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Temporal variations in the low-degree zonal harmonics of the Earth's gravitational field have recently been observed by satellite laser ranging. A host of geophysical processes contribute to these variations. The present paper studies quantitatively a prime contributor, atmospheric mass redistribution, using global surface pressure data from the European Centre for Medium Range Weather Forecasts for the period of 1980–1988. Seasonal variations are focused upon: We compute the annual and semiannual amplitudes and phases of the zonal $J_l$ coefficient with degree $l = 2–6$ with and without the oceanic inverted-barometer (IB) effect and obtain the predicted effects on the orbit nodal residuals $\Delta\Omega$ of Lageos and Starlette. These predicted $\Delta\Omega$ are then compared with the observed $\Delta\Omega$. It is found that the atmospheric influence, combined with the hydrological influence (from Chao and O'Connor, 1988), agree well with the Lageos observation for the annual term. The corresponding match appears poorer for Starlette. The comparisons for the semiannual term, however, show little agreement. The discrepancies can be attributed primarily to the uncertainty arising from the IB assumption, the systematic errors in the hydrological estimates and in the observations, as well as other geophysical processes that influence global mass distribution, particularly those in the oceans. Interannual and decadal variations, with their impact on the observations of "secular" changes are also discussed.

1. Introduction

Mass redistributions on or within the Earth will manifest themselves as temporal variations in the Earth's external gravitational field. Advances in the modern geodetic technique of satellite laser ranging have made it possible to detect these variations, particularly those in the low-degree zonal gravitational coefficients $J_l$ ($l$ being the degree) [Shum et al., 1987, 1988; Cheng et al., 1989, 1990]. Conversely, therefore, understanding the dynamics of the temporal gravitational variations becomes essential in orbit determination of geodetic satellites and ultimately in precise modeling of the Earth's gravitational field.

A host of geophysical processes of global mass redistribution are responsible for the gravitational variations. For example, the observed secular change in low-degree $\Delta J_l$ is believed to be a result of the postglacial rebound in the mantle [Yoder et al., 1983; Alexander, 1983; Rubincam, 1984], although mountain glaciers and polar ice sheets may also contribute substantially (Meier [1984] and Zwally [1989], but see also Douglas et al. [1990]). This paper mainly concerns the seasonal gravitational variation from atmospheric sources, although short-period as well as interannual variations will also be presented.

Seasonal gravitational variations arise from solar influences on the Earth's mass distribution; the primary ones are (1) seasonal redistribution of atmospheric mass; (2) seasonal redistribution of hydrospheric mass; and (3) solar $Sa$ and $Ssa$ tides raised in the solid Earth and oceans. This paper focuses on variation 1 and, to some extent, represents a revisit of the problem treated previously by Gutierrez and Wilson [1987], as will be discussed in section 4.

First, let us put the above mentioned three contributions in a qualitative, geophysical perspective. It is a fortuitous fact that, although having no physical connections, they are comparable in magnitude as far as the low-degree seasonal $\Delta J_l$'s are concerned. The latter due to variations 1 and 2 are of the order of $a\Delta m/M$, where $M$ is the mass of the Earth, $\Delta m$ is the seasonal amplitude of the total displaced mass from the mean state, and $a$ is a geographic factor of magnitude typically of a fraction of 1 (see equation (2) below). For variation 1 the annual $\Delta m$ is roughly 0.1% of the total atmospheric mass [Trenberth, 1981], or $\sim 5 \times 10^{15}$ kg. For variation 2 the contribution of the ocean is very poorly constrained: steric changes which constitute a large, but unknown, portion of the observed sea level variations have virtually no gravitational consequence. Seasonal wind-driven currents can be modeled from surface wind observations (J. Alexander, personal communication, 1990). However, the primary hydrospheric contribution probably comes from the continental hydrological constituents of annual snow and rain (minus evapotranspiration and runoff). They combine to give a $\Delta m$ of $\sim 3 \times 10^{15}$ kg for water, corresponding to an annual sea level variation of as large as 1 cm [e.g., Chao et al., 1987; Chao and O'Connor, 1988]. Thus, with $M = 6 \times 10^{24}$ kg, one expect low-degree annual $\Delta J_l$ to be of the order of $10^{-10}$ for both variations 1 and 2. The corresponding semiannual terms are somewhat smaller (cf. Chao and O'Connor, 1988).

