Ten-yearly Polar Motion Connected with Precipitation Changes over the North American and Eurasian Continents

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Abstract

A hydrologic excitation of ten-yearly polar motion appearing in the combined five-daily polar motion data of International Polar Motion Service (IPMS) and SPACE94 during 1962–1995 is discussed. The discussion is based on time-dependent exponential decay models of land water storage supplied by precipitation, in which the National Oceanic and Atmospheric Administration (NOAA) monthly gridded precipitation anomaly data for $5^\circ \times 5^\circ$ is used. The decay parameter of the land water storage τ_d is determined when square of the difference between the observed ten-yearly polar motion and calculated one becomes minimum. Amplitude in the calculated polar motion can explain about 36 % of that in observed ten-yearly polar motion when τ_d takes 3.7 months. The behavior of the calculated polar motion is similar to the observed one though the phase of the former leads that of the latter by about one year. When we do not consider the small phase difference, about 52 % of amplitude in the observed polar motion can be explained when τ_d takes 5.1 months, suggesting a globally averaged decay time of the land water storage. The seesaw-like changes of precipitation that excite efficiently the ten-yearly polar motion are also confirmed between the North American continent and the western region of the Eurasian continent. These facts suggest a conceptual model of the feedback system on decadal hydrologic cycle centered in the North Atlantic Ocean, in which the system consists of atmospheric variations connected with the North Atlantic Oscillation, seesaw-like sea level changes, and precipitation changes in the North American and Eurasian continents.

1. Introduction

The understanding and identification of decadal variations in the climate system are important for the detection and prediction of global warming, and its effect. The reason is that the rate of the decadal variations is the same as, or more than, that of global warming and its effects. Thus, the decadal variations are obstacles to verifications of global warming, and its effects, especially in the use of short data from several years to decades.

The mechanism of the decadal variations, has not been well known yet in contrast with the recent rapid development of understanding of dominant inter-annual variations, such as El Niño/Southern Oscillation (ENSO) phenomena (*e.g.*, Philander, 1990). This is because the decadal variations are produced by complicated interactions among

the various elements composing the climate system (*e.g.*, Latif, 1998), in addition to shortage of data for modeling the decadal variations.

In polar motion, known as the change in the Earth's instantaneous rotational axis with respect to the geographical coordinate system, a ten-yearly variation (hereafter, a ten-yearly polar motion), has been found by Kikuchi and Naito (1995). The polar motion, up to the decadal time-scales, is known to be excited by redistribution of mass, and changes of relative angular momentum in the atmosphere and hydrosphere (*e.g.*, Munk and MacDonald, 1960; Lambeck, 1980). Thus, the decadal variations in the atmosphere and hydrosphere can be understood by identifying the excitation sources for the ten-yearly polar motion, as one of the indicators of global-scale motion of fluids distributed on the Earth's surface.

Global-scale motions, showing so-called atmospheric teleconnection patterns (*e.g.*, Walker and Bliss, 1932; Wallace and Gutzler, 1981; Wallace, 1996) can be one of the major excitation sources for the polar motion. For example, Chao and Zhou

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(1999) have shown that an inter-annual polar motion has a connection with the North Atlantic Oscillation (NAO) known as a dominant teleconnection pattern in the mid- and high- latitudes in the Northern Hemisphere. It is represented by a seesaw-like variation of atmospheric pressure in winter, between the Icelandic Low and Azores High (*e.g.*, Walker and Bliss, 1932; Hurrell, 1995). The Southern Oscillation (SO), is also a well known teleconnection pattern in the tropical Pacific (Walker and Bliss, 1932). It is closely connected with the length of day (LOD) change (see Dickey *et al.*, 1992, for example). The relationship between SO and the polar motion on the inter-annual time-scale is weaker than that between NAO and the polar motion (Chao and Zhou, 1999; Naito *et al.*, 1999).

According to Earth rotation dynamics, a mass redistribution pattern with a spherical harmonic mode $Y_{2,1}$ (*i.e.*, degree 2, and order 1 mode of spherical harmonics) excites polar motion efficiently. Contributions of the mass redistribution to the polar motion become maximum at latitude $\pm 45^\circ$, and 0 at the poles and the equator, respectively. Most of the atmospheric pressure contribution from oceanic areas, is considered to be cancelled out on decadal time-scales, because of inverted barometer (IB) response of the oceans to atmospheric pressure changes. Considering land and ocean distribution of the Earth, we can easily imagine that the atmospheric pressure changes connected with NAO dominating in mid- and high- latitudes of the Northern Hemisphere, can excite polar motion more efficiently than those connected with SO, dominating in the equatorial region.

A ten-yearly variation has also been found in the winter (December to March) NAO index defined as the difference of normalized pressures between Lisbon Portugal, and Stykkisholmur Iceland, in which the variation has been remarkable since 1950 (Hurrell, 1995). The spectrum analyses of the NAO index has also confirmed that spectral power on the period of 6–10 years, becomes stronger toward the end of the period during 1865 and 1994 (Hurrell and van Loon, 1997).

Lately, Iwabuchi *et al.* (1997) have found common behavior between the ten-yearly polar motion and the ten-yearly variation in the winter NAO index. However, calculated polar motion, based on atmospheric pressure data in the Northern Hemisphere, was found to be less than one fifth of the observed polar motion. This suggests that a variation of land water storage, supplied by precipitation whose pattern is connected with the NAO index, is a possible excitation source for the ten-yearly polar motion.

