BIDECADAL AND PENTADECADAL CLIMATIC OSCILLATIONS OVER THE NORTH PACIFIC AND NORTH AMERICA

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Abstract. Climatic variations on bidecadal (about 20 years) and pentadecadal (50–70 years) timescales are analyzed in terms of their seasonal and regional dependencies in the North Pacific and North American sector. The both bidecadal and pentadecadal variations are evident in SLP fields associated with the strength changes of the Aleutian low. The bidecadal variability is evident only in winter both in the SLP and air-temperature fields, whereas the pentadecadal signal exists from winter to spring seasons in the SLP field and only in spring in the air-temperature field. The SLP structure of the pentadecadal variability is approximately unchanged through the present century. The bidecadal variability exhibits frequency decrease (period increase) from 1930 to 1950, and simultaneously the center of the variability migrated southward. The century scale modulation of the bidecadal signal is also supported by the fact that an out-of-phase relationship between the Aleutian low strength and air-temperature holds throughout the present century for the Alaska but only after the 1930s for the West Coast of midlatitude North America.

1. Introduction

Recent investigations have revealed that the Aleutian Low plays a dominant role in the climatic changes over the North Pacific and west coast of North America from interannual to interdecadal timescales. On the latter timescale, two oscillatory variations have been identified from the analyses of instrumentally observed data; one is bidecadal (~20 years) (*Royer* 1989, 1993, *White et al.* 1997, *Mann and Park* 1994, 1996) and the other is a 50–70 year variability (*Minobe* 1997, see Figure 1). The evidence of the latter timescale is also found in the tree-rings in the eighteenth and nineteenth centuries (*Minobe* 1997). Hereafter, for simplicity, the 50–70 year variability is referred to as pentadecadal variability. Three porality changes of the pentadecadal variability in the present century are known as climatic regime shifts in the mid 1920s, late 1940s and mid 1970s (*Kondo* 1988, *Minobe* 1997, *Mantua et al.* 1997). The last regime shift has attracted large attentions of climate researchers (e.g., *Nitta and Yamada* 1989, *Trenberth* 1990). The similarity between the shifts in the 1940s and 1970s were documented by several papers (e.g., *Yamamoto et al*. 1986, *Francis and Hare* 1994, *Dettinger and Cayan* 1995, *Zhang et al.* 1997).

For both timescales, it has been reported that the climatic changes significantly influence marine ecosystem. Evidence of the effect of the bidecadal variability on fishery resources is summarized by *Royer* (1989), and significant influences of the 50–70 year variability have been documented by *Francis and Hare* (1994) and *Mantua et al*. (1997) for salmon resources in Alaska and Pacific Northwest, and *Kodama* et al. (1995) and *Yasuda et al.* (1998) for Japanese fish catches. A Summary of the influence of the interdecadal regime shift in the 1970's are contained in *UNESCO* (1992), *Trenberth and Hurrel* (1994) and *Mantua et al.* (1997). Therefore, the bidecadal and pentadecadal variability are responsible for important socio-economic effects through the change of fishery resources in the North Pacific.

The important socio-economic effects of the interdecadal changes urged us to understand the nature of the interdecadal variability. Although the existence of the two major timescales has been reported in the North Pacific/North American sector, the features of these timescales have not been fully investigated. The purpose of this paper is, therefore, to clarify the characteristic of the variations of these two timescales. In particular, we focus our attention on similarities and differences of these two interdecadal timescales concerning their regional and seasonal distributions based on instrumental records mainly obtained in the present century.

In addition to the important socio-economic influences of the bidecadal and pentadecadal variations, investigations of these climatic variations are also important in order to understand the earth's climate system. Interdecadal climatic variations in the Pacific Ocean were considered to arise from the coupled atmosphere and ocean system (e.g., *Latif and Barnett* 1994, *Gu and Philander* 1997, *White and Cayan* 1998, *Jin* 1997). All these papers attributed mechanisms of the interdecadal variations to delay-negative feedback processes. Although there were differences between specific models treated in those papers, there were also important common features. In a delay-negative feedback model, the effect of the atmospheric forcing in a region (forcing region) is kept in the ocean, and resulted oceanic anomaly (usually temperature anomaly) moves to another region (feedback region), where the retained anomaly in the ocean feedbacks to the atmosphere. The feedback from the ocean to the atmosphere changes the polarity of the atmospheric anomaly in the forcing region, resulting in a periodic oscillation in the coupled ocean-atmosphere system. The period of the oscillation is tends to be a double of a delay time, for which the oceanic anomaly travels from the forcing region to the feedback region (e.g., *Gu and Philander* 1997). In other words, the ocean acts as a memory by connecting the forcing and feedback regions with a specific delay time, and determines the timescale of the variability. Different timescales suggests that different processes play significant roles as the oceanic memories, and the feedback regions to the atmosphere are also different.