For variation 3 the solar tides give rise to a $\Delta J_2$ in the Earth that can be evaluated from the solar tidal perturbation in the Earth's rotational rate [e.g., Lambeck, 1988], assuming a zero change in the trace of the Earth's inertia tensor [e.g., Chao and Gross, 1987, equation 43]. The $\Delta J_2$ ampli-
tude in the solid Earth due to the Sa (solar annual) tide is thus $1.5 \times 10^{-8}$, comparable to variations 1 and 2 above. The $\Delta J_2$ amplitude generated by the Ssa (solar semiannual) tide in the solid Earth is similarly found to be about 6 times larger. We note in passing that these quantities can be alternatively obtained from first principles [e.g., Melchior, 1983]: It can be shown that $\Delta J_2 = 1.5 k_2^2 Q$ due to the Sa tide and $\Delta J_2 \approx 0.75 k_2^2 Q$ due to the Ssa tide in the solid Earth, where $k_2 = 0.30$ is the Love number of the Earth, $e = 0.0167$ and $\delta = 0.41$ rad are the eccentricity and obliquity of the Earth’s orbit, respectively, and the factor $Q = (\text{Sun mass/Earth mass}) (\text{Earth radius/Sun-Earth distance})^2 = 2.1 \times 10^{-8}$. The $\Delta J_2$ due to the ocean tide is approximately an order of magnitude smaller than its solid Earth counterpart [e.g., Lambeck, 1988]. The ocean tidal $\Delta J_3$ is not much smaller than the ocean tidal $\Delta J_2$ given the ocean-continent geography [e.g., Schwiderski, 1980], while the solid Earth tidal $\Delta J_1$ for $l > 2$ are totally negligible.

Thus a first-order geophysical problem is to identify various contributions in the observations, thereby providing a geophysical “budget” for $\Delta J_l$. This would be possible in principle when equipped with our results for variation 1 together with previous results by, e.g., Chao and O’Connor [1988] for variation 2. In reality, however, it will not be feasible until the resolution of some key ambiguities stemming largely from our insufficient knowledge about the hydroscopic behavior. On the other hand, the result of the present study can shed light on the latter. We shall furthermore address the possible contamination in the observations of “secular” $\Delta J_l$ changes (mentioned above) from interannual and decadal meteorological fluctuations. All these will be discussed in section 4.

It should be stressed at the outset that the atmospheric fluctuation has been linked unequivocally to variations in the Earth’s rotation: It is found to be responsible for almost all of the length-of-day variation on a wide range of time scales [e.g., Rosen and Salstein, 1983; Chao, 1989; Rosen et al., 1990]. It can also explain some of the “rapid” polar motion [Barnes et al., 1983; Eubanks et al., 1988], contributing to the excitation of the Chandler wobble. More relevant to our present study is the fact that the atmospheric mass redistribution is a primary excitation source for the annual wobble in the polar motion [e.g., Munk and MacDonald, 1960; Wilson and Haubrich, 1976; Wahr, 1983]. However, important discrepancies exist between the atmospheric and the observed annual wobble excitation [Chao and Au, this issue], the source of which has yet to be identified. Contrary to the case for $\Delta J_1$, the hydrological contribution to the annual wobble is rather small owing to a double cancellation effect (that between the eastern and western hemispheric components and that between the snow and rain components), as demonstrated by Chao and O’Connor [1988]. The long-period tidal perturbations on the polar motion are also negligible [e.g., Lambeck, 1988]. These are fundamental differences between the zonal harmonics under study here and the tesseral harmonics of degree 2 and order 1, which is responsible for the polar motion. So no analogy exists between these two cases.

