Two Types of Baroclinic Life Cycles during the Southern Hemisphere Summer

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ABSTRACT

Baroclinic eddy life cycles of the Southern Hemisphere (SH) summer are investigated with NCEP–NCAR reanalysis data. A composite analysis is performed for the years 1980 through 2004. Individual life cycles are identified by local maxima in synoptic-scale eddy energy. Two types of baroclinic life cycles are examined, each defined by the strength of the barotropic energy conversion 2 days prior to the maximum baroclinic growth. For one life cycle, the barotropic conversion is anomalously weak before the maximum baroclinic growth; for the other, the barotropic conversion is anomalously strong. These two life cycles are referred to as the weak barotropic (WB) and strong barotropic (SB) life cycles.

The analyses for the WB life cycle find that a poleward anomalous wave activity flux is observed within the SH tropics and subtropics just before the initial growth of the synoptic-scale eddies. In contrast, the SB life cycle exhibits an equatorward anomalous wave activity flux prior to the initial wave development. For the WB life cycle, these changes in the wave activity flux are shown to induce a mean meridional circulation that weakens and broadens the midlatitude zonal mean jet and reduces the baroclinicity in the midlatitude lower troposphere. Opposite characteristics are observed for the SB life cycle. Since the eddy growth rate is found to be greater in the WB life cycle, these results suggest that the influences of the barotropic governor mechanism (a reduction in horizontal shear coinciding with more rapidly growing baroclinic eddies) and the midlatitude baroclinicity oppose each other at the beginning of the life cycle, with the former being dominant.

Both the WB and SB life cycles coincide with anomalous tropical convection. For the WB life cycle, there is a strengthening of the convection over the Maritime Continent, and for the SB life cycle there is a weakening in the convection over the same region. These results suggest that the two types of baroclinic life cycles are ultimately triggered by convection in the tropics.

1. Introduction

Throughout the past half century, beginning with Charney (1947) and Eady (1949), there has been steady advancement in our understanding of linear baroclinic instability [see Pierrehumbert and Swanson (1995) for an extensive review of this subject]. One aim of these studies was to better understand the dynamical processes that determine the three-dimensional spatial structure of growing midlatitude synoptic-scale waves. An extension of this research into the nonlinear realm, where both wave growth and wave decay were examined, was initiated by the analytical, weakly nonlinear studies of Pedlosky (1970, 1971). [See also Pedlosky (1982) and Warn and Gauthier (1989), where it was shown that the weakly nonlinear theory breaks down near the minimum critical shear because the basic wave is resonantly excited by its higher harmonics.] Although these studies provided much insight into wave–mean flow interaction for finite-amplitude baroclinic waves, their relevance for the atmosphere was limited by the absence of meridional shear of the basic state zonal-mean zonal wind.

Subsequent studies of the nonlinear growth and decay of baroclinic waves, usually referred to as baroclinic life cycles, used nonlinear, multilevel numerical models. This numerical modeling approach was adopted since the attainment of analytical solutions for a zonal mean flow with both meridional and vertical shear was not
found to be feasible. The most striking nonlinear features of baroclinic life cycles were found to occur during the decay stage of the life cycle (Gall 1976; Simmons and Hoskins 1978, 1980). After the initial wave growth, which occurred via the poleward eddy heat flux in the lower troposphere as in classical baroclinic instability, a substantial increase in the poleward eddy momentum flux was shown to take place in the upper troposphere. This eddy momentum flux, which accounted for the wave decay, was accompanied by a strong eastward acceleration of the zonal-mean zonal wind. From an energetics perspective (Lorenz 1955), the wave growth corresponds to an energy transfer from the zonal mean available to the eddy available potential energy, and the wave decay to an energy transfer from the eddy kinetic to the zonal-mean kinetic energy. This sequence in the energy conversions has often been referred to as baroclinic growth followed by barotropic decay (e.g., Held and Phillips 1987). In a study of baroclinic life cycles with a two-layer model, Feldstein and Held (1989) found that the waves decay barotropically when there is a critical level (where the phase speed of the waves matches that of the zonal-mean zonal wind) present in the upper layer. Otherwise, the waves decay baroclinically, as in the weakly nonlinear modeling studies of Pedlosky (1970, 1971). Feldstein and Held (1989) showed that wave breaking and irreversible mixing of potential vorticity take place within the vicinity of the model’s critical levels.

To examine the extent to which the model life cycle solutions apply to the atmosphere, Randel and Stanford (1985a,b) undertook an observational analysis of baroclinic life cycles that occurred during three consecutive Southern Hemisphere summer seasons. Their study verified that the numerical models had correctly simulated the observed energetics, as they found numerous life cycles that were characterized by baroclinic growth followed by barotropic decay.