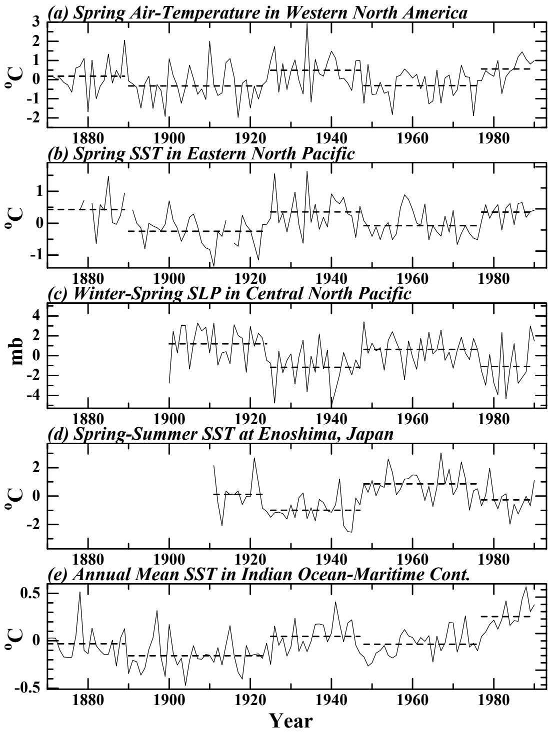


Figure 1. Time series of anomalies exhibiting coherent interdecadal climate changes (thin solid curve), with temporal averages of the anomalies for the periods 1870–1889, 1890–1924, 1925–1947, 1948–1976 and 1977–1990 (thick dashed lines). (a) Spring (Mar.–May) air-temperature anomalies in western North America averaged over 130°W–105°W, 30°N–55°N. The air-temperature anomaly is calculated relative to 1930–50 at each station, and then the anomalies are averaged spatially. (b) Spring SST anomalies in the eastern North Pacific averaged over 140°W–110°W, 30°N–55°N. The average is calculated when available grid points are more than 20% of total grid points in the spring of respective years. (c) Winter-spring (Dec.–May) SLP anomalies in the central North Pacific averaged over 160°E–140°W, 30°N–65°N. (d) Spring-summer (Mar.–Aug.) SST anomalies at Enoshima, Japan. (e) Annual mean SST anomalies in the Indian Ocean–maritime continent region averaged over 40°E–160°E, 15°S–15°N. All differences in the temporal average between successive periods are significant at the 95% confidence level in each time series. A stronger (weaker) Aleutian low results in warmer (colder) air-temperature and SST in the west coast of North America by the advection of warmer air, and colder (warmer) Enoshima SST at Enoshima probably due to southward anomalous penetration of the Oyashio along Japanese coast. A stronger (weaker) Aleutian low is accompanied by the warmer (colder) SST in the Indian Ocean/Martime continet region. *(after Minobe 1997).*

The rest of the present paper is organized as follows: in section 2, data and method of data analysis are described; the results are shown in section 3; and summary and discussion are given in section 4.

2. Data and analysis method

We examine the following two gridded datasets: sea-level pressure (SLP) and air temperatures. The monthly SLP data are the updated version from January 1899 to December 1997 of *Trenberth and Paolino* (1980). The seasonal air-temperature data from 1851 to 1993 were compiled by *Baker et al.* (1994) based on monthly temperature data collected as part of the *Global Historical Climate Network* (*Vose* et. al., 1992). Both SLP and air-temperature data are on 5°×5° grids.

Through the present paper, yearly time series are examined for each month or each season. The yearly time series of a variable for each season is obtained from averaging the variable for each season. Winter was defined as the months December-February, spring as March-May and so on.

We employ the Maximum Entropy Method (MEM) to identify the periodicity of signals. The significance of a MEM spectral peak is tested by using a Monte-Carlo Method. We generate 10,000 surrogate time series according to a red-noise model using a observed correlation coefficient with one-year lag. Then, the amplitude at the top 5 (10) % population of surrogate MEM spectra is identified as 95 (90) % confidence level. As in the case of Blackman-Turkey spectrum, for a larger lag-1 correlation, higher MEM spectral amplitude is required to be significant.

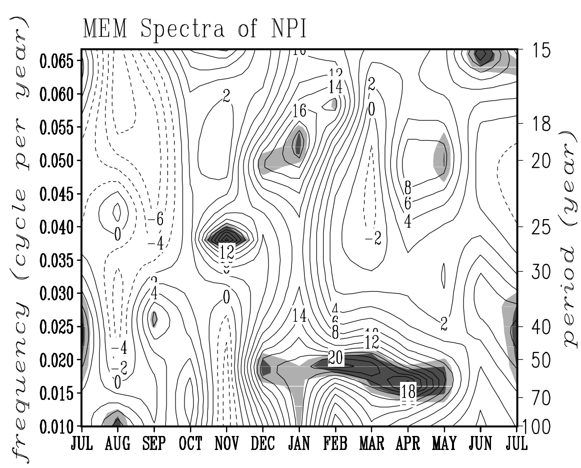


Figure 2. MEM spectra of NPI for each month. The contour indicates the spectral amplitude and the shade denotes the significance of the spectrum. The contour interval is 2 DB. Heavy and light shades indicate where the spectrum is significant at 90 and 95 % confidence limits, respectively.

