Interannual Seesaw between the Aleutian and Icelandic Lows. 
Part I: Seasonal Dependence and Life Cycle

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ABSTRACT

The seasonal dependence and life cycle of the well-known interannual seesawlike oscillation between the intensities of the surface Aleutian and Icelandic lows (AL and IL, respectively) are investigated, based on the National Meteorological Center operational analyses for the period from 1973 to 1994. It is found that the correlation between the AL and IL intensities is significantly negative only from February to mid-March. It is also found that the seesaw exhibits an equivalent barotropic structure within the troposphere. For this late-winter period an index is defined that measures the intensity difference between the two lows. A linear lag regression analysis between this index and circulation anomalies averaged in each of the nine 45-day periods from early winter to midspring reveals that the stationary AL and IL anomalies constituting the seesaw do not start developing simultaneously over the respective ocean basins in the course of a particular winter season. Rather, the seesaw formation is initiated by the amplification of the AL anomalies with wave-activity accumulation in early through midwinter. In midwinter, part of the wave activity accumulated over the North Pacific propagates across North America in the form of a stationary Rossby wave train, which appears to trigger the formation of stationary anomalies over the North Atlantic. The IL anomalies thus initiated amplify and then become matured by late winter through the persistent feedback forcing from migratory eddies around the Atlantic storm track, while the AL anomalies remain strong until late winter through the continual feedback forcing from the Pacific storm track. It is suggested that interannual variability in the IL intensity for late winter tends to be strongly influenced by the AL anomalies that develop over the North Pacific in early through midwinter. The AL–IL seesaw is robust in a sense that it is apparent even after the influence of El Niño–Southern Oscillation is statistically removed from the data, suggestive of the importance of midlatitude processes in the seesaw formation.

1. Introduction

The surface Aleutian and Icelandic lows (hereinafter referred to as AL and IL, respectively) are wintertime semipermanent low pressure cells residing over the respective northern basins of the North Pacific (NP) and North Atlantic (NA; Fig. 1). Wallace and Gutzler (1981) indicated that the variability in the AL and IL is closely related to the Pacific–North American (PNA) pattern and the North Atlantic oscillation (NAO), which are the two most prominent teleconnection patterns in the upper-level wintertime Northern Hemisphere (NH) circulation (Kushnir and Wallace 1989). The anomalous AL and IL and their associated teleconnection patterns aloft give rise to changes in stationary flow patterns and the activity of synoptic-scale migratory disturbances along the major storm tracks as well (e.g., Lau 1988; Hurrell 1995b). Thus, they likely exert significant climatic impacts over the two maritime regions on monthly to interannual timescales (van Loon and Rogers 1978;
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Fig. 1. Climatological-mean SLP (hPa) for Jan over the NH. AL and IL denote the Aleutian and Icelandic lows, respectively. Two regions are shaded where the lowest pressures were identified, separately, for defining their intensities.

2. Data and methods

Our analysis on the AL and IL intensities and their seasonal evolution is based upon the National Meteorological Center [currently the National Centers for Environmental Prediction (NCEP)] operational analyses obtained from the National Center for Atmospheric Research (NCAR) Data Library. We used sea level pressure (SLP), 250- and 500-hPa geopotential height (Z_{250} and Z_{500}), and 250-hPa horizontal wind (U_{250} and V_{250}).

Several previous studies pointed out that the AL and IL do not fluctuate independently over the individual ocean basins. Rather, a seawayslike oscillation has been known to exist between the AL and IL from one winter to another (e.g., Kutzbach 1970; van Loon and Rogers 1978; Wallace and Gutzler 1981; van Loon and Madden 1983). In the monthly surface pressure fields, a signature of the seesaw is hinted even in the weak zonally asymmetric component of their wintertime leading variability that Thompson and Wallace (1998, 2000) called the Arctic oscillation. An interbasin atmospheric link between the NP and NA including the AL±IL seesaw has also been hinted in other observational studies (e.g., Deser and Blackmon 1993; Kushnir 1994; Hurrell 1995a; Xie and Tanimoto 1998; Rodwell et al. 1999; Tanimoto and Xie 1999).

