RECENT RELATIVE SEA-LEVEL CHANGE IN
EASTERN NORTH AMERICA

by

BARBARA VANSTON BRAATZ
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Signature of
Author ______________________________ January 1987

Joint Program in Oceanography, Woods Hole Oceanographic
Institution/Massachusetts Institute of Technology

Certified by ______________________________ David G. Aubrey
Thesis Supervisor (WHOI)

Accepted
by ______________________________ William F. Brace
Chairman, Department Committee on Graduate Students

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Submitted to the Department of Earth, Atmospheric, and Planetary Sciences on 8 January 1987 in partial fulfillment of the requirements for the Degree of Master on Science in Earth, Atmospheric, and Planetary Sciences

ABSTRACT

Eigenanalysis of tide-gauge records between 1920 and 1983 in eastern North America reveals highly variable spatial and temporal patterns of relative sea-level change. Auxiliary data from numerical modeling suggest that much of the long wavelength (thousands of km) spatial patterns of sea-level change are due to post-glacial isostatic adjustment of the land surface. Filtering the isostatic component from the rates of relative sea-level movement yields residual rates that fluctuate about a coastal mean of 1.0-1.5 mm/yr during this 64-year time interval. This mean rate is within the range of previous estimates of the mean rate of eustatic rise in sea level during the past century. Some residual fluctuations (wavelengths of tens to hundreds of km) correlate with tilts of the land surface revealed by geodetic leveling transects, and appear to be related to regional geology (i.e., basement structures and tectonic provinces in Florida, Georgia, the Carolinas, and the Chesapeake Bay area; fault reactivation in northern New England and the Maritime Provinces). These results suggest that tide-gauge data can be used to determine neo-tectonic movements along this coastline. Analysis of the temporal patterns of relative sea-level change reveals a gradual increase in the rate of rise centered at about 1934, which may be due to steric expansion of the ocean. Broad peaks in the spectrum of temporal sea-level fluctuations at 3-, 6-, and 20-year periodicities (significant at the 80% level) may be a reflection of oceanographic, atmospheric, and astronomical forcing.

Thesis Supervisor: David G. Aubrey
Title: Associate Scientist
Woods Hole Oceanographic Institution

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INTRODUCTION

The CO_2 content of the atmosphere has risen approximately 20% since the Industrial Age (Hoffman et al., 1983). Most of this increase has been due to burning of fossil fuels and clearing of forested areas. Current projections estimate a doubling of pre-industrial age atmospheric CO_2 by 2050 to 2080 (Berger, 1984), and an associated increase in global average surface air temperature of 1.5-4.5°C (NRC, 1979) due to the "greenhouse" effect (tropospheric warming caused by increased absorption of long wave radiation emitted by the earth's surface). Such warming, by thermally expanding the oceans and melting continental and alpine glaciers, would cause a rapid eustatic (i.e., global, although not necessarily globally equivalent) sea-level rise. Estimates of this rise range from 38cm in the next century, to 5m in the next 40 years (Mercer, 1978; Hoffman et al., 1983; Revelle, 1983). These estimates, as well as estimates of other aspects of the atmospheric-biospheric-oceanic response to increased atmospheric CO_2 and other gases, are subject to much debate (e.g., Perry, 1983; Berger, 1984; Kondratyev, 1984). Adverse impacts of higher sea levels include shoreline retreat, erosion, flooding, destruction of coastal structures, and saltwater intrusion of bays, rivers, and aquifers.

To accurately predict future changes in sea level, we must assess the past response of the oceans to atmospheric change. If atmospheric CO_2 has increased by 20% since the Industrial Revolution, what has been the associated increase in eustatic sea level? Is this rate of sea-level rise increasing? Various estimates of this rate have been made (Table I). Agreement between the estimates is surprisingly good,

: Estimates of Mean Eustatic Sea-Level Increase (After Barnett, 1983)

	Rate (cm/century)	Method
nsson (1940)	>5	Cryologic Aspects (ice budget data from 6 glaciers, extrapolated for global
rg (1941)	11	Tide Gauges (69 stations, 1880's-~1939)[A]
(1950)	12-14	Tide Gauges (11 Dutch stations, 1832-1942)[B]
n (1958)	11.2±3.6	Tide Gauges (5 N. European & 1 Indian station, 1800's-1943)[C]
dge & Krebs (1962)	12	Tide Gauges (1900-1950)[D]
1980)	30	Tide Gauges (247 stations, ~1850-1979)[E]
et al. (1982)	12	Tide Gauges (193 stations, ~1900-1979)[F]
(1983)	15.1±1.5	Tide Gauges (9 stations, 1903-1969)[G]
(1984)	14.3±1.4	Tide Gauges (155 stations, ~1880-1980)[H]
& Emery (Aubrey,1985)	0-30	Tide Gauges[I]

ons in regions of known post-glacial uplift omitted

includes correction for "secular sinking of crust" determined from geologic, archaeologic, and leveling data

includes correction for secular relative sea-level movement

ons in regions of known post-glacial uplift, orogenic activity, and sediment compaction omitted

rate is based on values for continental stations which record relative sea-level rise, if the continental stations wh
ord a relative sea-level fall are also included, the resulting rate is ~15 cm/century

ons in regions of known post-glacial uplift, orogenic activity, and sediment compaction omitted; if corrections for s
ative sea level change, determined from ^{14}C dated shoreline indicators, are included, Gornitz et al. obtain a global
10 cm/century

station chosen as a "stable" regional representative to obtain equal-area weighted global average

equal-area weighting scheme, stations with anomalous vertical motion deleted

rates based on separate studies of different regions, over different intervals of time given the different data sets and methods of analysis. Results from some of these studies suggest that the rate of sea-level rise has increased during the past century (Thorarinsson, 1940; Gutenberg, 1941; Fairbridge and Krebs, 1962; Emery, 1980; Barnett, 1984), although the timing of this increase varies between studies. The most recent comparison of average global sea-level rise over different time periods (Barnett, 1984) revealed little or no trend between 1881-1920, but a positive trend between 1920-1980 (22.7 $\pm$ 2.3 cm per century for the period 1930-1980).

Tide-gauge data are complicated because coastlines along which sea level is measured do not maintain constant elevation. Factors such as isostatic adjustment due to glacial and hydrostatic loading and unloading, tectonic movements, sediment loading and flexure, pore-fluid removal, and sediment compaction affect relative sea levels with varying magnitudes over the globe. Oceanographic and atmospheric factors, such as the El Nino-Southern Oscillation, meanders and eddies of western boundary currents such as the Gulf Stream and the Kuroshio, and changes in wind stress and atmospheric pressure, also have varying effect in space and in time on sea-level elevation.

For these reasons, recent studies (Aubrey and Emery, 1983, 1986; Emery and Aubrey, 1985, 1986a,b) have examined relative sea-level (RSL) change on a regional basis. Results from these studies indicate that in some regions RSL is dominated by factors other than pure eustacy. In Japan, the spatial structure of RSL rise can be explained by subduction of the Pacific and Philippine plates beneath the Japanese Islands, and the higher frequencies of the temporal fluctuations in RSL probably are related to shifts in the position of the Kuroshio (Aubrey and Emery, 1986). Isostatic rebound clearly explains the concentric pattern of RSL lowering seen in the Fennoscandian region (Emery and Aubrey, 1985). RSL along the coast of the east Asian mainland appears to be controlled by geologic structure (Emery and Aubrey, 1986a). These studies demonstrate the variablity, both in time and in space, of the rate and magnitude of RSL change. To quantify the eustatic sea-level rise during the past century, the magnitude of these tectonic, isostatic, and oceanographic effects on RSL in different regions around the world must be determined.

Empirical orthogonal function analysis of tide-gauge data from the coastlines of the United States (Aubrey and Emery, 1983) revealed a highly variable spatial record of RSL change over the 40 year period 1940-1979, which is probably a result of a number of factors. Along the U.S. east coast, three distinct regions were identified, each with a consistent sea-level trend. From Pensacola to Cape Hatteras, rate of RSL rise increases; from Cape Hatteras to Boston the rate decreases; and from Boston to Eastport the rate increases again. Aubrey and Emery (1983) suggested isostatic rebound as an explanation for the trend in the most northern region, but found no tectonic or geologic reasons to explain the remainder of the east coast signal. A comparison between the mean annual RSL and depth-to-shelfbreak along the east coast was made with the hypothesis that if RSL trends had persisted for a long period of time, the trends would be reflected in the depth-to-shelfbreak. Lack of correlation between the two data sets led the authors to conclude that segmentation of RSL along the east coast may be due to oceanographic and atmospheric forcing.

A number of previous analyses of different data sets (geodetic leveling; archaeology; radiocarbon dating of terraces, beaches, and peat bogs; oceanographic temperature and salinity measurements) have demonstrated this spatial and temporal variability in rate of RSL change along the U.S. east coast (Wunsch, 1972; Winkler and Howard, 1977; Brown, 1978; Newman et al., 1980; Bloom, 1983; among many others). The abundance of these studies, as well as the good coverage by tide gauges along the U.S. east coast compared to other regions, justifies further analysis of RSL trends. The purpose of this study is to expand the U.S. east coast portion of Aubrey and Emery's (1983) analysis of U.S. tide-gauge records to include additional years of record and additional stations in the U.S. and Canada that were not previously analyzed, and then use related studies to determine how eustatic, isostatic, geologic, and oceanographic factors produce the spatial and temporal variability in RSL in eastern North America.

METHODS

Data for this study are yearly mean sea levels from tide-gauge stations, from 1920 through 1983, provided by the Permanent Service for Mean Sea Level (United Kingdom) and the National Ocean Survey (United States). Each sea-level series was plotted and visually inspected for discontinuities or jumps that could be the result of movement of the tide gauge or datum, or simply a data entry error. Stations located on large rivers also are suspect. Meade and Emery (1971) found that variations in annual river runoff can account for 7-21% of the total variation in sea level along the U.S. east coast. The entire record at Richmond, on the James River, was deleted for this reason. Stations having less than 20 years of data, from the period 1920-1983, inclusive, were also deleted, leaving a total of 44 stations to be analyzed (figure 1 and Table II). Aubrey and Emery (1983) analyzed 26 stations along the U.S. east coast over the 40-year time span between 1940 and 1979.

Empirical orthogonal function analysis, or eigenanalysis, was applied to sea-level records to determine the dominant temporal and spatial structure of the data. Advantages of eigenanalysis over averaging and linear or non-linear regression techniques are outlined in Aubrey and Emery (1983). Briefly, eigenanalysis separates a data set into orthogonal spatial and temporal modes that most efficiently describe the variability of the data set (see Appendix of Aubrey and Emery, 1986). No preconceived subjective models are fit to the data, as in regression techniques. Eigenanalysis usually is applied to overlapping time series (equal and concurrent duration), and any stations having gaps are excluded. In order to use all available data without

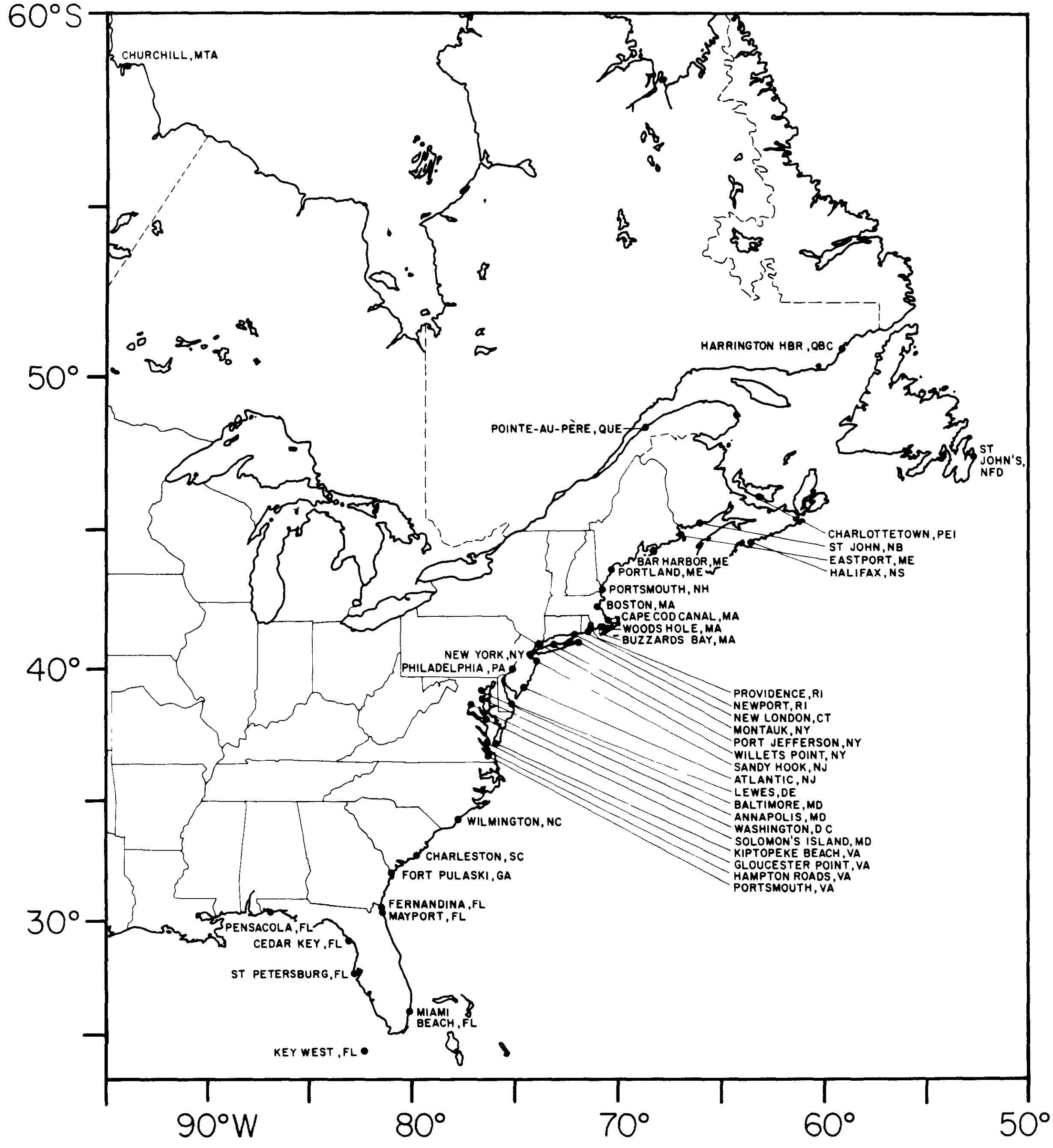


Fig. 1. Location map for tide-gauge stations.