2. Theory

The Earth’s external gravitational field is customarily expressed in a harmonic expansion in terms of the Stokes coefficients [e.g., Kaula, 1966]. The Stokes coefficients are directly proportional to the multipole moments of the Earth’s mass distribution [e.g., Chao and Gross, 1987]. Among them, the dimensionless “$J_l$” coefficients for the zonal harmonics are given by

$$J_l = \frac{1}{M R_1} \int \frac{\rho(r)r^l P_l(\cos \theta)}{dV} dV$$

(1)

In (1), $M$ and $R$ are the Earth’s total mass and mean radius; $\rho(r)$ is the mass density at position $r = (r, \theta, \lambda)$, where $r$ is the radial distance, $\theta$ is the colatitude, and $\lambda$ is the east longitude; $P_l$ is the Legendre function of degree $l$; and the integration is over the entire volume of the Earth (including the atmosphere) with volume element $dV = r^2 \sin \theta \, dr \, d\theta \, d\lambda$. Note that $J_l$ are not normalized; they are related to the normalized Stokes coefficients $C_{l0}$ by $J_1 = -(2l + 1)^{1/2} C_{l0}$.

Any temporal variation in $\rho(r)$ will, according to (1), give rise to a temporal variation in $J_1$, $\Delta J_1(t)$. For mass redistributions in the atmosphere (where $r = R$) in an Eulerian description and assuming a vertical hydrostatic profile for the atmosphere, it can be shown that

$$\Delta J_1(t) = -\frac{1 + k_1}{M g} \int \Delta \rho(\sigma, t) P_l(\cos \theta) d\sigma$$

(2)

where $g$ is the average gravitational acceleration, $\Delta \rho$ is the departure of the surface air pressure from the mean state, $\sigma$ is the solid angle representing $(\theta, \lambda)$, and the integration is over the unit sphere with surface element $d\sigma = \sin \theta \, d\theta \, d\lambda$. The Earth’s load Love numbers, $k_1$, are taken from Farrell [1972]: $k_1 = -0.31$, $k_2 = -0.20$, $k_3 = -0.13$, etc. The factor $(1 + k_1)$ accounts for the elastic yielding effect of the solid Earth under surface loading/unloading [Munk and MacDonald, 1960]. Note that $P_l$ serves as a geographical weighting function in the surface integral: $\Delta \rho$ is weighted favorably around the antinodal latitudes of $P_l$ and unfavorably around the nodal latitudes.

The response of the fluid ocean to the overlying atmospheric loading/unloading should also be properly treated in the evaluation of (2). The measured $\Delta \rho$ over oceanic regions applies directly only if the ocean were rigid. Otherwise it would be modified by the ocean’s response as felt by the solid Earth (at ocean bottom). If the ocean responds isostatically to changes in atmospheric pressure, sea level will be depressed as local atmospheric pressure rises. In the extreme case where the ocean can react completely isostatically in the fashion of a so-called inverted barometer (IB) behavior, the effective $\Delta \rho$ will be uniform over ocean (as if the oceans were “averaged out”), varying only with time in accordance with the total mass change over the entire ocean. This will greatly reduce the net $\Delta J_1$, considering the large ocean area. The IB hypothesis appears reasonable on seasonal time scales, but it has not been proven conclusively to be valid. Here, lacking pertinent knowledge, we will do the computation both ways: with and without the IB effect. Similar approach is taken by Chao and Au [this issue] in a polar motion study.

3. Data and Computation

The evaluation of equation (2) calls for global surface air pressure as the input data. We choose the modern meteo-
logical analysis data set from the European Centre for Medium Range Weather Forecasts (ECMWF). The data set consists of a set of meteorological variables, including the needed surface pressure. These pressure values are derived for the mean elevation (0 m for the sea level) at each grid point. The original data grid is 2.5° latitude by 2.5° longitude. The particular data set that we use has been reprocessed and interpolated onto a 2° latitude by 2.5° longitude grid, as described by Schubert et al. [1990]. The integration (2) over the geographic grid is performed using a linear interpolation scheme, i.e., the trapezoidal rule. The data are given twice daily at 0000 UT (Greenwich midnight) and 1200 UT (Greenwich noon); the corresponding pressure values differ only slightly. We take their mean to be the "daily" value, and 9-year series (1980–1988) of these daily data are employed.