So far, various decadal climate changes, that are prominent around the North Atlantic Ocean, the North American continent, and Eurasian continent have been found from observational evidences such

as sea level height (Groger and Plag, 1993; Unal and Ghil, 1995; Iwabuchi *et al.*, 1997), sea surface temperature (SST) (Deser and Blackmon, 1993), inclusive (land, air, and sea surface) temperature (Mann and Park, 1994, 1996), precipitation (Bradley *et al.*, 1987; Diaz *et al.*, 1989), and snow cover (Walland and Simmonds, 1997). The most probable excitation source for the ten-yearly polar motion is considered to be a change of land water storage, supplied by precipitation because the oceans are in isostatic equilibrium on decadal time-scales. According to Bradley *et al.* (1987) and Diaz *et al.* (1989), precipitation changes over continents on inter-annual to inter-decadal time-scales can be seen globally.

This study has two purposes. First is to investigate hydrologic excitations of the ten-yearly polar motion by using precipitation data and a time-dependent decay model of the land water storage supplied by precipitation. Second is to discuss the mechanism of the decadal variation having a center of action in the North Atlantic Ocean from a viewpoint of the Earth rotation dynamics. Section 2 describes formulation of wobble dynamics, reservoir models, and data. Section 3 shows the results of the calculations. Section 4 discusses relationship between calculated ten-yearly excitation poles and precipitation pattern over land, and proposes a conceptual model of a decadal hydrologic cycle centered in the North Atlantic Ocean. In Section 5, concluding remarks are given.

2. Formulation, modeling, and data

2.1 Polar motion and excitation function

Polar motion or wobble, variation of the Earth's rotational axis with respect to the Earth's figure axis, which is described by the dimensionless small quantity for equatorial angular velocity in complex form $\tilde{m} = m_1 + im_2$, where m_1 and m_2 are components for 0°E and 90°E , respectively, is given by

$$\frac{i}{\sigma_0} \frac{d\tilde{m}}{dt} + \tilde{m} = \tilde{\chi}, \quad (1)$$

where $\tilde{\chi} (= \chi_1 + i\chi_2)$ is the excitation function for polar motion, $\sigma_0 = 2\pi\{F_c(1 + 1/2Q_c)\}$ is complex-valued frequency of the Earth's free oscillation, called Chandler wobble; F_c is its frequency (0.843 cpy), and $1/Q_c$ is its factor indicating the dissipation of energy near F_c (*e.g.*, Munk and MacDonald, 1960; Wilson, 1985).

In Eq. (1), $\tilde{\chi}$ calculated from $\tilde{m}$ determined by geodetic observation, is generally called geodetic excitation pole. In general, Eq. (1) can be simplified to $\tilde{m} \cong \tilde{\chi}$ at frequencies much lower than F_c of 0.843 cpy (*e.g.*, Eubanks, 1993). On the other hand, $\tilde{\chi}$ calculated from geophysical observation is generally called the geophysical excitation pole. The geophysical excitation pole, due to mass redistribution of land water, is given by

$$\tilde{\chi} = 1.00 \frac{-1}{C - A} \int \rho H r^2 \cos \phi \sin \phi \exp(i\lambda) dS, \quad (2)$$

where the factor 1.00 on the right-hand side is a transfer function indicating Earth's deformation (see Barnes *et al.*, 1983; Eubanks, 1993, for example), H is height of water level with density ρ on the Earth's surface C and A are the Earth's axial and equatorial principal moments of inertia, respectively, r is mean radius of the Earth, ϕ and λ are latitude and longitude, respectively, and $dS = r^2 \cos \phi d\phi d\lambda$.

2.2 Reservoir model of land water storage

The terrestrial water balance equation, with land water storage $W(t)$ is generally written as

$$\frac{dW}{dt} = P - E - R_r - R_g, \quad (3)$$

where P , E , R_r , and R_g are precipitation, evaporation, river run-off, and groundwater run-off, respectively. In the parameters on the right-hand side, E , R_r , and R_g can hardly be estimated accurately, because no reliable data of these parameters, particularly in E and R_g , are available for discussions on the decadal time-scale.

Thus, this study assumes that E , R_r , and R_g can be expressed by only one parameter κ , so that the temporal integral in Eq. (3) is given by

$$W(t) = \sum_{\tau=0}^{N_{month}} \{P(t - \tau) \cdot \kappa(\tau)\}, \quad (4)$$

where unit of t and τ is month, $W(t)$ is computed by going back from $\tau = 0$ to $\tau = N_{month}$ (number of month), and κ is the decay rate of land water storage including effects of E , R_r , and R_g . To simply express the decay pattern in the land water storage, this study employs an exponential decay model defined by

$$\kappa(\tau) = \exp(-\tau/\tau_d), \quad (5)$$

where τ_d shows the decay constant for the land water storage supplied by precipitation. Eqs. (4) and (5) suggest that the land water supplied by precipitation at an arbitrary month decreases its $1/e$ just after τ_d months, where water within each grid is assumed to run off into the ocean directly.

In the model computation, it is necessary to consider conservation of water mass in the whole hydrosphere. To conserve water mass between land and ocean, we distribute the integrated water mass anomaly over the land, to the whole ocean equally. The resulting oceanic excitation for polar motion, however, has been confirmed to be negligible due to the large cancellation of the excitation in the globally distributed ocean. In addition, excitation of polar motion, caused by redistribution of water vapor in the atmosphere, is also very small (*e.g.*, Hinnov and Wilson, 1987; Chao *et al.*, 1987). Iwabuchi *et al.*

(1997) considered water vapor excitation in the atmospheric pressure excitation of the ten-yearly polar motion.

At $N_{month} = 0$, Eqs. (4) and (5) are simplified to $W(t) = P(t)$, which means that precipitation within one month runs off into ocean at the end of the month. Hereafter, we call this the non-reservoir model, though monthly precipitation data itself reflects accumulation of precipitation during one month.

