1. Introduction

La Niña (LN) and El Niño (EN) events greatly impact the global climate. The occurrence of these events in the tropical Pacific can cause unusual weather and climate all over the world (e.g., Enfield and Mestas-Núñez 2000). Therefore, it is still necessary to investigate in detail the LN and EN events. This study examined how the duration of LN and EN events is modulated by the Australian winter monsoon (AWM).

It is well known that the duration of LN events and that of EN events differ (Kessler 2002; Ohba and Ueda 2009; Okumura and Deser 2010; Wu et al.). Therefore, it is still necessary to investigate in detail the LN and EN events. This study examined how the duration of LN and EN events is modulated by the Australian winter monsoon (AWM).

Effects of the Australian Winter Monsoon on the Persistence of La Niña Events

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Abstract

La Niña (LN) events are generally longer than El Niño (EN) events. Using objective analysis data, we herein investigated the effects of the Australian winter monsoon (AWM) on prolonging LN events. Conventionally, EN events end through the eastward shift of the anomalous Walker circulation in the equatorial Pacific during March–August. In contrast, the stronger-than-usual AWM induced by LN anchors the upflow branch of anomalous Walker circulation in the Indonesian maritime continent (IMC). The strength of the AWM is controlled by the surface temperature difference between the IMC and the northern Australian continent (NAC). LN has a large impact on the decrease in surface temperature in the NAC through a decrease of the downward surface short-wave radiation flux and the increase in surface soil moisture in the NAC. In LN events, the strength of the AWM and the anomalous Walker circulation reinforce each other through the common convective ascending in and around the IMC, which may be termed LN–AWM feedback, prolonging the duration of LN events. During EN events, such feedback is weak so that EN events generally end in the period of March–August.

Keywords  La Niña; El Niño; asymmetry in duration; the Australian winter monsoon

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investigated if LN and EN events occurred independently of each other or were embedded in a cycle and found that there was a break in the cycle and that the LN events could remain for 2 years. As a physical mechanism extending LN events, Ohba and Ueda (2009) proposed an asymmetry in the anomalous atmospheric circulations in the western North Pacific between LN and EN events, which further resulted in an asymmetry in the anomalous zonal winds in the western equatorial Pacific controlling the termination of the events. Wu et al. (2010) pointed out that the asymmetry in the zonal distribution of equatorial Pacific SSTAs resulted in the asymmetry of anomalous atmospheric circulations in the western North Pacific between LN and EN events. That is, in LN events, the anomalous cyclonic circulation in the western North Pacific is located more westward compared with the counterpart, i.e., anomalous anti-cyclonic circulation, in EN events.

Okumura and Deser (2010) further proposed the following two factors regarding the difference in duration: one is the nonlinear response of local convective activity to tropical SSTAs, and the other is the asymmetry of the SSTA distribution in the equatorial Pacific between LN and EN events. In the equatorial Pacific, the negative SSTAs in LN events extend further westward (by 10–15°) than the positive SSTAs in EN events, and the difference in duration has been discussed in relation to the western Pacific oscillator (Weisberg and Wang 1997). Okumura et al. (2011) validated the above discussions through atmospheric general circulation model (GCM) experiments. Using a coupled ocean–atmosphere GCM, Ohba and Watanabe (2012) examined the contribution of the Indian Ocean to the persistence of LN and EN events and found that the effects of the Indian Ocean shortened the EN events. That is, EN events tend to end by summer with regularity, whereas LN events may persist longer.

In addition, DiNezio and Deser (2014) revealed that the asymmetry of the nonlinear relationship among thermocline depth, subsurface temperature gradient, and SSTAs in the central equatorial Pacific between LN and EN events resulted in longer LN than EN. Hu et al. (2014) proposed the following two conditions for the continuous appearance of LN events: one is that the first LN event is very large, and the other is that the oceanic Kelvin wave in the equatorial Pacific, which induces the end of the LN event, is weak or not forced after the boreal winter LN maximum with continuous easterly wind anomalies. Chen, M. et al. (2016) found that dynamical and thermodynamical forcing factors cause the difference in duration of LN and EN events. As an important indicator, they proposed the magnitude of the temporal changing rate of the SSTA in the central equatorial Pacific in the decaying year. Regarding the spatial patterns, many studies have documented the asymmetry between LN and EN events (Hoerling et al. 1997; Kang and Kug 2002; Wu et al. 2010; Chen, Z. et al. 2016, 2017; Tao et al. 2016, among others).

Focusing on the effects of the AWM, this study investigated the physical mechanisms related to why LN events tend to remain longer than EN events. With regard to the physical mechanisms, earlier studies, as introduced above, focused on basin-scale oceanic and/or atmospheric circulations in the equatorial region or in the western North Pacific. Moreover, studies discussing the effects of the AWM on lengthening LN events relative to EN events are limited.

The Australian continent is located closer than the Asian continent to the Indonesian maritime continent (IMC), where the world’s most vigorous convective activity is connected with the El Niño/Southern Oscillation (ENSO); however, the Australian continent is much smaller than the Asian continent. The AWM, which appears in the time period centered around boreal summer, may have a potential to modulate the duration of ENSO events through the modification of convective activity in the IMC during the onset or decaying stage of the ENSO. This study assesses this potential of the AWM.

To address the above issues, the remaining sections are organized as follows. Section 2 explains the data and methodology employed in this study. Section 3 provides the definition of AWM used in this study with the climatological seasonal means in atmospheric circulations in and around the IMC. Section 4 assesses how LN events persist and end, together with the associated zonal atmospheric circulations, through a comparison with EN events. Section 5 elucidates the
mechanisms of how the AWM controls the persistence of LN events by modifying the convective activity in and around the IMC. Finally, Section 6 summarizes the findings of this study and briefly discusses the remaining issues. In order to avoid confusion about the notation of seasons, a season is hereafter expressed by months or a series of abbreviations of the months, e.g., DJF for December, January, and February (unless otherwise noted).

