Decadal Climate Variability: Is There a Tidal Connection?

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ABSTRACT
A possible connection between oceanic tides and climate variability arises from modulations in tidally induced vertical mixing. The idea is reexamined here with emphasis on near-decadal time scales. Occasional extreme tides caused by unusually favorable alignments of the moon and sun are unlikely to influence decadal climate, since these tides are of short duration and, in fact, are barely larger than the typical spring tide near lunar perigee. The argument by Keeling and Whorf in favor of extreme tides is further handicapped by an insufficiently precise catalog of extreme tides. A more plausible connection between tides and near-decadal climate is through “harmonic beating” of nearby tidal spectral lines. The 18.6-yr modulation of diurnal tides is the most likely to be detectable. Possible evidence for this is reviewed. Some of the most promising candidates rely on temperature data in the vicinity of the North Pacific Ocean where diurnal tides are large, but definitive detection is hindered by the shortness of the time series. Paleoclimate temperature data deduced from tree rings are suggestive, but one of the best examples shows a phase reversal, which is evidence against a tidal connection.

1. Introduction
My title is borrowed from Munk et al. (2002), except their “millennial” is here “decadal.” The Munk et al. study was stimulated in large part by two papers by Keeling and Whorf (1997, 2000), who argued that global temperature data show significant inverse correlations with times of maximum tidal forces. In the Keeling–Whorf view, large tides induce enhanced oceanic vertical mixing, with associated perturbations to global surface temperatures and climate. Munk et al. took a dim view of any such tidal effect working at millennial time scales, but they concluded that the Keeling–Whorf proposal was “the most likely among unlikely candidates.”

The connection between tides and vertical mixing at shorter time scales is more firmly established, and possible climatic influences appear more conceivable. Tidal mixing in shallow seas is a common and well-studied phenomenon, with fortnightly oscillations arising from the spring–neap cycle apparent in many locations. Mixing in key regions, such as the Labrador Sea, is capable of influencing larger-scale ocean circulation, including deep-water formation (e.g., Lee et al. 2006). Moreover, the meridional overturning itself is possibly driven in part by deep-ocean tidal mixing. Munk and Wunsch (1998) lay out the main ideas and a very rough energy budget. Egbert and Ray (2000) show that the barotropic tide is energetically capable of depositing the requisite energy into the deep ocean, and many current studies of deep-ocean internal tides are attempting to unravel the mechanisms and energetics in better detail (e.g., Rudnick et al. 2003).

Notwithstanding the existence of tidally induced mixing, however, it is difficult to see how occasional extreme tides—or tidal “events,” as labeled by Keeling and Whorf (1997, hereinafter KW97)—which are of very short duration and only marginally larger than typical spring tides, could affect climate on decadal (or longer) time scales. Yet there continues to be a steady stream of papers published along these lines, many of them finding supposed correlations between occasional extreme tides and various decadal-scale climatic variables. Because the Keeling–Whorf study is one of the more influential of these efforts, its arguments are worth examining in greater depth. When that is done, as shown below, the KW97 hypothesis appears even weaker than one might have initially thought.

The material in sections 3 and 4 is developed primarily to address the KW97 paper. But those calculations...
naturally lend themselves to related connections, or possible connections, between tides and climate. To answer the question in our title with anything but a negative, the most promising of these connections is investigated, with a generously flexible definition of “climate” to include even very localized atmospheric effects. The most promising link involves “harmonic beating” (Munk et al. 2002) from the moon’s 18.6-yr nodal cycle and its modulation of short-period tides (Loder and Garrett 1978). It is quite analogous to the spring–neap fortnightly modulation readily observed in many shallow seas. Thus, in this case the physical mechanism via mixing between tides and decadal (or nearly bidecadal) variability appears on somewhat firmer ground, although details remain murky. Evidence for an 18.6-yr effect in historical climate records is intriguingly suggestive but by no means completely convincing. These matters are discussed in sections 5–7, primarily as a brief review of work by others. Because the climate literature abounds with misconceptions regarding the nodal cycle, typically as confusion between the nodal modulations of short-period (daily) tides and the 18.6-yr node tide itself, appendix B offers a short tutorial on the nodal regression and its tidal consequences.

2. Keeling–Whorf hypothesis

KW97 began with monthly mean, global surface temperature anomalies as described by Jones et al. (1999) and shown here in Fig. 1a. They filtered these data in several different ways to reveal variability in the near-decadal band. The filtered time series adopted here is similar to the one that KW97 generally emphasized in terms of correlations with tides. It is based on KW97’s maximum entropy spectrum analysis that suggested nine dominant frequencies in the period range between 6.05 to 31.4 yr. The monthly data are fit by least squares to sinusoids at these identified frequencies (plus a low-degree polynomial to absorb the main secular trends).

![Figure 1](image-url)
The resulting nine-term sinusoidal series is shown in Fig. 1b. The curve in Fig. 1b corresponds closely to the “low frequency/spectral” curve in KW97’s Fig. 7.

The vertical gray bars in Fig. 1b delineate the so-called tidal events of KW97. These are times identified as having unusually large tidal forces; they were extracted from a compilation by Wood (1986). As KW97 emphasized, there is some tendency for the tidal events to fall near times of global low temperatures, although there are also obvious exceptions. In an attempt to be more objective, a statistical test has here been devised, with details given in appendix A. The result is that the null hypothesis of complete independence between KW97’s tidal events and the temperature oscillations can be rejected at the 5% level but not at the 1% level.1

Because the entire KW97 hypothesis, as well as others of a similar nature, rests upon these coincidences between tidal event times and filtered temperature extremes, it is important to understand in detail how these tidal events are determined and how anomalous (or not) they actually are. This is discussed in the following section.

3. Times of maximum tidal force

The tidal events used by KW97, corresponding to times of unusually large tidal forces on the earth, are extracted from Table 16 of Wood (1986). They are not actually based on direct calculations of the tidal forces but rather on a particular angular velocity of the moon, which is taken as a proxy for the tidal force. Since the calculation of the actual force (or potential) is straightforward, there appears no reason to resort to such a proxy, especially since the identified events turn out to be sensitively dependent on the precision of the calculations. In this section, the times of maximum tidal force (or potential) are recomputed and it is shown how so-called extreme events are identified.

