study of significant events of atmospheric mass depletion from the Northern Hemisphere (NH) during the extended boreal winter indicated that Southeast Asian pressure surges were an important physical mechanism that acted to channel the atmospheric mass equatorward out of the NH on a rapid time scale. This study builds upon this finding and examines both the direct and indirect roles of Southeast Asian pressure surges for a particular event of dry atmospheric mass depletion from the NH. The focus of this study is on the enhanced interhemispheric interactions and associated Southern Hemisphere (SH) tropical and extratropical responses resulting from the pressure surges.

First, this study examines the conservation of dry atmospheric mass (i.e., the relationship between the dry meridional winds and the area-integrated dry air surface pressure) in the NCEP reanalysis for the 25 significant events of dry atmospheric mass depletion from the NH. Results indicate that the NCEP dry meridional winds are able to qualitatively capture the dry atmospheric mass evacuation from the NH. In a quantitative sense there is very good agreement between the wind and pressure data in the extratropics of both hemispheres. A distinct negative or southward bias in the NCEP vertically and zonally integrated dry meridional winds is apparent between 5° and 17.5°N. This southward bias was not present in the ECMWF Re-Analysis. The source of the southward bias in NCEP appears to result from a weaker analyzed ITCZ.

The particular case of dry atmospheric mass depletion from the NH examined in detail is associated with an intense pressure surge over Southeast Asia. A significant enhancement of convection in the monsoon trough region of northern Australia occurs roughly 4 days after the peak in intensity of the Siberian high. A low-level westerly wind burst develops in response to this enhanced zonal pressure gradient caused by the pressure surge as part of the onset of an active phase of the Australian summer monsoon. This study shows that three prominent anticyclonic circulations intensify in the SH extratropics, stretching from the south Indian Ocean to the South Pacific, beneath regions of upper-tropospheric dry atmospheric mass convergence, originating partly from the monsoon convection outflow. These anticyclonic circulations are regional manifestations of the dry atmospheric mass increase in the SH.

1. Introduction
Recently, Carrera and Gyakum (2003, hereafter CG03) examined the time evolution of the large-scale circulation associated with extreme events of dry atmospheric mass loss from the Northern Hemisphere (NH) on subseasonal time scales during the extended boreal winter season. The breakdown of the NH dry atmospheric mass was found to be a multiple-phenomenon event, involving explosive cyclogenesis in the Gulf of Alaska, and pressure surges occurring over both Southeast Asia and North America. Of particular importance, pressure surges over Southeast Asia and North America acted to channel the atmospheric mass equatorward out of the NH extratropics on a rapid time scale (~4–5 days).

There have been numerous studies related to pressure surges occurring over Southeast Asia, and China in particular, during the Asian winter monsoon (see Compo et al. 1999 for a review). As Compo et al. (1999) discuss, no precise definition of pressure surges exists owing to the fact that researchers have defined this phenomenon based upon the varying local impacts. However, general agreement exists on the prominent characteristics of pressure surges; namely, the surging of atmospheric mass southward over Southeast Asia and the decrease in the lower-tropospheric tempera-
tures over China combined with an increase in the northerly component of the near-surface wind field to the south of China (Slingo 1998; Compo et al. 1999). Important to this research are those studies related to the enhanced interhemispheric interactions and associated Southern Hemisphere (SH) tropical and extratropical responses resulting from the direct and indirect effects of the pressure surges (Williams 1981; Davidson et al. 1984; Love 1985a,b; Johnson et al. 1987; Kiladis et al. 1994; Suppiah and Wu 1998).

The onset of the Australian summer monsoon (ASM) has been linked to the occurrence of cold surges from Southeast Asia. Davidson et al. (1984) documented a link between divergent northerly surges emanating from subtropical anticyclones over Southeast Asia and the onset of the ASM in December 1978. A 12-yr composite study by Suppiah and Wu (1998) found that cold surges typically preceded the onset of the ASM by 5–10 days. On intraseasonal time scales, Love (1985a) showed that the effect of cold surges and the shifting of atmospheric mass equatorward was to increase the west-to-east near-equatorial pressure gradients over the tropical western Pacific. The strengthened pressure gradients led to an enhanced low-level westerly flow, which led to increased cyclonic vorticity in the monsoon trough region through the effects of lateral shear. Murakami and Sumi (1982) examined intraseasonal active and break phases of the SH monsoon for the 1978/1979 season. An important finding was that the vorticity advection by the northerly divergent winds was largely responsible for the intraseasonal changes in the zonal mean vorticity and hence the strength of the monsoon.

Kiladis and Weickmann (1992) found that, when convection peaked in the equatorial trough north of Australia (15°S–5°N, 140°–160°E), a prominent wave train was established after six days, with a pronounced upper-tropospheric ridge situated over the central South Pacific (see their Fig. 6b). Knutson and Weickmann (1987) found significant upper-level ridge development in the South Pacific when convection, associated with the intraseasonal oscillation, was enhanced over the equatorial trough north of Australia. Tyrrell et al. (1996) noted strong extratropical wave responses in the SH sub tropics within five days of the period of maximum tropical convection in the western tropical Pacific during the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE).

The objectives of this study are to build upon the work of CG03 to document the direct and indirect effects of Southeast Asian pressure surges for a case study of dry atmospheric loss from the NH. We will provide evidence that links the pressure surge over Southeast Asia to the dramatic onset of a convectively active phase of the ASM, as seen through the divergent atmospheric circulation. Using calculations of the dry atmospheric mass flux, we will document the atmospheric mass redistribution within the SH. Three prominent anticyclonic circulations intensify in the SH extratropics, stretching from the south Indian Ocean to the South Pacific, beneath regions of upper-tropospheric dry atmospheric mass convergence, originating partly from the monsoon convection outflow. These anticyclonic circulations are largely responsible for the dry atmospheric mass increase in the SH.

The outline of the paper is as follows. In section 2 we discuss the datasets used in this study and briefly describe the methodology used by CG03 to isolate significant events of NH dry atmospheric mass depletion. We also assess the quantitative skill of the divergent winds in the two reanalysis datasets used in this study. In section 3 we examine in detail a particular event of dry atmospheric loss from the NH associated with a pressure surge over Southeast Asia and the onset of a convectively active phase of the ASM. In section 4 we analyze the relationship between the diabatic heating anomalies associated with the onset of the active phase of the ASM and the redistribution of atmospheric mass within the SH extratropics. Finally a short summary is given in section 5.

2. Data and methodology

a. Data

We employ two sets of reanalysis data: the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996; hereafter referred to as the NCEP Reanalysis), and the 15-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis (ERA-15; Gibson et al. 1997). The pressure-level archive data, which have been interpolated from the original model grid to $2.5° \times 2.5°$ global grids, are used from both the NCEP and ERA-15 reanalyses.