2. Data and methodology

This study mainly used the dataset of the European Center for Medium-Range Weather Forecasts reanalysis (ERA)-Interim (Dee et al. 2011) for physical quantities at pressure levels and the surface. The assessed period is the 36 years from 1979 to 2014. The temporal resolution is monthly, and the spatial resolution is 0.75° longitude and 0.75° latitude over the whole globe. Vertically, 37 pressure levels are set from 1000 to 1 hPa. In addition, the dataset of the Global Precipitation Climatology Project was used for precipitation rate (Adler et al. 2003). The time period is again the 36 years from 1979 to 2014 with monthly data, and the spatial resolution is 2.5° longitude by 2.5° latitude over the entire globe. As indices of the ENSO, this study used the Oceanic Niño Index edited by the U.S. National Oceanic and Atmospheric Administration (NOAA), which is available at the following site: https://origin.cpc.ncep.noaa.gov/products/analysis_monitoring/ensostuff/ONI_v5.php.

To identify LN (EN) events, this study employed the following definition of NOAA: the 3-month running mean of SSTA averaged in the NINO3.4 region (170–120°W, 5°S–5°N) is consecutively below (above) the threshold of −0.5°C (+0.5°C) for 5 months or longer. Based on this definition, the DJF periods with a negative or positive maximum of LN and EN events were determined, and the onset (denoted by a value of 0) and the following (denoted by a value of 1) years were set by placing the DJF in the middle of 2 years. The next following year was denoted by a value of 2. Hereafter, the years of LN and EN events are indicated by the letters LN or EN and the numerals 0, 1, or 2, e.g., LN(0) or EN(1). Table 1 lists the years of LN and EN events, each of which includes 11 cases.

In this study, an anomaly is defined as the difference from the climatological mean of the long-term period of 1979–2014 (36 years). The statistical significance of correlation coefficients and of differences between two means was tested by the two-tailed t-test. A 3-month or longer mean is used as a seasonal mean in this study.

3. Definition of the Australian winter monsoon

In this study, the AWM is defined as a counterpart of the Australian summer monsoon (cf. Ramage 1971), although some previous work has identified it as a part of the southeast trade wind regime (cf. Fein and Stephens 1987). Figure 1 presents the seasonal means (MJJAS and NDJFM) of precipitation rate and horizontal wind at 10-m height in and around the IMC (Figs. 1a, b), as well as the differences between them (MJJAS–NDJFM; Fig. 1c). In the seasonal mean fields of MJJAS (Fig. 1a), the regions with a small precipitation rate extend northwestward from the northern Australian continent (NAC) with southeasterly winds in the Southern Hemisphere tropics, including the NAC.

The AWM region is determined by the differences between the seasonal mean fields of MJJAS (Fig. 1a) and of NDJFM (Fig. 1b) in precipitation rate and surface wind (MJJAS–NDJFM; Fig. 1c), which also define the region of the Australian summer monsoon.
The region of the AWM extends northwestward from the NAC to the IMC in the Southern Hemisphere tropics, where dry and rainy seasons clearly appear and the seasonal difference is large in the surface wind direction. In fact, these conditions conform to the definition of monsoon proposed by Ramage (1971). Note that the region of the AWM and the Australian summer monsoon is confined in the Southern Hemisphere tropics in this study, although the region continuously extends to the Northern Hemisphere beyond the equator. Wang et al. (2003) and other workers have already pointed out that the ENSO is a major control factor on the interannual variability of the Asian–Australian monsoon. In contrast to these earlier studies, the present study discusses the effects of the AWM on the ENSO, in particular, in relation to prolonging LN events.

4. Long La Niña events and the associated zonal atmospheric circulations

LN events are certainly longer than EN events. The difference in duration between LN and EN events is summarized in Fig. 2. For ease of comparison, the NINO3.4 SSTA is multiplied by −1 for the EN events.
in Fig. 2b. At the D(1)JF(2) between LN(1) and LN(2) (Fig. 2a), the SSTA remained negative in 9 of the 11 events, whereas it was positive in 8 of the 11 events at the D(1)JF(2) between EN(1) and EN(2) (Fig. 2b). The LN events exhibit a tendency to remain until LN(2), and the negative SSTA turned positive around JJA(2) of LN(2). On the other hand, many EN events end by JJA(1) of EN(1), which is 1 year earlier than the end of LN events.

The transition was surveyed for the periods of LN(1) and EN(1) in the seasonal-mean anomalies of SST (contour), precipitation rate (shading), and surface wind (vector) (Fig. 3). In D(0)JF(1) of LN(1) (Fig. 3a), a region with negative SSTAs extends in the central equatorial Pacific, and regions with positive SSTAs surround it in a horseshoe pattern. In and around the IMC, significant positive precipitation rate anomalies are observed with the confluence of large-scale zonal wind anomalies, indicating that the Walker circulation is enhanced in the equatorial region from the Indian Ocean to the Pacific. We may further observe the anomalous cold air flow to the IMC from the north along the eastern coast of the Asian continent.

The distribution of anomalies in the following MAM(1) (Fig. 3b) is similar to that in D(0)JF(1) (Fig. 3a) but with smaller absolute values. A closer inspection may find an increase in precipitation rate near the northern edge of the NAC and a westward shift of about 10° in the negative SSTAs in the central equatorial Pacific. The anomalous northerlies along the eastern coast of the Asian continent are weakened, and the latitude of the southern edge withdraws to around 15°N. Over the Indian Ocean, anomalous south-westerlies in the northern part and anomalous south-easterlies in the southern part are intensified in MAM(1) of LN(1), indicating the stronger-than-usual or earlier onset of the Indian summer monsoon. Until JJA(1) of LN(1) (Fig. 3c), these anomalies decrease the magnitude. It is noticeable that the positive precipitation rate anomalies remain in and around the IMC with anomalous easterlies in the central–western equatorial Pacific.