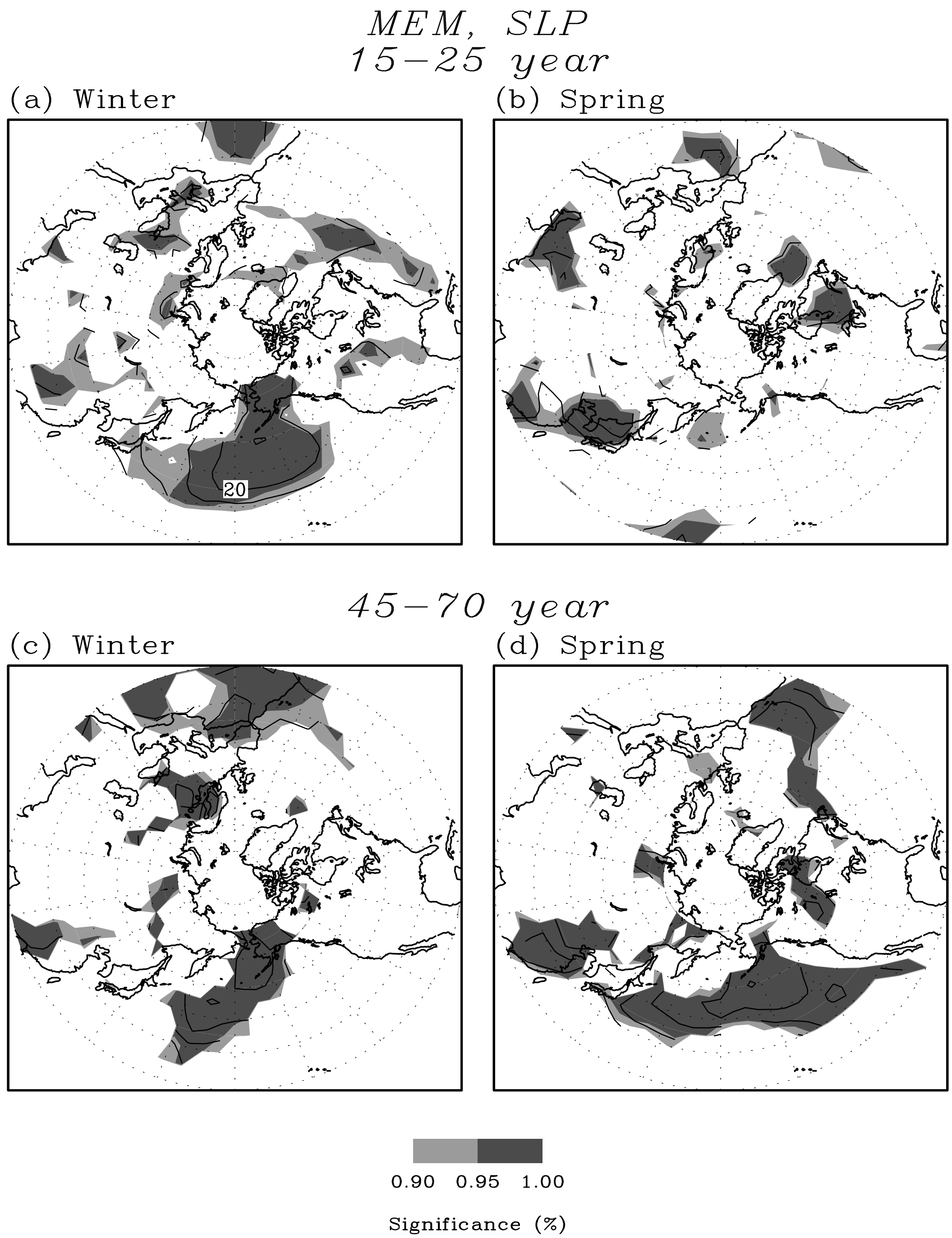


Figure 3. MEM spectra for winter (a and c), and spring (d and e) in bands of 15–25 year period (a and b), and 45–70 year period (c and d). The contour indicates the spectrum amplitude at the most significant spectrum peak, and the shade denotes the significance of the peak. The contour is drawn for the region where the significance is larger than 0.8. Contour interval is 5 DB.

3. Results

In order to examine the seasonality of the interdecadal variations of the Aleutian low, we identify the existence of significant spectral peaks in the North Pacific Index (NPI) for each month using the MEM. The NPI serves as a measure for the strength of the Aleutian low, and is defined as the SLP averaged over 160°E–140°W, 30°–65°N (*Trenberth and Hurrel*, 1994). The MEM spectra exhibit significant peaks around the period of 20 years in the months of December and January, with most of the spectral power concentrated in January (Figure 2). The other cluster of significant spectral

