North Atlantic sea-level and circulation

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Accepted 1986 January 16. Received 1986 January 9; in original form 1985 September 5

Summary. Monthly sea-levels from an extensive array of North Atlantic tide gauges (26°–50°N) are examined. The spatial scale of the sea-level variations, and the reasons for them, are discussed; one application of such a study is clearly in the design of a tide gauge network for monitoring eustatic changes of sea-level.

The spatial scale of the sea-level changes is large. There is a coherent sea-level signal which can be traced along the eastern boundary of the North Atlantic from Newlyn (50°N) to Tenerife (28°N). There are also two distinct groupings of tide gauges along the western boundary, separated by Cape Hatteras.

The contribution of local air pressure and wind stress is quantified at each gauge through multiple regression techniques and the gains are then interpreted in terms of recent theoretical and numerical modelling studies. For example, the gains suggest that the wind-forced boundary current along the Nova Scotian shelf is trapped to within about 16 km of the coast.

The influence of local meteorology cannot account for the large-scale modes of variability. The coherent signal along the eastern boundary is correlated with changes in the Sverdrup transport of the North Atlantic and hence the large-scale wind field. The two modes on the western boundary appear to be related to baroclinic boundary current variations.

The Newlyn sea-level record is finally ‘corrected’ for some of the above effects to illustrate the utility of such a residual series in the identification of eustatic changes and vertical crustal movement.

Key words: sea-level, circulation, climate, eustasy

1 Introduction

Interest in the rate of rise of global sea-level has been stimulated recently by predictions of a change in air temperature associated with the increasing concentration of atmospheric CO₂. Tide gauge records are a unique source of information concerning the sea-level rise this
century (e.g. Barnett 1983a), mainly because of their length and accuracy. Rossiter (1972) has shown that, given good datum control, an annual mean can be considered accurate to about 1 mm, certainly less than the variability due to meteorological forcing, steric changes etc. Hicks (1978) for example has shown that the standard deviation of detrended annual means ($\sigma_a$) is typically 3 cm along the western boundary of the North Atlantic. This implies that at least 35 years of data are required to determine a trend of 1 mm yr\(^{-1}\), with 95 per cent confidence, along this boundary. If $\sigma_a$ could be halved, by removing the effect of local wind etc., only 22 years of data would be required to detect this trend with the same degree of confidence. Clearly this procedure could be applied to short records from 'data poor' regions and so allow them to contribute usefully to the global picture of sea-level rise. Apart from improved linear trend estimates, an accelerating sea-level rise could also be more readily detected in the 'corrected' series.

Rossiter (1967) was one of the first to correct annual sea-levels in his study of secular trends on the north-west European Shelf. (See Lisitzin (1974) for an historical review of this topic.) Rossiter used combinations of air pressure stations to represent implicitly the joint effect of air pressure and wind forcing over shelf seas and the North Atlantic. Multiple regression techniques are also employed in this paper to model meteorological and density effects on North Atlantic monthly sea-level. One major difference between this study and Rossiter's is in the choice of 'independent' variables for the regression analysis: only those variables which correspond to a direct physical influence (e.g. local wind stress, wind-forced ocean circulation) have been included. The advantages are two-fold. Firstly it is possible to obtain useful oceanographic information on shelf/ocean circulation from the present regression models. Secondly, it is important to know what is being removed when forming the corrected series; many geophysical signals are dominated by low-frequency variations which could be blindly interpreted as the cause of the sea-level trend without a genuine physical connection.

In Section 2 the main features of North Atlantic sea-level variability are described. Particular attention is paid to the spatial scale of the changes along meridional boundaries and, for the first time, across the North Atlantic. It will be shown that, in contrast to the Pacific (e.g. Enfield & Allen 1980), the cross-correlation functions generally peak at zero lag and that an empirical orthogonal function analysis based on the correlation matrix is appropriate for a first-order description. The relevant forcing functions are discussed in Section 3 and a multiple regression analysis, with physical interpretation, is given in Section 4. The influence of both local (shelf) and basin-wide winds is quantified and the role of baroclinic boundary currents in producing sea-level changes along the western boundary is discussed. Finally, in Section 5, one of the long annual series (Newlyn) is corrected for some of the above influences to illustrate the usefulness of such residual series in identifying changes due to eustasy and vertical crustal movement.

2 Observed sea-level variability

Monthly sea-levels from 30 tide gauges, located on both the eastern and western boundaries of the North Atlantic, are analysed in this study. All the monthly data were recorded during the period 1950–75 although some stations had missing values. The latitude and longitude of each tide gauge, and the number of available monthly means, are given in Table 1. All data were obtained from the Permanent Service for Mean Sea-level.