While the above models were successful at capturing the energetic features of the observed life cycles, they did have important shortcomings that relate to the initial wave perturbation. For example, in the model calculations, the amplitude of the initial perturbation was always specified to be very small. The choice of such a small wave amplitude ensured that the perturbation would evolve into the fastest-growing normal mode. However, the specification of a small amplitude initial perturbation suppresses the occurrence of sizable eddy heat and momentum fluxes, which are always present in the atmosphere. These limitations on the amplitude of the initial perturbation make it difficult to identify the underlying dynamical processes that trigger baroclinic life cycles in the atmosphere. This issue can be highlighted by examining the composite baroclinic life cycle evolution (see Fig. 1). This composite is derived from those life cycles for which the maximum synoptic-scale eddy energy (Fig. 1b), during the Southern Hemisphere summer, exceeds one standard deviation. We focus on this particular season because of its relatively greater degree of zonal uniformity as compared to other seasons in either hemisphere and because that season is dominated by a single eddy-driven jet, rather than by both an eddy-driven and a subtropical jet, as in the winter season (Lee and Kim 2003). For this composite calculation, 69 events, or life cycles, are selected. The lag 0 day corresponds to local maximum in the synoptic-scale eddy energy. One striking feature displayed in

![Composite baroclinic life cycle evolution](image-url)
Figs. 1a and 1d is that there is a weakening in the total (all zonal wavenumbers) and synoptic-scale barotropic conversion that peaks at lag $-2$ days. (In this study, we will focus on the total barotropic conversion at negative lags. This is because, as will be discussed in section 4d, the total eddy momentum fluxes drive a mean meridional circulation that has a strong impact on the horizontal shear of the zonal mean wind and the meridional temperature gradient at the beginning of the life cycle.) This feature, which was also observed by Randel and Stanford (1985a,b), and in the wave packet study of Chang (2005), was not present in the model calculations of baroclinic life cycles, simply because the perturbation in the models was initiated with a very small amplitude; that is, the initial barotropic conversion was close to zero.

The occurrence of this weakening in the total barotropic conversion term reveals the importance of the eddy momentum fluxes at the beginning of the life cycle, as the reduction in the barotropic conversion term must arise primarily from a weakening in the poleward eddy momentum fluxes. Furthermore, because the zonal-mean zonal wind is driven by eddy momentum fluxes, this observation alludes to the possibility that there may be a preferred structure for the zonal mean flow at the start of the life cycle. Motivated by this observation, in this study we will investigate whether there is a preferred eddy momentum flux and zonal-mean wind structure at the beginning of baroclinic life cycles. As we will see, such structures do indeed occur. We will then investigate the dynamical processes that account for these preferred structures and present results that suggest that the barotropic governor mechanism of James and Gray (1986) plays a prominent role.

An important consideration for our analysis is the observation that the majority of Southern Hemisphere summer baroclinic life cycles are characterized by synoptic-scale waves that are largely confined to localized wave packets spanning about 120° longitude (Chang 2005). This observation may suggest that our focus should be on the dynamical processes that take place within wave packets. However, Chang (2005) also finds that the local extremum in the barotropic conversion at negatives lags, as described above (see Fig. 4 of that study), arises from energy conversions that take place outside of the wave packet. This contrasts the subsequent baroclinic and barotropic conversions, which occur primarily inside the wave packet. These findings therefore suggest that, although throughout most of the life cycle the energy and energy conversion are confined to localized wave packets, the triggering of life cycles occurs via processes that take place within the 240° longitude span that resides outside of the wave packet. Since we are investigating the full baroclinic life cycle, from the initial weakening of the barotropic conversion to the baroclinic growth and barotropic decay stages, we examine the energy and energy conversions integrated over the full Southern Hemisphere. Therefore, in this study, it may be helpful to keep in mind that the main contribution to the hemispheric energy conversions arises from either inside or outside the wave packet, depending on the stage of the baroclinic life cycle.

In section 2, the data and methodology are described. This is followed in section 3 by an investigation of the energetics for two types of baroclinic life cycles. In section 4 we examine the temporal evolution of the zonal mean flow, and in section 5 the impact of tropical convection. The discussion and conclusions follow in section 6.

2. Data and methodology

a. Data

We use daily (0000 UTC) National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data extending from 1 January 1980 to 31 December 2004 for our analysis. The horizontal resolution is truncated at rhomboidal 30, and there are 28 spaced sigma levels. For our analysis, all data are logarithmically interpolated onto 20 equally spaced pressure levels. As indicated in the introduction, the analyses are performed during the Southern Hemisphere summer, covering the months from December through February.

b. Methodology

In this study, baroclinic life cycles are examined with composites of different quantities, including zonal available potential energy, baroclinic and barotropic energy conversion terms, zonal-mean zonal wind, mass streamfunction, potential temperature, and Eliassen–Palm (EP) flux cross sections. All of the energetic terms are calculated following the primitive equation formulation in pressure coordinates. The mathematical expressions for these terms are shown in the appendix. All composites are based on a one standard deviation threshold for synoptic-scale eddy energy summed over zonal wavenumbers 4 through 7. Each of these zonal wavenumbers is found to exhibit similar characteristics. To ensure that consecutive life cycles do not overlap, for those life cycles that are separated by less than 15 days, the latter life cycle is discarded. As indicated in the introduction, the above criteria resulted in the selection of 69 life cycles. In this study, both total and anomalous values of different quantities are calculated. Anomalous values are obtained by subtracting the Southern Hemisphere summer climatological mean values.
As explained in the introduction, an examination of the dynamical processes associated with the local extremum in the total barotropic conversion term, before lag 0 days, is the main focus in this research. To investigate the characteristics of this extremum, we divide the life cycles into two separate composites based on the strength of the barotropic conversion term over the time interval from lag $-5$ days to lag $-1$ day. One composite comprises those life cycles with the weakest barotropic conversion over the lag $-5$ to lag $-1$ day time interval while the other composite comprises the strongest barotropic conversion over the same time interval. The former composite is referred to as the weak barotropic (WB) life cycle and the latter, the strong barotropic (SB) life cycle. For this study, the WB and SB composites each comprise 20 events. Together, these two composites contain slightly more than half the total number of cases.