TABLE II: Tide-gauge station locations; synthetic rates of RSL movement generated using eigenanalysis of tide-gauge data between 1920 and 1983; model estimates of isostatic adjustment from Peltier, 1986 (? = interpolation from contour map, otherwise directly from table, see text); residual rates of RSL movement equals synthetic rate minus model estimate.

STATION NAME	LOCATION (latitude, longitude)	SYNTHETIC RATE (mm/yr)	MODEL ESTIMATE (mm/yr)	RESIDUAL RATE (mm/yr)
Pensacola, FL	30°24.2'N, 87°12.8'W	2.0±.4	0.6?	1.4±.4
Cedar Key, FL	29°08.1'N, 83°01.9'W	1.2±.2	0.4?	0.8±.2
St. Petersburg, FL	27°46.4'N, 82°37.3'W	1.6±.3	0.2?	1.4±.3
Key West, FL	24°33.2'N, 81°48.5'W	1.9±.4	-0.1	2.0±.4
Miami Beach, FL	25°46.1'N, 80°07.9'W	1.6±.3	0.0	1.6±.3
Mayport, FL	30°23.6'N, 81°25.9'W	1.9±.4	0.3	1.6±.4
Fernandina, FL	30°40.3'N, 81°28.0'W	1.8±.3	0.4	1.4±.3
Fort Pulaski, GA	32°02.0'N, 80°54.1'W	2.1±.4	0.6	1.5±.4
Charleston, SC	32°46.9'N, 79°55.5'W	2.8±.5	0.8	2.0±.5
Wilmington, NC	34°13.6'N, 77°57.2'W	1.8±.3	1.2?	0.6±.3
Portsmouth, VA	36°49.3'N, 76°17.6'W	2.9±.5	1.7	1.2±.5
Hampton Roads, VA	36°56.8'N, 76°19.8'W	3.5±.7	1.8	1.7±.7
Gloucester Point, VA	37°14.8'N, 76°30.0'W	3.2±.6	1.8?	1.4±.6
Kiptopeke Beach, VA	37°10.0'N, 75°59.3'W	2.8±.5	1.8?	1.0±.5
Solomon's Island, MD	38°19.0'N, 76°27.2'W	2.4±.5	1.9	0.5±.5
Washington, D.C.	38°52.5'N, 77°01.4'W	2.6±.5	1.8	0.8±.5
Annapolis, MD	38°59.0'N, 76°28.8'W	2.6±.5	1.8	0.8±.5
Baltimore, MD	39°16.0'N, 76°34.7'W	2.8±.5	1.8	1.0±.5
Lewes, DE	38°46.9'N, 75°07.2'W	3.0±.6	2.1	0.9±.6
Atlantic City, NJ	39°21.3'N, 74°25.1'W	3.0±.6	2.1	0.9±.6
Philadelphia, PA	39°57.1'N, 75°08.4'W	2.6±.5	1.8	0.8±.5
Sandy Hook, NJ	40°28.0'N, 74°00.1'W	3.2±.6	1.7	1.5±.6
New York, NY	40°42.0'N, 74°05.5'W	2.7±.5	1.6	1.1±.5
Willets Point, NY	40°47.6'N, 73°46.9'W	2.2±.4	1.6	0.6±.4
Port Jefferson, NY	40°57.0'N, 73°04.6'W	2.8±.5	1.6?	1.2±.5
Montauk, NY	41°02.9'N, 71°57.6'W	1.7±.3	1.6?	0.1±.3
New London, CT	41°21.3'N, 72°05.2'W	1.8±.3	1.5	0.3±.3
Newport, RI	41°30.3'N, 71°19.6'W	2.1±.4	1.6	0.5±.4
Providence, RI	41°48.4'N, 71°24.1'W	1.9±.4	1.5	0.4±.4
Woods Hole, MA	41°31.5'N, 70°40.4'W	2.2±.4	1.7	0.5±.4
Buzzards Bay, MA	41°44.5'N, 70°37.1'W	0.8±.2	1.7?	-0.9±.2
Cape Cod Canal, MA	41°46.3'N, 70°30.4'W	1.8±.3	1.7?	0.1±.3
Boston, MA	42°21.3'N, 71°03.0'W	1.9±.4	1.2	0.7±.4
Portsmouth, NH	43°04.9'N, 70°44.7'W	1.6±.3	0.9	0.7±.3
Portland, ME	43°39.4'N, 70°14.8'W	2.2±.4	0.7	1.5±.4
Bar Harbor, ME	44°23.5'N, 68°12.3'W	2.0±.4	0.8?	1.2±.4
Eastport, ME	44°54.2'N, 66°59.1'W	2.2±.4	0.9	1.3±.4
St. John, NB	45°16.0'N, 66°04.0'W	2.1±.4	0.9?	1.2±.4
Halifax, NS	44°40.0'N, 63°35.0'W	2.6±.5	1.6?	1.0±.5
Pointe-au-Père, QUE	48°31.0'N, 68°28.0'W	0.5±.1	-2.2?	2.7±.1
Charlottetown, PEI	46°14.0'N, 63°70.0'W	1.4±.3	0.6?	0.8±.3
Harrington Hbr, QUE	50°30.0'N, 59°29.0'W	-0.1±.0	-1.4?	1.3±.0
St. John's, NFD	47°34.0'N, 52°43.0'W	1.1±.2	0.4?	0.7±.2
Churchill, MTA	58°46.0'N, 94°11.0'W	-2.2±.4	-10.0?	7.8±.4

interpolation or extrapolation, a modified eigenanalysis (Aubrey and Emery, 1986; Aubrey and Welch, in prep.) was employed. Station means were removed, and each station variance was set to unity before the eigenvalues and eigenvectors were calculated to minimize dominance by any single energetic sea-level station. Linear regressions of the 1920-1983 portions of each station's record were computed to determine anomalous yearly means and to compare with eigenanalysis results. RSL rates are presented as positive for RSL rise and negative for RSL fall. Finally, spectral analysis was used on the temporal eigenfunctions to determine the dominant frequencies of sea-level change.

RESULTS

Eigenanalysis of tide-gauge data yields both spatial and temporal functions ranked to explain sequentially diminishing amounts of the total variation in the records. Each spatial function has a corresponding temporal function which explains the same amount of the variance (figures 2 and 3). Most (81.2%) of the total sea-level variance is explained by the first three eigenfunctions (the first, second, and third functions account for 69.6%, 6.4%, and 5.2%, respectively, of the true variance of the data set).

An F-test was used to determine if the slopes of the temporal eigenfunctions were statistically significant at the 95% level. The first temporal function (figure 3) has a significant positive trend, of 0.0053/yr (in dimentionless units); the second function has no statistically significant trend; and the third function displays large amplitude fluctuations about a significant positive trend of 0.0084/yr. The first function shows a distinct change in slope, from a flat trend during the early portion of the time series, to a positive trend during the later portion. Based on the best fit (in a least-squares sense) of a two-segment line, this change in slope is centered at 1934/1935. Linear regression slopes for both the 1920-1934 (-0.0037 mm/yr) and for the 1935-1983 (0.0057) segments are significant at the 95% level. This change in slope of the first function could have been a result of either a low sampling density before 1934 or bias introduced by the eigenanalysis, so a comparison was made between the first temporal eigenfunction computed using only those stations in operation before 1935, and the first temporal eigenfunction computed using all of the

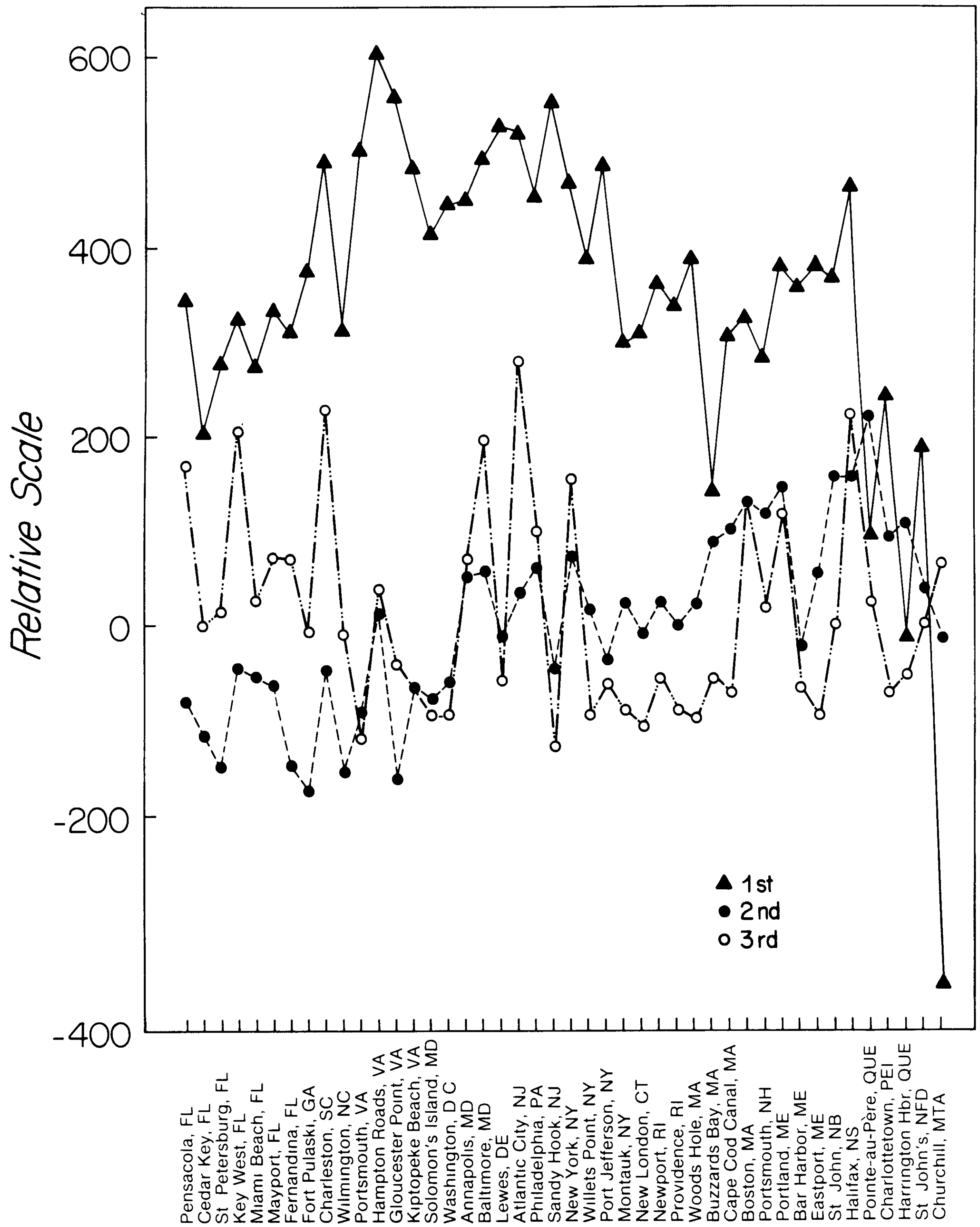


Fig. 2. First three spatial eigenfunctions for the 44 stations having records in excess of 20 years between 1920 and 1983, inclusive. These functions account for 69.6, 6.4, and 5.2%, respectively, of the true variance of the records.