2.1 Space scales

If attention is focused on the annual period then it is obvious that the spatial scale of sea-level changes will be truly global. A more interesting question is the scale of the aperiodic
Table 1. Positions of the tide gauges and number of monthly means used in this study. (All data were for the period 1950–75. The exact coverage can be obtained from publications of the Permanent Service for Mean Sea-level.) The next two columns list the standard deviations of deseasonalized monthly sea-level \( \sigma_1 \) and residual from the multiple regression model \( \sigma_2 \), both for the common period 1961–70 where data coverage permits. The final four columns list the amplitude and phase of the annual cycle in the observed sea-level and after correction for the effect of local air pressure and wind (i.e. \( c_{11}, c_{12} \) from (8)).

### North of Cape Hatteras

<table>
<thead>
<tr>
<th>STATION</th>
<th>LAT (°N)</th>
<th>LONG (°W)</th>
<th>MONTHS</th>
<th>( \sigma_1 ) (cm)</th>
<th>( \sigma_2 ) (cm)</th>
<th>H (cm)</th>
<th>g (cm)</th>
<th>e (_1)</th>
<th>e (_2)</th>
<th>e (_3)</th>
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### South of Cape Hatteras

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<th>LAT (°N)</th>
<th>LONG (°W)</th>
<th>MONTHS</th>
<th>( \sigma_1 ) (cm)</th>
<th>( \sigma_2 ) (cm)</th>
<th>H (cm)</th>
<th>g (cm)</th>
<th>e (_1)</th>
<th>e (_2)</th>
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### Eastern boundary

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<th>LONG (°W)</th>
<th>MONTHS</th>
<th>( \sigma_1 ) (cm)</th>
<th>( \sigma_2 ) (cm)</th>
<th>H (cm)</th>
<th>g (cm)</th>
<th>e (_1)</th>
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<td>219</td>
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<td>(4.5, -118)</td>
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Changes of monthly sea-level. To examine this, the seasonal oscillation was first removed from the 21 series that were continuous over the common period 1961–70 (Table 1). The deseasonalized series were then subjected to an empirical orthogonal function (EOF) analysis based on the correlation matrix. The first three eigenvectors or modes, \( e_i, i = 1, 3 \), account for 79 per cent of the total variance. Rather than present the eigenvectors, which are generally difficult to interpret for \( i > 1 \) due to the orthogonality constraint (i.e. \( e_i \cdot e_j = 0 \) for \( i \neq j \)), scatterplots of \( e_1 \) against \( e_2 \), and \( e_2 \) against \( e_3 \), are shown in Fig. 1. This display clearly shows that there are three natural groupings of North Atlantic tide gauges. One group corresponds to the eastern boundary gauges; the remaining two groups are to be found on the western boundary and are separated by Cape Hatteras. The split at Cape Hatteras immediately suggests that the Gulf Stream may play a role in defining the groupings.
Figure 1. Scatterplots of the first three modes ($e_i$, $i = 1, 3$) from an empirical orthogonal function analysis of 21 North Atlantic sea-level series, 1961–70. (The gauges are indicated in Table 1 by an entry for $e_1$ and $e_2$. Each record was deseasonalized prior to the correlation-based EOF analysis.) The first three modes account for 48, 19 and 12 per cent of the total standardized variance.

To examine the similarity of the sea-level variations within each of the three groups, squared coherency ($K^2$) and phase spectra were calculated for pairs of long series taken from the same group. All three coherency spectra (Fig. 2) were significantly different from zero at the 95 per cent level across the whole frequency band (0–0.5 cpm). The highest coherence was found within the South Atlantic Bight (Mayport–Charleston), particularly at the annual period and its harmonics. North of Cape Hatteras (Boston–Lewes) the highest coherence was at 6 months and the lowest frequencies where $K^2 \sim 0.8$. There was no evidence in the

Figure 2. Squared coherency spectra of monthly mean sea-level. The top row shows coherencies between groups based on representative tide gauges, i.e. eastern boundary (Newlyn), north of Hatteras (Boston) and south of Hatteras (Charleston), all for the same period 1950–75. The bottom row shows coherency within each of the groups, i.e. eastern boundary (Newlyn–Coruna, 1957–75), south of Hatteras (Charleston–Mayport, 1950–74) and north of Hatteras (Boston–Lewes, 1953–73). The horizontal line is the 95 per cent significance level for zero coherence.
phase spectra or cross-correlation functions (not shown, but see Thompson 1981) of the type of non-dispersive wave propagation found by Enfield & Allen (1980) in their sea-level analysis for the eastern boundary of the Pacific. The maximum correlations were at zero lag in contrast to the Enfield & Allen (1980) analysis which gave lags of ~2 month between the tropics and mid- to high latitudes. The lack of a well-defined lag structure justifies, to some extent, the present use of an EOF analysis based on the correlation matrix rather than cross-spectral matrix.