For the NCEP Reanalysis we use a subset of the data, extending over the 30-yr period from 1968 to 1997, owing to a problem with the NCEP Reanalysis surface pressure and mean sea level pressure data (encoding error; see http://www.cdc.noaa.gov/cdc/reanalysis/problems.shtml) prior to 1968. The complete ERA-15 data extend over a shorter time period from 1979 to 1993. Surface pressures are not provided as standard output in the ERA-15, and hence we calculate the surface pressures following the procedure of Trenberth et al. (1987).
To examine the time variations in tropical convection, we use an interpolated daily outgoing longwave radiation (OLR) dataset derived from National Oceanic and Atmospheric Administration (NOAA) satellites (Liebmann and Smith 1996). The data begin in June 1974 and have a global coverage at 2.5° resolution. Missing values are filled with a combination of spatial and temporal interpolation (Liebmann and Smith 1996). OLR data have been used extensively in the scientific literature as a proxy for deep tropical convection and rainfall (Arkin and Ardanuy 1989; Kiladis 1998). The observed OLR data have not been assimilated into the NCEP Reanalysis or the ERA-15 for the calculation of diabatic heating and vertical motion (Wheeler et al. 2000). Therefore, we can view the observed OLR data as an independent confirmation of the results from both the NCEP Reanalysis and the ERA-15.

b. Methodology

The methodology used to isolate the significant events of dry atmospheric loss from the NH is outlined in detail in CG03. A threshold-crossing procedure is applied to the time series of the rate of change of anomalous dry air mass for the NH from the NCEP Reanalysis. When the rate of change of anomalous dry air mass first becomes negative, we count the number of days until the rate of change becomes positive again and label this as an event.

The onset for the given event occurs when the rate of change of anomalous dry air mass first becomes negative, and is denoted $T_0$. The event duration is defined as the total number of days between the local maximum and local minimum in NH dry airmass anomalies. Finally, the event magnitude is defined as the difference in anomalous NH dry air mass between the local maximum and local minimum. Hence, for each event we can define a duration (days) and a magnitude (hPa). A significant event of NH dry atmospheric mass fall is defined as that subclass of events whose magnitude exceeds the 95th percentile. For reasons discussed in CG03 we focus upon those events during the extended boreal winter season comprising the months of October to March.

For the October to March period from 1968 to 1997, a total of 25 events were found with durations ranging from 6 to 10 days, with a most frequent duration of 9 days. These 25 events comprise our sample of significant events of NH dry air mass collapse. The onset for each event and the total NH anomalous dry air surface pressure fall (event magnitude) are given in Table 1 of CG03. For more details on the methodology the reader is referred to CG03.

c. Vertically integrated meridional transport of dry atmospheric mass

In this section we assess the quantitative skill of the NCEP-divergent winds by examining the time evolution of the dry atmospheric mass fluxes for the composite NH dry atmospheric mass fall event. In the appendix we formulate the relationship between area-integrated dry air surface pressure changes and the vertically and zonally integrated dry meridional winds at the boundary of the area. Figure 1 shows a latitude versus time plot of the composite anomalies of the vertically integrated dry meridional atmospheric mass flux [meridional component of $\mathbf{M}_d$ in (A4)], zonally integrated from 0° to 360°, for the 41-day period centered on the composite onset time $T_0$.

Between $T_{-6}$ and $T_{-1}$, a meridionally extensive zone of anomalous northward dry atmospheric mass flux extends from the mid- to high latitudes of the SH ($\sim 60^\circ$S) to the mid- to high latitudes of the NH ($\sim 60^\circ$–70°N), with peak values near 45°N. This time period corresponds to the rise (fall) of anomalous dry atmospheric mass in the NH (SH; see right panel of Fig. 1). Commencing at $T_{-1}$ and extending to $T_{+6}$ and $T_{+7}$, we see evidence of a dramatic onset of anomalous southward dry atmospheric mass fluxes extending from the high latitudes of the NH ($\sim 80^\circ$N) southward to the mid- to high latitudes of the SH ($\sim 60^\circ$S), coinciding with the evacuation of dry atmospheric mass from the NH. Highest magnitudes are found between 30° and 60°N at $T_{+2}$, with values peaking at the equator two days later ($T_{+4}$). For the composite, in a qualitative sense, the dry atmospheric mass evacuation from the NH is reflected in the anomalous dry meridional wind data. What is impressive about Fig. 1 is the suddenness with which the anomalous dry atmospheric mass fluxes change from northward to southward near $T_0$. This could be an artifact of the compositing, as individual cases do not exhibit such a pronounced transition.

To examine the performance or skill of the NCEP Reanalysis dry meridional winds, in a quantitative sense, as they relate to the dry atmospheric mass changes, we consider relation (A12) for various polar caps extending from the North Pole southward and the South Pole northward. In Figs. 2 and 3 we have plotted both the composite area-integrated dry air surface pressure changes for given polar caps [i.e., left-hand side of (A12)] along with the composite vertically and zonally integrated dry meridional winds for the given latitude wall bounding the polar cap [i.e., right-hand side of (A12)] for the 41-day period centered on the onset time for the composite event. In calculating the area-integrated rate of change of dry air surface pressures, a
second-order finite difference scheme was used over the given 3-day period. To be consistent we used a 1–2–1 weighted mean of daily averaged $M_{dy}$ calculated over the same 3-day period [see relation (A12)]. The statistical significance, at the 95% level, based upon a two-sided Students $t$ test for the difference of sample means is shown by the solid circles in Figs. 2–4.

For the 25-case composite, the area-integrated changes in dry air surface pressures for the NH are reflected to a high degree in the vertically and zonally integrated dry meridional winds, especially as the latitude wall is displaced poleward (Fig. 2). Also, as the polar cap is shifted poleward, the area-integrated dry air surface pressure changes are amplified, with both the rise prior to the onset and the decline during the event possessing greater magnitudes. Between 5° and 15°N the NCEP dry meridional winds demonstrate a negative bias (Figs. 2b–d), as reflected by the statistically significant differences between the vertically and zonally integrated dry meridional winds and the area-integrated dry air surface pressure changes. The vertically and zonally integrated dry meridional winds are negative (southward) throughout the entire period. The bias, however, appears to be systematic, as the NCEP wind data do capture the enhanced northward component (i.e., reduced negative values) prior to the onset, followed by the enhanced southward component (i.e., increased negative values) during the event.

For polar caps north of 20°N (Figs. 2e–h), the agreement between the wind and pressure data is remarkably good, especially for the time centered on the dry atmospheric mass fall (i.e., between $T_{-2}$ and $T_{+6}$). The NCEP meridional winds are able to capture the entire cycle of dry atmospheric mass contained within the northern polar caps, comprising the rise prior to the composite event and the subsequent fall during the event.