When $N_{month} > 0$, precipitation anomalies can be accumulated in each grid according to Eqs. (4) and (5). Hereafter, we call this the reservoir model. In this case, 120 months are adopted for the integration period N_{month} in Eq. (4). Errors originated in the round off effect of $N_{month} = 120$, are confirmed to be negligible till $\tau_d = 10.0$.

In the following analysis, all the calculated excitation poles are 60-month low-pass filtered, and detrended. The most optimal τ_d , to account for the ten-yearly polar motion, is chosen when the sum of the square of the difference between the observed and calculated excitation poles in χ_1 , and that in χ_2 , becomes minimum by varying τ_d every 0.1. Hereafter, we call this method minimum (O-C)². We compare the observed, and calculated, excitation poles based on mean amplitude of χ_1 and χ_2 in each excitation pole in the following discussions, where the amplitude of each component is determined by least-square fitting of ten-year sinusoids to each excitation pole from 1967 to 1988.

In the reservoir model, dependencies of τ_d on geographical conditions such as latitude and topography are not considered due to a shortage of data for evaluating the parameters. However, since precipitation over land close to the ocean, in general, runs off earlier than that in the inland area, we evaluate this effect based on two models whose grids depend on distances of less than 500 km and 1000 km from the nearest ocean, where 500 km is approximately equal to distance in latitude of the grid size of 5°. Hereafter, we call this the inland model.

2.3 Data

2.3.1 Polar motion data

This study employs a geodetic excitation function, calculated by Kikuchi and Naito (1995). The excitation function is based on two kinds of polar motion data. One is the International Polar Motion Service (IPMS) data for 1962–1975, and the other is the SPACE94 data for 1976–1995 (Gross, 1996), where the sampling interval of both data is five days. Note that the two data are based on different techniques. The former is based on optical observations, and the latter on space geodetic techniques. Although precision of the former is less than that of the latter (*e.g.*, Gross, 1992; McCarthy and Luzum, 1996), we regard the two data as identical, and continuous be-

Fig. 1. The ten-yearly motions of observed geodetic excitation pole (central curve, hereafter, *OP*) from the IPMS and SPACE94 data, and a calculated geophysical excitation pole (right upper curve) based on atmospheric pressure data set in the Northern Hemisphere. Both excitation poles have been 60-month low-pass filtered and detrended. Unit 0.2×10^{-7} of change in excitation pole corresponds to approximately 13 cm of the Earth's surface. The dotted arrow shows the principal axis determined by least-square fitting of 10-year sinusoids to *OP*. The direction of the oscillation in *OP* whose magnitude is about 40 cm is almost between 59°E and 126°W . After Iwabuchi *et al.* (1997).

cause they agree with each other in the overlapping period, with just slight discrepancies as mentioned by Kikuchi and Naito (1995). The earliest routine observations of polar motion under the International Latitude Service (ILS) are not used here, because of their low precision (*e.g.*, Eubanks, 1993; McCarthy and Luzum, 1996).

The 60-month low-pass filtered and detrended observed geodetic excitation pole (hereafter, referred to as *OP*) is shown in Fig. 1. The *OP* oscillates toward 59°E and 126°W . It is equivalent to an excitation pole due to a mass redistribution pattern with spherical harmonics $Y_{2,1}$ mode, having precipitation changes of 10 mm, whose center of action is the North American and the Eurasian continents. An excitation pole based on atmospheric pressure data, is also presented in the upper-right corner of Fig. 1, but its magnitude of oscillation is approximately one fifth of *OP*, and its direction of rotation is opposite to that of *OP* (Iwabuchi *et al.*, 1997).

2.3.2 Precipitation data

The monthly $5^\circ \times 5^\circ$ gridded precipitation anomaly data, compiled by the National Oceanic and Atmospheric Administration (NOAA) for 1950–1993, is used for calculating precipitation P in Eq. (3), where we assume the density of water to

be $1.0 \times 10^3 \text{ kg m}^{-3}$. The precipitation anomaly has been computed from the median for a period of 1951–1980 in each grid by assuming spatial homogeneity of precipitation within the grid. Figure 2 shows the distribution of the gridded data in this study, where land is divided into five regions; North America, South America, Eurasia, Africa, and Australia. A 90°E line in Fig. 2 indicates a boundary between the western and eastern regions of Eurasia defined in this study for discussions (hereafter, we call the two regions western Eurasia and eastern Eurasia, respectively). The blank grids in Fig. 2 are unused in the model computation, where continuous data from 1960 to 1985 are not available. There are precipitation data over approximately 450 grid areas during 1950–1988. From 1989 to 1991, however, the number decreases to 220.

When raw gridded precipitation data are used for the model computation in Eqs. (4) and (5), large variations of the contribution from one grid occurs for missing data because a bias of average precipitation in the grid, is not considered in the period. By contrast, when the precipitation anomaly data of NOAA is used, variation of the contribution in one grid to the excitation pole is small, even when data is missing, because average precipitation has

Fig. 2. Distribution map of the precipitation data for $5^\circ \times 5^\circ$ grid in continents. Blank grids are not used here because continuous data from 1960 to 1985 is not available. NA, SA, EU, AF, and AU denote North America, South America, Eurasia, Africa and Australia. 90°E line indicates a boundary between the western region and eastern region of Eurasia, EU(W) and EU(E), respectively, in this study.

already been removed as the median in each grid. Thus we choose to use the anomaly data.

Comparisons of the calculated geophysical excitation pole with the observed geodetic excitation pole begin at 1962. To avoid systematic biases, 11 years of the calculated geophysical excitation poles before 1962 are considered as the spin-up period. Hereafter, 60-month low-pass filtered, and detrended calculated geophysical excitation poles, are referred to as CP.