The distribution pattern of anomalies in D(0)JF(1) of EN(1) (Fig. 3d) is similar to that in D(0)JF(1) of LN(1) (Fig. 3a) but with opposite polarities, which may also be identified in the conventional EN event (e.g., Rasmusson and Carpenter 1982). In the following MAM(1) (Fig. 3e), the anomalies retain a similar...
pattern to that in D(0)JF(1) (Fig. 3d), although the magnitude of the anomalies decreases. By JJA(1) (Fig. 3f), the SSTAs change the polarity from positive to negative in the eastern equatorial Pacific, whereas positive SSTAs remain in the tropical Indian Ocean. It is interesting to note that the positive precipitation rate anomalies appear in the western IMC. In this season, the Asian summer monsoon is weaker than usual, with easterly wind anomalies from the Bay of Bengal to the tropical western North Pacific, which could further increase SST in this region by decreasing evaporation with smaller-than-usual wind speed. These changes in and around the IMC are sufficient to shift EN to LN. The distribution patterns of anomalies in JJA(1) are definitely not mirror images between LN(1) and EN(1) (Figs. 3c, f).

A descent (ascent) branch of anomalous Walker circulation generally moves eastward from the IMC to the central equatorial Pacific, whereas EN (LN) decays and changes to LN (EN) (cf. Figs. 3d–f). Figure 4 presents the equatorial longitude–height cross sections of vertical p-velocity anomalies in LN(1), EN(1), and
Fig. 4. Longitude–height cross sections of 3-month mean anomalies in vertical p-velocity (Pa s\(^{-1}\); contour and shading) averaged in 10°S–10°N for LN(1) (left) in (a) D(0)JF(1), (b) MAM(1), (c) JJA(1), and (d) SON(1); EN(1) (middle) in (e) D(0)JF(1), (f) MAM(1), (g) JJA(1), and (h) SON(1); and LN(2) (right) during (i) D(1)JF(2), (j) MAM(2), (k) JJA(2), and (l) SON(2). The contour interval is 0.1 \(\times 10^2\) Pa s\(^{-1}\). The scale of shading is presented at the bottom. The solid contours with shading indicate the regions with anomalous ascent. The dashed contours are for regions with anomalous descent. A simple nine-point smoothing has been applied to the anomalies. A vertical thick solid line is drawn at 150°E in each panel as an indicator of the eastern edge of IMC in the equatorial zone.
During MAM(2)–JJA(2) (Figs. 4j, k), the region with anomalous descent centered and its center is located around 150°E. At the same time, the region with anomalous ascent in the IMC shifts eastward by about 20°, which may indicate re-enhancement of LN’s anomalous Walker circulation in the region from 90°E to 150°W, which may amplify the region with positive precipitation rate anomalies from May, which seems to lead to the appearance of the region with anomalous westerlies. From September to December, the anomalous westerlies are strengthened in the equatorial Indian Ocean to the IMC. Thus, the reversal in the direction of circulation is reversed by September. In the equatorial Pacific, the positive precipitation rate anomalies are dominant in the region where westerly wind anomalies are dominant from January to May. Unlike the anomalous Walker circulation from January to May in LN(1) (Fig. 5a), the magnitudes of anomalous zonal winds were not comparable between the large positive in the equatorial Pacific and the small negative in the equatorial Indian Ocean in the first half-year. In the equatorial Indian Ocean, the eastward extension of the region with positive precipitation rate anomalies is observed from May, which seems to lead to the appearance of the region with anomalous westerlies. From September to December, the anomalous westerlies are strengthened in the equatorial Indian Ocean to the IMC. Thus, the reversal in the direction of the anomalous Walker circulation is completed.

In LN(2) (Fig. 5c), we identify the eastward shift of the region with positive precipitation rate anomalies
from the IMC to the equatorial Pacific in the second half-year. The duration of large easterly wind anomalies in the central–western equatorial Pacific is short and about a half of that in LN(1), i.e., about 3 months from January to March. The region with anomalous westerlies then extends eastward into the Pacific from the Indian Ocean. By December, the region with anomalous westerlies (easterlies) occupies the equatorial Pacific (Indian Ocean), and the anomalous Walker circulation is reversed.

5. Anomalous Australian winter monsoon and the persistence of rainfall in the Indonesian maritime continent

The persistence of rainfall in the IMC from MAM(1) to JJA(1) of LN(1), i.e., the longer convective latent heating in the region, seems to play a key role in elongating LN events to a greater extent than for EN events (cf. Figs. 3, 5). The AWM may possess the potential to reinforce the rainfall. This section investigates the potential of AWM, i.e., the relationship between the anomalous AWM and the persistence of rainfall in the IMC. Since the monsoon is first driven by the large-scale pressure difference between the ocean and the continent, which is induced by the surface temperature difference (Ramage 1971; Fein and Stephens 1987), this study examined the near surface temperature anomalies in the IMC–NAC region, as associated with LN and EN events (Fig. 6). The three columns in Fig. 6 indicate the seasonal transitions from DJF to JJA of the 2-m-height temperature (T2m) anomalies in LN(1), EN(1), and LN(2), respectively.