### 3. Energetics of the WB and SB life cycles

The WB composite life cycle is shown in Fig. 2. Consistent with our selection criterion, there is a well-defined local extremum in the barotropic conversion at lag $-2$ days. The composite SB life cycle is shown in Fig. 3. As can be seen, the extrema in the eddy energy and energy conversion terms are smaller than those of the WB life cycle (Fig. 2). Most noticeably, over the lag $-5$ through lag 0 day interval, the increase in synoptic-scale eddy energy is markedly greater in the WB life cycle, and the barotropic conversion for the SB life cycle is much stronger than that of the WB life cycle (Fig. 2). The case-to-case variability of the eddy energy and energy conversions for the WB and SB life cycles is highlighted by showing the one standard deviation spread both above and below the composite value at each lag. As can be inferred from Figs. 2 and 3, the variability of the life cycles is sufficiently large that not all WB and SB cycles closely resemble their respective composites. However, an examination of each of the individual life cycles did find that most WB cases undergo a weakening of the barotropic conversion at the start of the life cycle, whereas most SB cases do not.

### 4. Zonal-mean field

#### a. Zonal-mean zonal wind

We next examine whether there are preferred structures for the zonal mean flow at the beginning of the WB and SB life cycles. If such preferred zonal mean flow structures are indeed found, then the natural questions to ask include what processes excite these particular zonal mean flow structures and why these zonal mean flow structures are followed by large increases in eddy energy. We begin our evaluation of these questions by showing the anomalous lower tropospheric (925 hPa) zonal-mean zonal wind and horizontal wind shear for the WB and SB composites (see Fig. 4). The corresponding upper tropospheric zonal-mean zonal wind and horizontal shear anomalies resemble those at 925 hPa (not shown). In the upper troposphere, the zonal-mean zonal wind is primarily driven by the eddy momentum flux convergence, whereas in the lower troposphere the zonal-mean zonal wind is mainly driven by the Coriolis torque. We chose the 925-hPa level rather than a level closer to the surface because of the increasing lack of reliability that arises from the intersection of the isobaric surfaces with the topography. During the Southern Hemisphere summer, the climatological zonal-mean zonal wind is characterized by a single eddy-driven jet centered at about 48°S in the
upper troposphere. The latitude of this jet is denoted by the straight line in Fig. 4. (See the NCEP–NCAR Reanalysis Atlas from the NOAA/Earth System Research Laboratory Web site at http://www.cdc.noaa.gov/data/ncep_reanalysis.)

For the WB composite, beginning at lag $24$ days, the 925-hPa zonal-mean zonal wind (Fig. 4a) is weakened at the center of the jet and strengthened on both the equatorward and the poleward sides of the jet, with this increase being much greater on the poleward side. The anomaly of the absolute value of the 925-hPa horizontal shear, $\left| \frac{\partial u}{\partial y} \right|$, where the brackets specify a zonal mean and the prime denotes an anomaly (Fig. 4c), indicates that the absolute value of the total horizontal shear has declined on both sides of the jet.

In contrast, for the SB composite, prior to lag 0 days, the primary impact of the zonal wind anomalies (Fig. 4b) is to strengthen the zonal-mean zonal wind on the equatorward side of the jet and to weaken the zonal-mean zonal wind on the jet’s poleward side. Furthermore, an increase in the maximum speed of the jet is found. As can be seen, this corresponds mostly to an increase in the absolute value of the total horizontal shear on both sides of the jet (Fig. 4d). Therefore, we see that before lag 0 days, there is an overall decrease in the horizontal shear in the WB life cycle, and vice versa for the SB life cycle.

The case-to-case variability of the 925-hPa zonal-mean zonal wind and horizontal shear was also examined. The WB life cycles were found to be more robust; that is, almost all of the WB life cycles showed a reduction in the horizontal shear at the beginning of the life cycle. In contrast, even though the composite SB life cycle exhibited an increase in the horizontal shear, there were some life cycles for which the horizontal shear underwent little change and others for which it decreased.

### b. Synoptic eddy EP flux

Composite EP flux diagrams (Edmon et al. 1980) for the synoptic-scale eddies (zonal wavenumbers 4–7) for the WB and SB life cycles are shown in Figs. 5 and 6. The anomalous values of the EP flux vectors and their divergence are illustrated. For the WB life cycle (Fig. 5), over the time interval from lag $-4$ to lag $-2$ days, in the upper tropical and midlatitude troposphere, the anomalous EP flux vectors are poleward. This corresponds to a weakening of the poleward eddy momentum flux and, therefore, a deceleration of the midlatitude jet. At lag $-2$ days, baroclinic growth begins, as indicated by an intensification of the vertical component of the midlatitude EP flux vectors, particularly at lower levels. By lag $-1$ days, the vertical propagation of wave activity reaches the upper troposphere and spreads both poleward and equatorward, resulting in an acceleration of the zonal-mean zonal wind in midlatitudes. At lag 0 days, the baroclinic growth of the synoptic eddies reaches its maximum, and most of the meridional wave activity propagation is toward the equator. By lag $+1$ days, the vertical propagation of wave activity is very much weakened and the equatorward propagation continues to intensify. The key point in Fig. 5 is that the extremum in the barotropic conversion term at negative lags (Figs. 2a and 2d) can be understood as arising from the poleward anomalous EP fluxes in the upper troposphere. Because the corresponding anomalous EP flux divergence in the upper troposphere occurs between 10° and 20°S, it is to be anticipated that there exists an anomalous source of Rossby wave activity at lower latitudes.