The sea-level signals associated with each of the three groups are not necessarily uncorrelated. In fact $e_1$ spans Cape Hatteras (Fig. 1). To examine the between group similarity in more detail, coherency spectra for representative stations from different groups are shown in Fig. 2. As expected, coherence at the annual period is significant between groups. One notable feature in Fig. 2 is the high coherence between Boston and Charleston at periods greater than 6 yr. These low-frequency sea-level variations are coherent along the whole western boundary of the North Atlantic and presumably account for the first mode spanning Cape Hatteras ($e_1$, Fig. 1). In summary, both the EOF and spectral analysis indicate that the trans-Atlantic coherences are generally weak (apart from the annual cycle); the strongest coherences are found along the meridional boundaries of the North Atlantic.

2.2 POWER SPECTRA

Power spectra of long representative series from each of the three groups are shown in Fig. 3. The Charleston series is clearly the most energetic, partly due to the high energy levels at the annual period and its harmonics. The annual cycle at Boston is clearly much smaller. This difference is part of a well-defined, poleward decrease in the amplitude of the annual cycle of sea-level along the western boundary of the North Atlantic (Table 1). We will return to the reasons for this in Section 4. The power spectrum for Newlyn also has a well-defined peak at the annual period corresponding to an amplitude of about 6 cm (Table 1). The annual cycle of sea-level is somewhat variable along the eastern boundary (Table 1).

Ignoring the annual period and its harmonics, it is clear from Fig. 3 that the Charleston record is more energetic than the Boston record. Again this is part of a general poleward
decrease of variance along the western boundary (Table 1); the standard deviation of sea-
level, after removal of the seasonal variation (Jan' = Jan − Jan etc.), is ~ 6 cm in the South
Atlantic Bight compared to ~4 cm north of Hatteras. This result is in sharp contrast to the
eastern boundary where the variance increases polewards (Table 1).

3 Forcing functions

A very brief description of some of the more important influences on sea-level is given below
in order to motivate the regression analysis of Section 4 and help in its physical
interpretation.

3.1 AIR PRESSURE

The well-known ‘inverse barometer’ law relates local air pressure \( p_a \) and sea level \( \eta \) and is
given by

\[
\rho g \eta = \bar{p}_a - p_a,
\]

where \( \bar{p}_a \) is the average pressure over the world’s oceans. Pattullo et al. (1955) have shown
that \( \bar{p}_a \) has a surprisingly large mean annual range of 2.1 mb and included it in their study of
the seasonal oscillation of sea-level. It is unlikely however that \( \bar{p}_a \) has a significant trend and
its effect can probably be ignored on a decadal time-scale. [Fig. 3 of Bunger (1980) for
example shows that the average air pressure over the Atlantic (1948–72, 40°S–70°N) has a
trend of only 0.01 mb yr\(^{-1}\).] The time taken for a shelf sea to adjust to \( p_a \) changes is
complicated by stratification, shelf waves, etc. However the spindown time under bottom
friction is probably a determining factor on most tidally energetic shelves and this leads to a
response time of several days. In the following regression analysis it is assumed that (1) is
valid for monthly means and the sea-levels have been adjusted accordingly.

3.2 WIND STRESS

Both observation and theory confirm that wind stress can have an important effect on
coastal sea-level. Csanady (1982) discusses some simple analytic models for the steady
response of idealized shelf seas to wind forcing. For example the coastal sea-level response of
a wedge shelf sea to an longshore wind stress \( \tau^y \) can be expressed in the form

\[
\frac{\rho g \eta}{\tau^y} = \frac{fL}{r},
\]

where \( L \) is the cross-shelf scale of the wind-driven boundary current. This scale is a function
of the Coriolis parameter \( (f) \), linear bottom friction coefficient \( (r) \), bottom slope (assumed
uniform) and the spatial structure of \( \tau^y \). The time taken to achieve a steady state is again
complicated by stratification and shelf waves but it is generally much less than the present
averaging period of one month. [Wright et al. (1986) calculated an \( \epsilon \)-folding time of 20 hr
for the spin-up of their barotropic model of the Gulf of Maine.] For the purposes of this
study the response has been assumed quasi-steady on a monthly time-scale and the empiri-
cally determined gains of sea-level on longshore stress have been used to obtain estimates of
\( L \). This is described in the next section.
3.3 WIND-FORCED OCEAN CIRCULATION

Recent theoretical and numerical modelling studies show the initial barotropic response of a stratified, mid-latitude ocean to a change in surface stress is quasi-steady within about one month (e.g. Anderson et al. 1979). Away from the western boundary this response can be approximated by the bottom modified Sverdrup relationship

\[ J(\psi, f/h) = k \cdot \nabla \times (\tau/\rho h), \]  

where \( \psi \) is the stream function, \( h \) is the depth and \( J \) denotes a Jacobian. The associated sea-level slopes are given by

\[ gJ(h f/\eta) = w_e, \]  

where \( w_e \) is the Ekman pumping, i.e. \( k \cdot \nabla \times (\tau/\rho f) \). The sea surface topography can be determined, up to an arbitrary constant, by integrating (4) along \( f/h \) contours from the eastern boundary. To calculate the arbitrary constant of integration, the ocean is assumed closed and conservation of mass is applied, i.e.

\[ \bar{\eta} = 0. \]  

where overbar denotes a basin-wide average and \( \eta \) is measured relative to the undisturbed level. (Note that in (5) we ignore the relatively small contribution from the western boundary current region.) The longshore momentum equation at the eastern boundary, (4) and (5) then give the interior change in sea-level. If we assume for simplicity that \( h \) is constant and that wind set-up is negligible along the eastern boundary, then the sea-level along this boundary is also constant and given by

\[ \eta^e = \frac{f^2}{gh^2} (x w_e), \]  

where \( x = 0 \) on the western boundary and increases eastward. This is the response of eastern boundary sea-level to wind-forced circulation changes in this closed basin, subject to the above assumptions.