3. Calculated excitation functions from precipitation data

3.1 Non-reservoir model

As stated in the previous section, the non-reservoir model when $N_{\text{month}} = 0$ in Eqs. (4) and (5) means no accumulation of precipitation anomalies. Figure 3a shows ten-yearly variations in each component of OP (observed geodetic excitation pole), and Fig. 3b shows those of CP_0 (calculated geophysical excitation pole when $N_{\text{month}} = 0$). In Fig. 3b, GLOBAL means global contribution, which consists of the excitation pole calculated in each continent shown in the lower panel of Fig. 3b, and the oceanic contribution due to conservation of water mass.

OP and CP_0 show similar variations with each other. The peaks and troughs of the two excitation poles occur roughly in phase. The small discrepancies appear in the phases of CP_0 , which leads that of OP by approximately 10 months in χ_1 , and by approximately 14 months in χ_2 . The magnitude of variation in CP_0 , is fairly small comparing to that in OP. These suggest that the accumulations of the precipitation anomalies are essential to explain the observed excitation pole.

In Fig. 3b, variations in North America and in western Eurasia are similar to that in OP, especially in χ_2 . On the other hand, variations in other continents also show large magnitudes, comparing to those in North America and Eurasia. For example, the contribution of Australia is remarkably large around 1974, which corresponds to the unusual great volume of coastal rainfall in Australia shown in Diaz *et al.* (1989). Around 1974, the largest difference of behavior between CP_0 and OP appears.

3.2 Reservoir model

The reservoir model reflects effects of the accumulation of precipitation anomalies that depend on the decay constant τ_d in Eq. (5). The solid lines in Figs. 4a and 4b show OP, and a calculated excitation pole at $\tau_d = 3.7$ (hereafter, $CP_{3.7}$), respectively, where $CP_{3.7}$ is determined by the minimum $(O-C)^2$, when the phase difference of about 1 year between the observed, and calculated excitation poles is considered. The dashed lines show periodic motion of excitation poles obtained by least-square fitting of 10-year sinusoids to the excitation poles. The amplitude in $CP_{3.7}$ can explain 36 % of that in OP. $\tau_d = 3.7$ (months) means the globally averaged relaxation time of land water storage supplied by precipitation (time needed for monthly precipitation anomalies to decrease its $1/e$) is approximately 3.7 months under the reservoir model. Though $CP_{3.7}$ in Fig. 4b, and CP_0 in Fig. 3b, also show similar behavior with each other, the amplitude of $CP_{3.7}$ is three to four times larger than that of CP_0 . Behavior of excitation pole, calculated at each continent as $\tau_d = 3.7$, also agree well with those of CP_0 though it is not shown here.

Fig. 3. (a) Ten-yearly variations in χ_1 (0°E) and χ_2 (90°E) components of OP . (b) Same as (a), but for a calculated excitation pole based on “non-reservoir model” at $N_{\text{month}} = 0$ in Eqs. (4) and (5) (hereafter, CP_0), where *GLOBAL* shown in the upper panel indicates global contribution to the excitation pole and abbreviations (see Fig. 2) shown in the lower panel indicate each continental contribution. Note that scale in (b) is one fifth of that in (a) because of small contribution to the excitation pole in the non-reservoir model.

$NA+EU$ in Fig. 4b shows a calculated excitation pole at $\tau_d = 4.6$ in the minimum $(O-C)^2$, when only two major continental contributions of North America and Eurasia are considered (hereafter, $CP_{4.6}(NA+EU)$). The amplitude in

$CP_{4.6}(NA+EU)$ can explain 37 % of that in OP . The behavior is similar to that of $CP_{3.7}$, especially in χ_2 , in spite of considering only two continental contributions to the excitation pole. This suggests that OP seems to reflect variations mainly of the land

Fig. 4. (a) Ten-yearly variations in χ_1 and χ_2 components of *OP* (solid lines) and periodic motions obtained by least-square fitting of 10-year sinusoids to the ten-yearly variations from 1967 to 1988 (dashed lines). (b) Same as (a), but for calculated excitation poles based on "reservoir model" at decay parameter of land water storage supplied by precipitation $\tau_d = 3.7$ (hereafter, $CP_{3.7}$), and that at $\tau_d = 4.6$ when only two continental contributions of *NA* and *EU* are considered (hereafter, $CP_{4.6}(NA+EU)$). (c) Same as (b), but for at $\tau_d = 5.9$ and 6.7 , respectively, based on "inland model," where land area whose distance from the ocean is less than 500 km is not considered. These τ_d s are determined by the evaluation that the square of difference between observed and calculated excitation poles is minimum when the phase difference between the excitation poles is considered.

water storage in North America and Eurasia, due to their large capacity to reserve land water. The increase in τ_d of 0.9 months in $CP_{4.6}(NA+EU)$, as compared with that in $CP_{3.7}$ can be seen. In principle, it reflects asymmetric distribution patterns of precipitation changes between the Northern and Southern Hemisphere, because polar motion is excited by asymmetric mass redistribution pattern between both hemispheres.

Fig. 4c shows calculated excitation poles at minimum $(O-C)^2$ based on the inland model, where land area whose distance from the ocean is less than 500 km, is not considered in the model computation.

The minimum $(O-C)^2$ solution of another inland model, in which land area whose distance from the ocean is less than 1000 km is not considered, shows no convergence. The two τ_d s, based on global contribution and only two continental contributions of North America and Eurasia, are about two months larger than those in $CP_{3.7}$ and $CP_{4.6}(NA+EU)$ in Fig. 4b. The behavior in the excitation poles, based on the inland model, are similar to those in $CP_{3.7}$ and $CP_{4.6}(NA+EU)$ on decadal time-scales. On the other hand, intra-decadal variations appearing in Fig. 4b disappear in Fig. 4c. This simply suggests that effects on intra-decadal variations are

prominent in coastal area.