For the SB life cycle, in contrast to the WB life cycle, starting at lag $-4$ days (Fig. 6), the anomalous EP flux vectors at upper levels are seen to be oriented in an
equatorward direction and are located mostly on the equatorward side of the jet. This direction for the EP flux vectors suggests that there is an anomalous sink of Rossby wave activity in the tropics, which accounts for the absence of an extremum in the barotropic conversion term at negative lags (Fig. 3a) and results in a strengthening of the zonal-mean zonal wind on the equatorward side of the upper tropospheric jet. Prior to lag 0 days, the baroclinic growth in the lower troposphere remains weaker than that for the WB life cycle. At lag 0 and lag +1 days, the meridional component of the EP flux vectors, which are mostly directed equatorward, are also weaker than those for the WB life cycle. During these two lags, the WB and SB composites are qualitatively rather similar.

As suggested in the introduction, the barotropic governor concept may apply to the WB and SB life cycles. The WB life cycle, characterized by a broader jet with weakened horizontal shear at the beginning of its life cycle, has more intensified baroclinic growth in midlatitudes. Conversely, for the SB life cycle, which has a sharper jet and stronger horizontal shear at the start of its life cycle, the baroclinic growth is weaker. This relationship between the strength of the horizontal shear and the intensity of the baroclinic growth is consistent with the barotropic governor concept.

**FIG. 4.** Lag composites of the anomalous 925-hPa zonal-mean zonal wind $[u]\; ^\prime$ for the (a) WB life cycle and (b) SB life cycle and the anomaly of the absolute value of the 925-hPa horizontal wind shear $|\left[\frac{du}{dy}\right]|\; ^\prime$ for the (c) WB life cycle and (d) SB life cycle. The contour interval in (a), (b) is 0.2 m s$^{-1}$, and in (c), (d) is $5.0 \times 10^{-2}$ m s$^{-1}$. Solid contours are positive, dotted contours are negative, and the zero contour is omitted. Dark (light) shading indicates positive (negative) $\tau$ values that exceed the 95% confidence level. The straight line shown in all frames denotes the position of the center of the climatological midlatitude jet.
c. Planetary eddy EP flux

We also examined planetary-scale (zonal wavenumbers 1–3) EP fluxes. During the lag $-4$ to lag $-2$ day interval, composites of the anomalous planetary-scale EP flux indicate that in the subtropical upper troposphere there is a poleward anomalous wave activity flux during WB life cycles and an equatorward anomalous wave activity flux during SB life cycles (not shown), as with the synoptic-scale eddies. The primary difference between the planetary- and synoptic-scale EP fluxes at this stage of the life cycle was that the planetary-scale EP fluxes were found to have a smaller amplitude. These findings suggest that in the tropics there is anomalous source of planetary-scale Rossby wave activity during the WB life cycle and an anomalous sink of planetary-scale wave activity for the SB life cycle.

d. Mean meridional circulation

We next focus on the changes in the horizontal shear of the lower tropospheric, zonal-mean zonal wind at the
beginning of the WB and SB life cycles. As shown by Hartmann (2000) and consistent with the barotropic governor mechanism (James and Gray 1986; James 1987), it is the horizontal shear near the surface that most strongly impacts the eddies during baroclinic life cycles. Because the zonal-mean zonal wind in the lower troposphere is primarily influenced by the Coriolis torque associated with the mean meridional circulation and by surface friction, to investigate the changes in the lower level zonal-mean zonal winds we will examine the anomalous mass streamfunction, from which we can infer the Coriolis torque.

Figure 7 illustrates the anomalous mass streamfunction for the WB and SB life cycles. Positive values of the mass streamfunction correspond to a clockwise circulation and vice versa. In the WB life cycle, over the time interval from lag $-4$ to lag $-2$ days, the anomalous mass streamfunction in the Southern Hemisphere shows a weakening of the Hadley, Ferrel, and polar cells. These changes in the mean meridional circulation are consistent with the eddy driving shown in Fig. 5. By lag 0 days, the above changes in the Ferrel and Hadley cells have reversed, as both of these cells start to strengthen. These two cells continue to strengthen as the synoptic-scale
eddy energy increases, a typical feature of baroclinic instability.

The Coriolis torque implied by Fig. 7 is consistent with the changes in the 925-hPa zonal-mean zonal wind shown in Fig. 4. This is because a weakening of the Southern Hemisphere Ferrel cell before lag 0 days leads to a decrease in the zonal-mean zonal wind near the surface (Fig. 4a), which in turn reduces the horizontal shear of the surface zonal-mean zonal wind on both sides of the jet (Fig. 4c). Similarly, the weakening of the polar cell at lag −2 days is also consistent with the increase in the higher-latitude 925-hPa zonal-mean zonal wind (Fig. 4a), which contributes to a further decrease in the horizontal shear on the poleward side of the jet, as shown in Fig. 4c.