We assume a rigid, spherical earth. Figure 2 defines all the astronomical quantities required for the calculation of the maximum tidal potential, which occurs at the point \( P \), located along the great circle between the sublunar and subsolar points. The actual orientation of the earth is irrelevant for this purpose. (This is not the case if the potential is to be decomposed into diurnal and semidiurnal components, see section 5.) The total potential at point \( P \) is given by

\[
V(P) = V_\text{m}(P) + V_\text{s}(P),
\]

where the lunar potential is

\[
V_\text{m}(P) = \frac{GM_\text{m}}{R_m} \left( \frac{a}{R_m} \right)^2 P_2(\cos \delta) + \left( \frac{a}{R_m} \right)^3 P_3(\cos \delta)
\]

\[
+ \left( \frac{a}{R_m} \right)^4 P_4(\cos \delta)
\]

\[\text{(1)}\]

where \( G \) is the Newtonian constant, \( M_\text{m} \) is the lunar mass, \( a \) is the mean equatorial radius of the earth, and \( P_n(\mu) \) is a Legendre function of degree \( n \). The expansion in (1) is taken to the fourth degree, because, as discussed below, a proper analysis of KW97’s arguments requires that the equilibrium tide \( V(P)/g \) is computed with a precision of about 0.1 mm, and this generally requires a degree-4 expansion. Owing to the sun’s much greater distance, the solar tidal potential can be limited to just the first term:

\[
V_\text{s}(P) = GM_\text{s}R_s\gamma^3a^2P_\gamma[\cos(\psi - \delta)],
\]

\[\text{(2)}\]

with \( M_s \) being the solar mass.

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1 KW97 made the correlations somewhat better by adding three additional tidal events that were not times of especially large tidal forces, in the years 1904, 1922, and 1940. KW97 asserted that these were times of large semidiurnal tides, but they are not, see Table 2. Because these events appear to have been added ad hoc they are not used here.
In the following calculations the adopted constants are
\[
GM_m = 4.902798 \times 10^{12} \text{ m}^3 \text{ s}^{-2},
\]
\[
GM_s = 1.327124399 \times 10^{20} \text{ m}^3 \text{ s}^{-2}, \quad \text{and}
\]
\[
a = 6.37816 \times 10^6 \text{ m}.
\]

To convert potentials to equilibrium tidal elevations, use is made of the mean gravitational acceleration \( g \), taken as 9.82 m s\(^{-2}\).

The time-varying quantities \( R_m, R_s \), and \( \psi \) are determined from lunar and solar ephemerides. The precision required for the ephemerides is again dictated by required \( \pm 0.1\)-mm precision in the equilibrium tide \( V(P)/g \). When the tide is near peak, \( V \) is fairly insensitive to errors in angle; variational analysis suggests \( \psi \) is needed to only about 0.2° or 10'. Requirements on the lunar distance \( R_m \) are somewhat more stringent, about 25 km, and for the solar distance \( R_s \) only about 25,000 km. In modern terms these precisions are relatively crude and easy to achieve.

The lunar ephemeris adopted here is an approximation to the semianalytical theory of Chapront-Touzé and Chapront (1983), with a slight update of the mean elements that form the fundamental arguments in the series expansions (Meeus 1998). While the original series require several thousands of terms, Meeus’s reduced series consist of 50 periodic terms in the lunar longitude, 45 terms in latitude, and 31 terms in distance; the resulting lunar positions are reportedly accurate to about 10° in angle and 20 km in distance (Meeus 1998) during the twentieth century. Errors in longitude are somewhat larger when extending backward (or forward) in time by several centuries, owing primarily to a slightly incorrect constant (by about 2° century\(^{-2}\) according to modern estimates) for the moon’s secular tidal acceleration, but such errors are of no consequence for our purposes. The adopted solar series is based on Stumpff (1979, 1980), with three-dimensional coordinates accurate to better than 2000 km. The accuracy of \( R_m \) in the Meeus approximate ephemeris is marginal, so the most critical parts of our calculations are checked in two independent ways, to be discussed presently.

Because the lunar potential dominates the solar, the angle \( \delta \) in Fig. 2 is always confined to the interval [0°, 16°]. Its value is found by maximizing the potential \( V(P) \) over this interval by employing an algorithm that combines successive golden-section searches and parabolic interpolations (Brent 1973).

A continuous time series of \( V(P) \) is shown in Fig. 3 in terms of the equilibrium ocean tide \( V/g \). Not unexpectedly, the main oscillations are the near-fortnightly spring–neap cycle. Spring tides occur when \( \delta \) reaches a minimum, as the sun–moon angular distance \( \psi \) approaches either 0° (new moon) or 180° (full moon). When one spring tide is significantly larger than its successor, it is because either full or new moon is nearly aligned with the lunar perigee. That happens about twice a year as the earth–sun system rotates relative to the longitude of lunar perigee.

If only the local maxima (one per fortnight) of Fig. 3 are plotted as discrete points, the result (expanded for the entire 1950–60 decade) is shown in Fig. 4. The maxima are seen to oscillate back and forth between approximately 56 and 62 cm as the spring tides are in phase or in quadrature with perigee. An occasional spring tide at perigee may be a few millimeters larger than others, depending mostly on how closely full or new moon occurs to the perigee points.
ample, successive full (or new) moons generally jump along in longitude by 25° to 35°, and such increments normally fail to align precisely with perigee. The small differences between the spring tides are brought out more clearly in the bottom part of the figure, which is a zoom view of the extreme end of the y axis. Nine spring tides are seen to generate an equilibrium tide exceeding 62.3 cm, with the largest occurring on 0244 UTC 29 December 1955. This tide was one of the tidal events of KW97, but it must be emphasized that the tide was but a few centimeters larger than any other spring tide during the decade; it was hardly anomalous.

It is of interest to note that, while small values of $\psi$ tend to generate large tides, only one of the nine dates depicted in the bottom of Fig. 4 experienced an eclipse, 14 February 1953, which was a partial solar eclipse with $\psi$ approximately 1.66°. A moderately small $\psi$ of a few degrees can also yield a large $V(P)$ if the times are closely aligned with lunar and solar perigee. Since the latter occurs in the modern era near the beginning of January, all notably large $V(P)$ values occur in either December, January, or February.

Figure 5 is an extension of Fig. 4 to the entire timespan of the KW97 analysis. The extreme tides are again seen to be hardly remarkable, with all perigean spring tides occurring in a cloud less than 3 cm wide.

The zoom view in the lower diagram, emphasizing the very largest tides, is only 6 mm wide.

The circled times in Fig. 5 are those that correspond to times of KW97’s tidal events, as extracted from Wood (1986). As noted above, KW97 do not use actual tidal force or potential but rather a proxy for it based on a particular lunar angular velocity, the details of which need not concern us. The main point is that, at the precision with which KW97 use the data, this proxy is not sufficiently accurate. Many highest tides are missed; others selected are not especially large, at least in comparison. Table 1 gives the correct list of the 20 largest equilibrium tides over the 1860–2000 timespan. Consider as an example the events of 1961 and 1964. The computed values of the equilibrium tide $V(P)/g$ at the relevant times—37315.92361 (2210 UTC 16 January 1961) and 38748.20000 (0448 UTC 19 December 1964)—are 62.758 and 62.727 cm, respectively. In other words, the 1961 event, which KW97 ignore, is about 0.3 mm larger. The values of Wood’s angular velocity that KW97 adopted for these two times are 17.119° and 17.138° day$^{-1}$, respectively (Wood 1986). The 1964 event falls near one of the temperature dips in Fig. 1b, so its use strengthens KW97’s argument, whereas the larger 1961 event is closer to a temperature maximum, which weakens their argument. That such tiny submil-
limeter differences in equilibrium tide is of importance when testing KW97’s hypothesis is, it would seem, strong evidence against the entire proposition. Global temperatures are surely insensitive to whether the 1961 tide was 0.3 mm larger than the 1964 tide.