4. Discussion

4.1 Ten-yearly sinusoids

We have confirmed from spectral analysis that *OP* (observed geodetic excitation pole) in Fig. 4a shows maximum spectral power at the period of 10.2 years, and the calculated geophysical excitation poles of $CP_{3.7}$ and $CP_{4.6}(NA+EU)$ in Fig. 4b, show that at 10.4 and 9.8 years, respectively. Figure 5a shows two-dimensional periodic motion of the excitation poles determined by least-square fitting of 10-year sinusoids to χ_1 and χ_2 in *OP*, $CP_{3.7}$, and $CP_{4.6}(NA+EU)$ (see dashed lines of Figs. 4a and 4b). All the three excitation poles rotate prograde, and the direction of major axis in ellipsoid of *OP* pointing to North America, and Eurasia agrees nearly with those of the two calculated excitation poles.

On the other hand, in Fig. 5a, phases of the two calculated excitation poles ($CP_{3.7}$ and $CP_{4.6}(NA+EU)$) lead that of *OP* by about 1 year as mentioned before. We have confirmed that the phase differences do not decrease significantly, even if τ_d increases. For example, the phase differences become only 1–2 months smaller, even if τ_d becomes 20.0. It indicates the limits in the reservoir model used here in expressing the land water storage.

Note here that the phase lead of about 1 year in the calculated excitation poles to the observed one is small enough to discuss the ten-yearly polar motion, and the calculated and observed excitation poles show similar variations with each other as mentioned before. These suggest that some sophisticated models to delay run-off time of the land water storage supplied by precipitation, should be applied in the reservoir model to account for the phase differences, in which the movement of surface water including groundwater across grids is considered. However, we cannot evaluate such an effect precisely because of no reliable and comprehensive data. Satellite gravity measurements in the future will provide us with precise information for this kind of estimation. In addition, the evaluations, using the climate model coupled with the terrestrial hydrologic system, are considered as an alternative approach to evaluate mass redistribution of land water comprehensively.

The water masses restricted in the same place for several months to several years such as snow, ice, permafrost and hydrological reservoir, are considered to be plausible factors to explain the phase difference. In the above elements, a freezing effect of precipitation in the winter season, especially in North America and Eurasia, is one of the most plausible effects to preserve the land water for several months. Walland and Simmonds (1997) have shown decadal variations of snow covered area in

North America and Eurasia by using snow distribution data based on NOAA's National Environmental Satellite, Data, and Information Service (NESDIS). In their Fig. 3, we can see a decadal seesaw-like variation of snow covered area between the two continents. This suggests that redistribution of the snow mass may excite the ten-yearly polar motion.

It is difficult to estimate the snow mass accurately based on satellite image of the snow covered area. Then, we try to evaluate snow mass excitation for polar motion based on a simple freezing model of precipitation used in Hinnov and Wison (1987). Precipitation in one grid accumulates if surface air temperature in the grid is less than 0 degree Celsius, and runs off into the ocean if the surface temperature is more than 0 degree Celsius in the model. We made $5^\circ \times 5^\circ$ gridded data for precipitation and surface air temperature by averaging the station data of NOAA in each grid. The number of stations in the precipitation data decrease rapidly in 1989, and thus the possible period for discussion is limited. The magnitude of the calculated excitation pole is approximately one fifth of *OP* on decadal time-scales, and the calculated excitation pole is not so similar to *OP*, compared with the calculated excitation poles based on the reservoir model. Although we should consider that the resolution for the freezing model is too coarse to precisely evaluate the freezing effect of precipitation, the preliminary results suggest that snow is a minor contributor to polar motion on decadal time-scales.

In Figs. 4b and 5a, the amplitude in calculated excitation poles ($CP_{3.7}$ and $CP_{4.6}(NA+EU)$) can explain only about one third of that in the observed excitation pole. We can imagine that the phase difference of about 1 year between the observed and calculated excitation poles, causes the underestimation of the τ_d based on the minimum (O-C)² evaluation. Figure 5b shows periodic motion of excitation poles when the phase difference is not considered. The amplitude in the calculated excitation pole considering global contributions can explain about 52 % of that in the observed one at τ_d of 5.1 months. Thus 5.1 months is considered to reflect a globally averaged decay time of the land water storage, supplied by precipitation more accurately than τ_d of 3.7 months in Figs. 4b and 5a when the phase difference between the observed and calculated excitation poles is considered. The amplitude in the calculated excitation pole, when only two continental contributions of North America and Eurasia are considered, can explain about 68 % of that in the observed one at τ_d of 8.0 months. The shape of the ellipsoid, in the calculated excitation pole is similar to that in observed one. Those results suggest that land water storage, supplied by precipitation—especially in North America and Eurasia—is a major excitation source for the ten-yearly polar motion.

Fig. 5. (a) Rotational component of 10 years in OP (dotted line), $CP_{3.7}$ (thick solid line), and $CP_{4.6}(NA + EU)$ (thin solid line) based on the periodic motions in Figs. 4a and 4b, and that in calculated excitation pole based on atmospheric pressure data (small line in the center of the figure, see Fig. 1). The phase at every year and direction of the motion are shown with circles and arrows, respectively, where solid circles indicate 1967. (b) Same as (a), but at $\tau_d = 5.1$ of global contribution and at $\tau_d = 8.0$ when only two continental contributions of NA and EU to the excitation pole are considered. The τ_d s in (b) are determined by the evaluation that square of differences between observed and calculated excitation poles is minimum when the phase difference between the excitation poles are not considered.