During LN(1) (Figs. 6a–c), negative T2m anomalies prevail in the NAC. In D(0)JF(1) (Fig. 6a), the
NAC is covered by large negative anomalies, whereas the anomalies are negative and small around the IMC. By MAM(1) (Fig. 6b), the negative anomalies are amplified in the NAC, and the statistically significant regions extend northwestward from the NAC. As a result, a large difference can be measured in the T2m anomaly between the NAC and the IMC (IMC > NAC), which amplifies the AWM. By JJA(1) (Fig. 6c), the negative anomalies in the NAC become smaller, but the anomalous T2m contrast between the NAC and the IMC remains large.

During EN(1) (Figs. 6d–f), the distributions of
T2m anomalies in each season are similar to those during LN(1), although with opposite polarities. In D(0)JF(1) of EN(1) (Fig. 6d), the regions with positive T2m anomalies distribute in the NAC, and the statistically significant regions extend to the northwest of the NAC. In MAM(1) (Fig. 6e), the magnitude of positive T2m anomalies exceeds +0.5°C in the NAC. However, by JJA(1) (Fig. 6f), the regions with T2m anomalies over +0.5°C disappear in the NAC (except at the northern tips).

During LN(2) (Figs. 6g–i), the magnitude of the T2m anomalies is generally small in the IMC–NAC region, particularly in D(1)JF(2)–MAM(2) (Figs. 6g, h) when the upflow branch of the anomalous Walker circulation shifts eastward and passes over 150°E (Figs. 4i, j). In JJA(2) (Fig. 6i), we can observe large negative T2m anomalies in the NAC, although these are not statistically significant, and the LN events have already ended (Figs. 2a, 4k). It is worth noting that the differences in T2m anomalies are marked between the MAM(1) of LN(1) (Fig. 6b) and MAM(2) of LN(2) (Fig. 6h).

We hypothesized that the T2m anomalies in the NAC were primarily controlled by the net surface short-wave radiation (SSR) and the surface soil moisture (SSM). In fact, the interannual anomalies in Australian land surface temperature have been investigated in relation to interannual ENSO variability through modification of SSR and SSM (Power et al. 1998; Jones 1999; Jones and Trewin 2000). To identify the consistency between the anomalies in SSR and SSM and those in T2m (Fig. 6), this study assessed the 3-month mean fields of SSR and SSM anomalies during LN(1), EN(1), and LN(2) (Fig. 7).

In D(0)JF(1) of LN(1) (Fig. 7a), when LN reaches the mature phase (Fig. 2a), the downward SSR decreases almost the entire region from Australia to IMC (except for the western equatorial Pacific and the northern central part of Australia). The stronger-than-usual convection in and around the IMC during LN (Figs. 3a, 4a) probably causes the decrease in downward SSR. The positive SSR anomalies in the western equatorial Pacific are preferably yielded in a descent branch of anomalous Walker circulation to the east of 150°E (Fig. 4a). It is confirmed that the distribution of the SSR anomalies in Australia (Fig. 7a) is similar to that of the T2m anomalies (Fig. 6a). However, the response of T2m in and around the IMC seems to be small because of ocean surface mixing and the large ocean heat content. In eastern Australia, the SSM increases in response to the increase in precipitation rate in the South Pacific convergence zone (Fig. 3a).

A similar distribution pattern remains until MAM(1) (Fig. 7b) in the equatorial zonal region, whereas in the NAC, SSM locally increases and downward SSR decreases. It is evident that the precipitation rate consistently increased in the NAC in MAM(1) (Fig. 3b). Therefore, the significantly large decrease in T2m seemed to be fully developed in the NAC (Fig. 6b). By JJA(1) (Fig. 7c), the regions with negative SSR anomalies concentrate in and around the IMC, implying that the stronger-than-usual convection is locally retained over this region (Fig. 3c). On the other hand, SSR and SSM anomalies diminish in the NAC by this season, whereas the negative T2m anomalies are still large in the NAC (Fig. 6c). It is conceivable that factors other than SSR and SSM, such as upward long-wave radiation, surface sensible heat, and horizontal T2m advection, might have maintained the lower T2m in the NAC. In fact, we can expect an anomalous local Hadley circulation between the IMC and the Australian continent. The anomalous ascent (descent) of this circulation is located in the IMC (the Australian continent), and it yields sunny days and leads to anomalous cold and dry southerlies in the Australian continent in the austral winter. The anomalous local Hadley circulation between the IMC and the Australian continent will be discussed later.

The distributions of SSR and SSM anomalies in D(0)JF(1) of EN(1) (Fig. 7d) are similar to those in D(0)JF(1) of LN(1) (Fig. 7a), although with opposite polarities (except for the anomalies in the central NAC and large negative SSM anomalies (shading in red) in the western NAC (Fig. 7d)). This distribution again corresponds well to that of T2m anomalies in the NAC (Fig. 6d). In MAM(1) (Fig. 7e), the magnitudes of positive SSR and negative SSM anomalies in the NAC are small compared with the counterpart anomalies of LN(1) (Fig. 7b). In this season (Fig. 7e), we may observe that the regions with negative SSR anomalies extend from the west in the western part of the IMC–NAC region. By JJA(1) (Fig. 7f), the regions with negative SSR anomalies extend further eastward with the eastward shift of enhanced convection (Figs. 3f, 4g, 5b), although positive T2m anomalies are still distributed in the NAC (Fig. 6f).