One way to analyze the seasonal dependence and life cycle of the NP–NA link across North America is to examine the seasonal dependence and life cycle of the NP–NA link across North America. In the present study, we will investigate these aspects of the seesaw based upon observational data over 22 recent years, during which the seesaw signature was even more apparent than before.

The seasonal dependence of the correlation between the AL and IL intensities shown in Fig. 2 can be reproduced, by identifying the two low as the regional minima in the 31-day moving-averaged daily SLP maps, then calculating the correlation coefficient between their intensities for each calendar day directly from those time series and finally smoothing out the sequence with another 31-day moving average. The seasonal dependence of the correlation becomes less clear as the period for moving averaging is shortened.
period clearly indicates their out-of-phase relationship over the 22 winters (Fig. 3). To represent the strength and polarity of the seesaw, we defined the AL–IL index (AII) as the normalized IL-central pressure anomaly subtracted from the normalized AL-central pressure anomaly, both of which had been averaged over the peak period (Fig. 4a). Thus, positive and negative values of the AII correspond to the stronger IL with the weaker AL, and the stronger AL with the weaker IL, respectively. The AII, of course, exhibits extremely high correlation with each of the AL and IL intensities in the peak period (Table 1). This result reflects the fact that this seesaw relationship accounts for a major portion of the interannual variations observed in their intensities.

To elucidate the spatial structure of the AL–IL seesaw, we composited SLP anomalies for the peak period of five years with the most negative AII (i.e., 1981, 1983, 1984, 1986, and 1988) and other five years with the most positive AII (i.e., 1976, 1982, 1989, 1990, and 1992). We then subtracted the former composite from the latter (Fig. 5a). The seesaw is evident in this difference.
Fig. 5. Composite differences of (a) SLP (hPa) and (b) 250-hPa geopotential height (m) obtained by subtracting the composites for the five strongest negative AII years (stronger AL: 1981, 1983, 1984, 1986, and 1988) from those for the five strongest positive AII years (stronger IL: 1976, 1982, 1989, 1990, and 1992) based upon the 45-day mean for 31 Jan–16 Mar (peak period). Shaded lightly and heavily where the differences are significant at the 95% and 99% confidence levels, respectively.

ference composite map. A corresponding seesaw is also evident in the middle and upper troposphere for the peak period (Fig. 5b). The upper-level anomalies over the NP and NA associated with the AL–IL seesaw at the surface bear marked resemblance to the PNA pattern and NAO, respectively. In fact, the AII is strongly correlated with the PNA and NAO indices in late winter (Table 1; also Figs. 4a,b, and 4d). In recognition of the equivalent barotropic structure of the seesaw over the NA and NP as depicted in Fig. 5, we hereinafter focus mainly on the evolution of the upper-tropospheric seesaw that appears in the $Z_{250}$ field.

3. Life cycle of the AL–IL seesaw

To investigate the subseasonal evolution of the AL–IL seesaw, we prepared a $Z_{250}$ field averaged over each of the nine 45-day periods that are staggered equally by 15 days (17 November–31 December, 2 December–15 January, . . . , and 17 March–30 April). These nine periods correspond to early December, late December, early January, . . . , and early April. Figure 6 shows the linear lag regression coefficients of the mean $Z_{250}$ anomalies for each of the nine periods on the AII for the peak period. They represent a canonical subseasonal evolution of the interannual anomalies in $Z_{250}$ during the winter when the AL and IL are weaker and stronger than normal, respectively (i.e., positive AII winter). The corresponding evolution in the SLP field is depicted in Fig. 7 for three selected periods as indicated. Superimposed on the regression fields of the $Z_{250}$ anomalies with arrows is a wave-activity flux of Takaya and Nakamura (1997) based on the climatological-mean and anomalous fields of $U_{250}$, $V_{250}$, and $Z_{250}$. Theoretically, the flux is parallel to the local group velocity of a stationary Rossby wave train in the WKB sense, and it is divergent (convergent) where the wave train is strongly forced (dissipated). Therefore, it may be useful for representing a snapshot that illustrates qualitative aspects of the three-dimensional propagation of the wave train in the zonally varying climatological-mean flow. It should be noted that the sign is reversed for each of the anomalies in the regression maps (including those for anomalous storm track activities and their feedback forcing as discussed later) for the case with the stronger AL and weaker IL. No change is necessary, however, for the wave-activity flux. We have confirmed that the seasonal evolution of the seesaw as depicted in the linear regression maps is reproduced in the 45-day mean anomalies composited for the five years with the strongest positive and negative AII, separately (not shown).