Quantitative differences exist between the right- and left-hand sides of (A12) for the SH; however, in a composite sense, the high-frequency changes in the area-integrated dry air surface pressures are reflected in the NCEP dry meridional winds (Fig. 3). Note that we have reversed the sign of the zonally integrated dry meridional winds for display purposes. A positive (negative) value now indicates a southward (northward) component. Unlike the NH, the magnitude of the area-integrated dry air surface pressure changes diminishes as the latitude wall is displaced southward (Figs. 3f–h).

d. Southward bias in NCEP Reanalysis wind data

There are many reasons why relation (A12) is not satisfied exactly in the NCEP Reanalysis. First, we are using Green’s theorem (A10) in discrete form, which may not be exactly valid. Additionally, on the right-hand side of (A12), the results depend upon values at one particular latitude circle and undoubtedly are sensitive to the nature of the interpolation procedure from the spectral grid to the latitude/longitude grid. The

**Fig. 1.** Latitude–time plot of the composite anomalies of the vertically integrated dry meridional atmospheric mass flux ($10^9$ kg s $^{-1}$), zonally integrated from 0° to 360°; shaded according to the given scale. Dry air surface pressure anomalies (hPa) for the N (S) H in solid (dash) are shown in the panel on the right. Time is given on the ordinate with respect to the onset time ($T_0$).
The NCEP Reanalysis system is complex, including the NCEP global spectral model, the spectral statistical interpolation scheme, and an advanced quality-control system designed to detect errors during the preprocessing stages (Kalnay et al. 1996). In such data-sparse regions as the Northern and Southern Hemisphere Tropics and subtropics, the analysis depends heavily upon the first-guess field provided by the assimilating model. This in turn is influenced by the model physical parameterizations, especially those associated with moist physics (i.e., convective parameterization; Annamalai et al. 1999). With reference to the southward bias in the NCEP wind data shown in Fig. 2, the source of the error may exist with the intensity (weaker ITCZ) and latitudinal placement of the ITCZ (Trenberth and Guillemot 1998; Trenberth et al. 2000; Newman et al. 2000; Lim and Ho 2000).

In Fig. 4, we again consider relation (A12), and compare the results for the 10 common cases of NH dry atmospheric mass fall events between NCEP and ERA-15 (events between 1979 and 1993) for polar caps with latitude walls within the Tropics and subtropics of the NH, where the negative bias in NCEP is most severe (Fig. 2). The ERA-15 system provides specific humidity...

---

**Fig. 2.** Composite area-integrated dry air surface pressure changes ($10^9$ kg s$^{-1}$) for various polar caps bounded by the North Pole and the given latitude wall [thick solid; see relation (A12)]. Vertically and zonally integrated dry meridional winds for the given latitude wall (thin solid): (a) $0^\circ$, (b) $5^\circ$, (c) $10^\circ$, (d) $15^\circ$, (e) $20^\circ$, (f) $30^\circ$, (g) $40^\circ$, and (h) $45^\circ$ N. The abscissa denotes time, in days, with respect to $T_0$. The difference of the sample means, at each latitude and time, is significant at the 95% level if a solid circle is shown.
(q) values at each archived pressure level up to the 10-hPa level and these values were used in the calculation of $M_d$ in (A4). Values of specific humidity (q) above 300 hPa were not available in the NCEP Re-analysis, and thus we assumed a value of $q = 0$ in the calculation of $M_d$ in (A4). This effect, we argue, is minor as values of $q$ above 300 hPa account for only 0.2% of the total area-averaged surface pressure owing to vertically integrated water vapor for the globe (Trenberth et al. 1987). It is clear from Fig. 4 that there is better agreement between the area-integrated changes in dry air surface pressures and the vertically and zonally integrated dry meridional winds in the ERA-15, as compared to NCEP in the subtropical latitudes (e.g., 10°–15°N). For the ERA-15 data between latitudes 0° and 15°N (Figs. 4a–d'), the differences between the two sample means are not statistically significant. The solid lines, denoting the dry air surface pressure changes, are very similar between both reanalyses. However, unlike NCEP, the ERA-15 does not possess the pronounced southward bias in the meridional winds.

In the next section we examine the mechanism of Southeast Asian pressure surges, assessing both the direct and indirect forcing for the collapse of dry atmospheric mass from the NH. The framework involves a case study.
3. Case study

The event of NH dry atmospheric mass collapse that we will investigate in greater detail began on 3 March 1989 (denoted T₀) and lasted for nine days (event 19 as given in Table 1 of CG03). The event magnitude, as defined in section 2b, was 1.808 hPa, ranking it fifth among the subset of 16 significant events of NH dry atmospheric mass collapse that possessed a Southeast Asian pressure surge (see CG03). We chose to examine this event as there was a significant pressure surge occurring over Southeast Asia (Suppiah and Wu 1998). CG03 found that pressure surges over Southeast Asia were an important physical mechanism that acted to channel the atmospheric mass equatorward out of the NH extratropics on a rapid time scale during significant events of NH dry atmospheric mass collapse.

Owing to the superior skill of the ERA-15-divergent winds, as compared with NCEP, we utilize the ERA-15 data to analyze the case in detail. CG03 showed in their Fig. 5 that the time evolution of the composite sea level pressure (SLP) anomalies between the ERA-15 and NCEP are qualitatively very similar. Additionally, the evolution of the composite 500-hPa geopotential height anomalies between both reanalyses (not shown) was found to be very similar.

Fig. 4. Same as in Fig. 2, except for the 10 Northern Hemisphere cold season dry atmospheric mass fall events between 1979 and 1993. (a)–(d) Data are derived from the NCEP Reanalysis, and (a’)–(d’) from the ECMWF Re-Analysis. The position of the latitude wall for each polar cap is indicated in the top right of each panel.


a. SLP and 500-hPa geopotential height anomalies

In this section we present the time evolution of the SLP and 500-hPa geopotential height anomalies for the given event. Surface pressure is a more accurate measure of atmospheric mass (Trenberth 1981; Van den Dool and Saha 1993; Trenberth and Smith 2005) when compared with SLP, owing to the necessary addition or subtraction of atmospheric mass when deriving SLPs. However, the SLP field is more representative of the circulation, and hence we show the SLP field. The anomalies have been calculated based upon a weighted monthly climatology from the ERA-15 period of 1979–93. The objective is to compare the circulation anomalies for this individual case with those of the 25-case composite shown in Fig. 4 of CG03.

Similar to the 25-case composite, a substantial positive SLP anomaly occupies much of the North Pacific and western Canada prior to the event onset (Figs. 5a,b). The anomaly is locally in excess of 39 hPa and is associated with an intense surface anticyclone with central pressures in excess of 1050 hPa. The 500-hPa height anomaly field (Fig. 6a) shows that the positive height anomaly extends to upper levels. A positive/negative SLP anomaly couplet over Asia extends westward to Europe. The positive SLP anomaly is associated with the building of the Siberian high, a statistically significant feature of the 25-case composite.

At 500 hPa, the building of the Siberian high is accompanied by a strengthening ridge to the northwest near 40°E (Fig. 6a). Much of the area poleward of 60°S is associated with negative SLP anomalies. As an inset to each panel, we have plotted the anomalous NH and SH area-averaged dry air surface pressure anomalies. On 26 February (Fig. 5a), the anomalous NH and SH area-averaged dry atmospheric mass in the NH (SH) is at a local minimum (maximum) and is about to commence a rise (fall).