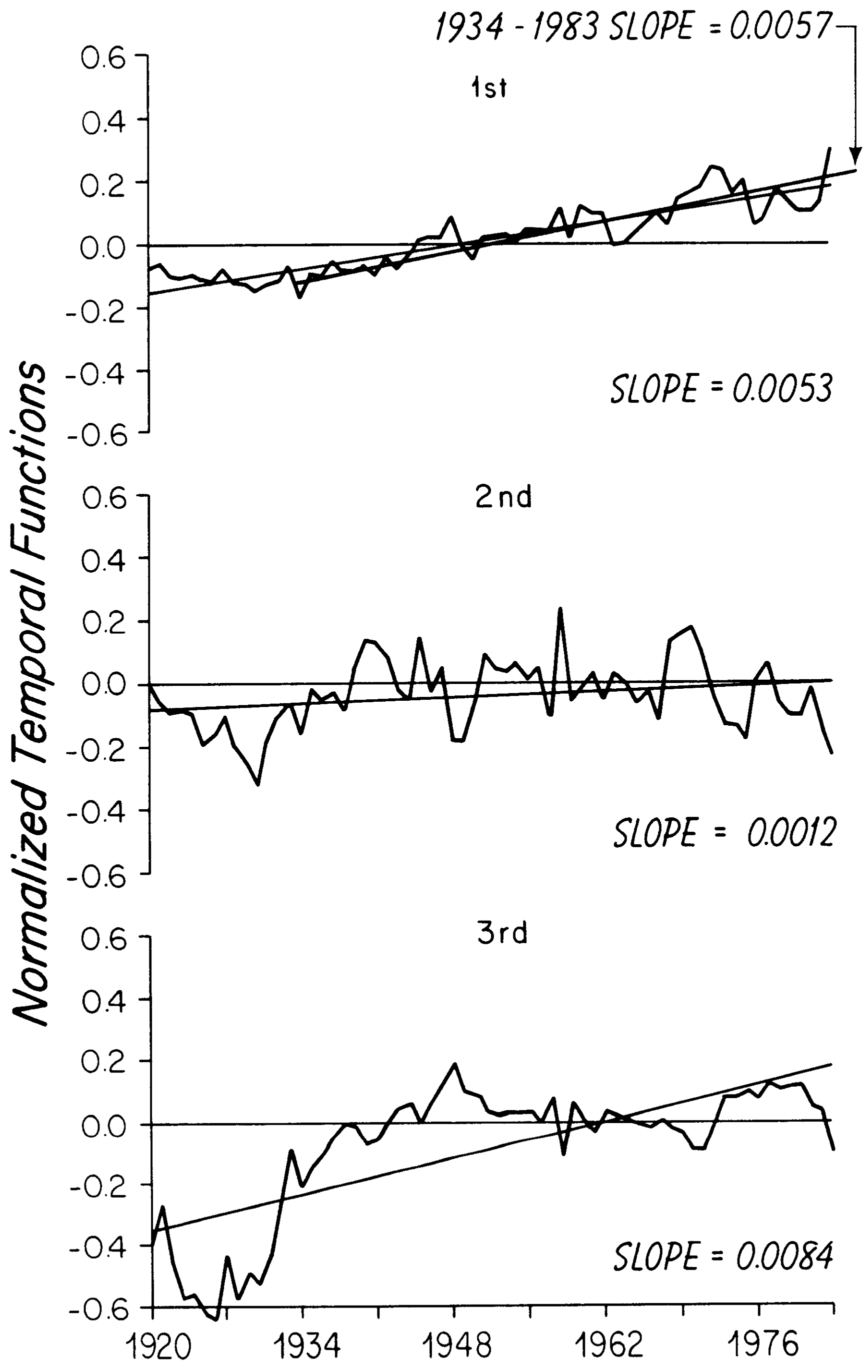


Fig. 3. First, second, and third temporal eigenfunctions. These functions account for 69.6, 6.4, and 5.2%, respectively, of the true variance of the records. Units are dimensionless and normalized. Based on a best-fit of a two-segment line, there is a change in trend in the first eigenfunction centered at 1934/1935. The 1920-1934 portion has a slope of -0.0037, and the 1935-1983 portion has a slope of 0.0057. This is not an artifact of lower sampling density between 1920 and 1934 or the analysis technique (see text).

data. This comparison yielded negligible differences (figure 4), indicating the change in slope centered around 1934/1935 is neither an artifact of a lower sampling density during the early period (26 of the 44 stations began operation before 1935) nor of bias introduced by the eigenanalysis, but represents an actual change in the rate of RSL movement for stations along the east coast of North America. This change in rate of RSL movement was not observed in Aubrey and Emery (1983) because they analyzed the U.S. east coast data from only 1940 to 1979.

Linear regression slopes of these temporal eigenfunctions were combined with the spatial eigenfunctions to produce synthetic rates of RSL movement for each station (figure 5). The 1935-1983 portion of the first temporal eigenfunction was used to produce the synthetic RSL curve. The slopes of the 1935-1983 portion of both the second and third temporal eigenfunctions were not significant. Error bars for the synthetic rates of RSL change for each station (Table II, figure 5) were derived from the expected deviation in slope of the 1935-1983 portion of the first temporal eigenfunction, using a t-test at the 95% significance level.

Synthetic RSL rates (figure 5) are highly variable along the coast. Rate of RSL rise increases northward from about 1.2 to 2.0 mm/yr at the Florida stations up to a maximum of 3.5 mm/yr at Hampton Roads. North of Hampton Roads, rate of RSL rise follows a general decreasing trend with fluctuations, to 1.6 mm/yr at Portsmouth, N.H. Stations north of Portsmouth, N.H., to Halifax, display an increase in rate of RSL rise. From Halifax to Churchill, the decrease in rate of RSL rise is dramatic,

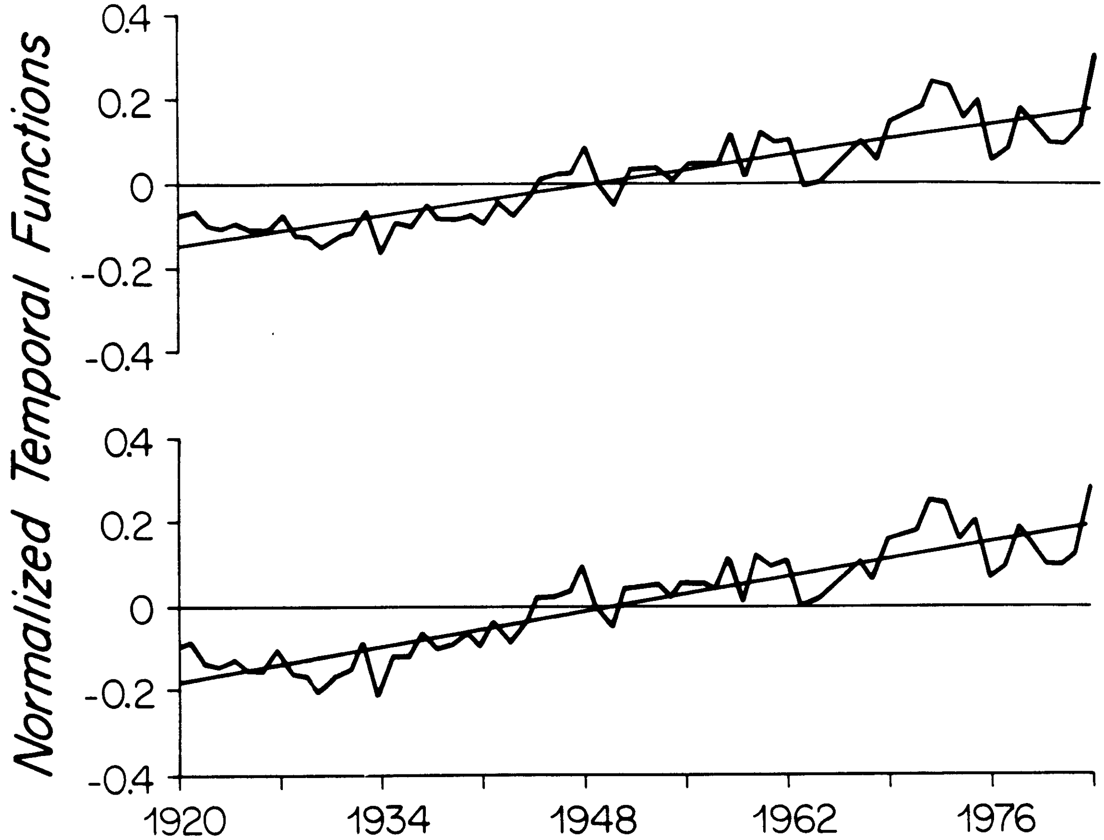


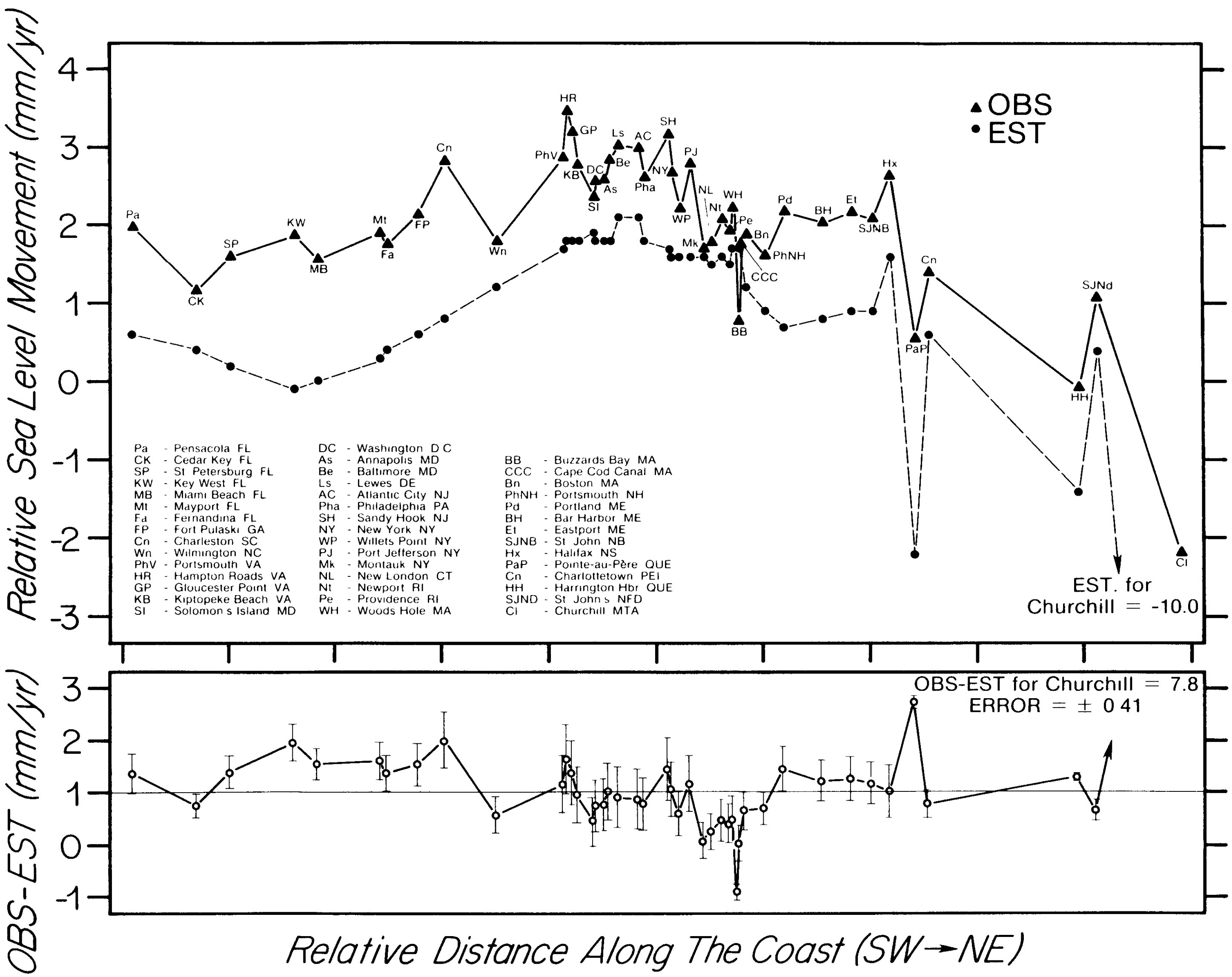
Fig. 4. Upper: The first temporal eigenfunction computed from all 44 stations used in this study.

Lower: The first temporal eigenfunction computed from the 26 stations which were in operation before 1935.

There is almost no difference between the two results indicating that the change in slope of the first temporal eigenfunction at 1934 is not an artifact of lower sampling density between 1920 and 1934, nor is it a bias resulting from the analysis (see text).

Fig. 5. Upper: Mean annual RSL movement for 1920-1980 from reconstructed eigenfunction data (synthetic rates, or OBS), and estimates of annual RSL movement due to post-glacial isostatic adjustment (EST, Peltier, 1986), plotted in a relative sense along the coastline. These relative distances are obtained by drawing perpendiculars from the stations to lines drawn approximately parallel to the coastline. From Pensacola to Key West, this line trends 146° measured clockwise from true north; from Key West to St. John, Newfoundland, the line trends 40° measured clockwise from true north. Churchill, located along the west central coast of the Hudson Bay, is placed at an arbitrary distance from St. John.

Lower: Residual annual RSL movement, i.e., synthetic (OBS) minus estimated isostatic adjustment (EST). Error bars are derived as discussed in the text.



Relative Sea Level Movement (mm/yr)
OBS
EST
Pa - Pensacola FL
CK - Cedar Key FL
SP - St Petersburg FL
KW - Key West FL
MB - Miami Beach FL
Mt - Mayport FL
Fa - Fernandina FL
FP - Fort Pulaski GA
Cn - Charleston SC
Wn - Wilmington NC
PhV - Portsmouth VA
HR - Hampton Roads VA
GP - Gloucester Point VA
KB - Kiptopeke Beach VA
SI - Solomon s Island MD
DC - Washington D C
As - Annapolis MD
Be - Baltimore MD
Ls - Lewes DE
AC - Atlantic City NJ
Pha - Philadelphia PA
SH - Sandy Hook NJ
NY - New York NY
WP - Willets Point NY
PJ - Port Jefferson NY
Mk - Montauk NY
NL - New London CT
Nt - Newport RI
Pe - Providence RI
WH - Woods Hole MA
BB - Buzzards Bay MA
CCC - Cape Cod Canal MA
Bn - Boston MA
PhNH - Portsmouth NH
Pd - Portland ME
BH - Bar Harbor ME
Et - Eastport ME
SJNB - St John NB
Hx - Halifax NS
PaP - Pointe-au-Père QUE
Cn - Charlottetown PEI
HH - Harrington Hbr QUE
SJND - St John s NFD
Cl - Churchill MTA
EST. for Churchill = -10.0
OBS-EST (mm/yr)
OBS-EST for Churchill = 7.8
ERROR = ± 0 41
Relative Distance Along The Coast (SW→NE)

from 2.6 mm/yr at Halifax to -2.2 mm/yr at Churchill, and highly variable. A comparison between the synthetic rates of RSL rise and linear regressions (figure 6) shows a good one-to-one correspondence, except at Churchill where the synthetic rate is less than half the regression trend.