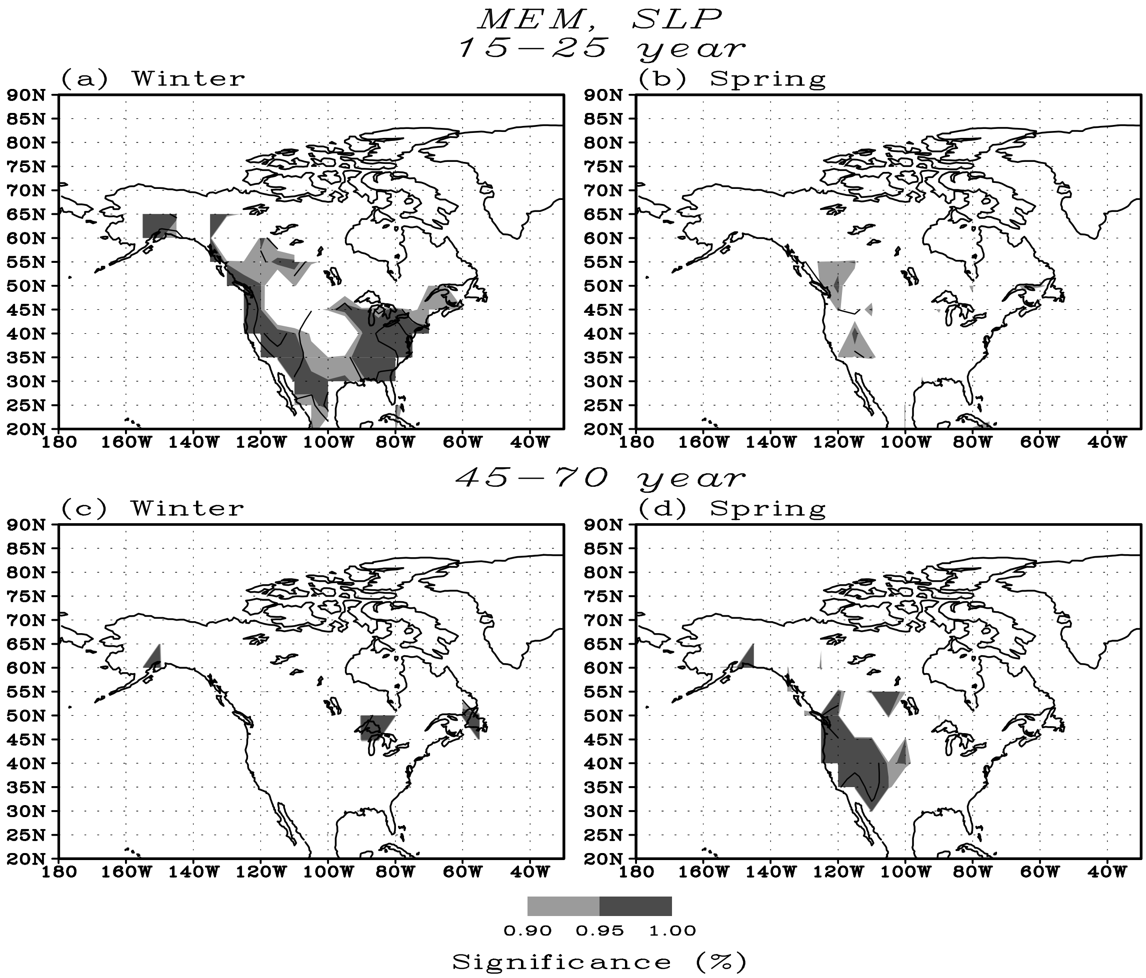


Figure 4. Same as Figure 3, but for the significance of the MEM spectra of the surface air-temperature at each grid point over North America.

peaks is in a period band of 50-70 years for the months from December to May. In short, bidecadal variability was observed only in wintertime, whereas pentadecadal variability was found both in winter and spring seasons. In addition to the bidecadal and pentadecadal signals, 25 year variability is observed in November. The 25 year signal is likely not to have a relationship with the bidecadal or pentadecadal variations from analyses of corresponding time series (not shown). We, thus, ignore this signal in the present paper, and confine our attention on the bidecadal and pentadecadal signals in winter and spring seasons.

In order to identify the spatial significance, MEM is performed at each grid point with winter and spring season data, respectively. The results of the "spectral power mapping" are summarized by the spectrum amplitudes and their significance in the frequency bands of 15−25 and 45−70 year periods (Figure 3). These band ranges are chosen so that the bidecadal and pentadecadal timescales are separated. The amplitude and significance of the spectra are shown for the most significant peak in each band. For the 15−25 year band, the spatial distributions of the MEM spectra indicate that the significant spectrum peak occurs in Aleutian region in winter. The bidecadal signal is the most well-organized in the central North Pacific. For the 45−70 year band, the significant signal observed both in winter and spring seasons, as expected from the analysis of the NPI shown in Figure 2.

In parallel to the above MEM assessment for the SLP, MEM spectra were calculated for the surface air-temperature at each grid point and in each season over North America (Figure 4). Again, significant signals on two interdecadal timescales are found with a distinction between winter and spring in North America. For the 15−25 year band, the significant peak is found on both sides of the continent. The significant temperature variability of 45−70 year period is seen in midlatitudes of western North America in spring with a further penetration toward the inside of the land than for the 10−25 year band. It is noteworthy, in contrast to the existence of the pentadecadal SLP signal in both winter and spring, the temperature signal on pentadecadal timescale is evident only in spring and absent in winter. The absence of the winter pentadecadal signal over North America might be resulted from the fact that the SLP signal in winter is more confined to the west over the North Pacific than in spring (Figure 3c,d). The springtime similarity between the 1930s and 1980s were also documented by *Parker et al*. (1994).