The short line in the center of Fig. 5a shows a rotational component at a ten-year period in the calculated excitation pole based on atmospheric pressure (also see Fig. 1). The contribution to the observed excitation pole is very small in contrast to the atmospheric pressure contribution to the annual wobble, though there are additional minor excitation sources of wind and land water storage to the annual wobble (*e.g.*, Kuehne and Wilson, 1991). Kuehne and Wilson (1991) compared the power spectrums of polar motion, excited by variations of the land water storage, with that excited by atmospheric pressure changes, and suggested that the variations of water storage become a dominant cause for the excitation in the period longer than several years, suggesting that our results are reasonable.

To detect what kinds of redistribution patterns of land water cause prograde rotation of the excitation poles shown in Fig. 5a, we show in Fig. 6 weighted amplitude $WA_p = A_p \sin 2\phi \cos \phi$, and phase ϕ in the ten-yearly precipitation changes obtained by least-square fitting of 10-year sinusoids in the precipitation changes from 1967 to 1988 in each grid, where A_p is the amplitude of ten-yearly precipitation changes. The epoch for the fitting is January, 1967, and $\sin 2\phi$ and $\cos \phi$ express weights resulting from dependency of latitude on polar motion excitation (*i.e.* see Eq. (2)), and the decrease

of area of a grid from equator to poles, respectively. The magnitude of the periodic precipitation changes at m (month) from the epoch is expressed by $A_p \sin(\omega m + \phi)$, where ω is $2\pi/120$ month $^{-1}$. The amplitude and phase in neighboring grids generally show similar pattern with each other. Relatively large amplitudes are distributed not only in the Northern Hemisphere, but also in the Southern Hemisphere.

The thick arrows above the map in Fig. 6, show partial contributions to the excitation pole, expressed from a partial sum of the vectors within $30^\circ \times 30^\circ$ grids between $30^\circ N$ and $60^\circ N$, shown by squares in the map of Fig. 6. The inverse phase pattern between North America, and western Eurasia is clearly seen, suggesting that the ten-yearly polar motion is excited effectively by the mass redistribution pattern of nearly $Y_{2,1}$ mode of spherical harmonics. Chao and O'Connor (1988) have shown that seasonal changes of snow and rainfall, both in the North American and Eurasian continents are of comparable size, so that a large cancellation between their contributions to polar motion is achieved on a seasonal time-scale. On the contrary, Fig. 6 shows that the two continental contributions strengthen the excitation pole, with each other on decadal time-scales.

Moreover, the partial contributions in the upper

Fig. 6. Map showing weighted amplitude WA_p (lengths of arrows) and phase ϕ in ten-yearly precipitation changes determined by least-square fitting of 10-year sinusoids to gridded precipitation data from 1967 to 1988, where the epoch for the fitting is January, 1967. The magnitude of the periodic precipitation changes at m (month) from the epoch is expressed by $A_p \sin(\omega m + \phi)$, where A_p is the amplitude of ten-yearly precipitation changes and ω is $2\pi/120 \text{ month}^{-1}$, respectively. The weighted amplitude is defined as $WA_p = A_p \sin 2\phi \cos \phi$, where $\sin 2\phi$ shows dependencies of latitude on polar motion excitation (*i.e.*, see Eq. (2)) and $\cos \phi$ expresses the decrease of area of a grid from equator to poles, showing dependency of latitude on polar motion excitation due to the grid size definition, respectively. The thick arrows above the map show partial contributions for the excitation pole expressed by the sum of the vectors within $30^\circ \times 30^\circ$ grids between 30°N and 60°N shown by squares in the map.

panel of Fig. 6 show a clockwise rotation of phase from east Eurasia to North America. This fact suggests that water-mass anomalies move eastward on decadal time-scales in mid-latitude where the westerly wind is prominent. The eastward movement of the water-mass anomalies, is considered to induce prograde rotation of the excitation poles show in Fig. 5a on decadal time-scales.

4.2 Hydrologic cycle connected with NAO

We have shown that the calculated geophysical excitation poles, based on the time-dependent exponential decay models of the land water storage supplied by precipitation, change similarly with the observed geodetic excitation pole on decadal time-scales. Also, Iwabuchi *et al.* (1997) have shown that the NAO index, and sea level height along both coastal regions in the North Atlantic Ocean, change similarly with the ten-yearly polar motions. These facts suggest a conceptual model of the feedback system on the decadal hydrologic cycle centered in the North Atlantic Ocean as shown in Fig. 7. The mechanism in the feedback system of the model, is described in the following.

Most of precipitation in North America and Eurasia runs out either into the North Atlantic Ocean, or into the Arctic Ocean due to mountainous regions in the western part of the North America, and in south-

ern and eastern parts of the Eurasia. Thus, precipitation changes over both continents cause freshwater anomalies, and salinity changes in the North Atlantic Ocean. We can imagine that the coastal sea level changes in the North Atlantic Ocean reflect the freshwater anomalies as one of the mechanisms to link precipitation and sea level changes in Fig. 7. To confirm the assumption is necessary by further researches.

The buoyant force fluctuations related to the salinity changes in the sub-surface ocean cause sea surface temperature (SST) anomalies induced by vertical mixing in the upper ocean in winter. In fact, decadal SST variations have been confirmed to be dominant in the North Atlantic Ocean (*e.g.*, Deser and Blackmon, 1993). Moreover, Houghton (1996) has considered that the decadal fluctuation in the North Atlantic Ocean is related to surface forcing and upper-ocean mixing, and that the local precipitation, river runoff, arctic ice, and freshwater outflow are potential sources for freshwater anomalies in the sub-surface ocean.