In December, upper-tropospheric anticyclonic anomalies corresponding to the weaker AL gradually develop over the eastern NP (Figs. 6a and 6b). Then, a PNA-like wave train emanates from the anticyclonic anomalies in January and early February (Figs. 6c and 6d), forming cyclonic and anticyclonic anomalies over western Canada and the southeastern United States, respectively. During the same period, cyclonic and anticyclonic centers also develop over the NA off Newfoundland to the south of Greenland and over northern Europe, respectively. The former corresponds to the cyclonic anomalies at the surface that will develop into

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The wave activity defined by Takaya and Nakamura (1997) is a particular form of pseudomomentum that is not exactly conserved in a zonally varying basic flow. Yet, as shown in detail by Takaya and Nakamura (2001), it is approximately conserved under several realistic assumptions that are qualitatively valid for the wintertime climatological-mean flow observed in the extratropical Northern Hemisphere.
the stronger IL by late winter (Fig. 7). A diagnosis based upon the wave-activity flux suggests that these anomalies are associated with another stationary Rossby wave train that may emanate from the leading edge of the PNA-like Rossby wave train. During February, the wave train across North America gradually weakens (Figs. 6e and 6f), while the anticyclonic anomalies associated with the weakened AL remain strong and amplify even further until they cover almost the entire NP by the peak period (Figs. 6f and 7b). At the same time, the cyclonic anomalies over the NA further develop, while gradually shifting northward until they become matured in the

Fig. 6. Maps of the linear lag regression coefficient between the AL–IL index (AII) for the peak period (31 Jan–16 Mar) and 250-hPa geopotential height ($Z_{250}$) averaged for each of nine 45-day periods that correspond to early Dec, late Dec, early Jan, ..., and early Apr. The coefficient corresponds to a local change in height (m) when the AII increases by its unit standard deviation. Areas of light and heavy shading indicate where the AII–$Z_{250}$ correlation is significant at the 90% and 95% confidence levels, respectively, based upon the $t$ statistic with 11 degrees of freedom (half of 22 yr) assumed for a conservative measure of the statistical significance. Arrows indicate the horizontal component of the wave activity flux ($m^2 s^{-2}$; scaled as at the bottom) formulated by Takaya and Nakamura (1997, 2001).
peak period. Through the above sequence the AL–IL seesaw becomes fully developed (Figs. 6f and 7b). The anticyclonic anomalies over the NP start decaying right after this period, whereas the NAO-like anomalies remain strong until early April (Figs. 6g–i and 7c).