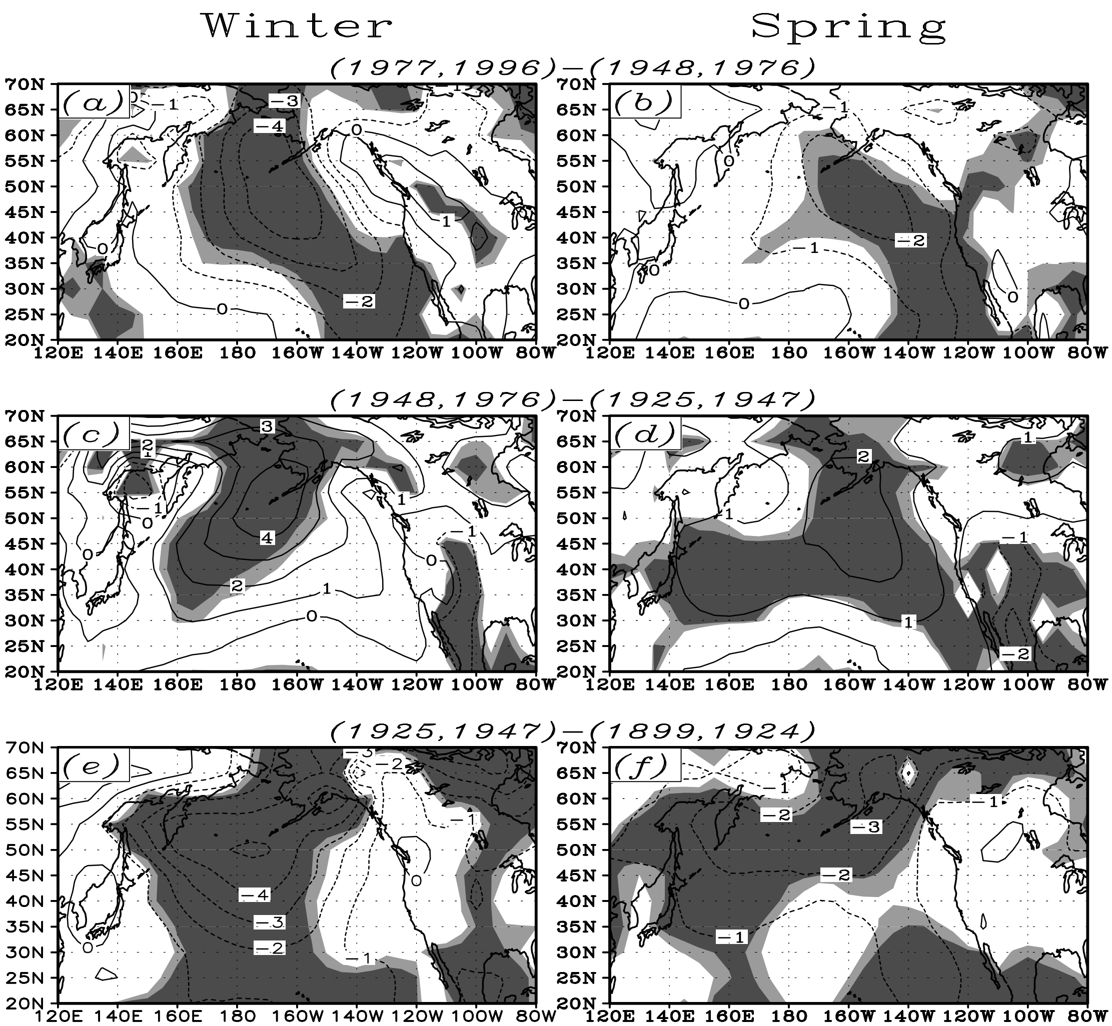


Figure 5. SLP difference between two successive periods in winter (left panels) and spring (right panels). The periods are defined as 1977–1996, 1948–1976, 1925–1947 and 1899–1924. The contour indicates the amplitude of the difference and the dens and weak shades indicate the regions where the difference is significant at the 95% and 90% confidence limit, respectively.

The sign reversals of the pentadecadal variability correspond to the regime shifts in 1920s, 1940s and 1970s (*Minobe* 1997, *Mantua et al.* 1997). Figure 5 shows the winter and spring SLP difference between two successive periods. Here use the identification in previous studies of regime shifts; the shifts were detected in 1924/25, 1947/48, 1976/77 (*Minobe* 1997). As shown in Figure 5, the regime shifts are detectable in both in winter and spring. Between two successive periods of a regime, the spatial distributions of SLP differences exhibit approximately the same patterns, with strongest anomalies over the central northern North Pacific and a weaker anomaly with the opposite sign over western North America. These patterns are related to the Pacific/North American teleconnection pattern in the atmospheric circulation aloft (*Wallace and Gutzler* 1981).

On the other hand, in consistent with the MEM assessment of the air-temperature, the air-temperature difference between two periods in midlatitude North America is significant mainly in spring, but not in winter (Figure 6). The relation between the SLP and air-temperature is a consequence of the strengthened (weakened) Aleutian low enhancing (reducing) the advection of warmer air onto the west coast of North America (*van Loon and Williams* 1976). The warming after the mid 1970s, which occurred over a wider area than the earlier three changes, might be due in part to the anomalous atmospheric circulation associated with the North Atlantic Oscillation (*Hurrel* 1995). Except for the widespread and amplified air-temperature difference for the regime shift in the 1970s, three shifts in the present century exhibit similar SLP and air-temperature distributions. Thus, the linkage in the changes between the Aleutian low strength and air-temperature in the west coast of North America well holds throughout the present century.

Although the spatial distribution of the pentadecadal variability are essentially the same through the present century, the bidecadal variability exhibits significant century scale modulations. Figure 7 shows the wavelet coefficients of the winter NPI. The frequency modulation feature is evident; the