Spectral analysis of the temporal eigenfunctions was made, after detrending, to determine dominant frequencies in time scales of RSL movement (figure 7). None of the spectral estimates are significant at the 90% level, but the first and second functions show slightly significant energy at the 80% level in the low frequencies. In the first function there are peaks at both 20- and 6-year periods, while in the second function, there is a broad peak between periods of 4 and 20 years, and a peak at about 3 years. The third function shows higher energy at the lower frequencies, but no significant peaks of energy at the 80% level. Short record length and nonstationary forcing combine to lower the significance of the spectral estimates.

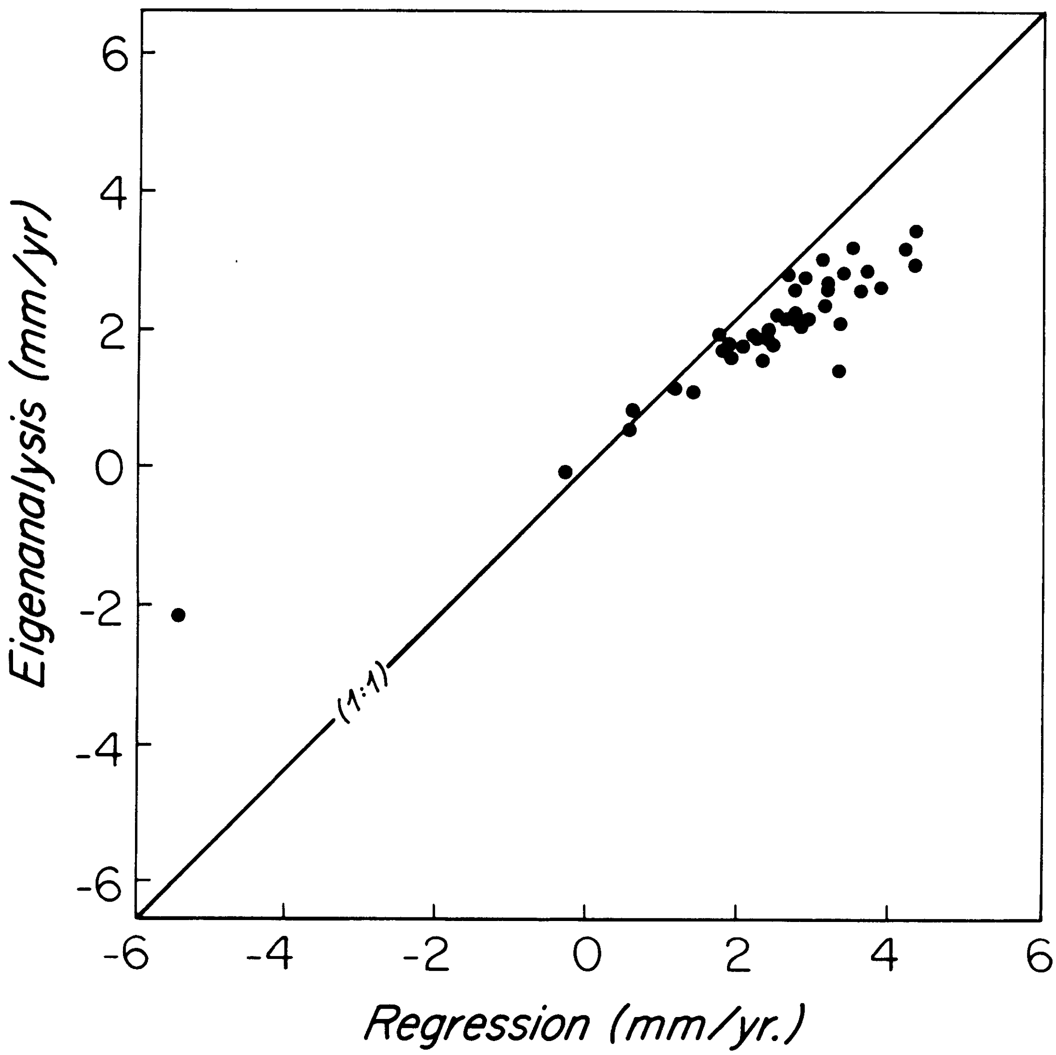


Figure 6. Scatter diagram of rates of change in RSL at individual stations obtained by both eigenanalysis and regression.

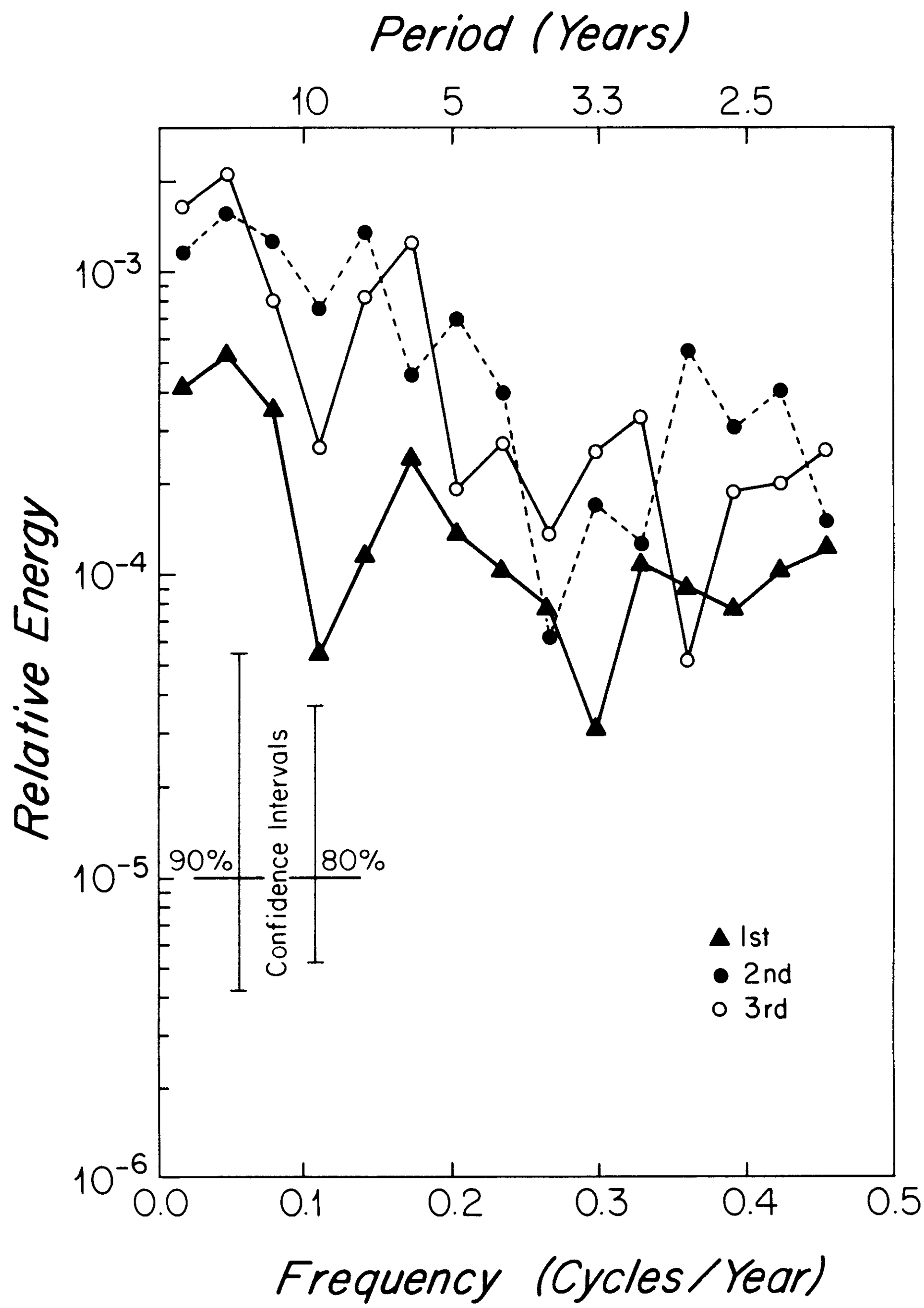


Fig. 7. Spectra of the first, second, and third temporal eigenfunctions. Estimates have four degrees of freedom. Energy units are relative. All three

DISCUSSION

SPATIAL FLUCTUATIONS

Spatial fluctuations in rates of RSL movement (figure 5) can be discussed in terms of two spatial scales: long-wavelength (order of thousands of km) and short-wavelength (order of tens to hundreds of km). Ice sheets that covered the Northern Hemisphere during the Pleistocene were large enough to cause elastic and plastic deformation of the lithosphere, having wavelengths of thousands of km. The maximum southeastern extent of these ice sheets along the east coast of North America is marked by Long Island, yet geologic evidence and modeling results indicate the isostatic effects of these ice sheets (crustal rebound from glacial depression, forebulge relaxation, hydrostatic depression) have extended great distances from the margins of ice sheets, and continue to influence vertical land motion today (e.g., Walcott, 1972a; Farrell and Clark, 1976; Clark et al., 1978; Quinlan and Beaumont, 1981). Shorter wavelength RSL movements could be caused by regional land warping due to tectonic movements, subsidence due to ground-water withdrawal or sediment loading, seismic activity, etc.

Oceanographic and Atmospheric Influence

Changes in the density of the water column, atmospheric pressure, wind velocity, and oceanic circulation all can affect tide-gauge measurements at different spatial scales. Strong interseasonal changes and small but measurable decadal changes in air pressure, wind stress, and Ekman upwelling over the North Atlantic have been documented (Thompson and Hazen, 1983). Unfortunately, studies of meteorological effects on sea-level variation have concentrated on the Pacific and have been over time scales too short to be useful in this study (e.g., Pattullo, 1960; Lisitzin and Pattullo, 1961). Similar problems exist with oceanographic data. Variations in transport of the Florida Current, the portion of the Gulf Stream system flowing through the Florida Straits northward to the point where the flow leaves the continental slope, could have a measurable effect on tide-gauge records south of Cape Hatteras (Iselin, 1940). However, studies of the transport of the Gulf Stream have also been on time scales too short to be applicable here (e.g., Niiler and Richardson, 1973). For these reasons, atmospheric and oceanographic effects on the spatial variation in RSL movement will not be discussed further.

Glacio-Isostatic Influence

Estimates of current glacio-isostatic vertical land movement at different tide-gauge stations can be obtained from RSL curves generated from ^{14}C dated shoreline indicators (e.g., Gornitz et al., 1982) and also from geophysical models of the response of the earth to glacial unloading and meltwater loading (e.g., Peltier, 1986). Both methods have their weaknesses. The ^{14}C method is constrained by a small amount of data from the last few thousand years. Therefore, linear rates based on a few data points over thousands of years must be used to represent the last 100 years or so of RSL change, when RSL curves are known to be non-linear (e.g., Bloom, 1967). Geophysical models can be adjusted to generate sea-level curves for shorter intervals of time, but the models are only as good as the parameters used to construct them, and many of these parameters are poorly constrained. There is considerable debate about the space and time scales of loading and unloading of the Pleistocene ice sheets, as well as the vertical structure and rheology of the earth. Variations in these parameters yield quite different model results (e.g., Clark et al., 1978; Peltier, 1984, 1986). Also, these models do not account for lateral inhomogeneities in the earth, such as variations in lithology and structure, so local records of RSL variation, as determined from ^{14}C dates, tide-gauge records, geodetic leveling, etc., can be expected to depart from model RSL curves. Another factor which is not accounted for in these models is a eustatic rise in sea level that may have occurred during the past century due to the "greenhouse" effect. Keeping these simplifications in mind, results of one of these models are used to assess the isostatic part (due to Pleistocene deglaciation) of RSL movement recorded by tide gauges.

In a series of papers over the past decade (e.g., Peltier and Andrews, 1976; Clark et al., 1978; Peltier et al., 1978; Peltier and Wu, 1982; Wu and Peltier, 1983, 1984; Peltier, 1984, 1986), models of earth rheology and structure have been successively refined by fitting these various models to different sets of geophysical data. Data consist of different manifestations of the earth's response to the loading and unloading of the Pleistocene ice sheets: postglacial (18 to 1K YBP) variations in relative sea level, free air gravity anomalies over present areas of rebound, and anomalies in the earth's rotation (non-tidal acceleration of rotation and secular drift of the rotation pole revealed by the International Latitude Service pole path). All of these models use a radially stratified viscoelastic representation of the earth. One of the most recent refinements of this representation used ^{14}C records of postglacial relative sea levels in North America and Europe, and strandline tilts in the Great Lakes (Peltier, 1986). The resultant model (elastic model 1066B with a 196.6 km thick lithosphere, upper mantle viscosity of 10^{21} Pa s and a lower mantle viscocity of 2 X 10^{21} Pa s) was then used to filter the glacio-isostatic portion of the recent (1940-1980) RSL signal recorded by 26 tide gauges along the U.S. east coast and 12 tide gauges along the U.S. west coast. The mean of the residual at the east coast stations (secular trend recorded by tide gauges minus the predicted present rate of RSL movement determined by modeling isostatic adjustment) was 1.1 mm/yr. If the residual were due to a purely eustatic sea-level signal, this residual would be fairly uniform along this coastline. However, the residual shows a great deal of variability about its mean, indicating that other mechanisms of RSL movement are influencing the tide-gauge records.