As a check of these calculations, the equilibrium tides for the 1961 and 1964 events have been recomputed by a completely independent method based on the harmonic decomposition of the tidal potential derived by Cartwright and Edden (1973). Their lunar ephemeris was based on several hundred terms from the Eckert et al. (1954) theory, so it is more precise than the simplified Meeus expansion used here. The Cartwright–Edden decomposition resulted in 505 spectral lines. Summation over all lines at the relevant times confirms that the 1961 tide was larger than the 1964 tide by a fraction of a millimeter.

Furthermore, a second check was made of the 20 largest tides tabulated in Table 1, including the two events of 1961 and 1964, by use of the numerical DE406 ephemerides of Standish (1990, 1998). For times around the present (late twentieth and early twenty-first centuries) these ephemerides are indisputable, with accuracies in the lunar distance $R_m$ less than 1 m owing to least squares fit to lunar laser ranging data. The Table 1 equilibrium tides change by amounts ranging from 0.03 to 0.12 mm (all with the same sign); two pairs of tides that are very close in Table 1 (1862 and 1930, 1875 and 1886) reverse their order, changes that are of order 0.03 mm, which are of no consequence whatsoever.

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Figure 6 corrects Fig. 1b by replacing the 15 KW97 times with the correct times of the 15 largest equilibrium tides. By examination, the correlation with decadal-scale temperature data appears less impressive. If these times of the 15 largest tides are used in the

<table>
<thead>
<tr>
<th>Ranking</th>
<th>Modified Julian date</th>
<th>Date</th>
<th>Time</th>
<th>$V(P)/g$ (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>8</td>
<td>01495.1937</td>
<td>21 Dec 1862</td>
<td>0439</td>
<td>62.798</td>
</tr>
<tr>
<td>5</td>
<td>03341.0021</td>
<td>10 Jan 1868</td>
<td>0003</td>
<td>62.824</td>
</tr>
<tr>
<td>17</td>
<td>06234.7778</td>
<td>12 Dec 1875</td>
<td>1841</td>
<td>62.742</td>
</tr>
<tr>
<td>6</td>
<td>08080.5833</td>
<td>31 Dec 1880</td>
<td>1400</td>
<td>62.819</td>
</tr>
<tr>
<td>16</td>
<td>09926.3917</td>
<td>20 Jan 1886</td>
<td>0923</td>
<td>62.741</td>
</tr>
<tr>
<td>7</td>
<td>12820.1667</td>
<td>23 Dec 1893</td>
<td>0401</td>
<td>62.813</td>
</tr>
<tr>
<td>11</td>
<td>14665.9736</td>
<td>11 Jan 1899</td>
<td>2321</td>
<td>62.789</td>
</tr>
<tr>
<td>4</td>
<td>19405.5569</td>
<td>4 Jan 1912</td>
<td>1322</td>
<td>62.834</td>
</tr>
<tr>
<td>13</td>
<td>21451.1438</td>
<td>26 Dec 1924</td>
<td>0327</td>
<td>62.774</td>
</tr>
<tr>
<td>9</td>
<td>25990.6465</td>
<td>14 Jan 1930</td>
<td>2243</td>
<td>62.801</td>
</tr>
<tr>
<td>12</td>
<td>30730.5333</td>
<td>6 Jan 1943</td>
<td>1248</td>
<td>62.789</td>
</tr>
<tr>
<td>19</td>
<td>32162.8118</td>
<td>8 Dec 1946</td>
<td>1929</td>
<td>62.723</td>
</tr>
<tr>
<td>20</td>
<td>32576.3354</td>
<td>26 Jan 1948</td>
<td>0803</td>
<td>62.723</td>
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<tr>
<td>2</td>
<td>35470.1139</td>
<td>29 Dec 1955</td>
<td>0244</td>
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</tr>
<tr>
<td>15</td>
<td>37315.9236</td>
<td>16 Jan 1961</td>
<td>2210</td>
<td>62.758</td>
</tr>
<tr>
<td>18</td>
<td>38748.2000</td>
<td>19 Dec 1964</td>
<td>0448</td>
<td>62.727</td>
</tr>
<tr>
<td>1</td>
<td>42055.5042</td>
<td>8 Jan 1974</td>
<td>1206</td>
<td>62.880</td>
</tr>
<tr>
<td>3</td>
<td>48640.8937</td>
<td>19 Jan 1992</td>
<td>2127</td>
<td>62.847</td>
</tr>
<tr>
<td>14</td>
<td>51534.6750</td>
<td>22 Dec 1999</td>
<td>1612</td>
<td>62.766</td>
</tr>
</tbody>
</table>
Rayleigh statistical test employed above, the result is a very small $R$ statistic, implying no support for a connection between the times of large tides and KW97’s filtered temperature time series.

4. Criticisms of “extreme tide” hypotheses

The hypothesis that occasional extreme oceanic tides can influence the earth’s climate at decadal (or longer) periods is advocated uncommonly often in the literature, with KW97 being one of the more serious investigations. Building on the discussion in the previous section, the following criticisms may be lodged against the idea:

1) The word “extreme” is misleading. The tides so labeled are seen to be only a few centimeters—or even a few millimeters or less—larger than typical spring tides, at least in terms of equilibrium theory. Of course, owing to various resonances, the real ocean tides can be several times equilibrium values, even on basin scales, and many times greater in places, such as the Bay of Fundy, Nova Scotia. Moreover, weather-driven effects, such as storm surges, can indeed generate extreme sea levels, with obvious system nonlinearities, such as spilling over sea walls. But storms coinciding with large spring tides are very isolated events (time wise and space wise), and places like the Bay of Fundy are isolated resonances. It is difficult to fathom how such isolated events could affect global climate at long time scales as KW97 advocate.

2) The extreme tides are large for only a very short time. As an example, consider the event of 8 January 1974, the largest equilibrium tide of the entire twentieth century. Plotting the continuous value of $V(P)/g$ for the days 7–9 January, at the same vertical scale as in Fig. 5, gives Fig. 7. Clearly the tide is large for only a few hours around midday, while before and afterward the value of $V(P)/g$ falls off the graph, back into the realm of typical values. Any physical mechanism causing decadal temperatures to be sensitive to such small tidal differences over such short durations must be unbelievably nonlinear.