Moving ahead to 28 February (Fig. 5b), the anticyclonic circulation anomaly in the North Pacific expands and intensifies, occupying the entire basin from the Asian coast eastward to the North American continent, with a pronounced omega-block structure in the 500-hPa geopotential height field (Fig. 6b) characteristic of intense blocking situations (Colucci et al. 1981). Also shown is the continued building of the Siberian high over central Asia, downstream of the intensifying upper-level ridge near 60°E (Fig. 6b). The building of the Siberian high and the strengthening of the positive atmospheric mass anomaly in the North Pacific are associated with a rise in dry atmospheric mass for the NH (inset of Fig. 5b). In the SH, the entire area poleward of 60°S is associated with negative SLP anomalies. The decrease in dry atmospheric mass for the SH is manifested as a weakening of the subtropical ridges in the south Indian Ocean and to the southeast of Australia combined with a deepening cyclonic disturbance to the south of Australia.

On 2 March (Fig. 6c), the time of maximum (minimum) dry atmospheric mass in the NH (SH), the Siberian high over central Asia has reached peak intensity, with a central pressure in excess of 1058 hPa (daily averaged), greater than 30 hPa above the long-term climatology. It is accompanied by a substantial ridge at 500 hPa near 80°E (Fig. 6c), with values locally in excess of 45 dam above the long-term climatology. Over the central subtropical Pacific, a cyclone (negative SLP anomaly) is intensifying (cf. Figs. 5b,c) to the south of the large anticyclonic anomaly. In the 25-case composite the breakdown of the zonally extensive positive atmospheric mass anomaly in the North Pacific occurs in conjunction with a deepening low pressure system in the region of the Gulf of Alaska (see Fig. 4 of CG03).

On 4 March (Figs. 5d and 6d), the dry atmospheric mass evacuation out of the NH has begun, and similar to the 25-case composite it is accompanied by southward surges of pressure over both Southeast Asia and North America. Over Asia the Siberian high has begun to weaken and expand in areal extent, while over North America the deepening cyclone in the Gulf of Alaska and the accompanying warm air advection to the north and east is associated with ridging at upper levels along the west coast of North America (Fig. 6d). The building of the upper-level ridge accompanying intense cyclogenesis in the Gulf of Alaska region has been noted in prior studies of North American cold air outbreaks (Dallavalle and Bosart 1975; Colucci and Davenport 1987; Colle and Mass 1995; Schultz et al. 1998).

On 6 March (Figs. 5e and 6e), the zonally extensive positive SLP anomaly over the North Pacific and North America seen on and prior to 4 March (Fig. 5d) has split into a meridionally extensive anticyclonic circulation over North America extending southward into Central America and a less intense and more localized anticyclonic disturbance over the North Pacific. A deep cyclone in the Gulf of Alaska occupies the zone between these two anomalies and appears to be important in breaking up the extensive positive atmospheric mass anomaly. The negative SLP anomaly associated with this cyclone can be traced back to 26 February over the subtropical central Pacific (Fig. 5a). Over central Asia, SLPs continue to fall. The rise in dry atmospheric mass in the SH is manifested as a buildup of atmospheric mass in the midlatitude ridge in the South Atlantic Ocean. Additionally, the extensive zone of negative SLP anomalies near 50°–60°S over the central and
Fig. 5. Sea level pressure anomalies from the ECWMF Re-Analysis for event of dry atmospheric loss from the NH between 3 and 12 Mar 1989. Contour interval (CI) is 5 hPa, with negative (positive) contours dashed (shaded). The zero contour has been omitted. Inset plots of the NH (solid) and SH (dash) dry air surface pressure anomalies (hPa), calculated from the ECWMF Re-Analysis. Vertical line denotes 3 Mar (onset date): (a) 26 Feb, (b) 28 Feb, (c) 2 Mar, (d) 4 Mar, (e) 6 Mar, (f) 8 Mar, (g) 10 Mar, and (h) 12 Mar 1989.
Fig. 6. Shown here are the 500-hPa geopotential height (thin solid contours) and anomalies (shaded) with CI of 12 (10) dam, respectively, from the ECWMF Re-Analysis for the event of dry atmospheric loss from the NH between 3 and 12 Mar 1989. Positive (negative) anomaly contours are solid (dashed). The inset plots are the same as in Fig. 5: (a) 26 Feb, (b) 28 Feb, (c) 2 Mar, (d) 4 Mar, (e) 6 Mar, (f) 8 Mar, (g) 10 Mar, and (h) 12 Mar 1989.
southeastern Pacific have reduced in magnitude from 4 to 6 March (Figs. 5d,e), indirectly contributing to the rise in SH dry atmospheric mass.

Five days after the start of the NH dry atmospheric mass collapse event atmospheric mass has begun to build over the region of the South Pacific Ocean (Fig. 5f). Similar results were also found for the 25-case composite. At 500 hPa (Fig. 6f), a broad ridge occupies the region of the South Pacific to the west of the date line, with a lower-latitude ridge extending southeastward.
Fig. 8. Dry atmospheric mass flux potential ($5 \times 10^9$ kg s$^{-1}$), along with the zonal and meridional components of the dry atmospheric mass flux potential (kg m$^{-1}$ s$^{-1}$). (a)–(f) The layer from the surface to 500 hPa and (a')–(f') the layer from 500 to 10 hPa are shown. Also plotted in each panel are the OLR anomalies (W m$^{-2}$; shaded): (a), (a') 28 Feb; (b), (b') 2 Mar; (c), (c') 4 Mar; (d), (d') 6 Mar; (e), (e') 8 Mar; and (f), (f') 10 Mar 1989.
from Australia. In the larger composite we also see evidence of a statistically significant ridge orientated northwest–southeast over the Australia–New Zealand sector five days after onset. On 10 (Figs. 5g and 6g) and 12 March (Figs. 5h and 6h), the SH dry atmospheric mass is at a maximum, and much of the buildup is manifested as atmospheric mass increases in the midlatitude ridges to the southeast and southwest of New Zealand and to the southwest of Australia in the south Indian Ocean. At 500 hPa (Figs. 6g,h), three prominent ridges overlay the positive atmospheric mass anomalies, developing from a primarily zonal flow to the south of Australia on 6 March 1989 (Fig. 6e).

b. Pressure surge and tropical convection anomalies

Longitude–time plots of OLR anomalies and 200-hPa velocity potential (χ) for the first three months of 1989 are shown in Fig. 7. The OLR anomalies were calculated by removing the local seasonal cycle defined as the mean plus the first four harmonics of the 25-yr period of 1979–2003. The divergent winds are directed from lower to higher values of χ, with their speed proportional to the horizontal gradient of χ. Knutson and Weickmann (1987) found the 150- and 200-hPa levels the most appropriate to examine the velocity potential related to tropical convection. In Fig. 7a, the OLR and velocity potential values are averaged between 5°N and 5°S, while in Fig. 7b the values are averaged between 5° and 15°S.

Beginning on or around 26 February, five days prior to the onset of the NH dry atmospheric mass depletion event, convection flares up over the Indian Ocean sector (Fig. 7a), albeit not as intense as in the middle of January. Between 26 February and 3 March a weak convective signal is seen to propagate eastward from the Indian Ocean toward the western Pacific. Of particular interest is the intense convection that commences on 6 March near 140°E (Fig. 7a). In the subtropical latitudes of the SH, the convection is even more pronounced, centered near 8 March (Fig. 7b). The region with OLR anomaly values less than −40 W m⁻² covers a vast area and agrees well with the velocity potential data (Fig. 7b). Keith et al. (1991) termed this period in early March the third active phase of the ASM, producing widespread heavy rainfall (Mills and Zhao 1991; Zhao and Mills 1991; Drosdowsky 1996; Suppiah and Wu 1998).