We have chosen the ECMWF data over a similar data set from the U.S. National Meteorological Center (NMC). Although going back further in time (1976), the NMC data suffer more from data gaps and inconsistencies (particularly those before 1980) resulting from operational and model changes in the analysis [Trenberth and Olson, 1988]. Changes of the latter kind also exist in the ECMWF data, but they are better documented and are of little consequence to the ECMWF pressure values. The ECMWF and similar data sets are quickly becoming widely used in general meteorological investigations and forecast experiments, as well as in geodynamics studies as, in fact, exemplified by the Earth rotation analyses mentioned in section 1.

4. RESULTS AND DISCUSSION

Figure 1 shows the computed 9-year series for ∆Jₙ, 𝑙 = 2 – 6: Figure 1a, for "non-IB" (without the IB effect); Figure 1b, for "IB" (with the IB effect). The mean values have been removed. They show prominent seasonal (annual and semiannual) signals, as well as interannual and decadal variations superimposed on short-period fluctuations.

4.1. Results: Seasonal Variations

We estimate the amplitude and phase of the seasonal components in each ∆Jₙ(𝑡) series as follows. First, we generate a composite "mean year" series by straight averaging of the 9 years. We then fit a linear combination of annual and semiannual sinusoids to the mean year series in the least squares sense. The results are given in Table 1. Here, following Chao and O'Connor [1988], the sinusoids are expressed as

\[ ∆J(t) = ∆J \sin (ωt + φ) \]  

where ω is the angular frequency (annual or semiannual), and t = 0 refers to January 1. The advantage of this expression is that a polar quantity with magnitude ∆J and phase angle φ, (∆J, φ), in the "sine-cosine" plane adds like a two-dimensional vector. This proves convenient in combining contributions of different origin. Note that the IB effect reduces the variation amplitudes roughly by a factor of 2; the resultant seasonal phase differences are moderate except for the annual ∆J₁. Table 1 also lists the estimated hydrological ∆Jₙ (𝑙 = 2 – 4) obtained by Chao and O'Connor [1988], which utilizes the snow accumulation data from satellite remote sensing [Chao et al., 1987] combined with rainfall data collected worldwide over many decades [Willmott et al., 1985].

4.2. Comparison of Seasonal Variations With Observations

We now wish to compare our results with observations of geodetic satellite orbits, for which we adopt Shum et al. [1987] for Lageos (courtesy of R. Eanes and M. Watkins) and Cheng et al. [1990] for Starlette. The solid Earth tides (see section 1) have been modeled and removed from these observations. Unfortunately, the individual ∆Jₙ are not directly observed. Instead, the primary observable is the satellite-specific orbit nodal residual, ∆Ω, which equals a linear combination of even-degree terms (∆J₂, ∆J₄, ∆J₆, ...) of the form

\[ \frac{d}{dt} \deltaΩ = \sum_{l=2}^{\text{even}} f_l \bar{a} l^{-l} ∆J_l \]  

where \( \bar{a} \) is the satellite semimajor axis in units of the Earth's mean radius and \( f_l \) is a numerical factor depending on the orbit elements (for details see, for example, Kaula [1966]). Terms in (4) cannot be separated until observations from a multitude of diverse satellite orbits are available. Thus, presently, we can only use Table 1 to predict ∆Ω for given satellites and compare them with the observed ∆Ω. The results for the annual (amplitude, phase) "vectors" are displayed in Figure 2, where amplitude is given in units of milliarcseconds (mas) and phase refers to the sine convention for January 1 (see above).