Keeling and Whorf acknowledge this when they write (KW97) “we do not wish to imply that single tidal events are likely to be responsible for modulating sea surface temperature worldwide.” However, they then

![Fig. 6. Same as in Fig. 1b except vertical gray bars here correspond to the correct times of 15 largest equilibrium tides, listed in Table 1, rather than the times used by Keeling and Whorf (1997).](image_url)

![Fig. 7. Time series of maximum equilibrium tide $V(P)/g$ for the tidal event of 8 Jan 1974, which is the peak value for the twentieth century. The vertical scale is consistent with the lower diagram in Fig. 5. The large tide was of relatively short duration; except for a few hours during midday of 8 Jan, it was otherwise no larger than other spring tides shown throughout Fig. 5.](image_url)
add that the events “may be important because they . . . could modulate temperatures by means of an ensemble of events over days or even years.” But there is no such ensemble. As Fig. 7 emphasizes, after only a few hours the tide returns to typical spring tide levels. Perhaps KW97 were thinking of the arcs they draw in their Fig. 6, connecting sequences of events separated by 18-yr intervals. Those arcs, however, are without meaning for the ocean, because the connected events simply fall within the general cloud of spring tides evident throughout the top panel of Fig. 5; the ocean is unaware of such connections.

For the specifics of the KW97 argument, we may add

3) The times of the KW97 tidal events have been either curiously selected, or (as in the case of the 1961–64 events) they have been mistakenly employed with precisions that the compilation by Wood (1986) cannot support. The latter relies on possibly weak nonlinearities to produce climate effects at frequencies that are the differences between two or more spectral lines of close frequency. The RC process is plainly the line of argument advanced by KW97 for tidal forcing of decadal variability, and it is vulnerable to the arguments summarized in the previous section. As noted by Munk et al. (2002), an HB process appears as a more promising avenue to connect tides and climate, because the perturbation persists over the entire period of the beating, unlike the very short durations of RC events. While Munk et al. had to resort to rather contrived combinations of integers to find possible HB frequencies within the millennial band, the decadal band (or most notably, bidecadal) is easy, since the moon’s node revolves once every 18.6 yr (appendix B), thus generating many candidate HB frequencies.

Now consider not the total tidal potential, as in the previous section, but separately the diurnal and semidiurnal potentials of second degree. By applying the addition theorem to (1) and (2), the lunar or solar potentials can be written as (e.g., Cartwright and Tayler 1971)

\[ V_{2m}(\theta, \lambda) = \frac{8\pi G M a^2}{5 R^3} (2 \cos^2 \epsilon - 3) \sin \theta \cos \theta e^{i \lambda} \]  

for the diurnal \((m = 1)\) or semidiurnal \((m = 2)\) band, where \((\theta, \lambda)\) is the geographical position on the earth’s surface where the potential is to be calculated, \(Y_{2m}^{(m)}(\Theta, \Lambda)\) is a complex spherical harmonic, and \((R, \Theta, \Lambda)\) is the distance and geographical coordinates of the sun or moon, as the case may be. The planetary positions are again determined by the ephemerides used in section 3.

A well-known implication of (3) that is worth emphasizing is the manner in which \(V_{2m}\) takes its local maxima. Because

\[ Y_{21}^{(1)}(\Theta, \Lambda) = \sqrt{(5/24\pi)^3} \sin \Theta \cos \Theta e^{i \Lambda} \quad \text{and} \quad Y_{21}^{(2)}(\Theta, \Lambda) = \sqrt{(5/96\pi)^3} \sin^2 \Theta e^{i 2 \Lambda} \]

applies to the diurnal and semidiurnal tides, respectively, the diurnal equilibrium tide is everywhere zero when the causative body (moon or sun) is over the equator \((\Theta = \pi/2)\). Diurnal tides in fact would not exist if the moon and sun remained always on the equator; they are maximized, at the expense of semidiurnal tides, when the moon and sun reach their maximum declinations off the equator. Unlike the diurnal tide the semidiurnal equilibrium tide is never identically zero everywhere, since \(\Theta\) never reaches 0 or \(\pi\).

Now as the simplest case of nonlinearity, consider the square of the combined lunar plus solar potentials, integrated over the earth’s surface, for which the integration can be performed analytically. (Note that physical arguments could well justify other powers of the poten-
The combined squared potentials take their largest values in the twentieth century on the dates listed in Table 2. But these “RC-type events” are not of direct interest for climate, for the same reasons that the Table 1 dates are not. It is instead of greater interest to average these squared potentials over time; Fig. 8 shows the result of daily averaging and Fig. 9 shows yearly.

The primary feature in the daily means is the spring–neap cycle. For the semidiurnal tides, this occurs as $M_2$ and $S_2$ constituents are in, and then out, of phase. As with the total potential in Fig. 3, neighboring peaks may differ in amplitude depending on whether the moon is near perigee or apogee, that is, on whether $N_2$ is also in phase. For the diurnal tides, the cycle occurs as $O_1$ and $K_1$ go in and out of phase. The diurnal and semidiurnal spring–neap cycles are not perfectly synchronized, because the frequencies of $O_1$ and $K_1$ differ by 2 cycles per month, while $M_2$ and $S_2$ differ by 2 cycles per month minus 2 cycles per year.

As the right part of Fig. 8 indicates, certain times have more pronounced spring tides than others. For the diurnal tides this occurs for a few days at times when the longitude of the lunar node is near zero (see appendix B) so that the moon’s declination is a maximum. Nodal modulation is not so evident for the semidiurnal tides in the figure; rather, maximum occurs for a few days every 4.4 yr. This is half the lunar perigee period.

FIG. 8. Mean square potential $|V_{2m}|^2$ over the globe for each day of 1900–40 for (top) diurnal ($m = 1$) and (bottom) semidiurnal ($m = 2$) tides. The left parts show in better detail the first 100 days of 1900; the near-fortnightly spring–neap cycle is evident.

TABLE 2. Date and times of largest diurnal and semidiurnal tidal potential in twentieth century along with mean square potential $|V_{2m}|^2$. Times are TDT.

| Modified Julian date | Date and time | $|V_{2m}|^2$ (m$^2$ s$^{-2}$) |
|----------------------|---------------|-------------------|
| Diurnal              |               |                   |
| 40209.8319           | 19 Dec 1968   | 1958              | 3.615 |
| 46794.8465           | 30 Dec 1986   | 2019              | 3.584 |
| 19405.2243           | 4 Jan 1912    | 0523              | 3.534 |
| 33624.8222           | 9 Dec 1950    | 1944              | 3.522 |
| 34008.1729           | 28 Dec 1951   | 0409              | 3.503 |
| 27423.1674           | 17 Dec 1933   | 0401              | 3.492 |
|                       |               |                   |
| Semidiurnal*         |               |                   |
| 40054.4951           | 8 Mar 1993    | 1153              | 5.464 |
| 21664.8903           | 12 Mar 1918   | 2122              | 5.440 |
| 15079.6847           | 1 Mar 1900    | 1626              | 5.426 |
| 16925.2646           | 21 Mar 1905   | 0621              | 5.422 |
| 39575.1486           | 26 Mar 1967   | 0334              | 5.418 |
| 44314.8708           | 16 Mar 1980   | 2054              | 5.408 |

* This list corrects dates in Cartwright (1974); see Ray and Cartwright (2007).
and the effect arises as perigee passes either spring or fall equinox, at which times both sun and moon are on the equator and the moon is at minimum distance.