The magnitudes of SSR and SSM anomalies in LN(2) (Figs. 7g–i) are generally smaller than those in LN(1) (Figs. 7a–c) and in EN(1) (Figs. 7d–f). However, the transition of the anomalies from D(1) JF(2) to JJA(2) in LN(2) (Figs. 7g–i) may be somewhat similar to that in EN(1) (Figs. 7d–f), although with opposite polarities. The similarity between EN(1)
Fig. 7. Three-month mean fields of the anomalies in surface short-wave radiation (J day$^{-1}$ m$^{-2}$; contour) and in surface soil moisture (m$^3$ m$^{-3}$ or no unit; shading) for LN(1) (left) in (a) D(0)JF(1), (b) MAM(1), and (c) JJA(1); EN(1) (middle) in (d) D(0)JF(1), (e) MAM(1), and (f) JJA(1); and LN(2) (right) in (g) D(1)JF(2), (h) MAM(2), and (i) JJA(2). In surface short-wave radiation, the downward direction is set as positive, and the anomalous positive (negative) is indicated by red (blue) solid (dotted) contours, where the contour interval is 2.0 J day$^{-1}$ m$^{-2}$. The zero contour is indicated by the black solid line. In the ERA-Interim dataset, surface soil moisture is estimated in a layer from the surface to the depth of 7 cm. Blue (red) shading on the land indicates the anomalously wet (dry) regions. The color scale is presented at the bottom.
(Figs. 7d–f) and LN(2) (Figs. 7g–i) may be induced by the eastward moving of the ascent or descent branch of anomalous Walker circulation (Figs. 4, 5). Thus, consistency was confirmed among the anomalies of T2m, SSR, and SSM.

We then identified the near surface temperature difference between the IMC and the NAC. To evaluate the asymmetry between LN(1) and EN(1) in the anomalous progress of the AWM, we examined the T2m difference between the IMC and the NAC. Figure 8 presents the time series of the T2m difference between the oceanic IMC (120–140°E, 15°S–0°) and the continental NAC (120–140°E, 30–15°S) (IMC–NAC), in which the difference was estimated using the continuous 3-month means of the T2m anomalies. The positive values in the period from FMA(1) to JAS(1), i.e., the austral fall and winter, may correspond to the stronger-than-usual AWM.

In LN(1) (dark gray), the positive values increase from D(0)JF(1) to MAM(1) and then decrease until JJA(1). The maximum value of MAM(1) exceeds 0.5°C and is statistically significant at the 10% level. The values of FMA(1) and AMJ(1) before and after are also large, although the latter value is not statistically significant. These positive values result from the large decrease of T2m in the NAC (Figs. 6b, c). On the other hand, in EN(1) (light gray), the values are negative, and the magnitudes are less than 0.2°C, in which the value in JAS(1) is almost zero. These negative values again result from the large increase of T2m in the NAC (Figs. 6e, f). However, asymmetry is clearly observed between the large positive values of LN(1) and the small negative values of EN(1) in the austral fall and winter from FMA(1) to JAS(1). In particular, the asymmetry is large in the earlier half around MAM(1), and the absolute value of LN(1) in this season is about five times larger than that of EN(1). The asymmetry in the near surface temperature difference between the IMC and the NAC may reflect the difference in duration between LN and EN events through the modification of AWM. In LN(2) (thick line), when the LN events decay, the values reach almost zero around FMA(2). The values become positively large in the following seasons but then decrease rapidly and change the polarity from positive to negative by JAS(2). It is reasonably expected that the anomalous development of the AWM was weak and its strength was rather neutral when the LN events decayed.

The modulation of AWM induced by the anomalous near surface temperature difference between the IMC and the NAC (Figs. 6, 8) was then examined in the anomaly fields of the 10-m-height meridional wind in and around the IMC (Fig. 9). In D(0)JF(1) of LN(1) (Fig. 9a), the anomalous southerlies (northerlies) appear in the eastern (western) NAC. The region with anomalous southerlies also expands to the north of 5°S. In addition, the regions with positive precipitation rate anomalies extend in and around the IMC, which reflects the enhanced convective activity in this region during LN’s mature phase (Figs. 2a, 3a). In the following MAM(1) (Fig. 9b), the anomalous southerlies are predominant in the NAC, which likely indicates the enhancement of a large-scale land breeze in the austral fall. We may reasonably expect an increase in positive precipitation rate anomalies near the coastline of the NAC with a system similar to a cold front (cf. Figs. 3b, 7b). In JJA(1) (Fig. 9c), the region with anomalous southerlies extends northward from the NAC to the IMC. Simultaneously, an anomalous meridional confluence is observed near 5°S with positive precipitation rate anomalies (Fig. 3c). The northward expansion of anomalous southerlies may be recognized as the northward advancement of the large-scale land breeze front. It is further suggested that the Hadley circulation is locally enhanced between the IMC and the NAC.

In D(0)JF(1) of EN(1) (Fig. 9d), we can observe a distribution pattern similar to that shown in D(0)JF(1) of LN(1) (Fig. 9a), although with opposite polarities.
However, this similarity disappears by MAM(1) (Fig. 9e) in the NAC, where large and significant anomalous northerlies are not observed. The small anomalous northerlies in the NAC are consistent with the small near-surface temperature difference between the IMC and the NAC (Fig. 8), suggesting that the strength of
the AWM was rather close to the climatological mean. By JJA(1) (Fig. 9f), a region with anomalous southerlies extending in the western tropical Pacific seems to move eastward, whereas the region with positive precipitation rate anomalies moves into the IMC from the west (Figs. 4f, g, 5b). The small but southerly wind anomalies distributing in the NAC were likely induced by the equatorial convective disturbances moving into the IMC from the west (Figs. 3f, 4g, 5b).

Throughout LN(2) (Figs. 9g–i), the meridional wind anomalies are generally small with no systematic distribution patterns. However, the distribution patterns and the transition may be somewhat similar to those in EN(1) (Figs. 9d–f), although with opposite polarities. In fact, close inspection indicates that the distribution pattern in D(1)JF(2) (Fig. 9g) is similar to that in D(0)JF(1) of LN(1) (Fig. 9a). In the following MAM(2) and JJA(2) (Figs. 9h, i), no large meridional wind anomalies appear in the NAC, suggesting that the strength of the AWM was close to the climatological mean.