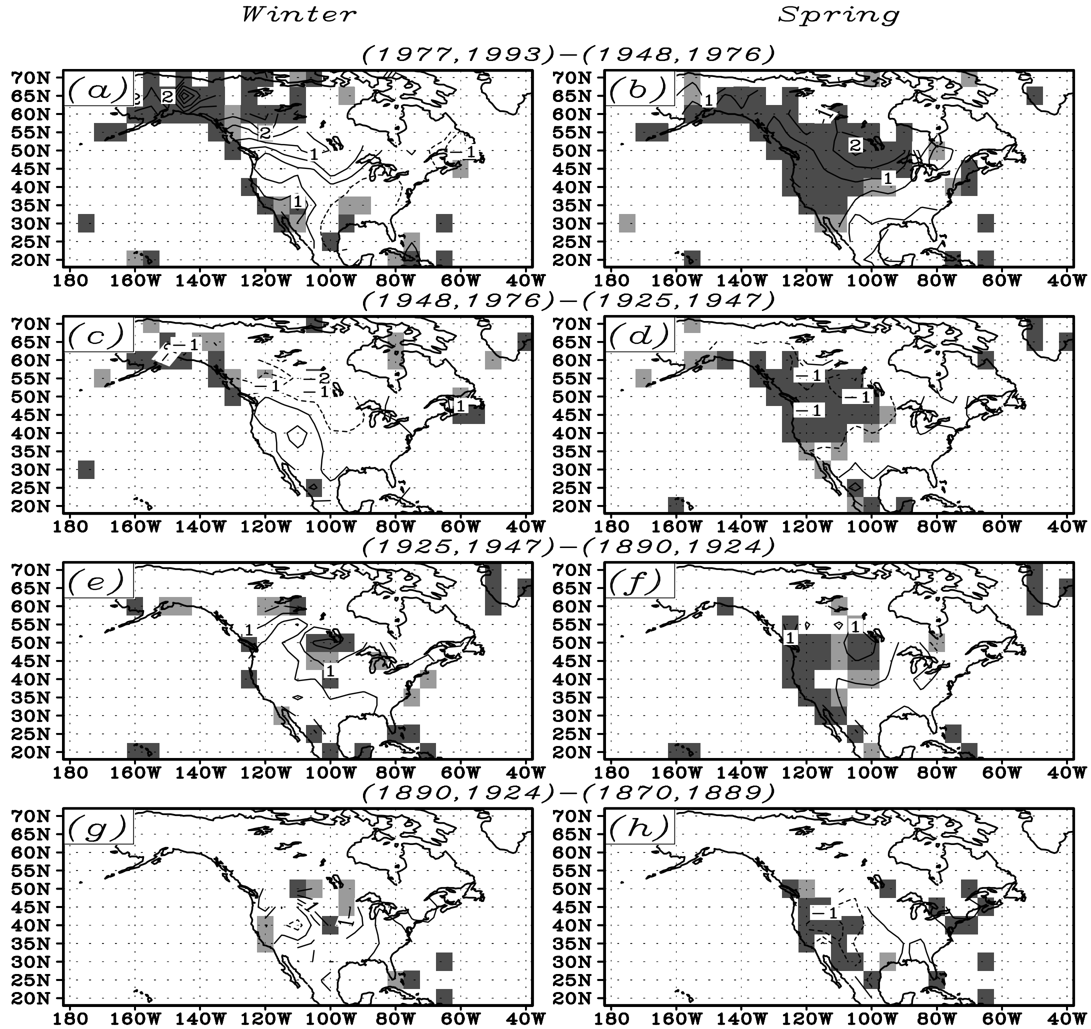


Figure 6. Same as Figure 5, but for the surface air-temperature. The periods are defined as 1977–1996, 1948–1976, 1926–1947, 1890–1925 and 1879–1899.

period in the first two decades of this century is about 14 years, and then increased rapidly from 1930 to 1950. The period in the last half of the record is about 19 years.

In order to identify the century-scale changes in the spatial distribution of the bidecadal variability, we examine the bidecadal filtered (10<yr<30) winter SLP field (Figure 8). In the first few decades, significant amplitudes are confined to the north of 60°N, then migrated toward the south from 1930 to 1950, which is approximately the same as the period of the period increase shown in Figure 7. The center of the variability reached about 45°N in 1950. After that the central latitude shifted further toward the south, and the center of the variability after 1980 locates about 40°N, which is about 10 degrees south of the center of the pentadecadal SLP signal shown in Figure 5.

It is interesting to see whether the southward migration of the bidecadal signal is consistent with other data than the SLP data, which are known to be less reliable for the first few decades of this century than for the recent decades (*Trenberth and Paolino* 1980). Thus, we compare the air-temperature data with the NPI. Figure 8 shows bidecadal-filtered (10<yr<30) air-temperatures averaged over Alaska (170–145°W, 50–50°N), western North America (125°–110°W, 35–50°N) and eastern North America (95–75°W, 30–40°N), respectively, along with the bidecadal-filtered NPI. The tightest linkage is found between the NPI and Alaska air-temperature. These two time series exhibit a prominent out-of-phase relationship throughout the present century. On the other hand, an out-of-phase linkage between the Aleutian low strength and air-temperature in western North America holds only after 1930s. The eastern North America air-temperature has an in-phase relation with the Aleutian low strength, though correspondence of peaks between the air-temperature and NPI in the first few decades is weaker than that after 1940s. These relationships between the NPI and air-temperature are consistent with the southward shift of the bidecadal variability in the SLP field. In the first few decades of this century, the bidecadal variability is mostly confined over the Alaska, accompanied by the coherent local air-temperature changes. In this period, however, the SLP anomaly over the North Pacific is weaker, and hence the relationship between the NPI and air-temperature in the west