For the yearly means (Fig. 9) the only variation clearly evident has a period of approximately 19 yr, which coincides with the regression of the lunar node. There is no indication of shorter period features, such as the 8.8-yr motion of perigee or the 4.4-yr half period noted for the peaks of Fig. 8; those peak daily tides occur on only a few days, which have little effect in the yearly average. [By taking a highly nonlinear function of the potential, rather than a simple square, it is possible that the 4.4-yr oscillation could be made to reappear. A possible geophysical example is sediment transport, for which only a few high tide days might be of significance; this seems the most likely explanation for the 4.4-yr bands reportedly detected in some corings (e.g., Berger et al. 2004).]

Squaring and averaging tidal potentials in this manner is an HB-type process, which, unlike an RC process, persists throughout the entire oscillation period. The frequencies of these oscillations would appear to be the most likely candidate frequencies for a nonlinear tidal process to affect climate. At relatively short time scales, we should therefore expect spring–neap oscillations and at longer time scales the 18.6-yr oscillations. It is conceivable that in some cases diurnal and semidiurnal tides are in just the right ratio that 9.3-yr oscillations could occur, since the 18.6-yr modulations of each tidal species is perfectly out of phase. Aside from this rather speculative exception, the only tide–climate effects at periods between a year and a century are likely to be those at 18.6 yr.

The relative strength of the 18.6-yr modulation of diurnal tides is in keeping with the discussion in appendix B. According to Table B1 the diurnal tide $K_1$ is modulated by $\pm 14\%$ and $O_1$ by $\pm 19\%$. In contrast, the dominant semidiurnal tide $M_2$ is modulated only $\pm 4\%$. (The semidiurnal $K_2$ does experience a large modulation, but $K_2$ itself is relatively weak.) The relative strength of the diurnal-band modulations is a most important point in any attempt to detect these effects.

6. Tide–climate scenarios

While KW97 advanced a purported link between tides and global temperatures, the remainder of this paper adopts a less ambitious definition for climatic effects, admitting almost any atmospheric variable that may be affected by oceanic mixing, possibly over a very localized region. Even under this more limited picture, a link with tides proves difficult to establish with any certainty.

Fig. 9. Same as in Fig. 8 except averaged over each year of the twentieth century. The 18.6-yr modulation is the main feature seen in both tides but relatively more important for diurnal tides.
Nonetheless, if tidal mixing is indeed an effective mechanism for bringing about tidally induced climate variations, at least two scenarios could be at work. The first (A) is that variations in oceanic tides directly modulate vertical mixing that brings colder water to the surface, thereby periodically cooling surface water and the overlying atmosphere (e.g., Loder and Garrett 1978; Keeling and Whorf 2000). Appropriate atmospheric teleconnections could conceivably extend the effect over larger regions. The second (B), stressed by Munk et al. (2002), is through variations in poleward heat transport associated with the ocean’s meridional overturning; to the extent that tides supply mechanical power for vertical mixing necessary to close the overturning loop, modulations of the tides could affect this transport, with associated climatic effects. A mixture of (A) and (B) is also conceivable: denser surface waters can enhance subduction and deep-water formation.

In fact, as is well known, the spring–neap oscillation, clearly evident in Fig. 8, can occur in sea surface temperatures and other surface phenomena through tidally generated mixing. This is mechanism (A) and there are many examples: Paden et al. (1991), Ffield and Gordon (1996), Rogachev et al. (2001), among many others. These examples are from relatively shallow waters, but the spring–neap cycle has also been observed, at least provisionally, in direct measurements of turbulent mixing in the deep Atlantic (Ledwell et al. 2000).

The tidal mixing mechanism so plainly evident at near-fortnightly periods should extend in principle to the 18.6-yr period. In such a case, however, it is more likely to be detected in diurnal tidal regimes rather than semidiurnal owing to the relatively weak modulation of the latter. In this regard, the Rogachev et al. (2001) study is noteworthy, because the fortnightly mixing they observe in the Okhotsk Sea is induced primarily by the diurnal \( K_1 \) and \( O_1 \) beating.

Concerning mechanism (B), several issues arise that suggest any modulations might be so weak as to remain undetectable. There is first the question of what fraction of the poleward heat flux may be attributed to the overturning circulation itself (Boccaletti et al. 2005; Saenko and Merryfield 2006), and whether a modulation of this is sufficient to influence the atmospheric state. A second question is whether the time scales associated with the overturning are too long to be affected by an 18.6-yr perturbation in the oceanic diffusivity; numerical models that include tidal mixing (e.g., Simmons et al. 2004) could presumably clarify this. A third point concerns the small magnitude of the modulation, since it must arise primarily from semidiurnal tides. Of the approximately 1 TW of tidal power that dissipates in the deep ocean, 75%–80% is from the \( M_2 \) constituent, 10%–15% from \( S_2 \), and probably less than 6% from all diurnal tides (Egbert and Ray 2003). Since the 18.6-yr modulation of \( M_2 \) is relatively small, and the modulation of \( S_2 \) is nonexistent, any tidally induced modulation of the overturning circulation must be fairly weak and therefore difficult to detect. A simple calculation based on the one-dimensional diffusion equation by Garrett (1979) shows that a 4% modulation of a mean vertical diffusivity (appropriate for \( M_2 \)) generates a perturbation in sea surface temperature of order 0.1°C.

The above discussion suggests that the most promising route for detecting a true tide–climate connection is to search for 18.6-yr modulations in regions of intense diurnal tides where mechanism (A) is likely to be at work. In such regions the primary fingerprint would be a general cooling effect from enhanced tidal mixing during periods when the longitude of the lunar node is near 0°. Figure 10 highlights (with relatively coarse resolution) those regions where the diurnal \( K_1 \) tide attains its largest current velocities. The boundaries of the North Pacific Ocean and the Antarctic Ocean are regions of strong currents. The most intense currents occur in the shallow waters of the Okhotsk, Bering, Yellow, South China, Java, Arafura, Ross, and Weddell Seas. Diurnal tides are weak throughout nearly the entire Atlantic Ocean; an important exception is the northern Labrador Sea and Baffin Bay region. See Foreman et al. (2006) for a recent discussion of the tidal energetics of the Bering Sea and its 18.6-yr modulation.