For Lageos annual ∆Ω, Figure 2a presents the predicted atmospheric vector of this study, the hydrological vector from Chao and O'Connor [1988], and their sum, for both cases of IB and non-IB. They are

\[ \deltaΩ(\text{atmospheric IB}) = (11.1 \text{ mas}, 134°) \]
\[ \deltaΩ(\text{atmospheric non-IB}) = (20.9 \text{ mas}, 115°) \]
\[ \deltaΩ(\text{hydrological}) = (9.5 \text{ mas}, 149°) \]
\[ \deltaΩ(\text{sum IB}) = (20.4 \text{ mas}, 141°) \]
\[ \deltaΩ(\text{sum non-IB}) = (29.2 \text{ mas}, 126°) \]
\[ \deltaΩ(\text{obs}) = (17.9 \text{ mas}, 122°) \]

The last quantity is a nominal "observed" vector obtained by averaging a 7-year (1980–1986) series from Shum et al. [1987]. It should be stressed that the year-to-year variation in this 7-year series is rather large: individual years can differ from the average by a factor of 2 or more. Hence the present ∆Ω(obs) vector is associated with a large uncertainty. In light of this, the general agreement of the ∆Ω(sum) vectors with the ∆Ω(obs) vector is remarkable. The ∆Ω(sum non-IB) agrees well with ∆Ω(obs) in phase but gives a considerably larger amplitude, whereas ∆Ω(sum IB) agrees better in amplitude but not as well in phase.

For Starlette annual ∆Ω, the corresponding vectors are

\[ \deltaΩ(\text{atmospheric IB}) = (85 \text{ mas}, -45°) \]
\[ \deltaΩ(\text{atmospheric non-IB}) = (117 \text{ mas}, -78°) \]
\[ \deltaΩ(\text{hydrological}) = (104 \text{ mas}, -25°) \]
Fig. 1. Temporal variation of the Earth's low-degree gravitational coefficients ($\Delta J_l$, $l = 2 - 6$, in units of $10^{-10}$) caused by atmospheric mass redistribution for 1980-1988: (a) without the inverted barometer effect (non-lB) for the ocean; (b) with the inverted barometer effect (lB).

\[ \delta\Omega(\text{sum IB}) = (186 \text{ mas}, -34^\circ) \]
\[ \delta\Omega(\text{sum non-IB}) = (198 \text{ mas}, -53^\circ) \]
\[ \delta\Omega(\text{obs}) = (236 \text{ mas}, -107^\circ) \]

The observed vector $\delta\Omega(\text{obs})$ is obtained by averaging a 9-year (1980-1988) series from Cheng et al. [1990]. Again, the year-to-year variation in this series is considerable (but not as large as with Lageos). The comparability, especially in amplitude, between the $\delta\Omega(\text{sum})$ vectors and $\delta\Omega(\text{obs})$, although much poorer than Lageos' above, is still worth noting (depicted in Figure 2b).

The corresponding comparisons for the semiannual variations, however, show little agreement. Only a qualitative
TABLE 1. The (Amplitude, Phase) of the Seasonal Variations of the Low-Degree Gravitational Field Due to Atmospheric Mass Redistribution With and Without the Oceanic IB Effect