A similar residual was calculated here, by subtracting Peltier's (1986) estimates of isostatic adjustment from synthetic rates of RSL movement constructed from eigenanalysis (figure 5). Present rates of isostatic adjustment are given in Peltier (1986) for only 26 of the 44 tide-gauge stations analyzed here; rates for the remaining 18 stations were interpolated from a contour map of predicted rates of present vertical glacio-isostatic motion for North America (Figure 12 in Peltier, 1986). Subtracting the isostatic portion of RSL movement from the RSL curve results in a residual RSL curve (figure 5), from which most of the long-wavelength component of the RSL curve has been removed (ignoring

Buzzards Bay, Pointe-au-Père, and Churchill, the filtering results in a reduction in variance by more than a factor of two). The equally-weighted average of the resultant residual is 1.2 mm/yr. Residual rates at Buzzards Bay, Pointe-au-Père, and Churchill are questionable, for reasons to be discussed later; removing these results yields an equally-weighted mean residual of 1.0 mm/yr. By averaging tide-gauge results from 8 regions of equal length along the coast, the distance-weighted mean residual rate of RSL rise is 1.5 mm/yr (1.2 mm/yr if residuals at Churchill, Pointe-au-Père, and Buzzards Bay are not included). This may be a more accurate estimate of mean residual RSL rise. Other weighting schemes, equally plausible and defensible, yield similar results.

If the variability is due to more local events along the coastline, i.e., seismic movements, subsidence and warping of the land surface, etc., one would expect fluctuations to occur about a mean close to zero (assuming Peltier's 1986 estimates of current isostatic adjustments are not uniformly low along the entire coastline). A non-zero mean residual rate of RSL movement that is close to previous estimates of the mean rate of eustatic sea-level rise over the past century (Table I) suggests that this mean residual rate of 1.0-1.5 mm/yr may approximate the eustatic component of RSL change. However, this suggestion is questionnable for three reasons: 1) The global tide-gauge network is heavily skewed towards the Northern Hemisphere, particularly the North Atlantic. Many long-term stations are in a zone of glacio-isostatic submergence (e.g., Walcott, 1972a). Therefore, the estimates shown in Table I may be biased by RSL rise due to land submergence, although most of the studies in Table I attempted some correction for this effect. 2) The mean residual rate of RSL change determined in this study may result from large-scale (margin-wide) tectonic movement of the coastline (i.e., subsidence of the Coastal Plain), although the change in rate of RSL rise, from a small negative to a larger positive rate (revealed by the first temporal eigenfunction), argues against a geologic cause (see section on temporal fluctuations). 3) The glacio-isostatic model may underpredict present rates of adjustment.

Geologic Influence

Shorter wavelength variations in RSL movement appearing as fluctuations about the mean residual rate may be due to lateral compositional inhomogeneities and structural discontinuities in the lithosphere not accounted for in Peltier's (1986) model (e.g., crustal thinning and faults). Geodetic leveling studies, geologic structure, and records of seismic activity along this coastline provide an indication of where these may be affecting local RSL.

Precise geodetic leveling surveys conducted during the past century by the National Geodetic Survey provide estimates of relative vertical land movements. Absolute rates of crustal movement can be determined only if the leveling net is tied into a fixed frame of reference, usually tide-gauge stations (e.g., Holdahl and Morrison, 1974; Vanicek, 1975, 1976). Leveled surfaces are warped to fit the tide-gauge data, which can introduce errors due to nonsynchronous leveling and tide-gauge data, as well as excessive smoothing (Brown and Oliver, 1976). For these reasons, and since it would be inconsistent to use leveling results fit to tide-gauge data to then analyze tide-gauge data, only relative movements determined by leveling studies are used here.

Leveling data are not without error. Constant relative movement over the time interval between relevelings is assumed, although various studies have shown that this assumption is not always correct (e.g., Brown and Oliver, 1976; Brown, 1978). Time frames of leveling studies must be matched to those of the tide-gauge data as closely as possible. Also, as discussed by Brown (1978), leveling surveys are susceptible to error from local effects (variations in atmospheric pressure, rapid pressure gradients, expansion or contraction of the ground due to temperature changes, frost heaving, and near-surface ground water variations), as well as systematic influences (unequal refraction, tidal attraction, ocean loading, and unequal lighting). Also, recently a magnetic error associated with the Zeiss Ni-1 level instrument, which was used between 1972 and 1979, was discovered in the U.S. leveling net (S. R. Holdahl, pers. comm.). On average, vertical error due to this instrument is 0.8 mm/km. Large (hundreds of km) north-south tilts in the U.S. leveling net, which indicate movement down to the north, are probably a result of this error and will not be used in this study.

Additional confusion results from the use of relative, instead of absolute, geodetic measurements since some vertical land movements due to isostatic adjustments will be apparent in relative geodetic leveling results and others will not. For example, if the vertical rise or fall of the land surface at a hypothetical station A is equivalent to that at another hypothetical station B over the time interval between levelings, then the relative movement recorded by the levelings between these two stations would be zero. The equivalent vertical movement at the two stations, however, will be apparent in the tide-gauge data, and if the movement is due to postglacial isostatic adjustment, should appear in Peltier's data. On the other hand, unequal vertical movement at stations A and B will be recorded as nonzero relative movement by geodetic levelings, the tide-gauge data, and Peltier's data (if applicable).

The geology of the study area (figure 8) also helps interpret local deviations from the coastal mean rate of RSL rise. In the north, Precambrian metamorphic and plutonic infracrustal rocks of the Canadian Shield are exposed (King, 1970, and references therein). These rocks are overlain by younger platform deposits to the south (the Interior Platform), and also to the north, along the southeastern side of Hudson Bay (the Hudson Bay Platform). Along the southeastern side of North America, and running out to sea to the northeast of Newfoundland, is the Appalachian foldbelt, composed of geosynclinal deposits that were highly deformed, faulted, and metamorphosed during the Taconian, Acadian, and Alleghenian orogenies of the Paleozoic era. Smaller units of Precambrian unaltered geosynclinal sedimentary and volcanic rocks, as well as metamorphic and plutonic infracrustal rocks, occur within this foldbelt. Less deformed, post-orogenic strata (Carboniferous and Triassic in age) are present in the Appalachian foldbelt as well. The Carboniferous rocks have undergone moderate to steep compressional folding; the Triassic rocks are tilted, warped, and block faulted. The Appalachian foldbelt is overlain along its southern and southeastern flanks by platform deposits of the Gulf and Atlantic Coastal Plains.

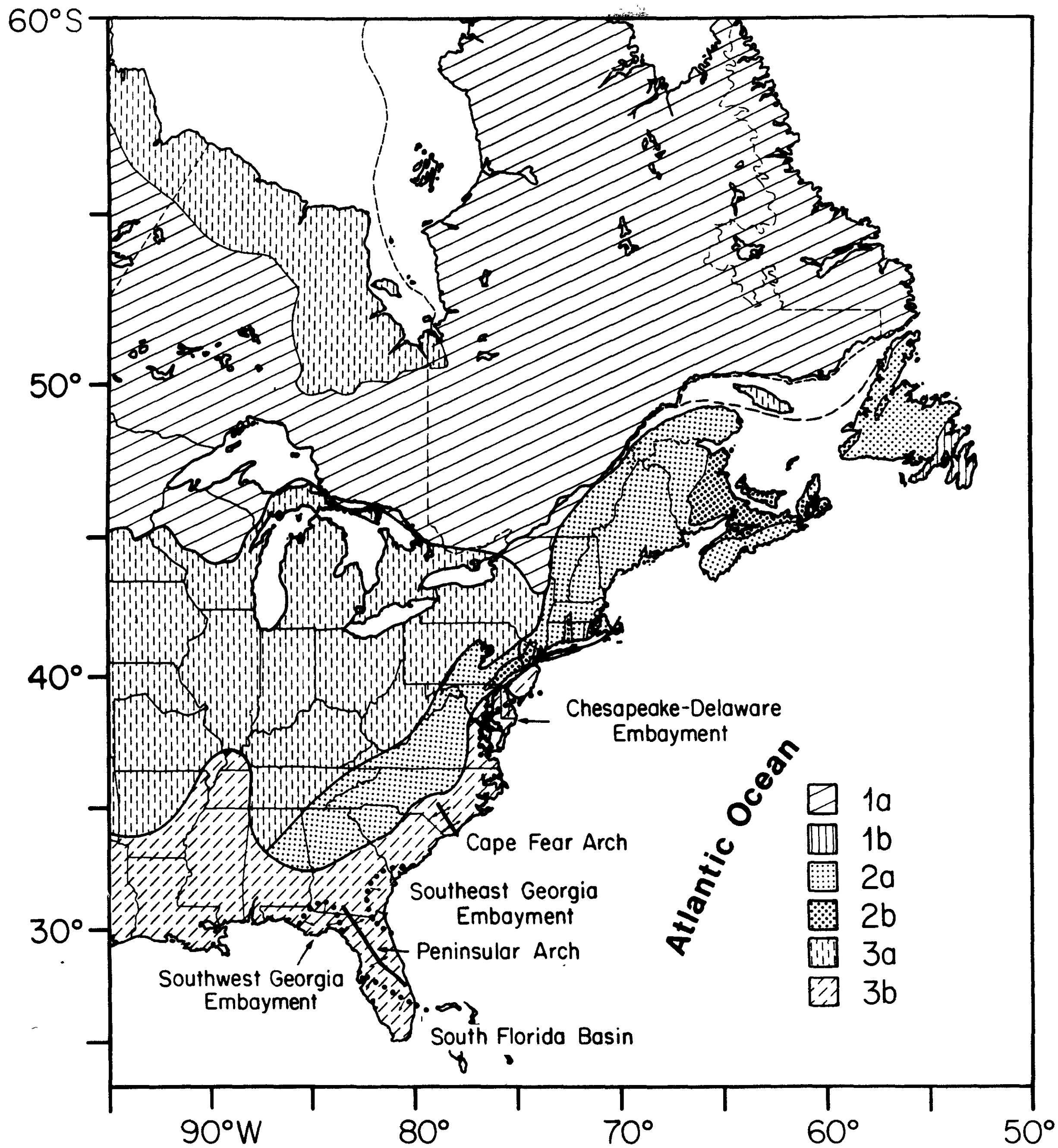


Figure 8. Tectonic Map of eastern North America, provinces and structural features discussed in text. Generalized from Murray (1961) and King (1969, 1970). Key: 1) Precambrian rocks (metamorphic and plutonic): a. Canadian Shield, b. Avalon terrane; 2) Appalachian foldbelt: a. geosynclinal deposits, b. post-orogenic deposits (Carboniferous and Triassic); 3) Paleozoic and younger platform deposits: a. on Precambrain basement (Interior and Hudson Bay Platforms), b. on Paleozoic basement (Gulf and Atlantic Coastal Plains).

These platform deposits are composed of thick sequences of Jurassic, Cretaceous, and Cenozoic marine and non-marine sediments.

From south to north, the tide-gauge stations from Pensacola (Florida) to Sandy Hook (New Jersey), as well as those on Long Island and Cape Cod, are located on the Coastal Plain. Tide-gauge stations between, and including, New York and Halifax (Nova Scotia), and also Pointe-au-Père (Quebec), are in the Appalachian foldbelt. Newport and Providence (Rhode Island), Boston (Massachusets), St. John (New Brunswick), and Charlottetown (Prince Edward Island) are in Carboniferous basins within the foldbelt. St. John's (Newfoundland) is on a terrane of Precambrian sedimentary and volcanic rocks, overlain by Cambrian and younger shallow-marine sediments (the Avalon terrane). Harrington Harbor (Quebec) is in the Canadian Shield, and the northernmost station, Churchill (Manitoba), is on Paleozoic and younger, relatively flat-lying, platform deposits overlying Precambrian basement (the Hudson Bay Platform).

Seismic data also can be used to interpret the variability in the rates of RSL movement along the coast. In areas of major earthquakes such as California and Japan, rapid horizontal and vertical crustal movements occur 10 to 20 years to a few hours prior to an earthquake and continue for as long as a few months after the earthquake (Scholz, 1972). Unfortunately, earthquakes and crustal displacements due to earthquakes are not as well understood in eastern North America (e.g., Kerr, 1981). Surface rupturing due to an earthquake has never been observed in this area, and leveling studies have not been made often enough to document movements due to specific earthquakes. Until recently, the density of seismograph stations in eastern North America has been low (spacing between stations typically ~300 km), so earthquake epicenters could not be determined with much accuracy and correlations between specific earthquakes and geologic or geophysical features have been difficult. However, zones of seismic activity following geological and geophysical trends have been delineated in eastern North America (e.g., Woollard, 1958; Fletcher et al., 1978; Sykes, 1978). Although a comparison between seismicity and land displacements determined by leveling along the entire U.S. east coast showed a poor correlation (Brown, 1978), a correlation between anomalous RSL movement recorded by tide-gauge data and a zone of high seismic activity may be significant.