7. Evidence of 18.6-yr effects

There is as yet no completely convincing demonstration of 18.6-yr tidal effects in any climate variable. Many have been suggested (see Lamb 1972 for a review of older literature), but none appears completely immune to criticism. For example, an inordinately large number of supposed detections have been based on a single autoregressive spectrum with no analysis of sensitivity to pole orders or other parameters known to cause spurious spectral peaks; as Ghil et al. (2002) emphasize, “cross testing with . . . other techniques is especially important” for these kinds of spectral estimates and this has rarely been done.

Moreover, with short time series it is difficult to distinguish true 18.6-yr periodicity from a broad bidecadal variability that may be completely unrelated to tides and tidal mixing (Mann and Park 1994). For example, the Pacific decadal oscillation reconstructed from tree-ring data by Biondi et al. (2001) contains an astonishingly strong bidecadal component centered roughly at period 23 yr, sufficiently close to 18.6 yr to contaminate
estimates from short time series. Even without external forcing at these periods, the intrinsic modes of the coupled ocean–atmosphere system display strong variance in this frequency band (e.g., Latif and Barnett 1996). Such difficulties apply equally to other purported periodicities in climate, such as the 11-yr solar cycle discussed insightfully by Garric and Huber (2003).

To distinguish a true periodicity from background requires a long time series, and these are scarce when one is interested in periods of 18.6 yr. Several recent suggestions merit rather more serious consideration (e.g., Loder and Garrett 1978; Yasuda et al. 2006). What follows is a very selective set of examples that show some intriguing evidence of tide–climate correlations at 18.6 yr. While all published examples can hardly be reviewed here, the following does satisfy the criterion of displaying general cooling trends in or near those regions where strong diurnal tides occur and at times when those tides are at their nodal maxima. All of the examples that follow appear to be regional climatic effects and, to the extent that tides are involved, can be attributed primarily to shallow sea tidal mixing. To my knowledge, no evidence currently exists for nodal modulations of the deep thermohaline circulation.

**a. British Columbia SST**

Probably the first paper showing real insight into the physics of the problem was by Loder and Garrett (1978). They developed a simple model for vertical mixing in coastal waters in terms of a depth-independent eddy diffusivity, and they showed that modulations in tidal currents may lead to variations in sea surface temperature (SST) dependent on the temperature differential between surface and sea floor. Loder and Garrett emphasized that, to the extent that the lunar node regression is involved, it is not the node tide itself causing SST effects but rather nodal modulations of short-period tides (see further discussion in appendix B).

Loder and Garrett (1978) presented six SST time series from the Pacific and Atlantic coasts of Canada, obtained (mainly) during the early to midtwentieth century, through 1968. These time series indeed showed bidecadal oscillations. Moreover, since the Canadian Atlantic tides are predominantly semidiurnal while the Pacific tides are mixed, which as noted in section 5 likely leads to modulation effects dominated by the diurnal tides, any nodal modulations in SST would cause the two coasts to be out of phase, and this was also observed in the data.

The west coast time series from 12 British Columbia lighthouses are here extended in Fig. 11. The main bidecadal oscillations noted by Loder and Garrett (1978) are still evident, although the 1972 lows are “late” relative to the moon’s node and the 1989 lows are not so evident on the eastern shore of Vancouver (which, of course, could stem from strictly local effects, such as continental runoff).

Note, however, the near coincidence of the temperature dips with times of La Niña, which occurred in 1950, 1955, 1971, 1974–75, 1989, and 2000 (e.g., Trenberth 1997); the double dip at 1950 and 1955 appears espe-
cially revealing. This again highlights the difficulty in separating 18.6-yr effects from other climatic patterns in a relatively short time series. Although the series are now roughly doubled in length in comparison with the original 1978 analysis, considerably longer time series are evidently needed for a more convincing case.

b. Sitka air temperature

One longer time series from a region just north of the lighthouse data is the Sitka, Alaska, air temperature data previously discussed by Royer (1993). Royer showed that, for the period 1970–92, these air temperatures are well correlated with Gulf of Alaska subsurface temperatures obtained at 60°N, 149°W.

The monthly Sitka air temperature anomalies are shown in Fig. 12a. Royer filtered these data with a simple Butterworth filter. Here a number of filtering experiments have been tried in order to emphasize the bidecadal signal. A promising approach is to use a robust filter, such as a median filter, to overcome the outlier-like, large, short-period excursions evident in the figure, followed by a simple low-pass filter; this is shown in Fig. 12b. The later part of the time series shows exceedingly good correlations between temperature dips and times of strong diurnal tides, but the correlation mostly breaks down before 1920. Whether this owes to inadequacies in the older temperature measurements or to the fact that the temperatures are simply independent of the moon’s node is unknown. The pre-1920 data nonetheless seem to spoil the argument for an 18.6-yr tide effect. The data suggest that one or more other mechanisms are at work, which are dominant before 1920.

c. Northwest subarctic Pacific

As noted above, the northwest Pacific Ocean and especially the Okhotsk Sea contain regions of intense diurnal tides. Nakamura et al. (2004) give persuasive arguments for the importance of diurnal tidal mixing in this region, which, if valid, should induce 18.6-yr modulations. Osafune and Yasuda (2006) and Yasuda et al. (2006) indeed find a number of suggestive examples. They find that periods of strong diurnal tides coincide with periods of higher surface salinity and density and deeper isopycnal surfaces near the Kuril Islands and the east coast of Japan. Models with such effects tend to generate similar changes in dense shelf water formation in the Okhotsk and subsequent changes in poleward heat transport along the Kuril chain (Yasuda et al. 2006).

Both data and theory here suggest the likely presence of 18.6-yr modulations in several climatic variables. In some instances the relative phases of variables appear to shift, with one leading and then lagging another (see Fig. 2 of Yasuda et al. 2006), and the time series themselves are relatively short, extending back to only 1930, a period over which the Sitka data also look strong. With these cautions noted, this northern Pacific case provides some of the strongest evidence in twentieth-century data for a tidal effect. Longer time series from this region will be invaluable.

d. Tree-ring indexes

Paleoclimate data can potentially provide clearer evidence by their considerably longer time series.

Cook et al. (1997) produced a drought index for the western United State based on tree-ring chronologies extending back to the year 1700. They note considerable variance in the bidecadal band (see also Mitchell et al. 1979). Cook et al. employ various statistical tests to uncover possible relationships with both the lunar nodal cycle and the 22-yr Hale magnetic cycle, which others have also attempted. The tests for the lunar cycle appear on the surface to be significant, but a phase reversal was noted around the year 1800.