To examine in more detail the temporal development of the precipitation rate anomalies and the northward extension of the region with anomalous southerlies, this study assessed the time–latitude cross section of the precipitation rate anomalies and the meridional wind anomalies averaged over the longitudes of the AWM, i.e., 120–140°E (Fig. 1c) (Fig. 10). In LN(1) (Fig. 10a), the region with positive precipitation rate anomalies moves northward from 17°S around the northern edge of the Australian continent to 5°S during March–June. Concurrently, anomalous southerlies (northerlies) are observed to the south (north) of the region with positive precipitation rate anomalies, where the resultant anomalous meridional confluence occurs. The positive precipitation rate anomalies remain large until October at around 5°S, during which time the anomalous meridional confluence was relatively large in this region. The large positive precipitation rate anomalies occurring at the latitudes of 0°–5°N during January–April seem to reflect the direct effects of stronger-than-usual Walker circulation induced in the mature phase of LN (Fig. 3a).

On the other hand, in EN(1), negative precipitation rate anomalies are observed at the latitudes of 0°–5°N during January–April (Fig. 10b); these are probably induced by the weaker-than-usual Walker circulation concurrent with the EN events (Fig. 3d). We may again identify the northward movement of the region with negative precipitation rate anomalies from the latitude of the northern edge of the Australian continent to 5°S during March–June. However, the magnitude of northerly wind anomalies appearing to the south was much smaller than that of southerly wind anomalies in LN(1) (Fig. 10a). In addition, no southerly wind anomalies appear to the north; that is, divergence of anomalous meridional winds does not occur around the region where the negative precipitation rate anomalies are predominant during March–June (Fig. 9e). In August, positive precipitation rate anomalies suddenly appear in the latitudes near the equator with the formation of confluence in the anomalous meridional winds (Fig. 9f). After this, an anomalous situation similar to that in LN(1) continues at least until December. The formations of anomalous meridional confluence and of positive precipitation rate anomalies in the equatorial region in August were likely induced by the convective disturbances moving into the IMC from the west in JJA(1) of EN(1) (Figs. 3f, 4g, 5b).

Figure 10 shows that the anomalous convective
system induced by the stronger-than-usual AWM moves northward from the northern edge of the NAC to the IMC during March–June in LN(1). This contributes to the maintenance of the anomalous updraft in and around the IMC, i.e., the ascent branch of anomalous Walker circulation, from May to July and beyond. In EN(1), the anomalous convective system comes into the IMC from the west, i.e., from the Indian Ocean, around August (Figs. 3f, 4g, 5b), which seems to overcome the effects of anomalous meridional winds induced by the weaker-than-usual AWM (Fig. 10b).

To identify the northward movement of the anomalous convective system in LN(1), we further assessed the latitude–height cross sections of the anomalous p-velocity averaged over 120–140°E using the 3-month means of JFM, MAM, MJJ, and JJA (Fig. 11). In JFM(1) of LN(1) (Fig. 11a), dominant anomalous upflows appear from the equator to 20°N in almost the entire troposphere, which reflects the stronger-than-usual Walker circulation in the mature phase of LN (Fig. 4a). Concurrently, another anomalous convective region occurs near the northern edge of the NAC around 15°S. By MAM(1) (Fig. 11b), the NAC’s anomalous convection is reinforced in the lower troposphere. Then, this anomalous convective system moves northward to the equator by MJJ(1) (Fig. 11c), whereas the region with anomalous upflows dominated in JFM(1) (Fig. 11a) at the equator–20°N moves slightly northward with weakening. In JJA(1) (Fig. 11d), a similar anomalous situation continues with further weakening in the northern convective system around 15°N. In LN(1), the convective system on the equator is replaced by that from the NAC during austral fall and winter.

The anomalous situation and its transition in EN(1) (Figs. 11e–h) are different from those in LN(1) (Figs. 11a–d). We cannot identify large downflow anomalies around the latitudes of the NAC, the northward movement, or the replacement of the equatorial downflow anomalies. Instead, the upflow anomalies of an equatorial convective system, which are embedded in the tropical latitudes with large downflow anomalies, are developed in the equatorial lower troposphere without meridional moving (cf. Figs. 4e–h).

It is interesting to compare the transitions in LN(1) (Figs. 11a–d) and those in LN(2) (Figs. 11i–l). The region with anomalous upflows in the equatorial latitudes in JFM(2) of LN(2) (Fig. 11i) gradually shifts northward to 15°N by JJA(2) (Fig. 11l). During this period (Figs. 11i–l), the region with equatorial anomalous downflows seems to move northward, but the anomalous upflows appearing between 10°S and 20°S fail to replace the equatorial anomalous downflows by northward moving. The anomalous upflows appearing between 10°S and 20°S are weaker than those in LN(1) (Figs. 11a–d), whereas the equatorial anomalous downflows are reinforced from the west (Figs. 4i–l). One significant difference between the transitions of LN(1) (Figs. 11a–d) and LN(2) (Figs. 11i–l) seems to be whether the anomalous convection near the equator was replaced by another convection from the south. LN and EN events are both ended by the anomalous equatorial disturbances moving into the IMC from the west (Figs. 4, 5).

The time development of meridional wind anomalies manifesting the strength of the AWM is further examined, together with the precipitation rate anomalies representing the convective activity in and around the IMC. Figure 12 indicates the longitude–time cross sections of the anomalies in precipitation rate and meridional wind at 10-m height for LN(1), EN(1), and LN(2). Since the precipitation rate anomalies are large near the northern edge of the region with meridional wind anomalies (Fig. 10), the latitudinal range for the meridional wind anomalies was set to 15°–5°S, whereas that for the precipitation rate anomalies was set to the equatorial region, i.e., 10°S–10°N. Only positive precipitation rate anomalies are indicated by shading.