The stationary anomalies observed over the NP and NA in the course of the development of the AL–IL seesaw must alter the storm track activity over the two ocean basins. Figure 8 shows a map of the linear lag regression coefficient between the AII for the peak period and anomalous upper-tropospheric storm track activity for the period from 16 January to 1 March, during which the AL and IL anomalies are both still developing prior to the peak period. In the figure, the instantaneous storm track activity is measured at each grid point as the instantaneous 250-hPa wind velocity across the local jet axis associated with synoptic-scale migratory eddies. Specifically, for every 12 h we extracted the unfiltered wind component perpendicular to the jet axis defined at each grid point in the $Z_{250}$ field smoothed by low-pass filtering with a cutoff period of 6 days. We then squared the wind component thus extracted at each time step, smoothed out this squared time series with the 6-day low-pass filtering, and finally took square root of it at each time step. The resultant daily fields that represent the storm track activity were averaged within the particular 45-day period before computing the regression map of Fig. 8. The figure indicates that the Pacific storm track becomes more active than its climatological mean to the north of the weakened AL and in the subtropics while it becomes less active in between (Lau 1988; Rogers 1990). This anomaly pattern corresponds to the stronger difference of the Pacific storm track and background westerlies. Within the NA sector, the most apparent signature is the substantial suppression of the storm track activity along the southeastern coast of the United States. This suppression is accompanied by the slight enhancement to the north where the westerly wind speed is above normal at the upper levels (Fig. 6e). These two anomalies are indicative of the northward shift of the Atlantic storm track, which occurs concomitantly with the suppression of the storm track activity in the vicinity of Greenland. These anomalous storm track activities are associated with the stationary anticyclonic anomalies over the southeastern United States as shown in Fig. 6e. The above result is consistent with Dickson and Namias (1976), who argued that the stationary height anomalies over the southeastern United States could influence the mean circulation around Greenland through such changes in the moving cyclone activities as suggested from Fig. 8. They showed that migratory surface cyclones tend to be less frequent in association with the northward shift of cyclone tracks when the southeastern United States is covered with the stationary anticyclonic anomalies. They also show that
positive anomalies in surface air temperature over the southeastern United States, when covered with the anticyclonic anomalies, act to reduce the surface baroclinicity along the coast, giving rise to the suppression of storm track activity there. Our result over the NA is also consistent with Lau (1988) and Rogers (1990).

Anomalously active transient eddies migrating along the storm tracks over the NA and NP, as shown in Fig. 8, exerts anomalous feedback forcing that is likely to contribute toward the amplification and maintenance of stationary anomalies (Lau 1988; Lau and Nath 1991; Hurrell 1995b). In order to investigate a seasonal evolution of the barotropic feedback forcing from the transient eddies at the 250-hPa level, we computed the linear lag regression coefficient between the AII for the peak period and the local anomalous feedback forcing for each of the nine 45-day periods. The anomalous forcing was evaluated as the low-pass-filtered $Z_{250}$ tendency induced by anomalous vorticity flux convergence associated with the transient eddies (Nakamura et al. 1997b). In the presence of the weakened surface AL, the above-normal diffusiveness of the westierlies imposes stronger deformation on incoming synoptic-scale eddies along the Pacific storm track (Fig. 8). This eddy deformation and the aforementioned northward shift of the storm track axis result in significant anticyclonic feedback forcing through the anomalous vorticity flux divergence upon the anticyclonic stationary anomalies in the NP throughout the winter (Fig. 9). This feedback tends to be strongest when the accumulated wave activity has been released toward the NA in midwinter (Figs. 6d and 9d). Yet, it still remains strong (Figs. 9e–h), continuously forcing the stationary anomalies over the NP even after the release of the wave activity. In contrast, no significant eddy feedback forcing is observed over the NA until January (Figs. 9a–d). It is after the development of the stationary cyclonic anomalies in late January, which may be regarded as a component of a Rossby wave train across the NA, that the cyclonic feedback forcing from the anomalous storm track activities over the NA becomes strong and statistically significant (Figs. 9e and 9f). Apparently, this forcing contributes to the anomalous deepening of the surface IL and NAO-like anomalies aloft, and it continues throughout the peak period and even until late March, acting to maintain these stationary anomalies (Figs. 9g and 9h). The feedback forcing from the anomalous storm tracks exerted on the primary centers of action of the PNA pattern and the NAO is not only statistically significant but also substantial in magnitude. Although height tendencies induced by the barotropic forcing associated with anomalous eddy vorticity flux tend to be offset by the influence of anomalous eddy heat flux by 50%–80% in the upper troposphere (Lau and Nath 1991), we still expect the total height tendencies as the net forcing of 2–5 m a day to the south of Greenland. The net tendency is enough to account for the observed amplification (20–30 m within a 15-day period) of the cyclonic anomalies there just before the peak period.