I would argue that a phase reversal between the lunar cycle and any climatological variable is prima facie evidence against a tide–climate relationship. If tidal mixing forces colder water to the surface, it cannot sud-
denly flip into another regime and start bringing up warmer water. Nor can a region that is primarily dominated by semidiurnal tides suddenly switch to diurnal. No realistic physical mechanism has been offered that is capable of establishing a “bistable phasing,” which has been advocated by some authors when they observe a phase reversal.

A more consistent coherence has been noticed by McKinnell and Crawford (2007) in western North America tree-ring data. They observe that periods of persistently warm temperatures occur only during periods when the diurnal tides are in the weak half of their 18.6-yr modulations (e.g., during the years 1939–42 and 1958–61 as seen in Fig. 9), whereas persistently cold temperatures can occur in any phase of the cycle. The tidal implication is that when diurnal tides are stronger than normal, mixing off of western Canada is more intense than usual, and no persistently warm period can occur. Weak diurnal tides do not guarantee persistent warm temperatures—other influences may prevent warming—so they are a necessary, but not sufficient, condition. This effect is seen to occur over a period extending back 400 yr and is therefore one of the most convincing examples to have been found, at least to the extent that tree rings are actually reflecting temperatures and not some other environmental variable.

Linderholm (2001) observes a significant 18.6-yr oscillation in tree-ring data from western Sweden. In the North Atlantic diurnal tides are relatively weak, so if any tidal modulation is at work, it must be via semidiurnal (primarily $M_2$) tides. Strongest mixing from semidiurnal tides occurs when the nodal longitude $N$ is near 180°, so warm temperatures occur when $N$ is near zero. This is indeed what Linderholm observes, although because the periodicity does not occur at all of his sites, he suggests a cause related to the lowering of the water table. This, it would seem, would be difficult for semidiurnal tides to achieve except over short 6-h periods, but some other mechanism may also be at work.

8. Concluding remarks

The difficult task of understanding the inherent variability, forces, interactions, and feedbacks in a broadband stochastic climate system often tempts us to attack simpler problems. Searching for simple periodicities in the climate system is extremely seductive, and, of course, extremely important if the periodicities truly exist. But pure line processes in the climate system must be rare. The one ready culprit of near-perfect spectral lines is the motions of astronomical bodies, so the search for periodicities understandably leads to investigations of tidal cycles and the related, but much longer period, Milankovitch cycles.

If the tides are contributing to decadal variability, it
is almost certainly not through the RC-type (repeat coincidences) process of occasional extreme tides advocated by Keeling and Whorf (1997) and a number of other authors. Extreme tides are very short duration events and, in reality, are hardly more extreme than typical spring tides near perigee. The KW97 study was further handicapped by reliance on a proxy for the tidal force, which was inadequate precisely for the task.

A more likely mechanism, as pointed out by Munk et al. (2002), is via an HB-type (harmonic beating) process, which acts nonlinearly on tidal currents, necessitating at the very least an analysis in terms of diurnal and semidiurnal tides rather than the complete tidal force. The most likely of these processes, and the one most likely to be sufficiently large to be detected, is that of the 18.6-yr modulation of diurnal tidal currents. By modulating vertical mixing tides can conceivably affect sea surface temperatures and thence other climatic variables—perhaps in regions strictly local to the mixing zone, or perhaps over wider regions if appropriate atmospheric teleconnections are excited. A modulation of the meridional overturning could have global consequences if the overturning responds to diffusivity modulations at these time scales; but any such tidal modulation would be caused principally by M₂, and because the nodal modulation of M₂ is small, any modulation of the overturning would be extremely difficult to detect in climate measurements.

Tidal mixing can vary, of course, as stratification and other ocean parameters vary, so we should not expect nodal modulations of climate variables to remain fixed over long time periods. Small changes in phase, for example, are to be anticipated. Phase reversals, however, are not. It is difficult to envision any physical scenario where tidal mixing cools the sea surface and overlying atmosphere and then suddenly reverses and begins a period of warming. It would seem that any phase reversal of a climate variable with respect to the lunar node is prima facie evidence against a relationship. One of the most promising examples of a nodal effect in a paleoclimate time series—the tree-ring series of Cook et al. (1997)—appears to suffer at least one phase reversal.

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APPENDIX A

Statistical Test

The correlation noted by KW97 between times of large tides and global temperature anomalies, as in Fig. 1b, can be subjected to a number of statistical tests, one of which is described here in further detail. Because the important factor is not temperature but rather the times of local temperature minima (or maximum temperature fall or rise or some similar characteristic) a relevant test should be based on local “phase” of the temperature. This can be done, for each tidal event, by mapping corresponding temperatures to a sinusoid with maxima and minima at times of nearest temperature maxima and minima. Phases corresponding to the KW97 events of Fig. 1 are shown in Fig. A1. Such mapped phases can be subjected to Rayleigh’s R test for uniformity or other criteria (e.g., Upton and Fingleton 1989).

If \( \varphi_i \) is the resultant local phase of the temperature anomaly for tidal event \( i \), Rayleigh’s R is given by

\[
R = \frac{1}{N} \left| \sum_{i=1}^{N} \cos \varphi_i + \sin \varphi_i \right|
\]

for \( N \) tidal events. Standard tables give critical values of \( R \) to test against a uniform distribution of phases. For probability \( \Pr(R > R_\alpha) = \alpha \), with \( N = 15 \), we have \( R_\alpha = 0.443, 0.542, \) and 0.649 for \( \alpha = 0.05, 0.01, \) and 0.001, respectively. A stricter test, and one somewhat more appropriate for the filtered temperatures used...
here, can be obtained by Monte Carlo sampling. If the filtered temperatures of KW97 (Fig. 1b) are Fourier transformed, and the phases, but not the Fourier amplitudes, are reset to uniformly distributed random numbers in the interval \([0, 2\pi] / H\text{9016}\), the resulting distribution for KW97’s tidal events is shown in Fig. A2. The value of \(R\) computed for the true temperature anomaly time series is 0.586. Hence, the null hypothesis that the tidal events are independent of the temperature data can be rejected at the 5% level (\(R\) / H11005 0.556) but not at the 1% level (\(R\) / H11005 0.647) for KW97’s events.

If the events, however, are taken to be the true times of maximum tidal force, as in Table 1 and Fig. 6, the \(R\) statistic is 0.274. In this case, the null hypothesis cannot be rejected. There is therefore no reason to hold that a relationship exists between the filtered temperatures of KW97 and times of maximum tidal force.

APPENDIX B

Regression of the Lunar Node

Misconceptions regarding the nodal cycle are so common in the climate literature that it seems some elementary description is called for. None of this material is new.