Also plotted in Fig. 7 are time series of the area-averaged SLP for two areas: one situated over northern Eurasia (40°–75°N, 85°–145°E; hereafter referred to as the northern box), the other area located over southern China (20°–30°N, 105°–120°E; hereafter referred to as the southern box). These boxes are the same as those used in CG03. To provide some perspective, the dashed
line in the bottom right panel of Fig. 7 shows a time series of the area-averaged SLP for the NH, highlighting this extreme event amid the larger winter season NH SLP evolution. Commencing 24 February, the area-averaged SLP in the northern box begins to rise and reaches a maximum on 2 March, a total rise of nearly 18 hPa, the most rapid increase throughout the entire 3-month period. This period of rise in area-averaged SLP is denoted by northern rise in Fig. 7. Considering that the SLP is averaged over such a large area, the rise of 18 hPa is impressive. We also note that the rise in area-averaged SLP that starts as convection is enhanced over the Indian Ocean sector near 90°E (Figs. 7a,b). At the time of maximum area-averaged SLP for the northern box, the region of intense convection has shifted eastward to the tropical western Pacific near 140°E (Fig. 7a).

The fall in area-averaged SLP for the northern box, from 2–11 March, denoted by northern fall in Fig. 7, is remarkable. The total fall is approximately 30 hPa in nine days, clearly the largest such fall during the entire 3-month period. Accompanying the fall is the outbreak of intense convection, both along the equator (Fig. 7a), and especially to the south over the northern part of Australia, starting 6 March (Fig. 7b). The surging of atmospheric mass southward is shown by the curve for the southern box (thin solid) in Fig. 7. The time of peak atmospheric mass in the southern box is shown by the dashed line occurs just prior to the outbreak of intense convection in the monsoon trough of northern Australia suggesting a possible role by the Southeast Asian pressure surge in the initiation of deep convection (Slingo 1998). The curves for the northern and southern boxes are very similar to the 25-case composite as seen in Fig. 8a of CG03.

The increase in the area and intensity of convective activity over the South China Sea, the Philippine Sea, and the Indochina coast following pressure surges is well documented (Chang and Lau 1980; Lau et al. 1983; Meehl et al. 1996; Compo et al. 1999; Slingo 1998; Garreaud 2001). The increased convection appears as part of an overall enhancement of the local Hadley circulation that leads to circulation anomalies in the extratropics of both hemispheres (Chang and Lau 1980; Chu and Park 2001). In the Northern Hemisphere extratropics the East Asian jet stream is found to strengthen and shift eastward largely in response to the Coriolis torque exerted on the enhanced northward-divergent meridional outflow (Lau et al. 1983; Chang and Lum 1985).

c. Local Hadley circulation intensification

To illustrate the intensification and southward movement of the local Hadley circulation over Southeast Asia we show in Fig. 8 the upper- and lower-tropospheric dry atmospheric mass flux potentials. The vertically integrated dry atmospheric mass flux $M_j$ can be partitioned into upper- and lower-layer contributions by partitioning the vertical integral in (A4). We consider two separate meridional dry atmospheric mass fluxes, the first from the surface to 500 hPa, the second from 500 to 10 hPa, with their sum equal to the total dry atmospheric mass flux for the given column. In a study of the atmospheric mass budget in the Tropics based upon the divergent wind, Newell et al. (1996) also used the 500-hPa level to partition the atmospheric mass flux into upper and lower levels. Also shown in Fig. 8 are the OLR anomalies.

On 28 February (Figs. 8a,a’) and 2 March (Figs. 8b,b’) the dry atmospheric mass circulation is prominent in the zonal direction. This so-called Walker circulation is clearly evident, spanning the entire tropical Pacific Ocean sector linking the region of lower- (upper-) tropospheric dry atmospheric mass convergence (divergence) over the western tropical Pacific to the region of lower- (upper-) tropospheric dry atmospheric mass divergence (convergence) over the tropical eastern Pacific. Another prominent local overturning circulation is observed in the Eastern Hemisphere linking the region of south-central Asia near 30°N to the heat source over the tropical western Pacific. This northwest–southeast-orientated overturning circulation is what is commonly referred to as the local East Asian Hadley cell (EAHC; Lau et al. 1983; Chu and Park 1984). Over the tropical western Pacific the negative OLR anomalies are small in magnitude, which supports the notion of a weak EAHC on 28 February (Figs. 8a,a’) and 2 March (Figs. 8b,b’).

The EAHC intensifies with time, notably between 4 and 10 March (Figs. 8c–f and 8c’–f’). The lower-tropospheric southward and southeastward dry atmospheric mass fluxes from central Asia toward the tropical western Pacific intensify dramatically (Figs. 8d–f). Much of the dry atmospheric mass flux emanates from the weakening anticyclone over China and coastal Asia (see Fig. 9). It is important to note that the lower-tropospheric convergence of dry atmospheric mass into the monsoon trough of northern Australia is asymmetric with the southward flux from the NH much more intense than with the northward flux from the SH (Figs. 8e,f). A similar enhancement of the lower-tropospheric southward cross-equatorial dry atmospheric mass flux occurs over tropical South America (Figs. 8d–f). Over Southeast Asia the northward return flow at upper levels also intensifies in conjunction with the low-level southward flow (Figs. 8d’–f’). On 8 (Figs. 8e,e’) and 10
March (Figs. 8f,f′) the EAHC is clearly the most vigorous dry atmospheric mass overturning circulation present over the globe.

The SLP analysis in Fig. 9 depicts twin cyclonic disturbances, one in each hemisphere, on 7 March in the Asian–Australian sector (Fig. 9a). After this time the cyclonic disturbance within the monsoon trough of the SH deepens (Figs. 9c,d), while the NH cyclone exhibits little SLP change. Owing to the asymmetric intensification, there must be a net atmospheric mass decrease in the atmospheric column over the SH low center. This implies that the intensified lower-tropospheric convergence of dry atmospheric mass into the monsoon trough must be accompanied by a more amplified upper-tropospheric dry atmospheric mass divergence above such as to provide a net divergence of dry atmospheric mass within the column. Only a portion of the divergent outflow flows northward back into the NH; the other portion spreads southward into the SH, leading to a net cross-equatorial exchange of dry atmospheric mass in this sector.

One can clearly see the southward migration of the center of convergence (divergence) of dry atmospheric mass at lower- (upper-) tropospheric levels north of Australia from 28 February (Figs. 8a,a′) to 10 March (Figs. 8f,f′). The southward displacement of the convergent center and the associated monsoon trough over northern Australia is characteristic of active phases of the ASM (Murakami and Sumi 1982; Chen et al. 1989). Note the pronounced meridional expansion of the zone of dry divergent atmospheric mass outflow at upper levels into the extratropics of the SH after 6 March (Figs. 8d–f and 8d′–f′).