In LN(1) (Fig. 12a), except for March, the meridional wind anomalies are positively large in the longitudes where the AWM prevails, i.e., 120–140°E (Fig. 1c). In particular, the anomalous southerlies are large in April–June and in the following several months. The increase in the anomalous southerlies from April reflects the stronger-than-usual AWM to the south of the IMC. It is interesting to note here that this enhancement in AWM is concurrent with the appearance of the second maximum in positive precipitation rate anomalies in the IMC in May–June (Figs. 5a, 10a). It is conceivable that the re-activation of convective activity in and around the IMC after March induced by the enhanced AWM contributes to the maintenance of the anomalous Walker circulation (Figs. 3b, c, 4b, c, 5a).

In EN(1) (Fig. 12b), the anomalous northerlies are predominant in the longitudes of the AWM from February to July; however, the magnitude was smaller than that of the anomalous southerlies in LN(1) (Fig. 12a). The large anomalous southerlies in the longitudes of 150°E–150°W from January to August are probably due to the meridional confluence associated with the equatorial anomalous westerlies in EN (Figs.
Fig. 11. Latitude–height cross sections of 3-month mean anomalies of vertical p-velocity (Pa s$^{-1}$; contour and shading) averaged in 120–140°E for LN(1) (left) in (a) JFM(1), (b) MAM(1), (c) MJJ(1), and (d) JJA(1); EN(1) (middle) in (e) JFM(1), (f) MAM(1), (g) MJJ(1), and (h) JJA(1); and LN(2) (right) in (i) JFM(2), (j) MAM(2), (k) MJJ(2), and (l) JJA(2). The contour interval is 0.2 × $10^2$ Pa s$^{-1}$. The scale of shading is presented at the bottom. The solid contour and shading are for the anomalous ascent, whereas the dashed contour is for the anomalous descent.
3d–f). The transition from EN to LN occurs around July when the region with positive precipitation rate anomalies expands to the IMC from the equatorial Indian Ocean.

The regions with anomalous southerlies and positive precipitation rate anomalies move eastward in LN(2), particularly in the latter half-year (Fig. 12c). In May–June, the anomalous southerlies are almost zero and much smaller than those in LN(1) (Fig. 12a). Then, the region with positive precipitation rate anomalies moves out of the IMC. By December, the southerly wind anomalies prevail all the way to the equatorial Pacific, and their distribution pattern is similar to that in January of EN(1) (Fig. 12b).

We further examined the detailed lag relationships among the anomalies in precipitation rate, 10-m-height meridional wind, and zonal wind at 850 hPa (Fig. 13). In LN(1) (Fig. 13a), the positive precipitation rate anomalies persist year-round and reach a maximum in June. The anomalous easterlies also remain throughout the year in the central equatorial Pacific, indicating the persistence of anomalous Walker circulation, although the magnitude decreases from February to July. These anomalous situations reflect the continuation of LN. In this year, the southerly wind anomalies reach their maximum in May, i.e., 1 month before the appearance of the maximum in positive precipitation rate anomalies. This supports the hypothesis that southerly wind anomalies prevailing in the region to the south induce the stronger-than-usual convective activity in the IMC. The large southerly wind anomalies during May–July represent the stronger-than-usual AWM,
although the positive anomalies temporarily become negative around March (cf. Fig. 12a).

In EN(1) (Fig. 13b), a systematic transition from EN to LN occurs around July. That is, the precipitation rate anomalies are changed from negative to positive, and the westerly wind anomalies become anomalous easterlies around July. The meridional wind anomalies also seem to change direction from northerly to southerly around this time, although the amplitude is small throughout the year. The magnitude of anomalies in LN(2) (Fig. 13c) is generally smaller than that in LN(1) (Fig. 13a) and in EN(1) (Fig. 13b). However, a closer inspection identifies the transition from LN to EN around July, i.e., changes in the anomalies of precipitation rate (positive to negative) and zonal wind (easterly to westerly). The meridional wind anomalies were almost zero throughout LN(2).

Finally, we diagnosed the processes connecting the
AWM to the persistence of LN events using the lag correlation technique (Fig. 14). Figure 14a is a scatter diagram of the T2m differences between the IMC and the NAC (120°–140°E, 15°S–0°) in MAM and the meridional wind at 10-m height in (120°–140°E, 15°–5°S) in AMJ (m s⁻¹; ordinate). Figure 14b is a scatter diagram of the meridional wind anomalies in AMJ (m s⁻¹; abscissa) and precipitation rate anomalies in MJJ (mm day⁻¹; ordinate). Both scatter diagrams are based on the 22 years of LN and EN events (Table 1). The small black dot (cross) represents the point of a LN (EN) event with its year, whereas the large black dot (cross) indicates the mean of the 11 LN (EN) events.

The lag correlation coefficient between the T2m differences in MAM(1) and the meridional wind anomalies in AMJ(1) is 0.46 and is statistically significant at the 5 % level (Fig. 14a). The positive correlation means that the southerly wind anomalies are enhanced when the T2m differences (IMC–NAC) are positively large, in which the latter leads the former. Almost half of the samples of LN(1) are in the first quadrant, whereas those of EN(1) are in the third quadrant, both of which contribute to the significant positive lag correlation in total. However, the locations of the means, i.e., large dot and cross, do not show point symmetry and do not form a straight line through the origin, indicating that the canonical LN and EN events were asymmetric in development. In particular, the relationship between the T2m differences in MAM(1) and the meridional wind anomalies in AMJ(1) is weak for EN events compared with those for LN events. It is interesting to note that only the LN events after 2000, except for an event in 2008, are in the first quadrant, which may reflect an interdecadal change in the relationship between the ENSO and the Asian–Australian monsoon (cf. Meehl and Arblaster 2011; Ashok et al. 2014).