The SB mass streamfunction composites show the opposite characteristics for negative lags. At the start of the life cycle, it can be seen that the Southern Hemisphere

![Fig. 7. Lag composites of the anomalous mass streamfunction for the WB life cycle at (a) lag −4, (c) lag −2, and (e) lag 0 days and for the SB life cycle at (b) lag −4 days, (d) lag −2, and (f) lag 0 days. Contour interval is 0.1 s⁻¹; solid (dotted) contours are positive (negative), and the zero contour is omitted. Dark (light) shading indicates positive (negative) t values that exceed the 95% confidence level.](image-url)
Hadley, Ferrel, and polar cells have all strengthened. With regard to its impact on the zonal-mean zonal wind, the strengthening of these cells enhances the surface Coriolis torque, which in turn leads to a stronger surface zonal-mean zonal wind on the equatorward side of the jet and to a weaker surface zonal-mean zonal wind on the jet’s poleward side, as shown in Fig. 4b. The subsequent eddy growth, through to lag + 1 day, coincides with an intensification of the Ferrel and Hadley cells, as is typical for baroclinic instability.

As discussed above, through the barotropic governor mechanism (James and Gray 1986; James 1987), the horizontal shear of the surface zonal-mean zonal wind can have a strong impact on the baroclinic growth of synoptic-scale eddies in midlatitudes. Moreover, according to the results of Hartmann (2000), it is the influence of the horizontal shear in the lower troposphere that has the strongest impact. From this perspective, the decrease of the horizontal shear near the surface in the WB life cycle is better conditioned for baroclinic growth. In other words, for the WB life cycle, these results suggest that a sequence of processes occurs in which the eddy fluxes alter the strength of the Ferrel and polar cells, which in turn modifies the lower tropospheric horizontal shear and thus the barotropic governor. Consistent with the barotropic governor mechanism, we see that the baroclinic growth in the WB life cycle (Fig. 2) is stronger than that in the SB life cycle (Fig. 3).

e. Potential temperature anomalies

We next examine the anomalous meridional potential temperature gradient in the lower troposphere during the WB and SB life cycles (see Fig. 8). For convenience, in the rest of this study we will refer to the meridional potential temperature gradient as the baroclinicity. We focus on this particular quantity because of its strong link to linear baroclinic instability (Charney 1947; Eady 1949).

For the WB life cycle, beginning at lag −3 days, the anomalous baroclinicity is mostly negative between 40° and 60°S (Fig. 8a), which corresponds to those latitudes where the initial baroclinic growth takes place. (The region of initial baroclinic growth can be identified by the upward-pointing EP flux vectors at lag −2 days in Fig. 5). These results suggest that the WB life cycle growth tends to occur where the baroclinicity is slightly weaker than that of the climatology. In contrast, for the SB life cycle, opposite characteristics are observed, as the anomalous baroclinicity is mostly positive (Fig. 8b) at those latitudes where the baroclinic growth takes place, which is located on the equatorward side of the jet between about 34° and 47°S (note the location of the upward-pointing EP flux vectors at lag −2 days in Fig. 6). Furthermore, at the end of both WB and SB life cycles, it can be seen that the anomalous baroclinicity becomes mostly positive throughout midlatitudes.

The robustness of the baroclinicity in the WB and SB life cycles was also examined. As with the horizontal shear, it was found that most WB life cycles exhibited a reduction in the baroclinicity at the start of the life cycle. In contrast, for the SB life cycles, the baroclinicity was found to be much more variable at the beginning of the life cycles.

The above results, together with those from earlier parts of this section, indicate that for most WB life cycles the baroclinic growth takes place when the horizontal shear of the zonal-mean zonal wind and the baroclinicity are both weaker than that of the climatology. The opposite characteristics are observed for the SB life cycle, as both the horizontal shear and the baroclinicity are stronger than that of the climatology in the majority of life cycles. These results allude to the possibility for baroclinic life cycles, during the Southern Hemisphere summer, that the baroclinic eddy growth is more strongly dependent on the horizontal zonal wind shear than on the meridional potential temperature gradient.

![Fig. 8. Lag composites of the anomalous 825-hPa meridional potential temperature gradient for the (a) WB life cycle and (b) SB life cycle. Contour interval is $2.0 \times 10^{-7}$ K m$^{-1}$; solid (dotted) contours are positive (negative), and the zero contour is omitted. Dark (light) shading indicates positive (negative) $t$ values that exceed the 95% confidence level.](image-url)
Based on baroclinic instability theory and the barotropic governor mechanism, one would expect that the strongest baroclinic life cycles would develop within a background flow that has both a weak horizontal shear of the zonal-mean zonal wind and a strong baroclinicity. Since such a background flow is not observed prior to the baroclinic wave growth, it would seem that there may be a dynamical constraint that suppresses the development of such a background flow. One possible explanation deals with the characteristics of eddy-induced mean meridional circulation, which alters both the surface horizontal wind shear and the meridional temperature gradient. For example, a mean meridional circulation that weakens the Coriolis torque in the lower troposphere, as in the WB life cycle (Fig. 7), also weakens the meridional temperature gradient via adiabatic warming (cooling) on the poleward (equatorward) side of the Ferrel cell. Analogous arguments apply to the SB life cycle in which the mean meridional circulation enhances both the lower tropospheric horizontal wind shear and the meridional temperature gradient across the Ferrel cell. Therefore, it is plausible that these characteristics of the eddy-induced mean meridional circulation constrain the relationship between the horizontal shear of the zonal wind and the meridional temperature gradient of the background flow prior to the development of strong baroclinic life cycles.