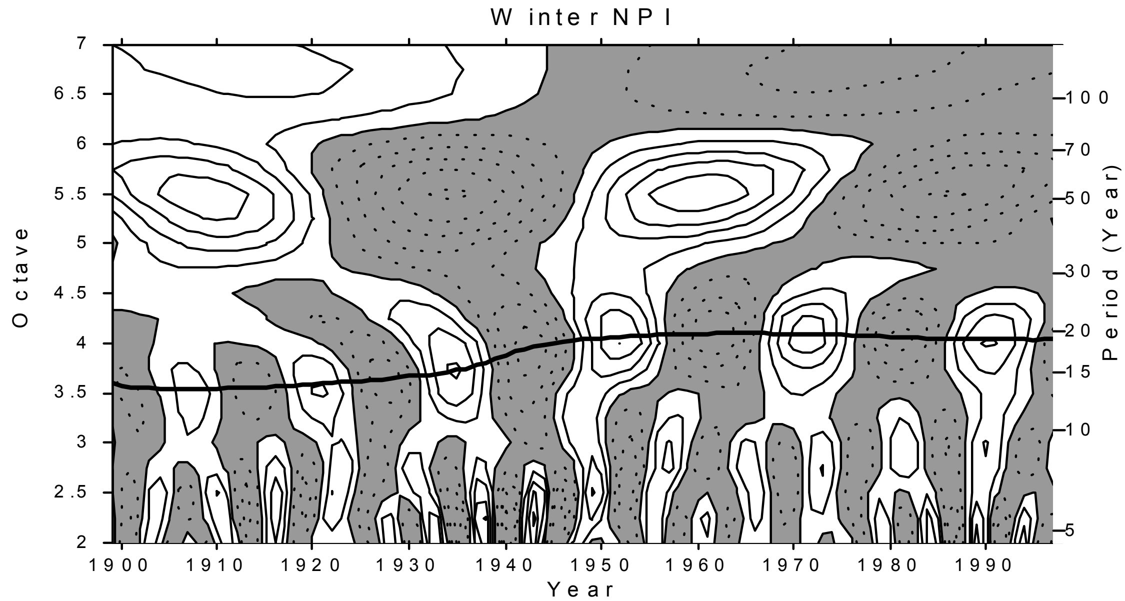
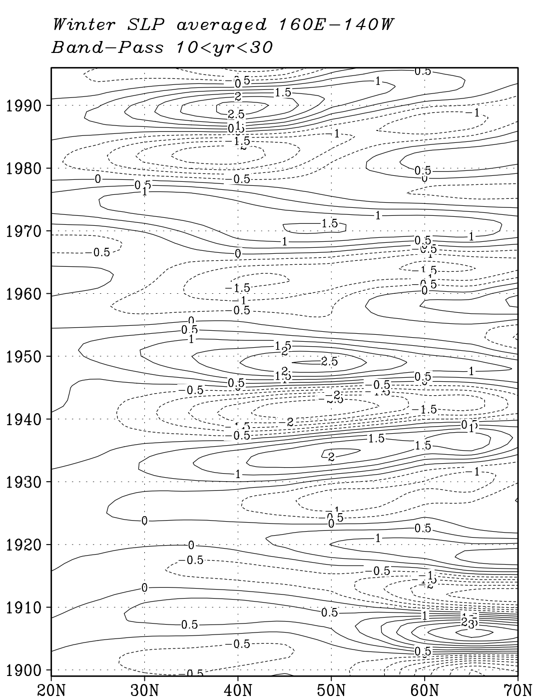


Figure 7. Real part of Morlet wavelet coefficients of winter NPI. Shade indicates the negative values of the amplitude. Thick curve indicates the center of the wavelet amplitude on bidecadal timescale. In order to reduce edge effects, time series of the winter NPI is extrapolated before wavelet transform by an autoregressive method (MEM) at the beginning and end by 20% of the record length. In the first few decades, the period is about 14 years, increased rapidly from 1930 to 1950. The period reaches its maximum of 19.5 years around 1960, and then slightly reduced to the present.

coast of North America did not hold. Consequently, the southward migration of the bidecadal signal is likely to be a real feature and not due to the change of the quality of the data.



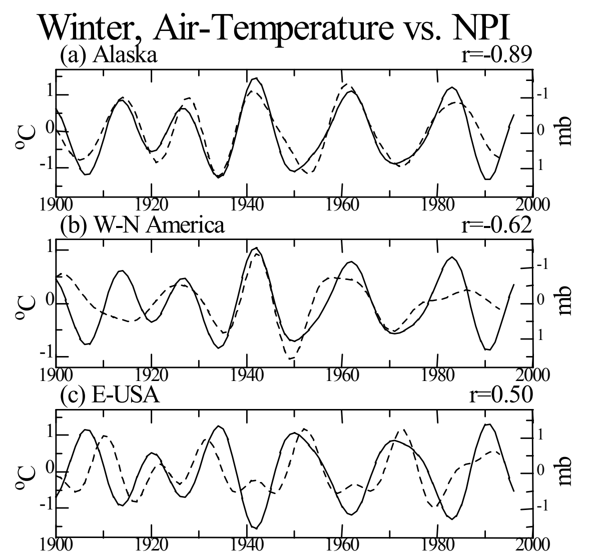


Figure 9. Band-pass filtered (10<yr<25) winter air-temperature (solid curves with left axis) and NPI (dash curves with right axis). The air-temperature is averaged over respective regions where the bidecadal variability is significant in Figure 4. For an easier comparison, the axis of the NPI in panels (a) and (b) are reversed (negative direction is toward the top of the page.) The numbers at the top-right corner of each panel indicates the correlation coefficients between the air-temperature and NPI.

Figure 8. Band-pass filtered (10<yr<30) winter SLP with a zonal average from 160°E–140°W.