Starting at the southwestern end of the residual RSL curve (figure 5), residual rates of RSL rise equal to or greater than 1 mm/yr occur between Pensacola, Florida and Kiptopeke Beach, Virginia except at Cedar Key, Florida (residual rate = 0.8 mm/yr) and Wilmington, North Carolina (residual rate = 0.6 mm/yr). Assuming the coastal mean residual rate of ~1.0 mm/yr represents a eustatic signal or a margin-wide tectonic signal, these fluctuations about the mean indicate that all the stations are undergoing subsidence, except Cedar Key and Wilmington, which are undergoing uplift. Analysis of leveling surveys perpendicular to the coasts shows generally consistent tilt downward toward the ocean along the Gulf and Atlantic Coastal Plains (Brown and Oliver, 1976). This oceanward increase in subsidence correlates in general with an oceanward increase in sediment thickness (Murray, 1961; Brown et al., 1972), although the tilt rates are too high to be explained by sediment compaction alone (Brown and Oliver, 1976). Nevertheless, if this tilt is real and not due to geodetic leveling error, it is consistent with most of the tide gauges along this coastline.

Analyses of leveling transects parallel to the coast, from Cedar Key, Florida southward to Key West, Florida and then northward to Charleston, South Carolina yield results grossly inconsistent with tide-gauge data (Holdahl and Morrison, 1974; Brown, 1978). However, the sense of movement revealed by residual RSL rates is in agreement with the sense of past movement revealed by basement structural features of the area. The tide gauge at Cedar Key, which shows a relative dip in the residual RSL curve (figure 5), is located just west of a basement structure known variously as the Central Georgia uplift, the Peninsular arch, and the Ocala uplift (figure 8). The Central Georgia uplift - Peninsular arch represents the late Paleozoic and Mesozoic axes of maximum upwarping; the Ocala uplift represents the center of Cenozoic upwarping, subparallel to and west of the Peninsular arch (Murray, 1961). Releveling data suggest this westward migration of the axis of maximum uplift in northern Florida continues today (Holdahl and Morrison, 1974; Brown and Oliver, 1976). This upwarping would explain the drop in the residual RSL curve at Cedar Key, relative to points north and south. The remaining stations in Florida, Georgia, and South Carolina are all located near or in structural lows. Pensacola is located west of the Southwest Georgia, or Apalachicola, embayment, a basin filled primarily with late Mesozoic and Cenozoic sediments (Murray, 1961). St. Petersburg, Key West, and Miami Beach are located in the South Florida basin, a post-Paleozoic northwest-southeast trending basin that dominates the structure of peninsular Florida south of the Ocala uplift (figure 8). Mayport, Fernandina, Fort Pulaski, and Charleston are located in the Southeast Georgia, or Savannah, embayment, an asymmetric shallow syncline, north of the Ocala uplift (figure 8). Deformation of ancient shorelines across the Southeast Georgia Basin suggests that warping of this area has continued through the Pleistocene (Winkler and Howard, 1977).

Beneath these basement features in Florida are older lithotectonic units bounded by basement hinge zones and fracture zones which resulted from early Mesozoic rifting of the North American from the South American/African plates. The basement hinge zone, which represents the block-faulted edge of the North American plate, separates Paleozoic and older crustal rocks to the northwest from Jurassic rifted crust beneath marginal basins to the southeast (Klitgord et al., 1984). This hinge zone intersects the west Florida coast at St. Petersburg. Key West and Miami are located in the Jurassic marginal rift basin to the southeast. Cutting across peninsular Florida, from southeast to northwest, is a broad Jurassic transform zone, that intersects the west Florida coast at Cedar Key. The Southwest and Southeast Georgia embayments to the north both overlie Jurassic sedimentary basins and late Triassic rift zones. All tide-gauge stations in Florida, Georgia, and South Carolina that record residual rates of RSL rise higher than the coastal mean are in regions underlain by Jurassic rift basins.

Some subsidence at Fort Pulaski also may be due to groudwater withdrawal. Leveling around Savannah, 20 km west of Fort Pulaski, indicates the land within a 5 mile radius of the city has subsided more than 20 mm between 1933 and 1955 (Davis et al., 1963). This subsidence has been due primarily to the decline in artesian pressure head which has resulted from groundwater withdrawal in the Savannah area (Davis et al., 1963). Groundwater withdrawal has been shown to cause substantial subsidence (tens to hundreds of millimeters) in many areas of the continental United States, but the effects are usually quite local (Chi and Reilinger, 1984). Subsidence around Savannah probably is due to compaction of the deposits comprising the artesian aquifer as well as underdrainage and compaction of the overlying confining unit (Davis et al., 1963). Fort Pulaski has undergone a concurrent decline in pressure head (20 feet between 1935 and 1955, compared to 40 to 100 feet around Savannah; Davis et al., 1963). Releveling between Savannah and Savannah Beach, along a line passing about 1.5 km south of the tide gauge at Fort Pulaski, indicates subsidence south of Fort Pulaski of ~10 mm between 1935 and 1955. However, this subsidence is local, affecting only about 11 km of the transect, which is on the opposite side of the Savannah River from the tide gauge.

At Charleston, South Carolina the rate of residual RSL rise is 0.4 to 0.6 mm/yr higher than the residual rates at the other stations in the Southeast Georgia embayment. Mayport and Fernandina are located on the southwestern flank of the syncline, Charleston on the northeastern flank. Either the northeastern flank is subsiding faster than the southwestern, resulting in a higher residual RSL rate at Charleston than at the other stations in this basin, or local subsidence is occurring around Charleston. The first option is unlikely since the southwestern flank is steeper and more pronounced than the northern and northeastern flanks (Murray, 1961). However, Charleston is located in a region of high seismic activity relative to the rest of the southeastern U.S. (e.g., Bollinger, 1973). The Charleston earthquake of 1886 (magnitude ~7; Kerr, 1981) is one of the few major U.S. earthquakes to have occurred outside the Pacific coast region. During the last century there have been two additional intervals of high seismicity (based on the number of earthquakes greater than magnitude 3): one between 1912 and 1917, and the other between 1958 and 1962 (Seeber et al., 1982). Many thrust and normal faults have been mapped in this area, but the fault or faults responsible for the earthquakes have not been identified, nor has fault displacement been measured (Bollinger, 1973; Kerr, 1981). Epicenters of the Charleston earthquakes lie along a zone extending from Charlestown northwest into Tennessee (Bollinger, 1972, 1973). Perhaps the subsidence indicated by residual RSL movement at the Charleston tide gauge is due to vertical crustal movement associated with stress release along this zone.

At Wilmington, North Carolina a sizeable relative drop in the residual RSL curve (figure 5) indicates local uplift. Wilmington lies along the crest of the Cape Fear arch, a northwest-southeast trending basement ridge extending under the continental shelf and slope at the southeast end, and gradually surfacing near the western edge of the Atlantic Coastal Plain at the northwest end (Bonini, 1955; Murray, 1961). Leveling transects indicate present dome-like uplift of this structural high centered offshore (Brown and Oliver, 1976; Brown, 1978). Brown (1978) suggests the movement recorded here by the relevelings may be associated with a Triassic basin located seaward of the Cape, near the inferred center of uplift. The basin, a zone of weakness in the crust, could be responding to contemporary regional stress.

Residual rates of RSL movement at tide-gauge stations around the Chesapeake Bay (Portsmouth, Virginia northward to Philadelphia, Pennsylvania) fluctuate about the coastal mean of 1.0 mm/yr. This entire region is deeply embayed, indicating long-term relative land subsidence and sea-level encroachment. The basement structure is dominated by the Chesapeake-Delaware embayment (figure 8), a major northwest-southeast trending feature underlain by Triassic basins (Bayley and Muehlberger, 1968; Klitgord and Hutchinson, 1985). Geodetic leveling from Norfolk, Virginia (next to Portsmouth) to Philadelphia indicates a broad downwarping in the Chesapeake region relative to points to the north and south (Brown, 1978). However, this leveling transect crosses the Delmarva Peninsula, from Norfolk to the southern tip of the peninsula (Kiptopeke Beach), and then northward to Philadelphia, while all of the tide-gauge stations in this region, except Kiptopeke Beach and Lewes, are on the opposite (western) side of the Chesapeake Bay. Residual rates of RSL movement at three stations located at the mouth of Chesapeake Bay (Portsmouth, Hampton Roads and Gloucester Point) suggest local subsidence, although not of the magnitude indicated by the leveling survey (residual RSL rates suggest subsidence of 0.2-0.7 mm/yr, while the leveling suggests relative subsidence of up to 3 mm/yr across the Delmarva Peninsula). Residual rates of RSL movement at the other tide-gauge stations in this region (Kiptopeke Beach to Philadelphia) indicate no anomalous rates of subsidence; in fact, there may be some uplift occurring at Solomon's Island. While tide gauges along this section record some of the highest rates of RSL rise along the east coast of North America, this region is also the area of the highest present rates of isostatic forebulge relaxation (figure 5; Peltier, 1986). All RSL movement recorded by tide gauges between Kiptopeke Beach and Philadelphia (except at Solomon's Island) can be accounted for by isostatic relaxation and a hypothesized eustatic sea-level rise. These results indicate the large tilt rates of the Chesapeake Bay region recorded by leveling surveys are due primarily to high rates of isostatic adjustment, although the magnitude of movement indicated by the leveling is high compared with tide-gauge results.

The residual rate at Sandy Hook suggests local subsidence, in addition to isostatic adjustment and eustatic sea-level rise. Recent publications have stated that the Sandy Hook data is erroneous or unreliable due to localized subsidence (Hicks, 1972; Brown, 1978), although no evidence is given for these statements. Hicks (1972) states that Sandy Hook is a station of "independently known local subsidence" but gives no support for this statement. Brown (1978) states that the Sandy Hook tide gauge is "unreliable" since it is "affected by very localized, non-tectonic subsidence" while citing Hicks (1972). Brown (1978) also suggests local groundwater withdrawal and rapid spit build-up nearby are probably responsible for subsidence at Sandy Hook, citing Minard (1969). However, although Minard (1969) states that the Sandy Hook quadrangle is a region of heavy industrial use of groundwater and is an active recurved spit, no mention is made of local subsidence.

Heavy industrial use of groundwater in this region may be causing localized subsidence, although leveling surveys in areas of equivalent water withdrawal along the eastern U.S. coastline show no effects of fluid withdrawal except at Savannah, Georgia. The amount of sediment compaction and surface subsidence due to fluid withdrawal is difficult to predict. The location of the tide gauge at Sandy Hook, on the northwestern side of a spit composed primarily of fine-to-coarse sand with relatively thin layers of clay and clayey sand (Minard, 1969, see Geologic Map and Section), argues against land subsidence due to groundwater withdrawal.

Residual rates of RSL movement at tide-gauge stations from New York northward to Portsmouth, New Hampshire are all lower than the coastal mean of 1.0 mm/yr, except New York and Port Jefferson, which are both close to the coastal mean (1.1 and 1.2 mm/yr, respectively; Table II, figure 5). These results indicate that isostatic adjustment, plus a eustatic sea-level rise, explain the RSL movement recorded by tide gauges at New York and Port Jefferson, while at Willets Point, Montauk, New London, Newport, Providence, Woods Hole, Buzzards Bay, Cape Cod Canal, Boston, and Portsmouth crustal uplift, in addition to isostatic adjustment and eustatic sea-level rise, may be occurring.

The large discrepancy in the rates of residual RSL movement (1.1 mm/yr) between the two stations on Long Island (Port Jefferson and Montauk) is surprising since both Montauk and Port Jefferson are underlain by Upper Cretaceous and Quaternary coastal plain deposits that show no evidence of deformation (Fisher et al., 1971). Residual rates suggest that except for post-glacial adjustments, Port Jefferson is stable while Montauk is undergoing uplift. High resolution seismic reflection profiles across the New York Bight have delineated a north-northeast trending fault (the New York Bight fault) which may continue northward underneath Long Island (Hutchinson and Grow, 1985). The fault runs along the western edge of a Mesozoic rift basin (the New York Bight basin), mapped using magnetic data (Klitgord and Hutchinson, 1985). This basin appears to continue northward across Long Island and Long Island Sound, and connect with the Hartford basin onshore.

Displacement along the New York Bight fault is down to the west. This sense of movement is consistent with the differential movement between Port Jefferson and Montauk, although evidence for Quaternary activity along the fault is ambiguous. An earthquake (magnitude ~2) was recorded near Port Jefferson during the period 1928-1959 (Smith, 1966) and there was another of about the same magnitude centered ~10 km to the west of the fault in the New York Bight in 1976 (Kafka et al., 1985). However, the trend of the fault and the basin indicates that they would pass very close to Port Jefferson, so any vertical motion associated with these features would occur near Port Jefferson instead of near Montauk, as indicated by the rates of residual RSL rise at these two stations.

Residual RSL rates at the remaining stations between New York and Portsmouth are all below the coastal mean. Uplift recorded at Willets Point may be related to seismic activity which has been historically and instrumentally recorded in the greater New York city area (Smith, 1966; Kafka et al., 1985). However, if seismic activity is affecting RSL movement at Willets Point, it is not clear why the station at New York would not be affected as well since both stations are located on the same geologic "province" of crystalline Paleozoic and Precambrian rocks known as the Manhattan Prong (Kafka et al., 1985).