The geometry is pictured in Fig. B1. There are three fundamental planes: the ecliptic, the earth’s equator, and the lunar orbit. The equator is tilted to the ecliptic by 23.439° (the obliquity of the ecliptic). The moon’s orbit plane is tilted approximately 5° to the ecliptic. Neither of these quantities is constant; the dominant variations are a very slow decrease in the obliquity of about 47° per century and a small oscillation of the lunar inclination of amplitude 8° with approximately a half-year period.

The moon’s ascending node is the point where the moon crosses from south to north of the ecliptic. The mean longitude of the ascending node, denoted \(N\), is reckoned relative to the vernal equinox, one of two points where the earth’s equator intersects the ecliptic. Owing primarily to gravitational attraction by the sun, the moon’s orbit plane precesses in a retrograde sense so that \(dN/dt\) is negative. The period of precession is 18.61 yr. The longitude \(N\) may be evaluated by the formula (Meeus 1998)

\[
N(T) = 125.0445 - 1934.1363T + 0.00208T^2 + T^3/467440
\]

in degrees, where \(T\) is the number of Julian centuries since noon 1 January 2000. During the twentieth century, \(N\) crossed 0° on the following dates:

- 27 May 1913
- 6 January 1932
- 17 August 1950
- 29 March 1969
- 8 November 1987
The term is to be dis-

Nodal modulations of major constituents.

<table>
<thead>
<tr>
<th>Tide</th>
<th>Doodson No.</th>
<th>$\omega$ (° h$^{-1}$)</th>
<th>$\mathcal{H}$ (cm)</th>
<th>$\mathcal{P}$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M_0$</td>
<td>055.555</td>
<td>0.000000</td>
<td>31.45</td>
<td>100.0</td>
</tr>
<tr>
<td></td>
<td>055.565</td>
<td>0.002206</td>
<td>2.79</td>
<td>8.9</td>
</tr>
<tr>
<td></td>
<td>055.575</td>
<td>0.004412</td>
<td>0.03</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td>145.535</td>
<td>13.938623</td>
<td>0.15</td>
<td>0.6</td>
</tr>
<tr>
<td></td>
<td>145.545</td>
<td>13.940829</td>
<td>4.94</td>
<td>18.8</td>
</tr>
<tr>
<td>$O_1$</td>
<td>145.555</td>
<td></td>
<td>26.22</td>
<td>100.0</td>
</tr>
<tr>
<td></td>
<td>165.545</td>
<td>15.038862</td>
<td>0.73</td>
<td>2.0</td>
</tr>
<tr>
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</tr>
<tr>
<td></td>
<td>255.545</td>
<td>28.981898</td>
<td>2.36</td>
<td>3.7</td>
</tr>
<tr>
<td>$M_2$</td>
<td>255.555</td>
<td>28.984104</td>
<td>63.19</td>
<td>100.0</td>
</tr>
<tr>
<td></td>
<td>275.545</td>
<td>30.079931</td>
<td>0.10</td>
<td>1.2</td>
</tr>
<tr>
<td></td>
<td>275.555</td>
<td>30.082137</td>
<td>8.00</td>
<td>100.0</td>
</tr>
<tr>
<td>$K_2$</td>
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<td>2.38</td>
<td>29.8</td>
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<td>30.086550</td>
<td>0.26</td>
<td>3.2</td>
</tr>
</tbody>
</table>

The lunar orbit precession has a very large effect on the moon’s declination (its angle above or below the equator), maximizing the declination when $N = 0^\circ$ and minimizing when $N = 180^\circ$ (bottom half of Fig. B1). This has, in turn, a large effect on the character of the lunar tidal force, since a large declination tends to maximize diurnal forces at the expense of semi diurnal and vice versa. The effect of this nodal modulation on individual tidal constituents is well understood and is routinely allowed for in tidal analyses of sea level data (e.g., Doodson 1928). Table B1, which is based on the harmonic decomposition of the tidal potential by Cartwright and Edden (1973), shows in detail how the major tidal constituents are split by the nodal modulation, generating additional lines whose frequencies differ from the primary frequencies by 1 or 2 cycles per 18.6 yr. The rightmost column of Table B1 gives the relative amplitudes of these sidelines with respect to the primary lines. Of the major lunar tides, the largest effect occurs for the diurnal tides $O_1$ and $K_1$; the relative modulation of the semi diurnal $K_2$ is actually larger, but $K_2$ itself is relatively weak in comparison with $M_2$. Nodal effects are therefore of greatest importance in the diurnal band, a point consistent with the discussion in section 5.

Included in Table B1 is the permanent tide $M_0$, a permanent component of the earth’s flattening induced by the presence of the sun and moon. The nodal modulation of the permanent tide gives rise to the so-called node tide, or nodal tide, of period 18.6 yr and Doodson number 055.565. Like all long-period tides the ocean response to forcing at this low frequency is very close to equilibrium, with a spatial dependence dominated by the degree-2 zonal harmonic $Y_2^0(\theta, \lambda)$. The node tide is small, with maximum amplitude (at the North Pole) approximately 15 mm (Agnew and Farrell 1978). Such small tide is almost completely buried in the ocean’s large background variability. It has been unequivocally detected in tide gauge data only by stacking long time series from several hundred stations to arrive at a mean global response (Trupin and Wahr 1990). That response was found to be equilibrium with an uncertainty of about 20%.

It is highly unlikely that the small node tide, with vanishingly small associated current velocities, can induce perceptible climatic effects. To the extent that the lunar node regression is affecting climate, it must instead be through the nodal modulations of the much larger short-period (diurnal and semi diurnal) tides, as discussed in the main body of this paper. Loder and Garrett (1978) emphasized this point clearly, but the idea has unfortunately been lost on many subsequent authors who enter into extended, but irrelevant, discussions of the node tide itself.

Several of the lines in Table B1 have sidelines that are second harmonics of the nodal regression, thus generating 9.3-yr modulations. This includes $M_0$. The resulting 9.3-yr tide is extremely tiny, two orders of magnitude smaller than the 18.6-yr tide. In fact, all 9.3-yr terms in Table B1 are very small. It seems, therefore, unlikely that any tidal effects could be detected in the ocean at this period, with the possible exception of the case mentioned at the end of section 5, which depends on a certain ratio of diurnal/semi diurnal tide and must thus be rare. Attributing tidal influences as the cause of observed signals with periods near 9.3 yr (e.g., Berger and von Rad 2002) should be considered with some suspicion.

A final point concerning the nodal regression concerns terminology. In the recent climate literature any tidal effects of period 18.6 yr are commonly said to arise from “the luni-solar oscillation.” The term is to be discouraged. The tidal forces generated by the motion of the moon’s node are all lunar in origin and all are computed from $V_{ni}$; nodal effects in $V_n$ are negligible. The only solar involvement arises because it is primarily the sun’s torque causing the orbit precession. But this is only one of a very large number of perturbations that the sun induces in the moon’s motion (Cook 1988).
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