Figure 9 shows the anomalous zonal available potential energy (ZAPE) for both WB and SB life cycles, calculated over the entire Southern Hemisphere. Interestingly, at the beginning of the WB life cycle, over the lag $-5$ to lag $-3$ day interval, the anomalous ZAPE declines. Analogously, over the same time interval the ZAPE in the SB life cycle increases. Even though the 95% confidence level is not exceeded during this time interval, these changes in the anomalous ZAPE are nevertheless consistent with the above description of the impact of the eddy-induced mean meridional circulation on the meridional potential temperature gradient. Given the physical consistency among Figs. 7–9, it is thus unlikely that the anomalous ZAPE values over the lag $-5$ to lag $-3$ day interval occur by chance. For both life cycles, after lag $-2$ days, once the baroclinic growth becomes well established, the anomalous ZAPE steadily declines. By lag $+4$ days, this trend in the anomalous ZAPE reverses, and positive values are attained by lag $+8$ days. An examination of the time-averaged anomalous mass streamfunction between lag $+2$ and lag $+8$ days (Fig. 10) suggests that the occurrence of positive anomalous ZAPE at the end of both life cycles may be explained by the persistence of anomalously strong Ferrel cells (the positive contours centered at 53° and 45°S in Figs. 10a and 10b, respectively). A comparison between Figs. 8 and 10 supports this viewpoint, as it can be seen that the latitudinal range of the strengthened Ferrel cells coincides with that of the positive anomalous baroclinicity over the lag $+2$ to lag $+8$ day time interval. Since the Ferrel cell is primarily eddy driven, these findings suggest that it is the eddy momentum fluxes at the end of both life cycles that drive the positive anomalous ZAPE.

These findings on the restoration of the baroclinicity at the end of the life cycles may have implications for the self-maintenance of the midlatitude jet, as proposed by Robinson (2006). In that two-level model study, it was found that under particular parameter settings, baroclinic eddies can maintain the meridional temperature gradient associated with the jet. As the radiative relaxation time scale in the atmosphere is about 30 days, much longer than that for the rapid restoration of the baroclinicity as shown in Figs. 8 and 9, the findings of this study suggest...
that this jet maintenance, for the Southern Hemisphere summer, may be accomplished by the eddy-induced mean meridional circulation associated with the barotropic decay stage of baroclinic life cycles.

f. Middle tercile

The middle tercile, which comprises those intermediate cases that do not satisfy our criteria for WB and SB life cycles, is also examined. For the energetics, anomalous zonal-mean zonal wind, EP fluxes, and mass streamfunction fields, the middle tercile composites more closely resemble those for the WB life cycle than for the SB life cycle. The initial weakening of the barotropic conversion is smaller than that of the WB life cycle, as are the subsequent baroclinic growth and barotropic decay. In contrast to these variables, the composite anomalous baroclinicity resembles neither life cycle as it is weak and poorly organized. An examination of the case-to-case variability also indicates that, with the exception of the baroclinicity, these results are robust.

5. Tropical convection

An important factor for distinguishing between the WB and SB life cycles involves the direction of the anomalous EP flux vector in the tropical and midlatitude upper troposphere during the lag \(-4\) to \(-2\) day interval. For the WB life cycle, the anomalous EP flux is poleward, and for the SB life cycle this flux is equatorward. As discussed in sections 4b and 4c, these EP flux properties suggest that in the WB life cycle there may be an anomalous Rossby wave source near \(10^\circ\)S in the upper troposphere and in the SB life cycle there may be an anomalous sink in the same location. This latitude is close to that of the local maximum in climatological convection observed over the tropical western Pacific and Indian Oceans, as measured by both outgoing longwave radiation (OLR) and precipitation, during the Southern Hemisphere summer. (In the tropics, the majority of the deep convection is confined to the western Pacific and Indian Oceans, with additional smaller regions of intense convection occurring over the Amazon and Congo basins. Throughout the rest of the tropics, the climatology exhibits relatively little deep convection. Over the region spanning the tropical western Pacific and Indian Oceans, the convection over Indonesia is strongest. Again, see http://www.cdc.noaa.gov/data/ncep_reanalysis/.) Because localized tropical convection drives poleward propagating Rossby waves (e.g., Hoskins and Karoly 1981) and thus a poleward EP flux, the poleward anomalous EP flux for the WB life cycle may be due to an intensification of the convection over the tropical western Pacific and Indian Oceans. Analogous arguments for the SB life cycle suggest that the anomalous equatorward EP flux in the upper troposphere may be associated with a weakening in the convection over the same region.

To examine the possible impact of tropical convection on the WB and SB baroclinic life cycles, we calculate composites of anomalous OLR as a function of longitude and lag, averaged from \(15^\circ\)S to \(5^\circ\)N (see Fig. 11). Negative OLR values indicate regions of enhanced convection. For the WB life cycle, at lag \(-20\) days, a broad region of enhanced convection is centered over the western Indian Ocean, and by lag \(-6\) days this eastward propagating zone of convection extends across Indonesia into the western Pacific Ocean. Since, as discussed above, the climatological convection is strongest over the tropical western Pacific and Indian Oceans, these results indicate that the early stages of the WB life cycle do, indeed,
coincide with an increase in local tropical convection. Opposite characteristics in the tropical convection are observed for the SB life cycle. At lag $-20$ days, there is a region of weakened convection propagating eastward from the western Indian Ocean. By lag $-6$ days, this region of weaker convection is centered over the eastern Indian Ocean and Indonesia. Such characteristics are consistent with a weakening in the local convection prior to the development of the SB life cycle.