On 8 (Figs. 8e,e′) and 10 March (Figs. 8f,f′) we can clearly see the southeastward extension over Australia of the lower- (upper-) tropospheric convergent (divergent) dry atmospheric mass flux associated with the deepening surface trough (Figs. 9a–d). Dry atmospheric mass converges into the deepening surface trough in the lower troposphere and diverges at upper levels, redistributing atmospheric mass to the south and south-east of Australia. A pronounced southwestward upper-tropospheric dry atmospheric mass flux, emanating from the convection over northwestern Australia, converges in the south Indian Ocean. Consistent with the deepening surface pressure trough, the dry atmospheric mass divergence in the upper troposphere exceeds the convergence in the lower troposphere.

Fig. 9. Time evolution of daily averaged SLP, CI 2 hPa, and outgoing longwave radiation anomalies (W m−2), with negative values shaded according to the given scale. All data are from the ECMWF Re-Analysis: (a) 7 Mar, (b) 8 Mar, (c) 9 Mar, (d) 10 Mar 1989.
4. Role of diabatic heating in dry atmospheric mass redistribution

In this section we examine the role of the diabatic heating anomalies associated with the onset of an active phase of the ASM in redistributing the atmospheric mass. Both observational (Hurrell and Vincent 1990; Berbery and Nogués-Peagél 1993; Tyrrell et al. 1996) and modeling studies (Matthews et al. 1996; Renwick and Revell 1999) support the notion of a strong linkage between convection over the summer monsoon region and extratropical responses in the SH on intraseasonal time scales.

In the following we make use of the concept of potential vorticity (PV; Hoskins et al. 1985). Following Bosart and Lackmann (1995), PV in isobaric coordinates in a hydrostatic atmosphere can be approximated to a very high degree as

\[ \text{PV} = -g \frac{\partial \theta}{\partial p} \left( f + \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) + g \left( \frac{\partial v}{\partial p} \frac{\partial \theta}{\partial x} - \frac{\partial u}{\partial p} \frac{\partial \theta}{\partial y} \right), \]

(1)

where \( g \) is the gravitational acceleration, \( \theta \) is the potential temperature, and \( u \) and \( v \) are the zonal and meridional wind components, respectively. The pressure derivatives in (1) are evaluated using centered finite differences. The units of PV are \( \text{m}^2 \text{s}^{-1} \text{K kg}^{-1} \) and values are typically plotted in potential vorticity units (PVU), where 1 PVU is equal to \( 10^{-6} \text{ m}^2 \text{s}^{-1} \text{K kg}^{-1} \).

PV is materially conserved for frictionless, adiabatic flow making it a suitable variable for the study of diabatic processes. The material surface of constant PV (generally close to 2 PVUs in magnitude) separating tropospheric air from stratospheric air is denoted the dynamic tropopause (DT; Morgan and Nielsen-Gammam 1998). If potential temperature is interpolated onto the DT \( (\theta_f) \), it is materially conserved and should change locally only through advection. Hence, large changes in \( \theta_f \), which cannot be explained by advection, result from the influence of vertical gradients in diabatic heating. Accordingly, regions of PV nonconservation are immediately apparent when visualizing a sequence of DT maps.

a. Dynamic tropopause analysis

Figure 10 shows \( \theta_f \) and winds interpolated onto the DT, defined as the location of the \( -2 \)-PVU surface. Recall that PV is typically negative in the SH. The location of the \( -2 \)-PVU surface was found by searching upward (toward lower pressure) from 700 hPa for the first occurrence of PV values of \( -2 \) PVUs. Given the pressure of the DT, values of \( \theta_f, u, \) and \( v \) were linearly interpolated with respect to pressure onto the DT surface.

Higher (lower) values of \( \theta_f \) reflect higher (lower) elevations of the DT surface. In Fig. 10, \( \theta_f \) generally decreases poleward reflecting the lower tropopause heights in the extratropical and higher latitudes. Across the entire domain, the gradients of \( \theta_f \) are amplified in two principal zones: in the extratropics near \( 55^\circ \text{S} \), and in the subtropics in the vicinity of \( 25^\circ-30^\circ \text{S} \). These amplified gradients are regions where the DT is steeply sloped and reflect the locations of the polar and subtropical jets, respectively. The winds interpolated onto the DT generally follow the contours of \( \theta_f \). The higher-latitude jet is zonal in character between 4–6 March (Figs. 10a–c); however, significant wavelike perturbations develop thereafter.

Between 7 (Fig. 10d) and 10 March (Fig. 10g) a dramatic warming of the DT is seen over southeastern Australia stretching southeastward toward New Zealand. By 11 March (Fig. 10h) the spatial extent of the DT warming is impressive. Referring back to Figs. 8e,f and 8e′,f′ it can be seen that the zone of DT warming coincides with the large region of negative OLR anomalies, stretching from Australia southeastward to New Zealand, which is indicative of the intense convection and associated latent heat release associated with this trough system. In the vicinity of the DT warming, the winds generally flow along the \( \theta_f \) contours and clearly cannot account for the southeastward spread of the \( \theta_f \) contours simply through advection. The configuration of the winds on the DT and the presence of deep convection suggest that the DT warming to the southeast of Australia results from diabatic heating, although we cannot completely discount the influence of other processes, such as wave-mixing processes, in contributing to the nonconservation of \( \theta_f \).

b. Dry atmospheric mass redistribution

The evolution of the OLR anomalies and the upper-tropospheric dry atmospheric mass fluxes has been discussed previously (see Fig. 8). The goal here is to relate their evolutions to the deep-layer upper-tropospheric PV field and the resulting SLP field, which is a reflection of the entire atmospheric mass within the column. In Fig. 11 we show the 100–500-hPa layer-averaged PV, OLR anomalies (as in Fig. 9), upper-tropospheric (500–10 hPa) zonal and meridional components of the dry atmospheric mass flux potential, and SLP.

The development of a pronounced upper-tropospheric ridge over southeast Australia (40°S, 160°E) extending to New Zealand can be seen in the PV field after 8 March (Fig. 11a). The development occurs slightly to the east of the enhanced convection.
near 37.5°S, 142.5°E. We outlined in the previous section that the lifting of the DT (i.e., upper-tropospheric ridging) in this region results largely from the influences of diabatic heating. As the ridge intensifies a simultaneous amplification of an upper-tropospheric trough occurs downstream near 20°S, 160°E. This form of downstream development linked to outbreaks of convection over the tropical western Pacific has been shown to be important for the initiation of active phases of the South Pacific convergence zone (SPCZ; Matthews et al. 1996). Enhanced convective activity is seen downstream of the upper-tropospheric trough near
25°S and 180°E (denoted C1) commencing 9 March (Fig. 11b).