<table>
<thead>
<tr>
<th></th>
<th>Atmosphere (Non-IB)</th>
<th>Atmosphere (IB)</th>
<th>Hydrological Estimates*</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Annual</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$J_2$</td>
<td>(2.54, $-157^\circ$)</td>
<td>(1.41, $-129^\circ$)</td>
<td>(1.36, $-116^\circ$)</td>
</tr>
<tr>
<td>$J_3$</td>
<td>(1.54, $73^\circ$)</td>
<td>(2.34, $86^\circ$)</td>
<td>(1.42, $-126^\circ$)</td>
</tr>
<tr>
<td>$J_4$</td>
<td>(1.52, $-155^\circ$)</td>
<td>(0.82, $-177^\circ$)</td>
<td>(0.36, $-177^\circ$)</td>
</tr>
<tr>
<td>$J_5$</td>
<td>(1.38, $-143^\circ$)</td>
<td>(0.86, $72^\circ$)</td>
<td></td>
</tr>
<tr>
<td>$J_6$</td>
<td>(1.01, $-102^\circ$)</td>
<td>(0.61, $-113^\circ$)</td>
<td></td>
</tr>
<tr>
<td>$C_{22}$</td>
<td>(1.19, $-89^\circ$)</td>
<td>(0.43, $-87^\circ$)</td>
<td></td>
</tr>
<tr>
<td>$S_{22}$</td>
<td>(0.46, $123^\circ$)</td>
<td>(0.38, $94^\circ$)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Semiannual</td>
<td></td>
<td></td>
</tr>
<tr>
<td>$J_2$</td>
<td>(0.27, $-122^\circ$)</td>
<td>(0.22, $-152^\circ$)</td>
<td>(0.52, $92^\circ$)</td>
</tr>
<tr>
<td>$J_3$</td>
<td>(1.33, $93^\circ$)</td>
<td>(0.69, $103^\circ$)</td>
<td>(0.36, $69^\circ$)</td>
</tr>
<tr>
<td>$J_4$</td>
<td>(1.12, $-101^\circ$)</td>
<td>(0.63, $-106^\circ$)</td>
<td>(0.22, $47^\circ$)</td>
</tr>
<tr>
<td>$J_5$</td>
<td>(0.44, $62^\circ$)</td>
<td>(0.33, $65^\circ$)</td>
<td></td>
</tr>
<tr>
<td>$J_6$</td>
<td>(0.67, $-28^\circ$)</td>
<td>(0.48, $-66^\circ$)</td>
<td></td>
</tr>
<tr>
<td>$C_{22}$</td>
<td>(0.10, $-147^\circ$)</td>
<td>(0.13, $64^\circ$)</td>
<td></td>
</tr>
<tr>
<td>$S_{22}$</td>
<td>(0.049, $-137^\circ$)</td>
<td>(0.069, $154^\circ$)</td>
<td></td>
</tr>
</tbody>
</table>

Amplitudes in units of $10^{-10}$ and phase in degrees with respect to the sine convention with $t = 0$ on January 1.
*From Chao and O'Connor [1988].

4.3. Discussion: Seasonal Variations

Several major factors contribute to the above discrepancies. Let us discuss the following contributions: (1) uncertainties in our atmospheric estimates; (2) uncertainties in the hydrological estimates of Chao and O'Connor [1988]; (3) systematic errors in the observed values; and (4) the presence of other geophysical processes that influence the seasonal global mass distribution.

For contribution 1, we first note that we have only included the (even) $\Delta J_l$ terms up to $l = 6$ in our $\delta\Omega$ (atmospheric) estimates. The omission of higher degree terms is inconsequential for Lageos given its high altitude but may introduce a considerable error for Starlette which is of relatively low altitude (see equation (4)). Cheng et al. [1989] suggest inclusion of terms up to $l = 16$ or even higher for Starlette $\delta\Omega$ to be accurate to within a few percent.

A more substantial uncertainty emanates from the large effect of the oceanic IB hypothesis (see section 2) on $\Delta J_l$ estimates, especially in terms of the amplitude. Thus it is important to determine the extent to which the IB effect is valid in reality on the time scale of interest. This problem remains unresolved to date.

Gutierrez and Wilson [1987] have previously computed the atmospheric gravitational influence on the orbit parameters of Lageos and Starlette. They used mean monthly air pressure data compiled by the National Meteorological Center (NMC) for 1958–1973. Their results for the annual $\delta\Omega$ (atmospheric) (IB or non-IB) for both satellites agree fairly well with our results: their amplitudes are smaller than ours by some 20%, whereas the phases agree to within 10°. How-

![Diagram of Lageos and Starlette](image-url)
ever, the differences for the semiannual term are quite large both in amplitude and phase, as with the differences in the odd degree terms. These presumably reflect the difference in the source atmospheric data.