For both WB and SB life cycles, the above OLR characteristics bear a resemblance to those of the Madden–Julian oscillation (MJO) (Madden and Julian 1971, 1972). To more carefully investigate a possible connection between the life cycles and the MJO, we examine the two-dimensional MJO vector index of Wheeler and Hendon (2004). This MJO index is based on the first two principal components of the combined OLR, 200-hPa zonal wind, and 850-hPa zonal wind fields at all longitudes between $15^\circ$S and $15^\circ$N. (Their MJO index is available online at http://www.bom.gov.au/bmrc/clfor/cfstaff/matw/maproom/RMM.) We adopt a null hypothesis that 50% of the WB and SB life cycles coincide with MJO events. This null hypothesis is based on the observations of Pohl and Matthews (2007), who find that the MJO is active about 50% of the time. Statistical significance is examined with the binomial distribution. We define an MJO event to have taken place whenever the 10-day, time-averaged, amplitude of the MJO index exceeds a threshold value of 1.24, which corresponds to the time-mean MJO amplitude for the entire time period of this study. It was found that 9 WB life cycles and 10 SB life cycles coincide with MJO events. These results suggest that the null hypotheses should not be rejected and indicate that both the WB and the SB life cycles do not coincide with an increase in the frequency of MJO events. The sensitivity to the amplitude threshold value, the number of days that make up the time average, and the application of filtering (Pohl and Matthews 2007) were also examined, and the essential results were unchanged.
The case-to-case variability in the longitudinal structure of the eddies associated with the WB and SB life cycles is also examined. This variability is investigated by evaluating both the potential temperature at the tropopause (the 2-PVU surface; 1 PVU = 10^{-6} m^2 s^{-1} K kg^{-1}) and the 300-hPa transient eddy kinetic energy. The examination of the 2-PVU potential temperature field showed that most WB and SB life cycles are associated with a single localized baroclinic wave packet (not shown). These wave packets, which propagate eastward and comprise two or three synoptic-scale troughs and ridges, were found to occur at most longitudes, as in Lee and Held (1993) and Chang (2005).

The range of longitudes occupied by these wave packets is summarized by illustrating the composite transient eddy kinetic energy for the WB and SB life cycles (Fig. 12). As can be seen, enhanced eddy kinetic energy for both WB and SB life cycles is observed to span most longitudes. Furthermore, to the extent that storm tracks in the Southern Hemisphere summer are associated with baroclinic life cycles, it would be expected that the Southern Hemisphere summer storm track should also span most longitudes. As shown in Chang (1999) and Rao et al. (2002), this is indeed the case. Rao et al. (2002) find that the Southern Hemisphere storm track is most intense near the date line to the southeast of Australia, the same location as the eddy kinetic energy maximum associated with the SB life cycles.

The above results make a link between tropical convection and anomalous wave activity propagation in the upper troposphere. These results suggest that for the WB life cycle enhanced tropical convection excites poleward wave activity propagation. When this wave activity reaches midlatitudes, the corresponding eddy fluxes drive a mean meridional circulation that weakens the horizontal shear of the zonal-mean zonal wind and the meridional temperature gradient in the lower mid-latitude troposphere. This, in turn, generates the background flow conditions that favor the development of the WB baroclinic life cycle. Similarly, for the SB life cycle, the above results suggest that weakened tropical convection results in an anomalous equatorward wave activity propagation. The corresponding eddy fluxes induce a mean meridional circulation that has an enhanced horizontal zonal wind shear and meridional temperature gradient, conditions that favor the development of the SB baroclinic life cycle.

6. Discussion and conclusions

In this study, we perform composite analyses with NCEP–NCAR reanalysis data to investigate baroclinic life cycles that occur during the Southern Hemisphere summer. Two types of life cycles are identified, based on the strength of the total barotropic conversion two days before the maximum baroclinic conversion. We refer to those baroclinic life cycles with an anomalous weak barotropic conversion as the weak barotropic (WB) life cycle and to those life cycles with an anomalously strong barotropic conversion as the strong barotropic (SB) life cycle.

![Figure 12](image-url)
The main characteristics of the WB and SB life cycles are summarized with a schematic diagram (see Fig. 13).

Several days before the start of the WB life cycle (Fig. 13a), enhanced convection is observed over Indonesia, the region in the tropics with the most intense time-mean convection. This is followed by a poleward anomalous wave activity flux in the subtropics (step 1 in Fig. 13a) and then a weakening and broadening of the jet in the upper troposphere (step 2 in Fig. 13a). The poleward anomalous wave activity flux also induces anomalously weak Hadley, Ferrel, and polar cells, which reduce the meridional temperature gradient through adiabatic warming and cooling and weaken the horizontal shear of the zonal-mean zonal wind in the lower troposphere through the impact of the Coriolis torque. According to the barotropic governor mechanism, the reduced horizontal zonal wind shear enhances baroclinic growth. Consistently, the ensuing WB life cycle exhibits stronger baroclinic growth than in the SB life cycle.