Thermal energy fluctuations in the upper ocean related to the SST anomalies, propagate to the atmosphere quickly through the transportation of latent heat, due to phase change of water. The decadal change of the NAO index reflects the resulting at-

Fig. 7. A conceptual model showing a decadal hydrologic cycle of the feedback system centered in the North Atlantic Ocean, in which water plays important role in the system. Thick arrows and dotted lines indicate the direction of action and existence of the correlation, respectively. The observed ten-yearly polar motion can be excited by redistribution of the land water storage supplied by precipitation in the North American continent and western region of the Eurasian continent.

mospheric condition. The winter sea ice anomaly in the Davis Strait/Labrador Sea region pointed out by Deser and Blackmon (1993), for example, may reflect the decadal SST changes, and SST-induced atmospheric changes. It is also one of the important sources for the freshwater anomaly in the sub-surface ocean.

Changes of moisture transport patterns, related to those of the NAO index, cause changes in precipitation patterns. The precipitation changes then trigger new cycles of the feedback system. Here, an eastward movement of precipitation anomalies in North America and Eurasia (*i.e.*, Fig. 6) excites the ten-yearly polar motion, and plays an important role in determining the time-scale in the feedback system. The changes of precipitation pattern may also affect the decadal variation in the NAO teleconnection pattern, caused by exchange of thermal energy through phase changes of water between the atmosphere and the land surface.

5. Summary

We have discussed hydrologic excitation of the ten-yearly polar motion by using the reservoir model with the exponential decay of land water storage supplied by precipitation and NOAA's $5^\circ \times 5^\circ$ gridded precipitation anomaly data. When the decay parameter of land water storage supplied by precipitation, τ_d , is equal to 3.7 months, the sum of squares of the difference between observed and calculated excitation poles becomes minimum, and the amplitude in the obtained calculated excitation pole can explain 36 % of the observed. Although the amplitude of the calculated excitation pole is not enough

to fully explain the observed excitation pole, peaks and troughs in both excitation poles show similar behavior with each other, with small phase lead of about one year in the calculated respect to the observed one. When the phase difference is not considered, the amplitude in calculated excitation pole can explain 52 % of that in the observed excitation pole at τ_d of 5.1 months, suggesting a globally averaged decay time of the land water storage supplied by precipitation, and the amplitude in calculated excitation pole with considering contributions of only North America and Eurasian can explain 68 % of that in the observed one at $\tau_d = 8.0$ months. These facts indicate that the major excitation source for the ten-yearly polar motion can be land water storage supplied by precipitation, especially in the North American and Eurasian continents.

We also proposed a conceptual model of a feedback system in the decadal hydrologic cycle, centered in the the North Atlantic Ocean, in which the system consists of decadal variations in the NAO index, coastal sea level changes, and precipitation changes in the North American and Eurasian continents. An eastward movement of precipitation anomalies in the North American and Eurasian continents play an important role in determining the time-scale in the feedback system. However, we need to wait for more comprehensive hydrologic data to verify the above model.