As far as contribution 2 is concerned, the inadequacy of Chao and O'Connor's [1988] estimate is threefold: (1) large uncertainty in conventional data for rainfall, (2) exclusion of the (unknown) seasonal ice mass balance on Greenland and Antarctica, and (3) omission of terms with \( l > 4 \) in computing \( \Delta \), especially for the low-altitude Starlette. All these may have considerably degraded the \( \Delta \) (hydrological) estimate.

Furthermore, for the semiannual term, the combined (atmospheric + hydrological) estimate is highly prone to errors in the two constituents because they are comparable in magnitude but roughly opposite in phase (see Table 1). The same is true for the annual \( \Delta J_3 \).

In contrast to the qualitative argument given in section 1, Gutierrez and Wilson [1987] yielded much smaller \( \Delta \) (hydrological) estimates than ours derived from Chao and O'Connor [1988]. The source of this discrepancy, again, presumably lies in the large difference in the data used in the two studies, particularly for the snow mass: Gutierrez and Wilson [1987] used the conventional monthly climatology data, whereas Chao and O'Connor [1988] used modern satellite remote sensing data.

For contribution 3, it is noted that even the Lageos and Starlette values are not consistent between them (R. Eanes and M. K. Cheng, personal communication, 1990). Progresses in improving the orbit observation and modeling can be expected to reduce the systematic errors in the future.

Contribution 4, of course, is of the most geophysical interest. As discussed earlier, perhaps the largest "missing piece" lies in the ocean. The seasonal mass redistribution in the ocean is poorly known, not only with respect to the IB behavior, but also in the tidal and the wind-driven components. One wishes that, in conjunction with dealing with the problems associated with contributions 1–3, the oceanic contribution can be determined in an unambiguous manner. This would lead to meaningful constraints on the behavior of the oceans on seasonal time scale.

4.4. Interannual and Decadal Variations

The gravitational variations of tidal origins should be identical from year to year. Hence any observed interannual variation is presumably meteorological in origin. Indeed, interannual variations are evident in Figure 1. Note the unequivocal signatures of the 1982–1983 and the 1986–1987 El Niño episodes, particularly in \( \Delta J_2 \) and \( \Delta J_6 \). It would be interesting to conduct a year-by-year comparison of the \( \Delta J_1 \) variations with those observed, similar to, although less robust than, our above analysis with respect to the seasonal variations. Perhaps more important, however, is the presence of the \( \Delta J_1 \) variations of decadal time scale that are also evident in Figure 1. They have amplitudes of the order of \( 10^{-10} \), which is comparable to those predicted and reportedly observed for the secular postglacial rebound [Yoder et al., 1983; Alexander, 1983; Rubincam, 1984]. It is unclear how much of these observations, spanning only a few years, has actually been "contaminated" by the decadal meteorological variations (which, of course, should also include the poorly known hydroospheric contributions.) This obviously needs to be investigated as it would have significant impact on realistic constraints for global rheology models of the Earth [e.g., Peltier, 1983; Rubincam, 1984].

4.5. Hemispheric Contributions and Other Harmonic Components

We have also computed \( \Delta J_l \) for the northern and southern hemispheres individually in order to identify the hemispheric contributions (results not shown). Typically, the southern hemispheric component is noticeably larger than its northern counterpart. The annual phase difference is about 6 months for \( \Delta J_2 \) but much smaller for higher harmonics, indicating the contrast among the modes of the hemispheric meteorology.

Although we have only focused on the zonal coefficients, one can, if so desire, readily compute the variations of other harmonic coefficients using formulas given by, for example, Chao et al. [1987]. For example, we have computed the variation in \( C_{22} \) and \( S_{22} \), the normalized sectoral Stokes coefficients of degree 2 and order 2. The results are also listed in Table 1. They correspond to a seasonal oscillation of a few arc seconds in the orientation of the two equatorial principal axes (cf. Chao and Gross, 1987, p. 585), which is probably too small to be detected geodetically (F. Lerch, private communication, 1990). Similarly, one can compute the two tesseral coefficients of degree 2 and order 1, which give the polar motion excitation due to the atmospheric mass redistribution. That result has been reported by Chao and Au [this issue].

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