If all the dissipation is assumed to occur in a narrow western boundary current, then the sea-level head in deep water along the western boundary can be shown to be

\[ \frac{\partial \eta^w}{\partial y} = \frac{\tau^y}{\rho gh} + \int_0^w k \cdot \nabla \times \left( \frac{\tau}{\rho gh} \right) dx. \]  

Again \( h \) is assumed constant but the results can be readily generalized to include bottom topography. Note (7) is independent of the form of dissipation.

How big are the changes predicted by (6) and (7)? Clearly the results depend on \( h \). If variations in \( \tau \) are slow compared to the time taken for a baroclinic Rossby wave to cross the mid-latitude ocean (i.e. decades) then we can approximate \( h \) by the mean thermocline depth and bottom topography plays no part (Anderson et al. 1979). For periods much less than this (i.e. months), the appropriate \( h \) is the ocean depth. The sea-level response therefore depends on the frequency of wind forcing. Power spectra of \( w_e \) were calculated for \((55^\circ N, 35^\circ W)\) and \((35^\circ N, 35^\circ W)\) using monthly values determined in the manner outlined by Thompson & Hazen (1983). The spectra were white, apart from an annual peak. The standard deviations of the monthly changes were 15 and 8 \( \times 10^{-7} \) m s\(^{-1}\) at \(55^\circ N\) and \(35^\circ N\) respectively. Thus assuming typical values of \( w_e = 5 \times 10^{-7} \) m s\(^{-1}\) and \( W = 5000 \) km, then \( \eta^e = 2 \) cm \((h = 4 \) km, initial barotropic\) and 8 cm \((h = 1 \) km, final baroclinic\). Thus we
Figure 4. Difference in the summer (July–September) and spring (April–June) sea surface topographies calculated by Csanady (1979, figs 4 and 7) from hydrographic data under the assumption that the along-isobath geostrophic velocity at the bottom is zero.

would not expect the large-scale wind field to contribute much to the monthly $\eta^e$ variability but it could become increasingly important on longer time-scales. The corresponding sea surface slopes along the western boundary are $0.6 \times 10^{-8}$ (initial barotropic) and $2.4 \times 10^{-8}$ (final baroclinic) if we take the same values for $W$ and $w_e$ and assume $\tau^r$ is 0.1 Pa.

3.4 THERMOHALINE CHANGES

Fluctuations in the heat and salt content of the top 200 m of the ocean are responsible for a pronounced seasonal oscillation in sea-level (Gill & Nüiler 1973; Pattullo et al. 1955). The amplitude in the deep water adjacent to the Mid-Atlantic Bight and Scotian Shelf is about 8 cm. Csanady (1979) has extrapolated the deep ocean steric field to the coast of North America under the assumption that the geostrophic velocity is zero at the seafloor. His topography for spring shows a strong surface (geostrophic) current which follows the 1000 m isobath; the difference between summer and spring shows that this current has a strong seasonal variation. This difference in sea-level (Fig. 4) is consistent with Gill & Nüiler (1973) and Pattullo et al. (1955) in deep water but suggests that the deep ocean amplitude (~10 cm) is strongly attenuated at the coast (~2 cm mid-Atlantic Bight, 0 cm Gulf of Maine).

The influence of thermohaline changes is not limited to the annual period or top 200 m. Recently Roemmich & Wunsch (1984) have identified decadal changes in the large-scale temperature field of the North Atlantic. The observed warming of the ocean between 700 and 3000 m, across 24°N and 36°N, would result in a thermal expansion of several centimetres. Roemmich (1984) has shown that Bermuda sea-level does indeed reflect such changes in the density field. On a larger spatial scale, Barnett (1983b) has examined slow changes of dynamic height in the major oceans (0–1000 m, early 1900s to date) but could not find a significant global trend.

In the next section it will be shown that there is significant interannual variability in the
salinity of the shelf waters along the western boundary of the North Atlantic. This can also have a significant effect on the low frequency sea-level changes along this boundary.

4 Multiple regression analysis

Multiple regression techniques are used here to help explain some of the observed features of sea-level variability in terms of the forcing functions of Section 4.