4. ENSO influence on the seesaw

It is well known that the interannual variability in the wintertime atmospheric circulation over the NP is strongly influenced by ENSO events (e.g., Bjerknes 1969; Horel and Wallace 1981; Lau and Nath 1994). In fact, the PNA index is significantly correlated with the Southern Oscillation index (SOI) in late winter (Table 1; see also Figs. 4b and 4c). In order to assess the ENSO influence upon the formation of the AL–IL seesaw, we first computed the linear lag regression coefficient between a given variable at each grid point and the SOI over the 22-yr period (1973–94) for each of the nine 45-day periods. The resultant regression maps represent typical anomalies of the variable as a remote ENSO influence during a La Niña winter. They are compared with the typical anomalies associated with the formation of the AL–IL seesaw that are reproduced in Figs. 10a–c and 11a–c. As mentioned in section 3, the sign should be reversed for each of those anomalies, in discussing the situation for an El Niño winter. Again, no such change is necessary for the wave-activity flux.

Typical ENSO-related anomalies in $Z_{250}$ as depicted in their linear regression maps with the SOI are characterized by significant anticyclonic and cyclonic anomalies over the eastern part of the NP and central Canada, respectively (Figs. 10d–f). The overall anomaly pattern exhibits a certain level of similarity to that of the AL–IL seesaw. Though slightly shifted eastward, the anticyclonic anomalies over the NP are nearly as strong as the corresponding seesaw-related anomalies with only 20% reduction in magnitude. Likewise, the feedback forcing from the associated anomalous activities of the NP storm track (Figs. 11d–f) is almost as strong as the corresponding forcing during the seesaw formation (Figs. 11a–c). As pointed out by Hoerling and Ting (1994), the involvement of the storm track is very important in maintaining the stationary atmospheric response to ENSO over the NP. However, in the ENSO response the stationary anomalies and feedback forcing from the storm track over the NA are both substantially weaker than those associated with the seesaw formation. The stationary cyclonic anomalies to the south of Greenland are marginally significant, suggesting that part of the remote influence of ENSO may reach into the NA. Yet, they are only half in magnitude of the corresponding anomalies associated with the seesaw. Furthermore, the stationary anomaly pattern over the NA–European sector associated with the seesaw resembles the NAO (Fig. 10b), but the anomaly pattern in the ENSO response over the same sector does not. In fact, the cor-
relation between the AII and SOI in late winter, though not negligible, is not statistically significant (Table 1).  

Although the above results may indicate the weak ENSO influence on the wintertime circulation over the NA, as pointed out by N.-C. Lau (2001, manuscript submitted to J. Climate), they suggest that midlatitude processes may be of primary importance in the for-
mation of the AL–IL seesaw. To explore this possibility further, we attempted to remove the ENSO signal from the atmospheric fields, following Zhang et al. (1996) and Honda et al. (1999). Specifically, the “hypothetical ENSO response” at each grid point was defined for each of the nine 45-day periods as a product of the local regression coefficient (as plotted in Figs. 10d–f or 11d–f) and the SOI value for a particular winter. Then, the subtraction of the “hypothetical response” from the observed anomaly field yielded what may be called the “ENSO-removed” field. Midlatitude processes must be dominant in the ENSO-removed field that is statistically

Fig. 10. As in Fig. 6, but for 250-hPa height anomalies linearly regressed against AII for the peak period corresponding to late Feb (the top row: the standard case reproduced from Figs. 6d, 6f, and 6h), those linearly regressed against the SOI for individual 45-day periods (the middle row) and the “ENSO-removed case” (the bottom row). The columns correspond to late Jan (1 Jan±14 Feb; left), late Feb (31 Jan±16 Mar; middle), and late Mar (2 Mar±15 Apr; right). In the ENSO-removed case the ENSO influence has been removed by using the SOI-regressed height anomalies shown in the middle row for each of the 45-day periods. See text for more details.
Fig. 11. As in Fig. 10, but for the linear lag regression coefficients of 250-hPa geopotential height tendency (m day$^{-1}$) induced solely by the anomalous vorticity flux convergence associated with the migratory eddies along the storm track. The flux was evaluated from 8-day high-pass filtered wind field at the 250-hPa level, as in Nakamura et al. (1997b).

independent of the ENSO signal. Finally, we repeated the same analysis as in section 3, linearly regressing the ENSO-removed fields of $Z_{250}$ and the storm track feedback forcing for each of the 45-day periods against the AII defined for the peak period.