Leveling transects between Willets Point and Portsmouth suggest a crustal tilt downward toward the northeast (Brown, 1978), as do RSL rates at the eastern end of this section of coastline (New London to Portsmouth). The strongest signature in the leveling transect is a sharp change in relative crustal movement as the leveling profile crosses the Hartford basin (Brown, 1978), a Triassic graben filled with sedimentary rocks dipping to the east. Leveling results indicate the same sense of movement. Historical and instrumentally-recorded earthquakes have been reported in the vicinity of this graben, particularly along the eastern border fault (Yang and Aggarwal, 1981). Seismic activity also has been noted in southeastern Connecticut, attributed to strain release along the Honey Hill thrust fault, a northward dipping zone of highly strained rock (Lundgren and Ebblin, 1972; Block et al., 1979). The sense of motion is southeastward thrusting of the northern blocks. Measurements of offsets along a single thrust plane suggest recently active faulting (Block et al., 1979). New London is located ~50 km east of the eastern (normal) border fault of the Hartford basin, and ~15 km south of the Honey Hill thrust fault. Crustal movement due to strain release along these fault systems may be affecting RSL at New London.

A sharp drop in the residual rate at Buzzards Bay, Massachusetts (to -0.9 mm/yr) indicates an anomalous amount of uplift here relative to surrounding tide gauges (figure 5). This anomalous result is probably due to the short record length at this station (20 years of data, the minimum used in this study). Such a short record may not record all the time scales of RSL motion along the coastline, particularly the long period fluctuations indicated by spectral analysis (figure 7) to be an important part of the temporal structure of the data. However, Cape Cod Canal, located approximately 10 km northeast of the Buzzards Bay station, on the opposite side of the Cape Cod Canal, also has a very short, concurrent record (20 years). It is not clear why the rates of RSL movement at these two stations are so different. Tide-gauge records at both stations are smooth, with no extreme outliers. Unfortunately, no detailed analyses of leveling transects in this region exist.

Between Boston and Portland, residual rates of RSL rise increase from 0.7 mm/yr at Boston and Portsmouth, to 1.5 mm/yr at Portland, suggesting crustal uplift at Boston and Portsmouth, and crustal subsidence at Portland. Vertical motion at these stations may be associated with seismic activity. Boston and Portsmouth are located in a zone of seismic activity similar to that passing through Charleston, South Carolina. This seismic belt trends northwest from Boston and Portsmouth through Ottawa, Ontario (Sbar and Sykes, 1973), although seismic activity may not be continuous along this zone (e.g., Yang and Aggarwal, 1981). Relative crustal movement between Portsmouth and Portland is also apparent in leveling transects, although leveling suggests a higher differential. Leveling between Portsmouth and Portland in 1923 and 1966 shows Portsmouth rose relative to Portland at a rate of 1.1 mm/yr (Brown, 1978), while tide-gauge data suggest Portsmouth rose relative to Portland at a lower rate of 0.6 mm/yr.

Rates of RSL rise from Portland northeast to Halifax, Nova Scotia are fairly constant between Portland and St. John (2.0-2.2 mm/yr), increasing at Halifax to 2.6 mm/yr (Table II, figure 5). Residual rates of RSL rise at these stations reveal that isostatic adjustment and a eustatic sea-level rise, or coastal submergence, fully account for the RSL movement at Halifax, whereas between Portland and St. John, northeasterly decreasing amounts of subsidence are occurring, from 0.5 mm/yr at Portland to 0.2 mm/yr at St. John. Previous studies of sea-level trends in Maine have considered the rate of RSL rise in the northernmost Maine station of Eastport anomalously high (Hicks, 1972; Anderson et al., 1984). However, regional trends of RSL along the northeast coast of North America determined using the most recent tide-gauge data demonstrate the movement recorded at Eastport is not anomalously high, but is consistent with data points to the north and south.

The fairly constant rates of RSL rise in Maine are not reflected in geodetic leveling studies, which indicate that northeastern Maine is subsiding relative to southern Maine (Brown, 1978; Tyler and Ladd, 1980; Anderson et al., 1984). Anderson et al. (1984) state that releveling data show Eastport to be subsiding by as much as 9 mm/yr. However, this rate, taken from Tyler and Ladd (1980), is the rate of movement at Eastport relative to Calais (~40 km northwest of Eastport), and therefore cannot be interpreted as an absolute rate. Geodetic leveling between Portland and Eastport indicates Eastport subsided relative to Portland at a rate of 1.8 mm/yr between 1926 and 1966/67, and at a rate of 7.8 mm/yr between 1942 and 1966/67 (Brown, 1978). Levelings between

Bangor (~65 km northwest of Bar Harbor) and Calais show Calais subsided with respect to Bangor at an average rate of 1 mm/yr between 1926 and 1966/67, and at 8 mm/yr between 1942 and 1966/67 (Brown, 1978). Anderson et al. (1984) suggest that this increase in rate of crustal movement may be related to a post-1940 increase in New England seismic activity (Shakal and Toksoz, 1977).

Since no definitive evidence of Cenozoic faulting has been found in Maine, it is not possible to unravel the discrepancy between the leveling and the tide-gauge data. However, both data sets indicate crustal movement in Maine and New Brunswick, possibly due to reactivation of Triassic fault systems (apparent in the leveling transects across the Hartford basin). The Gulf of Maine and the Bay of Fundy are heavily dissected with Triassic horsts and grabens, similar tectonically to Triassic grabens on land (Ballard and Uchupi, 1975). In situ measurements of stress in this region suggest a greatest principal stress direction similar to that inferred from Triassic-Jurassic block faulting in the Gulf of Maine (Anderson et al., 1984). The coastline surrounding the Gulf of Maine and Bay of Fundy has been, and is, seismically active (Smith, 1966; Lepage and Johnson, 1983; Anderson et al., 1984). Stress release along these fault systems also may be causing crustal movement from Portland to New Brunswick.

The remaining tide-gauge station situated in the Appalachian foldbelt is Pointe-au-Père, Quebec. Disjoint, releveled segments from northern New Brunswick to Halifax indicate Pointe-au-Père is rising relative to St. John and Halifax (Vanicek, 1976; Grant, 1980; note that Vanicek has warped the releveled network to fit tide-gauge data).

However, broad intertidal platforms cut into sedimentary rocks near Pointe-au-Père (Grant, 1980) indicate RSL has been static at this location for some time. Crustal deformation due to seismic activity may be occurring at Pointe-au-Père, an area with a high density of recorded earthquakes (Smith, 1966; York and Oliver, 1976). The historically most active seismic region in eastern Canada is centered ~150 km southwest of Pointe-au-Père (the Charlevoix region northeast of Quebec City, Basham et al., 1979). However, it is unlikely that seismic activity has caused ~2.7 mm/yr of RSL rise over the past 64 years. Peltier's (1986) estimate of the current rate of isostatic adjustment for Pointe-au-Père (-2.2 mm/yr) seems excessive, particularly since the tide gauge and local geomorphology indicate high amounts of uplift have not been occurring here. Also, the residual rate at Pointe-au-Père (2.7 mm/yr) is higher than Peltier's estimate of isostatic adjustment. It is unlikely that non-isostatic crustal movement is occurring here at a rate which is greater than the amount attributable to isostatic adjustment, since Pointe-au-Père is located in the region once covered by Pleistocene ice sheets.

Rates of residual RSL movement at Charlottetown, Harrington Harbor, and St. John's fluctuate within 0.3 mm/yr of the coastal mean. Charlottetown is located in a post-orogenic (Carboniferous) basin in the Appalachian Province; Harrington Harbor is located on Precambrian infracrustal rocks of the Canadian Shield; and St. John's is located on the Avalon terrane (believed to be a separate crustal block from the rest of Newfoundland, unrelated to Appalachian tectonics; Keen et al., 1986). Residual rates of RSL rise suggest that in addition to isostatic adjustment and a eustatic sea-level rise, Charlottetown and St. John are undergoing slight subsidence, while Harrington Harbor is undergoing slight uplift. Unfortunately, geodetic leveling transects are sparse in Canada; systematic relevelings are sparser still (Lambert and Vanicek, 1979). Most analyses of relevelings in eastern Canada have warped the leveling data to fit tide-gauge trends (e.g., Vanicek, 1976), so the results cannot be used as an independent means to determine vertical crustal motion.

The residual rate of RSL rise at Churchill, located on the Hudson Bay Platform, suggests that in addition to glacio-isostatic adjustment and eustatic sea-level rise, crustal uplift of 7.8 mm/yr is occurring. As at Pointe-au-Père, this rate seems unreasonably high. Glacio-isostatic adjustment should be dominating any vertical crustal motion here. Churchill is located 500-1000 km northwest of what has been the center of glacio-isostatic uplift in Canada over the past 7000 years (Walcott, 1972b), and free-air gravity anomalies in this region are close to the largest in eastern Canada (~30-40 milligals over Churchill). These anomalies are probably due to incomplete recovery from ice-load depression (Walcott, 1970, 1972b), indicating that the area of largest over-compensation in Canada, and by inference, the area of highest isostatic adjustment, is located here.

Some of the discrepancy between tide-gauge results and Peltier's results at Churchill may be due to eigenanalysis. A comparison of the results of eigenanalysis and linear regression (figure 6) indicates a one-to-one correspondence does not exist at Churchill. The discrepancy is partly an artifact of eigenanalysis. Most (95%) of the stations with positive rates of RSL movement lie below the line of 1:1 correspondance on the scatter plot (figure 7); the one station with a high negative rate of RSL movement (Churchill) lies substantially above this line. The offset of these two data sets about the line of 1:1 correspondence indicates RSL movements recorded at Churchill are not coherent (in time) with the rest of the east coast data. However, a residual rate of RSL rise at Churchill determined using linear regression (4.6 mm/yr) is still less than half of the estimate of isostatic adjustment (10 mm/yr). The average rate of RSL fall at Churchill over the past 2000 years, determined from radiocarbon-dated organic materials (Walcott, 1972b), is the same rate as Peltier's (1986) estimate. However, RSL curves derived from both radiocarbon-dated materials and geophysical models indicate the rates of glacio-isostatic adjustment have slowed with time. If tide-gauge data at Churchill are accurate, results indicate RSL fall has slowed here over the past 2000 years and Peltier's estimate of present glacio-isostatic adjustment is too large.

TEMPORAL FLUCTUATIONS

Linear Trends

The first temporal eigenfunction (figure 3) documents a change in the rate of RSL movement along the east coast of North America, from a shallow, decreasing trend (just barely significant) to a steeper, increasing trend. This rate change is centered around 1934. Analysis of tide-gauge records of western North America revealed a similar change in slope of the first temporal eigenfunction, from a nearly flat trend before 1935, followed by an increasing trend (Emery and Aubrey, 1986b). Barnett's (1984) analysis of global tide-gauge data also showed an increase in rate of RSL rise centered around 1930. It is difficult to imagine any geologic or tectonic processes that could explain this change in rate of RSL movement. Given the varied geologic structure of this coastline it is difficult to conceive of any coherent tectonic movement that could cause this change in slope, although its occurrence in tide-gauge data from both coasts of North America suggests that isostatic adjustments may be involved since the effect of the Pleistocene ice sheets was continent-wide.

Oceanographic factors may also be involved in this change in slope. A reoccupation (1957, 1981) of two zonal North Atlantic oceanographic sections (24°30'N and 36°6'N) documented a significant warming of an ocean-wide band between 700 and 3000 m depths (Roemmich and Wunsch, 1984). This warming resulted in a 2-3 cm steric expansion of the water column between 700 to 3000 m. If this expansion were basin-wide, it would explain a large part of the RSL rise along the east coast of North America. Unfortunately, no similar reoccupations are available to determine if this steric expansion was global. The consistency of the mid-depth warming observed between 24° and 36°N as well as its magnitude (roughly an order of magnitude greater than the perturbation of temperature profiles attributable to short time-scale eddies) suggest that the signal is not due to eddy noise. Moreover, the depth at which warming occurred implies local air-sea interactions were not important since the characteristics of mid-depth and deep waters are controlled by processes at middle and high latitudes.

To determine the time scale of the observed temperature changes at 24° and 36°N, Roemmich and Wunsch (1984) analyzed data from the Panulirus station (Bermuda, 32°N) for comparison. These data consist of temperature and salinity measurements taken twice monthly between 1954 and 1981, the longest continuous time series of hydrographic measurements available. Roemmich and Wunsch (1984) examined the depth change (with time) of the 4°C isotherm, because of its large observed displacement (100 m) between 1957 and 1981 at 24° and 36°N. The mean annual depths of the 4°C isotherm at the Panulirus station showed a clear trend, increasing approximately 100m over the past 25 years, supporting their hypothesis that the data at 24° and 36°N are part of a long-term (in an oceanographic sense) warming in the oceans. In a later study, Roemmich (1985) showed the warming trend below 700m at 24° and 36°N was also apparent at the Panulirus station (below 1000 m) over the 22 years of record.

Based on the results of the reoccupation study, Roemmich and Wunsch (1984) suggested steric expansion of the oceans is responsible for some or all of the observed RSL rise occuring over the globe during the past century (Table I). However, another study, in which hydrographic data from various regions of the globe were analyzed, found no significant trends in dynamic height over the past century (Barnett, 1983). Only data from the upper ocean was examined, however, and the studies discussed previously (Roemmich and Wunsch, 1984; Roemmich, 1985) have shown that long-term steric changes in the ocean are not confined to the upper layers of the ocean. More long-term, shallow- to deep-water, oceanographic data are clearly needed before a steric rise in sea level over the past century can be identified unambiguously.