The dynamical characteristics of the SB life cycle are found to be opposite to those of the WB life cycle (Fig. 13b). Several days prior to the beginning of the SB life cycle, a reduction in convection is observed over Indonesia. This is followed by an equatorward anomalous wave activity flux in the subtropics and then a strengthening of the jet in the upper troposphere on its equatorward side and a weakening of the jet on its poleward side. This results in an overall increase in the horizontal zonal wind shear. The equatorward anomalous wave activity flux induces anomalously strong Hadley, Ferrel, and polar cells that enhance the meridional temperature gradient and increase the horizontal shear in the lower troposphere. The resulting life cycle, with its weaker baroclinic growth and barotropic decay compared to that in the WB life cycle, is also consistent with the dominance of the barotropic governor mechanism.

As shown in Fig. 13a, the more rapidly growing WB life cycle is characterized at the beginning of its life cycle by both an anomalous weak horizontal zonal wind...
shear and an anomalously weak meridional temperature gradient. In contrast, the more slowly growing SB life cycle (Fig. 13b) features an anomalously strong horizontal shear and meridional temperature gradient. These results highlight one of the main findings of this study, which suggests that baroclinic life cycles of the Southern Hemisphere summer are controlled primarily by the horizontal shear of the zonal wind, that is, the barotropic governor mechanism (James and Gray 1986), and not the meridional temperature gradient, as one may anticipate from standard linear baroclinic instability theory (Charney 1947; Eady 1949). Further support for the dominating role of the barotropic governor mechanism is the observation that the more rapidly growing WB life cycle develops in a background flow with a meridional temperature gradient that is slightly weaker than that of the climatology.

Most studies of baroclinic life cycles have used idealized models. The initial perturbation in these studies consisted of very small amplitude disturbances. In contrast, the results of the present study indicate that in the atmosphere, during the Southern Hemisphere summer, large amplitude eddy fluxes present at the beginning of the life cycle play a crucial role. As described above, these eddy fluxes, through the mean meridional circulation that they induce, can establish the background zonal mean flow upon which subsequent eddy growth takes place.

Because many observational and modeling studies during the past decade focused on the occurrence of anticyclonic and cyclonic wave breaking (breaking waves with northeast–southwest tilt or northwest–southeast tilt, respectively, for the Northern Hemisphere; e.g., Akahori and Yoden 1997; Esler and Haynes 1999; Hartmann 2000; Martius et al. 2007; Riviere and Orlanski 2007; Thorncroft et al. 1993), we investigated whether these two types of wave breaking correspond to the WB and SB life cycles. Following the approach of Riviere and Orlanski (2007), who inferred the type of wave breaking from the direction of the eddy momentum flux, we conclude that both WB and SB life cycles are characterized by anticyclonic wave breaking.

This study was limited to the Southern Hemisphere summer because that particular season exhibits the largest zonal symmetry in its time mean flow. An interesting question for future research is to investigate the extent to which the findings of this study apply to baroclinic cycles during other seasons in the Southern Hemisphere, and the Northern Hemisphere, where large-amplitude stationary eddies are present.

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APPENDIX

Energies

Our calculation of the energy and conversion terms follow Krishnamurti et al. (1998). Every term calculated is based on the primitive equation set:

\[
\begin{align*}
ZAPE & = \int_M \frac{\alpha' \theta_z}{2\sigma} \, dm \\
ZKE & = \int_M \frac{u'^2 + v'^2}{2} \, dm \\
EAPE & = \int_M \frac{\alpha' \theta}{2\sigma} \, dm \\
EKE & = \int_M \frac{u'^2 + v'^2}{2} \, dm \\
BCC & = \int_M \frac{\alpha' z}{\sigma \alpha \cos \phi} \frac{\partial}{\partial \phi} \left( \frac{\alpha' \theta v u}{\cos \phi} \right) \cos \phi \, dm \\
BTC & = -\int_M \left[ \frac{u' v' \cos \phi}{a} \cos \phi \frac{\partial}{\partial \phi} \left( \frac{\tilde{u}}{\cos \phi} \right) + v' \frac{\partial \tilde{\alpha}}{\partial \phi} \right] \, dm \\
& - \tilde{u}' \tan \phi \frac{\partial}{\partial \phi} + \tilde{u}' \left( \frac{\partial}{\partial \phi} \right) \tilde{u}' + \tilde{v}' \left( \frac{\partial}{\partial \phi} \right) \tilde{v}' \, dm
\end{align*}
\]

in which

- ZAPE: zonal available potential energy, ZKE: zonal kinetic energy;
- EAPE: eddy available potential energy, EKE: eddy kinetic energy;
- BCC: baroclinic conversion term, BTC: barotropic conversion term;
- \(M\): integration is over the entire Southern Hemisphere;
- \(u\): zonal velocity, \(v\): meridional velocity;
- \(\alpha\): specific volume;
- \(dm = \frac{\alpha' \theta}{g} \, d\lambda d\phi dp\);
- \(\tilde{X}\): zonal mean of \(X\),
- \(X\): deviation from the zonal mean;


\[ \bar{\sigma} = -\frac{\alpha}{\vartheta} \frac{\partial}{\partial \varphi} \] is the mean static stability;

- \( \bar{\sigma} \) is the specific volume averaged on a pressure surface;

- \( \alpha' = \alpha'_{Z} + \alpha'_{E} \), where \( \alpha'_{Z} \) is the zonal average of \( \alpha' \) and \( \alpha'_{E} \) is the departure of \( \alpha' \) from \( \alpha'_{Z} \);

- \( u_{\psi} \) represents the rotational part of \( v \).

REFERENCES