The evolution of the AL–IL seesaw in the ENSO-removed $Z_{250}$ anomalies (Figs. 10g–i) is similar to the actual evolution shown in Figs. 10a–c with respect to their spatial structures. In the ENSO-removed field for late winter, the seesaw signature is apparent with the cyclonic and anticyclonic anomalies over the NA and NP, respectively, that are comparable in magnitude to one another ($-65$ m vs $+70$ m; Fig. 10h), as we found in the observed seesaw-related anomalies ($-90$ m vs $+105$ m; Fig. 10b). Unlike in the ENSO response, an NAO-like anomaly pattern emerges over the NA–European sector as actually observed. The only major effect the removal of the ENSO influence imposes is sig-
significant reduction in the magnitudes of the anomalies. The anomalies are generally weaker by 20%–40% in magnitude than the actual anomalies associated with the seesaw. The removal of the ENSO influence yields the most distinct weakening in the stationary anomalies over western Canada in late January and the associated northeastward wave-activity propagation from the NP. The storm track forcing over the NP is also weakened substantially. In contrast, the removal yields any substantial weakening neither in the eastward wave-activity propagation across North America along the subtropical jet (Fig. 10g) nor in the feedback forcing from the NA storm track (Fig. 11h). The latter is suggestive of the primary contribution from the anomalous NA storm track toward the development of the IL anomalies.

The overall results presented in this section reveal the primary importance of midlatitude processes in the formation of the AL–IL seesaw, despite the fact that part of the remote ENSO influence over the NP is reaching into the NA. Our results are robust in a sense that no significant difference was found when the SOI was defined for each of the months from November to March or for each of the 3-month periods of November–January, December–February, January–March, and February–April.

5. Summary and discussion

It became apparent through our analysis that the interannual seesawlike oscillation between the AL and IL intensities distinctly appears only in late winter (February to mid-March; Fig. 2) and that the AL and IL anomalies that constitute the seesaw do not start developing simultaneously over the NP and NA in the course of a particular winter season (Figs. 6 and 7). Rather, the seesaw formation is triggered by the propagation of wave activity accumulated over the NP in early through midwinter toward the NA in the form of a PNA-like Rossby wave train across North America. The feedback forcing from anomalous local storm track activities contributes to the wave-activity accumulation over the NP. It continues to do so even after part of the accumulated wave activity is released toward downstream in midwinter, acting to maintain the AL anomalies until early spring. The IL anomalies are not evident until midwinter, when the stationary anomalies develop over the southeastern United States at the leading edge of the PNA-like wave train. The IL anomalies appear to be initiated as a component of another wave train across the NA emanating from the southeastern United States (Figs. 6c and 6d). The IL anomalies thus form through the feedback forcing from the enhanced (suppressed) storm track activities in the presence of the intensified (weakened) westerlies along the NA jet throughout late winter (Figs. 8 and 9). In this manner, the feedback contributes to the seesaw development also over the NA, in spite of the weakening of the Rossby wave train across North America after January. The linkage between the stationary anomalies over the southeastern United States and NA through the anomalous activities of wave cyclones was argued by Dickson and Namias (1976). Our study substantiates their argument with an evaluation of the feedback forcing from migratory eddies and suggests that the particular linkage could be a key process in the formation of the AL–IL seesaw. We found through our analysis in section 3 that part of the ENSO influence tends to reach into the NA. Nevertheless, it is rather weak and the correlation between the AII and SOI does not reach the significant level. Rather, the apparent signature of the AL–IL seesaw in the ENSO-removed field suggests the primary importance of midlatitude processes in the formation of the seesaw as mentioned above.