The following model has been fitted to each series

$$\rho g \eta + p_a = a_T x + b_T y + c_{11} \cos (\omega_1 t) + c_{12} \sin (\omega_1 t) + c_{21} \cos (\omega_2 t) + c_{22} \sin (\omega_2 t) + \epsilon, \quad (8)$$

where $$\omega_1 = 2\pi/12 \text{ month}$$ and $$\omega_2 = 2\pi/6 \text{ month}$$.

Note that the influence of air pressure has been assumed to follow (1) and has been removed by adding the local air pressure and sea-level to obtain the total pressure ($$\rho g \eta + p_a$$). Local air pressure could of course be included as a forcing term in (8). However the spatial scale of monthly pressure patterns is large and $$p_a$$ could alias the influence of more distant winds. This would complicate our physical interpretation of the model’s coefficients. The influence of local wind stress ($$\tau^x, \tau^y$$) has been modelled by $$(a_T x + b_T y)$$ where $$(a, b)$$ are regression coefficients to be determined from the data. This form assumes a quasi-steady response to monthly mean winds (no lags) and a sinusoidal dependence on wind direction, i.e. the direction of maximum sea-level response is given by $$\tan^{-1}(b/a)$$ with zero response to winds perpendicular to this direction. [In contrast to the results of Noble & Butman (1979) there was no evidence in the monthly sea-levels of an asymmetrical dependence on the direction of wind forcing.] The periodic terms in (8) represent the component of the seasonal cycle which is not forced by local wind stress or air pressure. The periodic terms thus model the influence of seasonal density changes both on and off the continental shelf. Unfortunately there were insufficient hydrographic data to include such density effects explicitly in (8). All the air pressures, wind stresses and Ekman upwellings used in the regression analysis were estimated in the manner outlined by Thompson & Hazen (1983).

4.1 Eastern Boundary

There is insufficient space in this paper for a detailed discussion of all of the wind gains and seasonal cycles along the eastern boundary. In summary the wind gains for the NW European continental shelf were in reasonable agreement with the numerical modelling study of Pingree & Griffiths (1980) and a higher frequency study of the Newlyn record by Pugh & Thompson (1986). Along the Portuguese coast the maximum sea-level response was for a northward wind stress and the gains were consistent with a coastally trapped boundary layer of about 20 km width. [A more detailed description of the gains and seasonal cycles along the Portuguese coast, and in the Western Mediterranean, can be found in a study of the seasonal variation of flow through the Strait of Gibraltar by Bormans, Garrett & Thompson (1986).]

4.1.1 Seasonal cycle

Correcting the annual cycle of sea-level for the effect of local air pressure and wind with (8) reduced the scatter in the amplitudes and phases along the eastern boundary (Table 1). The mean amplitude (~4 cm, see Table 1) is close to the amplitude of the steric oscillation in this region (Gill & Niler 1973). There is however a significant difference in the phase of the
coastal sea-level and steric height. This could be due to a nearshore distortion of the density field which was not resolved by the coarse resolution of Gill & Niler (1973).

4.1.2 Residuals

The power spectrum of the representative series for the eastern boundary (Newlyn) is shown in Fig. 3. The removal of air pressure and wind effects has significantly reduced the variance of the record, particularly at the higher frequencies. Their combined effect can also account, in large part, for the poleward increase in variance along the eastern boundary (Table 1).

EOF analysis of the residuals from the regression analysis showed that the tide gauge groupings remained basically unchanged after removing the seasonal cycle and the influence of local meteorology, i.e. there remained one group for the eastern boundary of the North Atlantic and two for the western boundary, separated by Cape Hatteras. To illustrate, coherency spectra are shown in Fig. 5 for stations pairs from the same and different groups. (The format of Fig. 5 and station pairs are identical to the sea-level coherency spectra shown in Fig. 2.) The only significant coherence between Newlyn–Charleston is at harmonics of the annual cycle, i.e. 4 and 3 month. The highest coherence between Newlyn and Boston is at 14.7 month, the period of the pole tide. Apart from these periods, coherences across the North Atlantic are generally insignificant. The coherence between Newlyn and Coruna (same group) is reduced on removal of the effect of local air pressure and wind but remains significant at most frequencies.

In an attempt to explain the eastern boundary signal, a time series of $\eta^e$ was calculated using (6) and the 3-month mean Ekman upwelling fields of Thompson & Hazen (1983). (The ocean was assumed closed at 30°N, 60°N; the depth was taken to be 4 km.) The coherence and gain between $\eta^e$ and the Newlyn residuals are shown in Fig. 6. Encouragingly the gain increases with decreasing frequency in accord with the discussion of Section 3. The coherence also generally increases with decreasing frequency; the slight reduction at the lowest frequencies may be due to the quasi-linear trend in the Newlyn record which is associated with eustatic changes and land movement (Rossiter 1967).

![Figure 5](https://academic.oup.com/gji/article-abstract/87/1/15/729553)

Figure 5. As for Fig. 2 but for the residuals ($e$) from the multiple regression model, (8).
Thus it appears that the wind-driven circulation of the North Atlantic does influence sea-level along the eastern boundary. Note that the gain (Fig. 6) will transform a white \( \eta^e \) spectrum into a red \( \eta^e \) spectrum and so allow the meteorology to make a significant contribution to the interannual changes of sea-level. This point is discussed further in Section 5.