On 8 (Fig. 11a) and 9 March (Fig. 11b) a channel of intense dry atmospheric mass outflow extends south-eastwards from the anomalous convection over eastern Australia and converges in the vicinity of New Zealand (45°S, 180°E). In Fig. 11 the SLP contours are shown at an interval of 4 hPa for values greater than 1020 hPa. It can be seen that beneath the area of upper-tropospheric dry atmospheric mass convergence a surface anticyclone is building (denoted H1). The anticyclone, initially located at approximately 45°S, 180°E on 8 March (Fig. 11a) has a central pressure of 1021 hPa, which increases to 1027 hPa by 9 March (Fig. 11b) as the system propagates eastward. By 10 (Fig. 11c) and 11 March (Fig. 11d) the anticyclone has moved to the east-
ern Pacific and represents the positive SLP anomaly seen on 10 March (Fig. 5g) near 50°S, 150°W and 12 March (Fig. 5h) near 45°S, 130°W. It should be pointed out that on 8 (Fig. 11a) and 9 March (Fig. 11b) the southward dry atmospheric mass outflow from the intensifying convection downstream of the upper-tropospheric trough near 25°S, 180°E (C1) also contributes to the dry atmospheric mass convergence in the vicinity of H1.

A second channel of upper-tropospheric dry atmospheric mass divergence is seen as early as 8 March (Fig. 11a). The outflow extends southward from the intensifying anomalous convection to the southeast of Australia (37.5°S, 142.5°E). The dry atmospheric mass converges to the south and southeast in the vicinity of the building of the anticyclone located south of Tasmania (47.5°S, 150°E; denoted H2 in Fig. 11) on 10 March (Fig. 11b). This anticyclone (H2) propagates eastward

Fig. 11. (Continued)
and by 12 March (Fig. 11e) is situated to the southeast of New Zealand and represents a pronounced positive SLP anomaly (see Fig. 5h). From 9 to 11 March the central pressure of H2 increases from 1022 to 1026 hPa. On 12 March (Fig. 9d), the positive SLP anomaly can also be seen curving northwestward around New Zealand. This is consistent with the large upper-tropospheric convergence of dry atmospheric mass seen on 10 (Fig. 11c) and 11 March (Fig. 11d) near 35°S, 170°E. The anomalous convection downstream of the upper-tropospheric trough near 25°S, 180°E (C1) has intensified from 8 March and contributes to the upper-tropospheric dry atmospheric mass convergence into this region to the northwest of New Zealand.

Results from Fig. 5 show that on 12 March, the peak time of dry atmospheric mass in the SH, a large positive SLP anomaly occupies the region to the south and southwest of Australia (50°S, 100°E). This SLP anomaly is associated with a strong anticyclone and is denoted H3 in Fig. 11. A pronounced channel of dry atmospheric mass outflow extends southwestward from the deepening monsoon depression and region of intense convection over northwest Australia and contributes to the building of the anticyclone (H3; Figs. 11b–e). Between 9 and 12 March the central SLP of H3 increases from 1022 to 1030 hPa.

In Fig. 11 only the upper-tropospheric dry atmospheric mass flux was presented. From Fig. 8 it is evident that the lower-tropospheric dry atmospheric flux is directed oppositely to that of the upper troposphere and it may be wondered whether both dry atmospheric mass fluxes cancel in the vertical. The evidence that the upper-tropospheric dry atmospheric mass convergence in the vicinity of the anticyclones exceeds the divergence at lower levels comes from the increasing SLP values of the respective anticyclonic systems (H1, H2, and H3). SLP is a reflection of the atmospheric mass within the entire column.

5. Summary

This study built upon the prior work of CG03, examining the physical mechanism of Southeast Asian pressure surges in relation to the depletion of dry atmospheric mass from the NH. Unlike prior studies, which utilized pressure data, this study used the divergent winds to depict the redistribution of atmospheric mass. A comparison of the dry divergent winds with the area-integrated dry air surface pressure data indicated that the NCEP Reanalysis–divergent winds possessed a bias in the NH subtropics, most likely associated with a weaker ITCZ. The ERA-15-divergent wind data did not possess this bias, which justified their use for the detailed case study.
For the detailed case study, we chose an event of NH atmospheric mass depletion that possessed a significant pressure surge over Southeast Asia. The SLP analysis combined with the OLR data were strongly suggestive of enhanced interhemispheric interaction associated with the Southeast Asian pressure surge. We were able to show that the pressure surge was linked to a strengthening of the local East Asian Hadley circulation in association with the onset of an active phase of the ASM. The importance of the diabatic heating anomaly was to redistribute the dry atmospheric mass in the divergent outflow of the upper troposphere. We showed that three prominent anticyclonic circulations, within the SH extratropics stretching from the south Indian Ocean eastward to the South Pacific, intensified in the regions where the dry atmospheric mass converged in the upper troposphere. The pressure surge over Southeast Asia is not the sole physical process that ultimately leads to the anticyclonic developments within the Southern Hemisphere extratropics, but we argue that the pressure surge has played an important role through its influence upon the outbreak of atmospheric convection within the monsoon trough of northern Australia.

This particular case of dry atmospheric mass loss from the NH was chosen based upon the occurrence of a significant pressure surge over Southeast Asia. A question arises as to the robustness of the tropical response (i.e., convection) resulting from the tropical–extratropical interaction and the Southern Hemisphere response to Southeast Asian pressure surges (Compo et al. 1999). Compo et al. (1999) have shown in their study of East Asian winter monsoon pressure surges that the SH response to the surges is not as robust as the convective and circulation signals in the NH tropical regions. The SH response is complex and more work needs to be done to ascertain the individual contributions to the anticyclonic development in the SH.

Acknowledgments. The NCEP Reanalysis data were provided by NOAA/CiRES Climate Diagnostics Center, Boulder, Colorado, from their Web site (available online at http://www.cdc.noaa.gov), and we acknowledge their support. The ECMWF Re-Analysis data were acquired from the Mass Storage System at the National Center for Atmospheric Research (NCAR).

Financial support for this research was provided by the Atmospheric Environment Service of the government of Canada through a postgraduate fellowship for one year, and also by the Natural Science and Engineering Research Council of Canada (NSERC) through a two-year postgraduate fellowship. Subsequent support was provided by grants from NSERC and from the Meteorological Service of Canada and an NSERC network research grant on the Mackenzie GEWEX Project.

We thank the three anonymous reviewers for the helpful and insightful comments that have improved the manuscript.