We have confirmed that the AL–IL seesaw can also be identified in a 40-yr period from 1958 to 1997, using the NCEP–NCAR reanalyses. In the present study we confined ourselves to the 22-yr period from 1973 to 1994, during which the signature of the seesaw was more apparent than before. We focused on this period, to elucidate dynamical processes involved in the seesaw formation in the clearest manner as possible. To further assess the robustness of the seesaw, its reproducibility was examined by evaluating the correlation between the AL and IL intensities for two independent 11-yr datasets that consist of odd-numbered (1973, 1975, . . . , 1993) and even-numbered (1974, 1976, . . . , 1994) years, separately. A significantly high correlation was found between the AL and IL intensities in the peak period for both datasets (Table 2), which demonstrates the robustness of the seesaw. The same examination was then repeated for the ENSO-removed SLP fields. The statistical significance is somewhat lower for the odd-numbered years than for the even-numbered years. Nevertheless, the correlation exhibits relatively high significance in each of the cases (Table 2), suggestive of the robustness of the result obtained in section 4.

Caution must be exercised when interpreting the time

1 In recognition of the group velocity of stationary Rossby waves being perpendicular to their phase lines, the difference we found in the wave propagation over North America between the ENSO response and ENSO-removed anomalies seems consistent with Ting et al. (1996). They found that phase lines of a stationary wave pattern over the PNA sector in association with a “midlatitude mode” and remote response to ENSO are meridionally oriented and tilted from northwest to southeast, respectively. This consistency provides another support for the importance of midlatitude processes in the formation of the AL–IL seesaw.

5 No significant bias was found in the SOI toward the warm or cold phase of ENSO for any of the even- and odd-numbered years over our analysis period. For example, the means of the normalized SOI for even- and odd-numbered years for late winter (Fig. 4c) are −0.05 and 0.05, respectively. Therefore, our particular assessment of the robustness of the seesaw is not subject to the ENSO-related bias.
geopotential height for the area 40°–80° is almost independent from the two adjacent years in each of the sets.

for 22 yr. Nine degrees of freedom were assigned for each set of the anomalies over the southeastern United States. The lat-

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evolution of the wave trains in the 45-day mean anomaly fields (Figs. 6 and 10). The group velocity of stationary

Steady Rossby waves is nearly twice as fast as the background westerly wind speed. Therefore, a series of anomaly

centers associated with the wave trains should form within a few days and their decay should also be rapid unless they keep being forced, for instance, by anom-

alous storm track activities as indicated in our analysis. Although wave activity at a particular location may un-

dergo a rapid increase on the quick passage of the leading edge of the stationary Rossby wave train, it may make no significant contribution to the 45-day mean budget. However, it does not necessarily mean that the role of the wave-activity propagation from the NP to NA is unimportant. Since the stationary Rossby wave train propagates through a waveguide along the upper-tropospheric westerly jet, the propagation acts to change mean baroclinicity of the jet and thereby tends to cause anomalous activities of synoptic eddies moving along a storm track. The anomalous storm track activities thus generated start imposing continual backfire forcing on the stationary anomalies as soon as they are formed behind the leading edge of the wave train. Therefore, in the 45-day mean budget of wave activity, its diver-

gence tends to be balanced with storm track feedback forcing. Nevertheless, it is the propagation of the sta-

tionsary Rossby wave train that triggers the persistent feedback forcing.

At this stage, we are not certain whether the wave train across the NA, if it actually exists, is excited through a nonlinear reflection of the incident stationary Rossby wave train around its critical latitude or through interactions with the lower-boundary conditions several weeks after the incidence of the wave train from the NP. A brief inspection suggests that the former mechanism is less likely to occur, as the critical line (i.e., a zero-wind line in the climatological-mean state at the 250-

hPa level) is located farther south of the southernmost anomalies over the southeastern United States. The latter mechanism may be more likely, as the height anom-

ies at the leading edge of the PNA-like wave train may alter the heat release from the warm ocean surface of the Gulf Stream. Of course, analysis based on the horizontal component of the wave activity flux in the upper troposphere alone cannot reveal its importance relative to the storm track forcings. Mechanisms where-

by the NA wave train emanates from the leading edge of the PNA–wave train need to be investigated in future studies.