4.2 Western Boundary

Blaha (1984) has recently presented a thorough analysis of the monthly sea-level variability observed in the South Atlantic Bight. The following discussion will therefore focus on the stations north of Cape Hatteras.

4.2.1 Local wind effect

Wind gains from the regression model \((a, b)\) are shown in Fig. 7. To illustrate the type of information that can be extracted from this figure, consider the Scotian Shelf which is relatively straight and to which Csanady's idealized models are relevant. The longshore gain at Halifax implies that the cross-shelf scale of the wind-forced coastal boundary current is about 16 km \( (L, \text{see (2)}) \). A typical value of \( 5 \times 10^{-4} \) m s\(^{-1} \) was used for \( r \) (see Csanady 1982). This width is in reasonable agreement with the value of 23 km obtained from Csanady's 'box-car' forcing model (Csanady 1982, equation 6.53) if we assume (i) the same \( r \) as above, (ii) the longshore wind forcing starts at the deep Laurentian Channel, the natural 'upstream' boundary, (iii) the bottom slope is \( 5 \times 10^{-3} \), a value representative of the inshore bottom topography felt by the boundary current. (The inflow/outflow which feeds the frictional boundary layer is simply the Ekman flux in this model for longshore wind forcing.)

The gains for the Gulf of Maine have a stronger onshore component than the Scotian Shelf, presumably the result of enhanced wind set-up in this wide, semi-enclosed sea. Results
Figure 7. Wind gains (a, b) of sea-level on local wind stress from (8). The tide gauge positions are given in Table 1. Several of the gains have been omitted to avoid cluttering but they conform to the overall pattern. All the gains are significantly different from zero at the 95 per cent level except Miami.

from a recent numerical modelling study of the Gulf of Maine agree favourably with Fig. 7 [see Wright et al. (1986) for a detailed comparison.]

4.2.2 Seasonal cycle

The annual cycle of sea-level which is not forced by local wind or air pressure is given in Table 1. The amplitude and phase are less scattered than those for the observed sea-levels. There is a September maximum in both the mid-Atlantic Bight and Gulf of Maine. The amplitude is about 4 cm in the mid-Atlantic Bight and about 2 cm in the Gulf of Maine. These weak seasonal cycles are in favourable agreement with the change in coastal sea-level, from spring to summer, predicted by Csanady (1979), i.e. 2 cm in the mid-Atlantic Bight and 0 cm in the Gulf of Maine (Fig. 4). Both Csanady’s results and this analysis agree on a small amplitude at the coast. Thus our sea-level data provide some confirmation of the upper slope current which was estimated by Csanady under the major assumption that the geostrophic bottom velocity was zero. The westward attenuation and phase propagation of the annual cycle along the Scotian Shelf (Halifax–Yarmouth, see Table 1) are consonant with the seasonal freshwater discharge from the Gulf of St Lawrence (Drinkwater, Petrie & Sutcliffe 1979). The maximum westward flow in winter would correspond (geostrophically) to an increased coastal sea-level as observed. (The influence of the freshwater discharge is evident in Csanady’s difference topography shown in Fig. 4.)

4.2.3 Residuals

Power spectra of residuals from the representative tide gauges are shown in Fig. 3. It is clear that the contribution of local wind and air pressure is small at low frequencies. One notable feature in the Boston residual spectrum is the small peak at the pole tide frequency. Miller & Wunsch (1973) also detected a weak pole tide in the Boston monthly sea-level record but did
not attempt to reduce the background noise by removing the variations coherent with the meteorology. This analysis suggests that such a procedure would significantly improve the chances of detecting such a small signal.

The standard deviations of the residuals ($\sigma_2$, Table 1) show that the most energetic stations are still in the South Atlantic Bight, even though the influence of local meteorology has been removed ($\sigma_2 \sim 5$ cm). Further north, $\sigma_2 \sim 3-4$ cm. The proportion of sea-level variance which can be accounted for by local air pressure and wind ($1 - \sigma_2^2/\sigma_1^2$) is everywhere less than 42 per cent along the western boundary, in contrast to 80 per cent at Newlyn on the eastern boundary (Table 1). Clearly some important influence has been omitted from the multiple regression model.

Coherency spectra for the western boundary gauges are shown in Fig. 5. Coherences within the same group are reduced but generally remain significant after removal of air pressure and wind effects. Coherence between the two groups also remains high at the lowest frequencies showing that the definition of the groups still breaks down at these frequencies.