APPENDIX

Dry Atmospheric Mass Redistribution and Vertically Integrated Dry Meridional Winds

In this section we formulate the relationship between area-integrated dry air surface pressure and the vertically and zonally integrated dry meridional winds at the boundary of the area. The mass of the atmosphere as represented by surface pressure is simply the arithmetic sum of the surface pressure owing to dry air and the surface pressure owing to the vertically integrated water vapor. This can be written as

\[ P_s = P_d + P_w. \]  

where \( P_s \) represents the surface pressure, \( P_d \) the surface pressure owing to dry air, and \( P_w \) the surface pressure owing to vertically integrated water vapor. The surface pressure tendency can be written as

\[ \frac{\partial P_s}{\partial t} = \frac{\partial P_d}{\partial t} + \frac{\partial P_w}{\partial t}. \]

The surface pressure tendency owing to dry air, \( \frac{\partial P_d}{\partial t} \), is given by the divergence of the vertically integrated dry airmass flux (Trenberth 1991; Chen et al. 1997).

\[ \frac{\partial P_d}{\partial t} = -g \nabla \cdot \mathbf{M}_d, \]

where \( g \) is gravitational acceleration and \( \mathbf{M}_d \) is the vertically integrated dry airmass flux given by

\[ \mathbf{M}_d = \frac{1}{g} \int_{p_{\text{top}}(x,y)}^{p_d(x,y)} (1 - q) \mathbf{v} \, dp, \]

where \( \mathbf{v} \) represents the horizontal wind vector, \( q \) the specific humidity, \( p_d(x,y) \) the surface pressure, and \( p_{\text{top}}(x,y) \) the pressure at the top of the given atmospheric column (which should be zero). Note that the surface pressure \( (p_s) \) is taken as a function of \( x \) and \( y \). In many previous studies, the surface pressure has been assumed to be spatially uniform, an incorrect assumption that can introduce significant errors in the atmospheric mass budget near steep orography (Trenberth 1991; Van den Dool and Saha 1993; Trenberth and Guillemot 1995). Denoting a globally integrated quantity as \( \int_{\text{GB}} \) and a hemispherically integrated quantity as \( \int_{\text{hs}} \).
as \([\Pi]\), following the notation of Chen et al. (1997), we can write the conservation of dry airmass for the globe as

\[
\frac{\partial[p_d]{\Pi}_{GH}}{\partial t} = \frac{\partial[p_d]{\Pi}_{NH}}{\partial t} + \frac{\partial[p_d]{\Pi}_{SH}}{\partial t} = 0. \tag{A5}
\]

But we know from (A3) that

\[
\frac{\partial[p_d]{\Pi}_{NH}}{\partial t} = -g(\nabla \cdot \mathbf{M}_d)_{NH}, \tag{A6}
\]

and

\[
\frac{\partial[p_d]{\Pi}_{SH}}{\partial t} = -g(\nabla \cdot \mathbf{M}_d)_{SH}, \tag{A7}
\]

which, when combining (A5), (A6), and (A7), leads to

\[
\frac{\partial[p_d]{\Pi}_{NH}}{\partial t} = g(\nabla \cdot \mathbf{M}_d)_{SH}. \tag{A8}
\]

From (A8) we conclude that the NH dry airmass can change only if there is a net flux of dry airmass into or out of the hemisphere, and there are no source or sink terms. This net flux must come through the boundary given by the equator. If we now replace the \([\Pi]\) by integrals we can rewrite (A6) as

\[
\int \int_{\text{NHArea}} \frac{\partial p_d}{\partial t} dA = \int \int_{\text{NHArea}} (\nabla \cdot \mathbf{M}_d) dA.
\]

Green’s theorem in vector calculus states that

\[
\int \int_R (\mathbf{v} \cdot \mathbf{n}) dA = \oint_C \mathbf{v} \cdot dS, \tag{A10}
\]

where \(R\) denotes the region bounded by the curve \(C\) and \(\mathbf{n}\) is a unit vector pointing normally outward from the boundary of the curve \(C\). When considering the Northern Hemisphere, the bounding curve for the area is the equator, and hence we can replace (A9) with the following:

\[
\int \int_{\text{NHArea}} \frac{\partial p_d}{\partial t} dA = -g \oint_{\text{Equator}} \mathbf{M}_d \cdot \mathbf{n} dS.
\]

\[
\int \int_{\text{NHArea}} \frac{\partial p_d}{\partial t} dA = g \oint_{\text{Equator}} \mathbf{M}_d dS. \tag{A11}
\]

Given that the bounding curve is that of the equator, the unit vector \(\mathbf{n}\) is directed southward into the Southern Hemisphere \([\mathbf{n}=(0, -1)]\), and (A11) reduces to

\[
\int \int_{\text{NHArea}} \frac{\partial p_d}{\partial t} dA = g \oint_{\text{Equator}} \mathbf{M}_d dS. \tag{A12}
\]

where \(M_d\) represents the meridional component of the vertically integrated dry airmass flux. From (A12), we have a direct relationship between the vertically and zonally integrated dry meridional winds at the equator and the time rate of change of the Northern Hemisphere dry atmospheric mass. The relation expressed in (A12) is valid for any so-called “polar cap” bounded by a given latitude wall, where the integral on the left-hand side is over the area bounded by the polar cap and the given latitude wall, and the line integral on the right-hand side is replaced by a line integral over the given latitude wall.

The vertically integrated dry airmass flux \(\mathbf{M}_d\) is a two-dimensional vector, and from Helmholtz theorem (Holton 1992, p. 386) can be expressed in terms of a streamfunction \(\psi\) and velocity potential \(\chi\) as follows:

\[
\mathbf{M}_d = k \times \nabla \psi + \nabla \chi = \mathbf{M}_d' + \mathbf{M}_d''. \tag{A13}
\]

In (A13) \(\mathbf{M}_d'\) and \(\mathbf{M}_d''\) are the rotational and divergent components, respectively, of the vertically integrated dry airmass flux. The divergence of the rotational component of the vertically integrated dry mass flux is zero (i.e., \(\nabla \cdot \mathbf{M}_d' = 0\), and so only the divergent component of \(\mathbf{M}_d\) contributes a nonzero value to the right-hand side of (A9). Independent calculations of the right-hand side of (A11) using \(\mathbf{M}_d\) from NCEP show that it is approximately zero.

To calculate daily average values of \(\mathbf{M}_d\) we first calculated \(\mathbf{M}_d\) at 6-h intervals (4 times daily) as in (A4) using a pressure-weighted trapezoidal rule following Trenberth (1991), with an upper limit of integration of 10 hPa [i.e., \(p_{top}(x, y) = 10\) hPa in (A4)]. Above 300 hPa, where no specific humidity values are given in the NCEP Reanalysis, we assumed a value of 0 (i.e., \(q = 0\)) in the calculation of \(\mathbf{M}_d\). This effect is minor as values of \(q\) above 300 hPa account for only 0.2% of the total area-averaged surface pressure owing to vertically integrated water vapor for the globe (Trenberth et al. 1987). The 4-times-daily values were subsequently averaged to arrive at a daily averaged value of \(\mathbf{M}_d\). To evaluate \(\mathbf{M}_d''\) we solved a Poisson equation of the form

\[
\nabla^2 \chi_M = \nabla \cdot \mathbf{M}_d', \tag{A14}
\]

where \(\chi_M\) represents the dry atmospheric mass potential function, using a technique known as cyclic reduction (Swarztrauber and Sweet 1975). The routine is similar to that used by Chen (1985) and Weickmann and Khalsa (1990). The Poisson equation was evaluated over a global domain, where no lateral boundary conditions were necessary, and the potential function \(\chi_M\) was uniquely determined with the exception of a constant of integration. The constant of integration is ar-
bitrary, as the gradient of $\chi_M$ determines the zonal and meridional components of the divergent vector.

**REFERENCES**


Morgan, M. C., and J. W. Nielsen-Gammon, 1998: Using tropo-