4. Summary and Discussion

We have shown that the two dominant interdecadal timescales, about 20 years (bidecadal) and 50–70 years (pentadecadal), exhibit significant regional and seasonal differences in their distributions. The bidecadal variability is found in the wintertime SLP in the central North Pacific and also in the winter air-temperature over North America. The air-temperature in western (eastern) North America is high (low) when the Aleutian low is strong. The pentadecadal signal is evident in SLP both in the winter and spring, but the corresponding air-temperature variability is found only in the springtime air-temperature in western North America. The bidecadal variability exhibits significant century-scale modulations. In the first three decades of this century, period was about 14 years and center of the variability located to the north of 60°N. From 1930 to 1950, the period increased to about 19 years and the center of action migrated to about 50°N. The period exhibit slight decrease after 1960, while the center of variability further moved toward the south to about 40°N in the 1980s. The southward shift beginning in 1930 is consistent with fact that the out-of-phase relationship between the air-temperature and Aleutian low strength holds throughout the present century in Alaska, but after 1930s in midlatitude western North America.

Bidecadal variability in winter season in North Pacific climate has been documented in several papers in a regional or global analyses (*Royer* 1989, 1993, *Mann and Park* 1994, 1996, *Lau and Wang* 1996, *White et al.* 1997). The results of the present analyses are consistent with previous studies, but indicate that the bidecadal variability is limited to the winter season in SLP and surface air-temperature over the North Pacific and North American sectors. Furthermore, the century-scale modulation of the bidecadal variability is fully documented by the present paper.

Although global 65–75 year climate variability has been identified in some studies (*Schlesinger and Ramankutty* 1994, *Mann and Park* 1996), oscillatory variability at this frequencies appears to be significantly different from the pentadecadal variability documented in the present paper. First, the 50–70 variability has the range of the periodicity from 50 years to 70 years in the eighteenth and nineteenth centuries, but tends to the 50 year variability in the present century. The MEM spectra for the NPI averaged from winter to spring has the peak at a period of 51 years, which is somewhat shorter than the global 65–75 year periodicity. The global 65–75 year time series of *Shelesinger and Ramankutty* (1994) and *Mann and Park* (1996) exhibit a sign reversal in the mid 1920s as does the pentadecadal variability in the North Pacific. However, the next sign reversal identified in the global 65-75 year variability occurred around 1960, with no sign reversal after that to the present. Conversely the pentadecadal variability exhibits sign reversals in the 1940s and 1970s. Probably, the global analyses keys on a signal over the North Atlantic and northwestern Eurasian continent, and hence this signal does not capture interdecadal variations mainly limited to the North Pacific sector.

The difference between the bidecadal and pentadecadal oscillations must be taken into account for a planning of paleoclimate reconstructions. For example, *Minobe* (1997) showed that the reconstructed spring air-temperature over western North America exhibits the 50–70 year variability at least last three centuries. The results of the present paper indicates that this region is a most suitable region for capture the pentadecadal variability in North American continent linked with the Aleutian low strength. However, the bidecadal variability in the Aleutian low is likely to reflect more properly to the air-temperature in Alaska than in the western North America. Therefore, for a specific timescale of the variability, one should choose proper paleoclimate proxies in terms of their regional and seasonal dependencies.

Although it is clear that further researches are required for the interdecadal climate changes, there might be arguments about what kind of researches are required. I consider that the researches can be classified into two groups. One group consists of purely physical researches, which are conducted mainly to understand the mechanism of the climatic changes and to enable to predict climate variations in future. For the interdecadal climatic changes in the Pacific Ocean, several hypotheses were proposed. The most important differences between them is the *memory* of the climatic changes: *Latif and Barnett* (1994) and *Jin* (1997) attributed the memory of the variability to the oceanic Rossby waves; *White and Cayan* (1998) hypothesized the SST anomaly is retained more than 10 years and is advected by oceanic surface circulations; and *Gu and Philander* (1997) proposed that equatorward advection of a subducted water mass from the midlatitude North Pacific is the agent of interdecadal variability. These hypotheses are now examined intensively by analyzing the observed data and results of coupled GCMs or oceanic GCMs. A key element of the examinations is to identify the *memory* of the climate system. Fir the pentadecadal variability, a hypothesis has not been proposed, and is hoped to be investigated by data analyses and also by modeling studies.

The other group of researches is to identify the influences of the climate changes, especially for the changes exerting socio-economic importance. In particular, researches for marine ecosystem would be the focus of attentions. For those evaluations, closer corporations among physical oceanographers, meteorologists and ecosystem researchers are required. In a physics-oriented research, some interdecadal signals might be ignored, in the case where that signal seems to be somewhat *minor*, evaluated from the point of view of the physical oceanography or meteorology. However, the nonlinearity in a ecosystem and different sensitivity to the climatic changes in different regions and seasons might cause the *minor* signal to be meaningful for a ecosystem. For example, the influence of the climate variability to the marine ecosystem would be different on depending on whether the Aleutian low is strong in winter or spring seasons, with the latter being the season of blooms of plankton. Furthermore, physical oceanographers tend to be interested in broadly distributed changes in ocean interior, while ecosystem researchers might be more interested in narrower coastal regions. If investigations for the coastal regions are important, we need to analyze the oceanic variations (i.e., upper layer temperature studied by many physical oceanographers) with the finer resolution than those used for the basin-scale analyses. For a successful design of this class of researches, exchange of information between physical oceanographers and ecosystem researchers is necessary.

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