What causes these large-scale residual variations along the western boundary? Given the EOF split at Cape Hatteras and the enhanced residual variance in the South Atlantic Bight,

![Figure 8](https://academic.oup.com/gji/article-abstract/87/1/15/729553/8715729553)

Figure 8. The heavy, smooth line is the lowpassed monthly sea-level at the tide gauge indicated. (A Cartwright filter was used. It eliminated (passed unchanged) variations with periods less (greater) than 18 (124) month. Some missing monthly values for Wilmington and Kiptopeke Beach were interpolated but they did not influence the curves significantly.) The other curves are surface density variations recorded at the following nearby US lightships: from top to bottom, Boston, Ambrose, Chesapeake and Frying Pan Shoals. To highlight the aperiodic changes, the seasonal cycle has been removed from each density series; to facilitate comparison with the sea-level records, density increases with decreasing ordinate.
an obvious possibility is fluctuations in the Gulf Stream. I attempted to relate the residuals
to fluctuations in the Sverdrup transport across appropriate $f/h$ contours (i.e. transports
were calculated from (3) using North Atlantic bathymetry). No significant relationships
were found. I was also unable to explain the difference in Mid and South Atlantic Bight
sea-levels by the pressure head from (7). In short, no clear relationship could be found
between the residual variations along the western boundary and the large-scale wind field
over the North Atlantic. It is likely that fluctuations in the Gulf Stream do contribute to the
sea-level variability in the South Atlantic Bight, particularly as its influence has been so
clearly demonstrated on shorter time-scales (Maul et al. 1985). If this is so, then monthly
changes in the Gulf Stream do not appear to be forced by the bottom modified Sverdrup
transport of the North Atlantic or the local wind field.

It is also possible that the upper slope current may be contributing to the residual
variability north of Hatteras. The reasons are simply that (i) this current appears to have a
significant influence on the seasonal oscillation of coastal sea-level, and (ii) the position of
the shelf/slope boundary, which may reflect changes in the deeper density field and hence
the current, is known to vary on time-scales greater than a month (Smith & Petrie 1982).

Some statistically significant relationships were obtained between sea-level residuals and
shelf density. (Surface temperature and salinity data from US lightships were extracted from
the US Coast Guard Reports for the period, 1956–71.) Using data for the well-mixed time
of year (October–March), correlations between monthly sea-level and density were found to
be typically $-0.3$. This value is so low that it seems unlikely that local hydrographic changes
are the cause of the coherent monthly residual variations. It does appear however that
interannual variability in nearshore hydrography may well be contributing to the coherent,
low frequency sea-level variations along the western boundary. Low-passed sea-level
variations from four western boundary tide gauges are shown in Fig. 8 along with the density
variations recorded at nearby US lightships. It is clear that the surface density exhibits large-
scale, low frequency variations which are similar to the sea-levels. A rough gain would be
about $-10 \text{ cm (kg m}^{-3})^{-1}$. This is consistent with the gain predicted from the approach of
Csanyi (1979) if the density anomaly is assumed to be uniform through the water column
and to extend out to the 100 m isobath.

5 Secular changes of sea-level

Wind stress and air pressure are only small contributors to the interannual sea-level
variability along the western boundary of the North Atlantic (Fig. 3). Baroclinic boundary
currents may be important but their low frequency “cut-off” is not obvious at this stage.
Nearshore variations in density appear to be the cause of large ($\pm 5$ cm) sea-level changes on a
decal time-scale along the western boundary (Fig. 8). Clearly this effect warrants further
study if these long sea-level records are to be used for determining vertical crustal movement
and eustatic changes. In particular it would appear to be worthwhile, in a future analysis, to
collect a more comprehensive set of density measurements (including subsurface data) and
develop a diagnostic circulation model to remove the influence of changing shelf hydrography
from the coastal sea-level records.

Annual sea-level variations along the eastern boundary of the North Atlantic, from
Newlyn ($50^\circ$N) to Tenerife ($28^\circ$N), are shown in Fig. 9 for the overlap period 1915–75.
(All data were obtained from the publications of the Permanent Service for Mean Sea Level.
Only annual means based on averages of hourly heights, and referenced to a stable bench-
mark were used.) The three long series in Fig. 9 clearly show a gradual rise of sea-level over
the last 60 yr. The shorter records are dominated by interannual variability and the slow
Figure 9. Annual sea-level variations along the eastern boundary of the North Atlantic for the overlap period, 1915–75.

Figure 10. Annual mean sea-level at Newlyn, before and after removal of the effect of local $p_d$, $\tau$ and $\eta^e$ with a multiple regression model. The linear trends (± standard error) are $1.0 \pm 0.5$ and $1.4 \pm 0.2$ mm yr$^{-1}$ before and after correction respectively.
upward trend is more difficult to define. However it is clear that the interannual variations are coherent over large space scales and thus perhaps removable. The analysis of Section 4 shows that both local and basin-wide winds exert a significant influence on low frequency changes of sea-level along the eastern boundary. To illustrate how improved trend estimates can be obtained from short sea-level records the combined influence of local wind, air pressure and $\eta^*$ has been removed from the Newlyn record with a multiple regression model. (Annual means for the extended period 1950–80 were used.) The marked reduction in the variability about the trend is clear from Fig. 10. The trends in the observed and residual records are $1.0 \pm 0.5$ and $1.4 \pm 0.2$ mm yr$^{-1}$ respectively. The standard errors clearly indicate the increased confidence that can be placed in the latter estimate. Perhaps more important than the reduced standard error is that a change in the trend could be identified more readily in the residual, rather than observed, series (Fig. 10). There is no evidence for an increasing rate of rise in the Newlyn record.