A seasonal evolution pattern similar to that shown in Fig. 6 can be obtained even when the 45-day mean Z_{250}
anomalies are linearly regressed either on the IL intensity (Fig. 3) or on the NAO index (Fig. 4d) for the peak period (not shown). In fact, the indices of the IL intensity and NAO are significantly correlated with the AII in the peak period (Table 1). This result suggests that the formation of the interannual IL anomalies and of the NAO both observed in late winter is not strictly a local phenomenon within the NA. Rather, the formation appears to be associated with the wave-activity propa-

gation from the NP in the form of a PNA-like wave train in midwinter and with the continual feedback forc-

ing of the AL and IL anomalies by the anomalous storm track activities. Our study is suggestive of the possibility that the remote influence from the NP prior to the for-

mation of the NAO-like anomalies constituting the see-

saw may disturb the atmosphere–ocean coupling over the NA that has been built up during late autumn through midwinter. In fact, a piece of supporting evidence was given by Y. Kushnir (1999, personal communication), who found an interesting seasonal tendency in the cor-

relation coefficient between the NAO-like anomalies and underlying SST anomalies, as identified in the lead-

ing mode of the coupled atmosphere–ocean interannual variability over the NA. The correlation first becomes significant in November and increases until January, but it suddenly drops considerably in February, just before the formation of the AL–IL seesaw, and bounces back again in March.

Frequent occurrences of strong negative values of the AII in the early to mid-1980s (Fig. 4a) may manifest a tendency of the strong AL associated with the decadal variability in the coupled atmosphere–ocean system over the NP (Nitta and Yamada 1989; Trenberth 1990; Trenberth and Hurrell 1994; Zhang et al. 1997). Recent studies indicate the existence of an atmosphere–SST coupled mode over the extratropical NP, which is statisti-

cally independent of ENSO (Deser and Blackmon 1995; Zhang et al. 1996; Nakamura et al. 1997a). Nakamura et al. and Nakamura and Yamagata (1999) indeed suggested that the anomalous AL and PNA pattern aloft associated with the decadal variability are likely to be forced by SST anomalies in the subarctic frontal zone located in the extratropical western NP. Peng et al. (1997) and Peng and Whitaker (1999) showed through their GCM experiments that these SST anomalies in the midlatitude NP may possibly excite PNA-like anomalies by changing the NP storm track activities. In the NA

### Table 2. Correlation statistics between the AL and IL intensities a at the 250-hPa level for the peak period (31 Jan–16 Mar) for 1973–94.

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<th>ENSO-removed</th>
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<tr>
<td>All 22 yr (1973, 1974, 1975, ..., 1994)</td>
<td>−0.79** −0.73*</td>
</tr>
<tr>
<td>Odd-numbered years (1973, 1975, ..., 1993)</td>
<td>−0.80** −0.59*</td>
</tr>
<tr>
<td>Even-numbered years (1974, 1976, ..., 1994)</td>
<td>−0.89** −0.91**</td>
</tr>
</tbody>
</table>

** *, * Significant at the 99% and 95% confidence levels, respectively, based on the t statistic with 11 degrees of freedom assigned for 22 yr. Nine degrees of freedom were assigned for each set of the odd- and even-numbered years, in recognition that a particular year is almost independent from the two adjacent years in each of the sets.

a The AL and IL intensities are defined as the area mean 250-hPa geopotential height for the area 40°–55°N and 150°E–150°W for the AL, and 50°–70°N and 70°–10°W for the IL, respectively.

b Slightly misses the 95% confidence level but exceeds the 90% confidence level.
sector, it has also been shown through GCM experiments (Palmer and Sun 1985; Rodwell et al. 1999) that SST anomalies in the midlatitude western NA may also change the position and intensity of the IL. In their studies, Palmer and Sun focused on the emanation of a stationary Rossby wave train across the NA, whereas Rodwell et al. emphasized the role of anomalous storm track activities in the formation of the NAO. How the anomalous AL and IL and their associated wave train generate SST anomalies underneath and how they, in turn, exert feedback on those atmospheric anomalies are important subjects for future studies.

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REFERENCES