The lag correlation coefficient between the meridional wind anomalies in AMJ(1) and the precipitation rate anomalies in MJJ(1) is 0.48, which is again statistically significant at the 5 % level (Fig. 14b). The positive correlation means that the anomalous AWM can adjust the precipitation or convective activity in the IMC. However, it is worth noting that the samples for EN(1) are scattered in the second and third quad-
rants, whereas almost all samples for LN(1), except for events in 1989 and 1996, are located in the first quadrant. This result indicates that in the AMJ(1) of LN(1), the AWM is stronger than usual and increases precipitation in the IMC in the following MJJ(1), whereas in the AMJ(1) of EN(1), the AWM is clearly weaker than usual but may not control precipitation or convective activity in the IMC in the following MJJ(1). In fact, the locations of the two means again do not show point symmetry and do not form a straight line through the origin. Precipitation in the IMC in MJJ(1) of EN(1) seems to be controlled by disturbances moving into the IMC from the Indian Ocean (Figs. 3e, f, 4f, g, 5b). Through these processes, the modification of the AWM may prolong only LN events, resulting in the asymmetry in periods between LN and EN events.

6. Summary

This study has proposed a mechanism by which the AWM can prolong LN events relative to EN events. The seasonal processes making such a difference can be summarized as follows. In DJF of the mature phase of LN, precipitation is increased in and around the IMC–NAC region with an upward branch of the anomalous Walker circulation (Figs. 3a, 4a). The increased precipitation then augments SSM in the NAC, which decreases surface temperature through increasing evaporation (Figs. 6a, 7a; Power et al. 1998; Jones 1999; Jones and Trewin 2000). The decrease in downward SSR also contributes to the lowering of the surface temperature in the NAC (Fig. 7a). During DJF, the enhanced Walker circulation dynamically maintains relatively high SSTs around the IMC, with easterly wind anomalies in the equatorial Pacific (Fig. 6a; Rasmussson and Carpenter 1982; Enfield and Mestas-Nuñez 2000). Therefore, a large surface temperature difference is established between the IMC and the NAC (IMC > NAC) (Fig. 6a), which induces an anomalous surface pressure gradient (i.e., IMC < NAC). Because of the difference in heat content between the land and the ocean, the surface temperature difference is largely determined by the land surface temperature of the NAC (Fig. 6a). In DJF of the mature phase of EN, the opposite processes are advanced, and the surface temperature difference is reversed from that of LN (IMC < NAC) (Fig. 6d).

The following processes and the differences between the LN and EN events are schematically summarized in Fig. 15. In MAM of LN(1) (Fig. 15a), the surface temperature anomaly is negatively large in the NAC. The AWM of LN(1) is strengthened by the anomalous pressure gradient between the IMC and the NAC, which is induced by their surface temperature difference, e.g., an enhanced large-scale land breeze. The stronger-than-usual AWM induces the enhancement of convective activity in and around the IMC (Figs. 9b, 10a) and may further decrease the surface temperature in the NAC through a local positive feedback embedded in the AWM, i.e., for the stronger-than-usual meridional circulation, increasing anomalous southerlies in the NAC, reinforcing surface evaporation, and lowering surface temperature in the NAC. This positive feedback of the AWM, which recalls wind–evaporation–SST feedback (Xie and Philander 1994), may be activated only in LN(1), leading to the duration asymmetry between the LN and EN events.

In JJA of LN(1) (Fig. 15b), the effects of AWM overcome those of the anomalous descent disturbances that move into the IMC from the Indian Ocean (Figs. 4b, c, 5a). The anomalous descent disturbances might have forced the anomalous Walker circulation to reverse its direction and shift eastward, leading to the end of the LN event. However, in actual fact, the convective activity in and around the IMC enhanced by the meridional AWM circulation maintains the stronger-than-usual Walker circulation in the equatorial Pacific. In MAM–JJA of LN(2), the effects of the AWM are weak, and the LN events are then ended by the effects of anomalous descent disturbances moving into the IMC from the Indian Ocean (Figs. 4j, k, 5c). However, in some LN events, we find a 3-year (or longer) persistence (e.g., LN event in 1998–2001). In such long LN events, the effects of the AWM may still be active even in LN(2) through long-lived positive AWM feedback.

In MAM of EN(1) (Fig. 15c), the surface temperature anomaly is positive in the NAC, and the surface temperature difference between the IMC and the NAC (IMC–NAC) is moderately negative (IMC < NAC) (Fig. 8). Thus, the AWM is almost neutral in MAM of EN(1) (Fig. 9e). In addition, the IMC–NAC region exhibits no correlation between precipitation and weakness of the AWM for this season (Fig. 14b). By JJA of EN(1) (Fig. 15d), the effects of anomalous ascent disturbances moving into the IMC from the Indian Ocean change the direction of anomalous Walker circulation to that of the LN, leading to the end of the EN event.

The stronger-than-usual AWM in LN events is a factor maintaining the convective activity in and around the IMC. The anomalous AWM may first be activated by LN itself through the anomalous Walker
circulation, but the activated AWM then prolongs LN events by enhancing the convective activity in and around the IMC. The mutual activation of convective activity in and around the IMC by LN and the AWM, which may be termed LN–AWM feedback, may yield the duration asymmetry between LN and EN events. In EN events, cooperation with the AWM is weak owing to the drier conditions in and around the IMC.

The abovementioned selectivity, or the nonlinear switching between LN and EN, may produce the predictability barrier of the ENSO in MAM. This begs the further question: besides the ENSO, are there other controls on the strength of the AWM? To address this question, we require more detailed quantitative evaluations of the mechanisms proposed in this study through comparison with the findings from earlier studies. To approach these questions, GCM experiments that artificially remove the Australian continent may be effective; these are presently underway.

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