6 Discussion

What has been learnt about the sea-level and circulation of the North Atlantic? Ekman pumping of the North Atlantic may be causing significant changes in sea-level along the eastern boundary. Clearly more work is required to check this hypothesis because of assumptions required to calculate the $\eta^*$ series, e.g. the closing of the ocean at 30°N. For example, Ekman pumping data for the North Atlantic could be compared with changes in the observed density field, perhaps using ventilated thermocline theory (Luyten, Podlosky & Stommel 1983). Sea-level differences between island stations (e.g. Bermuda) and the eastern boundary could also be compared with Ekman pumping.

Local wind stress is a significant contributor to the sea-level variance at all tide gauges except Miami. It cannot however account for the tide gauge groupings along the western boundary. Along the relatively uniform Scotian shelf the longshore wind gain implies that the coastal boundary current is trapped to within about 16 km of the coast. This value is in good agreement with the inclined beach model of Csanady (1982) if the longshore wind forcing is assumed to start (upstream) at the deep Laurentian Channel. In the Gulf of Maine the gains are in favourable agreement with a numerical modelling study (Wright et al. 1986). The most useful oceanographic application of such empirically determined gains is probably the provision of such checks on numerical and analytical models. Ideally the gains should be based on hourly data and made frequency dependent. This can lead to estimates of the spin-down time of shelf circulation (Garrett, Majaess & Toulany 1985) and possibly the influence of non-local winds.

Changes in the intensity of the Gulf Stream are known to dominate the seasonal oscillation of Miami sea-level; it is also probable that the Gulf Stream makes a significant contribution to the aperiodic variability at Miami and the South Atlantic Bight as a whole. If we assume that the coastal sea-levels are indeed a measure of the intensity of the Gulf Stream, then our failure to relate them to bottom modified Sverdrup transport requires some other forcing mechanism for monthly fluctuations in the Gulf Stream. North of Cape Hatteras, the seasonal oscillation of sea-level again indicates the importance of baroclinic current variations, specifically the upper slope current identified by Csanady (1979) from hydrographic data. Smith & Petrie (1982) have recently shown that the surface position of the shelf/slope boundary along the Scotian slope exhibits submonthly onshore translations which appear related to changes in the longshore current in deep water and at the shelf break. This encourages us to speculate that aperiodic variations in the upper slope current may make a significant contribution to the monthly sea-level variability north of Hatteras. A
comparison of coastal sea-level with some of the long current records now available for the shelf break and slope is required.

Long and precise tide gauge records will probably continue to play a key role in determining the rate of rise of global sea-level. One of the objectives of this paper has been to show how more reliable trends can be obtained by first removing meteorological effects from the tide gauge records. We have seen that meteorological forcing is relatively unimportant to the interannual changes of sea-level along the western boundary. Nearshore salinity variations appear to be particularly important on a decadal time-scale; further work in the development of an appropriate diagnostic model and hydrographic data set is required before their effect can be adequately removed from the coastal sea-level records. Fluctuations in the Gulf Stream, and perhaps the upper slope current, probably contribute to the sea-level variability but their low frequency cut-off is difficult to quantify at this stage. Thus in the absence of appropriate independent variables for the regression analysis it has not been possible to 'correct' the western boundary sea-level series. [Meade & Emery (1971) have shown that river discharge can account for only 7–13 per cent of the annual sea level variance.]

The variance of the Newlyn record (eastern boundary) was reduced significantly by removing the influence of local wind, air pressure and Ekman pumping over the North Atlantic. The standard error of the trend was halved from 0.4 to 0.2 mm yr\(^{-1}\). Perhaps more important than the reduced error bars on the trend is the increased possibility of detecting changes in the trend from the corrected series. Given the present concern about an accelerating rise of sea-level, and the shortage of long series from important areas of the globe, it appears worthwhile to consider developing similar regression models, and perhaps diagnostic circulation models, for other strategic locations. A note of caution should be introduced here; it is important to know what is being removed. Many geophysical series have red spectra and could be included in a regression analysis to "explain" the sea-level trend. Studies based on shorter averaging periods (1 month or shorter) are often useful in understanding exactly what is being modelled and ultimately removed from the sea-level record.

Acknowledgments

I would first like to thank the Permanent Service for Mean Sea-level for providing all the sea-level data used in this study. Both Chris Garrett and Adrian Gill made some useful suggestions during the course of this work for which I am grateful.

References


K. R. Thompson


Roemmich, D., 1984. Thermal variability of the ocean on long time scales and large space scales (abstract), *EOS*, 65, 831.