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References

- Barnes, R.T.H., R. Hide, A.A. White and C.A. Wilson, 1983: Atmospheric Angular Momentum Fluctuation, Length-of-Day Changes and Polar Motion. *Proc. E. Soc. Lond.*, **A387**, 31–73.
- Bradley, R.S., H.F. Diaz, J.K. Eischeid, P.D. Jones, P.M. Kelly and C.M. Goodess, 1987: Precipitation Fluctuations over Northern Hemisphere Land Areas Since the Mid-19th Century. *Science*, **237**, 171–175.
- Chao, B.F. and W.P. O'Connor, A.T.C. Chang, D.K. Hall and J.L. Foster, 1987: Snow Load Effect on the Earth's Rotation and Gravitational Field, 1979–1985. *J. Geophys. Res.*, **92**, 9415–9422.
- Chao, B.F. and W.P. O'Connor, 1988: Global Surface-water-induced Seasonal Variations in the Earth's Rotation and Gravitational Field. *Geophys. J.*, **94**, 263–270.
- Chao, B.F. and Y.-H. Zhou, 1999: Meteorological Excitation of Interannual Polar Motion by the North Atlantic Oscillation, *Geodynamics*, **27** 61–73.
- Deser, C. and M.L. Blackmon, 1993: Surface Climate Variations over the North Atlantic Ocean during Winter: 1900–1989. *J. Climate*, **6**, 1743–1753.
- Diaz, H. F., R. S. Bradley and J. K. Eischeid, 1989: Precipitation Fluctuations Over Global Land Areas Since the Late 1800's. *J. Geophys. Res.*, **94**, 1195–1210.
- Dickey, J.O., S.L. Marcus and R. Hide, 1992: Global Propagation of Interannual Fluctuations in Atmospheric Angular Momentum. *Nature*, **357**, 484–488.
- Eubanks, T.M., 1993: Variations in the Orientation of the Earth, in Contributions of Space Geodesy to Geodynamics: Earth Dynamics, Geodynamics Series, 24, pp. 1–54, (Eds). Smith, D.E. and Turcotte, D.L., American Geophysical Union, Washington, DC.
- Gröger, M. and H.-P. Plag, 1993: Estimations of a Global Sea Level Trend: Limitations from the Structure of the PSMSL Global Sea Level Data Set. *Global and Planetary Change*, **8**, 161–179.
- Gross, R.S., 1992: A Combination of Earth Orientation Data: SPACE91, *IERS Technical Note 11*, Observatoire de Paris, 113, 1992
- Gross, R.S., 1996: Combinations of Earth Orientation Measurements: SPACE94, COMB94 and POLE94. *J. Geophys. Res.*, **101**, 8729–8740.
- Hinnov, L.A. and C.R. Wilson, 1987: An Estimate of the Water Storage Contribution to the Excitation of Polar Motion. *Geophys. J. R. astr. Soc.*, **88**, 437–459.
- Houghton, R.W., 1996: Subsurface Quasi-Decadal Fluctuations in the North Atlantic. *J. Climate*, **9**, 1363–1373.
- Hurrell, J.W., 1995: Decadal Trends in the North Atlantic Oscillation: Regional Temperatures and Precipitation. *Science*, **269**, 676–679.
- Hurrell, J.W. and H. van Loon, 1997: Decadal Variations in Climate Associated with the North Atlantic Oscillation. *Climate Change*, **36**, 301–326.
- Iwabuchi, T., K. Heki and I. Naito, 1997: Decadal Polar Motion Connected with Atmospheric Pressure and Sea Level Patterns over the North Atlantic Ocean, International Association of Geodesy Symposia 117. Segawa *et al.* (Eds.) Gravity, Geoid and Marine Geodesy, Springer-Verlag, 666–673.
- Kikuchi, N. and I. Naito, 1995: Notes on Behaviors of the Polar Motion Based on ILS, IPMS and SPACE94. *J. Geodetic Society of Japan*, **41**, 239–249.
- Kuehne, J. and C.R. Wilson, 1991: Terrestrial Water Storage and Polar Motion. *J. Geophys. Res.*, **96**, 4337–4345.
- Lambeck, K., 1980: The Earth's Variable Rotation: Geophysical Causes and Consequences, Cambridge University Press, Cambridge, 449pp.
- Latif, M., 1998: Dynamics of Interdecadal Variability in Coupled Ocean-Atmosphere Models. *J. Climate*, **11**, 602–624.
- Mann, M.E. and J. Park, 1994: Global-Scale Modes of Surface Temperature Variability on Interannual to Century Timescale. *J. Geophys. Res.*, **99**, 25, 819–833.
- Mann, M.E. and J. Park, 1996: Joint Spatiotemporal Modes of Surface Temperature and Sea Level Pressure Variability in the Northern Hemisphere during the Last Century. *J. Climate*, **9**, 2137–2162.
- McCarthy, D.D. and B.J. Luzum, 1996: Path of the Mean Rotational Pole from 1899 to 1994. *Geophys. J. Int.*, **125**, 623–629.
- Munk, W.H. and G.J.F. MacDonald, 1960: *The Rotation of the Earth*. Cambridge University Press, Cambridge, 323pp.
- Naito, I., Y.-H. Zhou, M. Sugi, R. Kawamura and N. Sato, 1999, to be submitted to *J. Meteor. Soc. Japan*, 1999.
- Philander, S.G., 1990: *El Niño, La Niña, and the Southern Oscillation*. Academic Press, New York, 293pp.
- Unal, Y.S. and M. Ghil, 1995: Interannual and Interdecadal Oscillation Patterns in Sea Level. *Clim. Dyn.*, **11**, 255–278.
- Walker, G.T. and E.W. Bliss, 1932: World Weather V., *Mem. Roy. Meteor. Soc.*, **4**, 53–84.
- Wallace, J.M. and D.S. Gutzler, 1981: Teleconnections in the Geopotential Height Field during the Northern Hemisphere Winter. *Mon. Wea. Rev.*, **109**, 784–812.
- Wallace, J.M., 1996: Observed Climatic Variability: Spatial Structure, Decadal Climate Variability: Dynamics and Predictability, 44, Anderson, D.L.T. and J. Willebrand (Eds.), Global Environmental Change, NATO ASI Series, Springer, 31–81.
- Walland, D.J. and I. Simmonds, 1997: North American and Eurasian Snow Cover Co-Variability. *Tellus*, **49A**, 503–512.
- Wilson, C.R., 1985: Discrete Polar Motion Equations. *Geophys. J. R. astr. Soc.*, **80**, 551–554.

北アメリカ大陸とユーラシア大陸における 降水量変動と関連する極運動の10年変動

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5日サンプリングで得られる IPMS (International Polar Motion Service) と SPACE94 を結合した 1962–1995 年の極運動データに見られる 10 年変動に対して、極運動の陸水変動による励起が議論された。議論は、降水により供給される陸水貯留量の指数関数的減衰を仮定したモデルを、NOAA (National Oceanic and Atmospheric Administration) の $5^\circ \times 5^\circ$ 分解能の月降水量偏差データに適用し、観測および計算に基づく極運動の差の二乗が最小となるときの降水により供給される陸水貯留量のグローバルな減衰パラメータ τ_d を決定することによりなされた。計算に基づく極運動の振幅は、 τ_d が 3.7 ヶ月のとき、極運動の 10 年変動の振幅を約 36 % を説明可能であった。このとき、計算に基づく極運動の位相は、観測に基づくそれより 1 年先行するものの、両者の変動は類似した振る舞いを示した。この位相差を考慮しなければ、 τ_d が 5.1 ヶ月のとき、極運動の 10 年変動の振幅の約 52 % が説明可能となる。この減衰時間は、降水により供給される陸水貯留量のグローバルな減衰時間を示唆する。北アメリカ大陸とユーラシア大陸西部の間には、極運動の 10 年変動を効率的に励起させる降水量のシーソーのような変動も確認された。これらの事実に基づき、極運動の立場から、北大西洋を作用点にもつ水循環の 10 年変動に関するフィードバックシステムの概念モデルが提案される。このモデルは、北大西洋振動と関係する大気変動、北大西洋における東西のシーソーのような海面変動、そして北米大陸とユーラシア大陸の降水量変動からなる。