Periodic Fluctuations

Spectral analysis of the temporal eigenfunctions reveals significant energy at periods of about 6 and 20 years in the first function (which accounts for 69.6% of the sea-level variance), and at 3 years and in a broad band between periods of 4 to 20 years in the second function (which accounts for 64% of the sea-level variance). The highest energy in the third function is in the lower frequencies, but there are no significant peaks. These periods are suggestive of oceanographic and atmospheric processes. Unfortunately, there are few long-term oceanographic time series that can be analyzed for dominant periodicities longer than one year. A 9- to 10-year periodicity is apparent (by visual inspection) in the 27 years of record of the mean annual depth of the 4°C isotherm at the Panulirus station (Roemmich and Wunsch, 1984), while five-year running mean averages of surface temperature at the Panulirus station between 1957 and 1967 display a 6-year half-cycle of a 12-year period (Pocklington, 1972). Both of these series are too short and too local to reach any significant conclusions. However, they do suggest that more than one oceanographic process is responsible for the broad bands of energy in the 6- to 20-year periodicities indicated by spectral analysis.

The high energy around a 20-year periodicity may be a reflection of the 18.6 lunar nodal cycle (a wobble of the plane of the moon's orbit around the earth). This cycle has been shown to affect annual mean sea-level curves, although rather weakly (Kaye and Stuckey, 1973).

The shorter 3-year periodicity in the spectrum of the first eigenfunction, significant at the 80% level, may be related to the Southern Oscillation. This alternation of regional pressure anomalies between the Indian and Southeast Pacific Oceans has a period of 2- to 3-years (Fairbridge and Krebs, 1962). However, even though global mean annual sea-level oscillations have been correlated with the Southern Oscillation (Fairbridge and Krebs, 1962), teleconnections between the Pacific and the Atlantic Oceans are not understood. This 3-year periodicity has appeared in spectra of RSL movement (along both the east and west coasts of North America) that have been corrected for barometric pressure fluctuations (Vanicek, 1978), suggesting the 3-year periodicity is due to some phenomena other than the Southern Oscillation.

CONCLUSIONS

Eigenanalysis of tide-gauge data along the east coast of North America reveals a highly variable record of RSL change. The segmentation of this coastline observed by Aubrey and Emery (1983) is a result of more than isostatic and oceanographic factors. Estimates of RSL movement based on numerical models of post-glacial isostatic adjustments indicate more than half of the variance of RSL movement determined from the eigenanalysis of the tide-gauge data is due to isostatic adjustment. The mean of the residual RSL movement along the entire coastline (tide-gauge rates minus isostatic adjustment rates), 1.0-1.5 mm/yr, may approximate the present eustatic sea-level rise. Leveling surveys, geologic structure, and seismic data indicate that much of the shorter wavelength features of RSL movement are due to various amounts of crustal movement. These results suggest that tide-gauge data, in addition to providing information on movements of the sea surface, also can be used as an indicator of neo-tectonic movements along the coastline.

Tide-gauge stations from Florida northward to New Jersey are located along the Gulf and Atlantic Coastal Plains, across which geodetic leveling surveys have shown a consistent tilt downward toward the ocean. This subsidence is reflected in all tide-gauge results from Pensacola to Gloucester Point, except at Cedar Key and Wilmington. Uplift is occuring at these two stations, both located in structural highs (the Pensacola arch and the Cape Fear arch, respectively). Pensacola, St. Petersburg, Key West, Miami Beach, Mayport, Fernandina, Fort Pulaski, and Charleston are all located in, or near, structural lows (the Southwest Georgia embayment, the South Florida basin and the Southeast Georgia embayment), and are underlain by Jurassic rift basins. Seismic activity may also be contributing to the subsidence recorded at Charleston. High rates of RSL rise recorded by tide gauges around the Chesapeake Bay, as well as the high rates of crustal subsidence measured by leveling transects in this region, are primarily due to high rates of isostatic adjustment. Tectonic subsidence also appears to be occurring at the mouth of the bay.

The remaining stations, except Harrington Harbor and Churchill, are located in the northern Appalachian foldbelt. Residual rates of RSL movement at these stations suggest crustal movements here are associated with preexisting faults and basins, particularly ones that were active during the Mesozoic. Seismic activity in New England demonstrates contemporary stress release, which may be related to crustal movement recorded by tide gauges. Different rates of RSL movement across Long Island (between Port Jefferson and Montauk) are probably due to movement associated with the New York Bight fault and basin (structures which appear to be related to the Hartford basin on land). Subsidence along the New York, Connecticut, and Rhode Island coasts may be related to movement along existing faults and the Hartford basin. A major seismic zone intersects the coast at Boston and Portsmouth (New Hampshire). Subsidence at these stations, as well as uplift to the north at Portland, may be related to this seismic activity. Subsidence at Pointe-au-Père, located in another zone of densely recorded earthquakes, may also be related to seismic activity. Subsidence around the Gulf of Maine and the Bay of Fundy (in Maine and the Maritime Provinces) may be associated with movement along Triassic horsts and grabens which dissect the offshore region. RSL movement at the remaining stations is dominated by isostatic adjustment; tide-gauge data suggest that estimates of current glacio-isostatic adjustment at Pointe-au-Père and Churchill (Peltier, 1986) are too high.

The temporal structure of the data shows that an increase in rate of RSL rise occurred along the east coast of North America centered near 1934. This increase may be due to a change in rate of isostatic adjustment over the continent, or steric expansion of the mid- to deep depths of the oceans. Further analysis, as well as more oceanographic data, are needed to evaluate these suggestions. Dominant periodicities of sea-level movement are indicative of oceanographic, atmospheric, and lunar cycles.

REFERENCES

Anderson, W.A., Kelley, J.T., Thompson, W.B., Borns, H.W., Jr., Sanger, D., Smith, D.C., Tyler, D.A., Anderson, R.S., Bridges, A.E., Crossen, K.J., Ladd, J.W., Andersen, B.G., and Lee, F.T., 1984, Crustal warping in coastal Maine: Geology, v. 12, p. 677-680.

Aubrey, D.G., 1985, Recent sea levels from tide gauges: problems and prognosis, in Glaciers, Ice Sheets, and Sea Level: Effect of a CO_2-Induced Climatic Change: National Academy Press, Washington, D.C., p. 73-91.

Aubrey, D.G., and Emery, K.O., 1983, Eigenanalysis of recent United States sea levels: Continental Shelf Research, v. 2, p. 21-33.

Aubrey, D.G., and Emery, K.O., 1986, Relative sea levels of Japan from tide-gauge records: Geol. Soc. Amer. Bull., v. 97, p.194-205.

Aubrey, D.G., and Welch, S.W., in prep., Objective methods for analysis of gappy data.

Ballard, R.D., and Uchupi, E., 1975, Triassic rift structure in Gulf of Maine: Amer. Assoc. Petrol. Geol. Bull., v. 59, no. 7, p. 1041-1072.

Barnett, T.P., 1983, Recent changes in sea level and their possible causes: Climate Change, v. 5, p. 15-38.

Barnett, T.P., 1984, The estimation of "global" sea level change: A problem of uniqueness: J. Geophys. Res., v. 89, p. 7980-7988.

Bayley, R.W. and Muehlberger, W.R., 1968, Basement Rock Map of the United States. U.S. Geol. Survey, Washington, D.C.

Berger, A., 1984, Book Reviews of Carbon Dioxide and Climate: A Second Assessment and The Long-Term Impacts of Increasing Atmospheric Carbon Dioxide Levels: EOS, v. 65, p. 53-54.

Block, J.W., Clement, R.C., Lew, L.R., and de Boer, J., 1979, Recent thrust faulting in southeastern Connecticut: Geology, v. 7, p. 79-82.

Bloom, A.L., 1967, Pleistocene shorelines: A new test of isostasy: Geol. Soc. Amer. Bull., v. 78, p. 1477-1494.

Bloom, A.L., 1983, Sea Level and coastal changes, in Wright, H.E., Jr., ed., Late-Quaternary Environments of the United States, The Holocene: University of Minnesota Press, Minneapolis, v. 2, p. 42-51.

Bollinger, G.A., 1972, Historical and recent seismic activity in South Carolina: Bull. Seis. Soc. Amer., v. 62, p. 851-864.

Bollinger, G.A., 1973, Seismicity of the southeastern United States: Bull. Seis. Soc. Amer., v. 63, p. 1785-1808.

Bonini, W.E., 1955, Seismic-refraction studies of geologic structure in North Carolina and South Carolina: Geol. Soc. Amer. Bull., v. 66, p. 1532-1533, abstract.

Brown, L.D., 1978, Recent vertical crustal movement along the east coast of the United States: Tectonophysics, v. 44, p. 205-231.

Brown, P.M., Miller, J.A., and Swain, F.M., 1972, Structural and stratigraphic framework, and spatial distribution of permeability of the Atlantic Coastal Plain, North Carolina to New York: U. S. Geol. Survey Prof. Paper 796, Washington, D.C., 79pp.

Brown, L.D., and Oliver, J.E., 1976, Vertical crustal movements from leveling data and their relation to geologic structure in the eastern United States: Rev. Geophys. Space Phys., v. 14, p. 13-35.

Chi, S.C., and Reilinger, R. E., 1984, Geodetic evidence for subsidence due to groundwater withdrawl in many parts of the United States of America: J. Hydrol., v. 67, p. 155-182.

Clark, J.A., Farrell, W.E., and Peltier, W.R., 1978, Global changes in post-glacial sea level: a numerical calculation: Quat. Res., v. 9, p. 265-287.

Davis, G.H., Small, J.B., and Counts, H.B., 1963, Land subsidence related to decline of artesian pressure in the Ocala Limestone at Savannah, Georgia, in Trask, P.D., and Kiersch, G.A. eds., Engineering Geology Case Histories, Geol. Soc. Amer., Boulder, no. 4, p. 1-8.

Emery, K.O., 1980, Relative sea levels from tide-gauge records: Proc. Natl. Acad. Sci., v. 77, p. 6968-6972.

Emery, K.O., and Aubrey, D.G., 1985, Glacial rebound and relative sea levels in Europe from tide-gauge records: Tectonophysics, v. 120, p.239-255.

Emery, K.O., and Aubrey, D.G., 1986a, Relative sea-level changes from tide-gauge records of eastern Asia mainland: Mar. Geol., in press.

Emery, K.O. and Aubrey, D.G., 1986b, Relative sea-level change from tide-gauge records of western North America: submitted to J. Geophys. Res.

Fairbridge, R. W., and Krebs, O. A., Jr., 1962, Sea level and the Southern Oscillation: Geophys. J. R. Astr. Soc., v. 6, p. 532-545.

Farrell, W.E., and Clark, J.A., 1976, On postglacial sea level: Geophys. J. R. Astr. Soc., v. 46, p. 647-667.

Fisher, D.W., Isachsen, Y.W., and Rickard, L.V., 1971, Generalized tectonic-metamorphic map of New York: New York State Museum and Science Service, Map and Chart Series No. 15.

Fletcher, J.B., Sbar, M.L., and Sykes, L.R., 1978, Seismic trends and travel-time residuals in eastern North America and their tectonic implications: Geol. Soc. Amer. Bull., v. 89, p.1656-1676.

Gornitz, V., Lebedeff, S., and Hansen, J., 1982, Global sea level trend in the past century: Science, v. 215, p. 1611-1614.

Grant, D.R., 1980, Quaternary sea-level change in Atlantic Canada as an indication of crustal deleveling, in Mörner, N.-A., ed., Earth Rheology, Isostasy, and Eustasy, New York, John Wiley and Sons, Ltd., p. 201-214.

Gutenberg, B., 1941, Changes in sea level, postglacial uplift and mobility of the earth's interior: Geol. Soc. Amer. Bull., v. 52, p. 721-772.

Hicks, S.D., 1972, Vertical crustal movements from sea level measurements along the east coast of the United States: J. Geophys. Res., v. 77, p. 5930-5934.

Hoffman, J.S., Keyes, D., and Titus, J.G., 1983, Projecting future sea level rise, methodology, estimates to the year 2100, and research needs: U.S. EPA 230-09-007, 121 p.

Holdahl, S. R., and Morrison, N.L., 1974, Regional investigations of vertical crustal movements in the U.S., using precise relevelings and mareograph data: Tectonophysics, v. 23, p. 373-390.

Hutchinson, D.R., and Grow, J.A., 1985, New York Bight fault: Geol. Soc. Amer. Bull.,v. 96, p.975-989.

Iselin, C.O'D., 1940, Preliminary report on long-period variations in the transport of the Gulf Stream System: Papers in Phys. Ocean. and Meteor., v. 8 No. 1, 40 pp.

Kafka, A.L., Schlesinger-Miller, E.A. , and Barstow, N.L., 1985, Earthquake activity in the greater New York City area: Magnitudes, seismicity, and geologic structures: Bull. Seis. Soc. Amer., v. 75, no. 5, p. 1285-1300.

Kaye, C.A., and Stuckey, G.W., 1973, Nodal tidal cycle of 18.6 yr: Its importance in sea-level curves of the east coast of the United States and its value in explaining long-term sea-level changes: Geology, v. 1, p. 141-144.

Keen, C.E., Keen, M.J., Nichols, B., Reid, I., Stockmal, G.S., Colman-Sadd, S. P., O'Brian, S.J., Miller, H., Quinlan, G., Williams, H., and Wright, J., 1986, Deep seismic reflection profile across the northern Appalachians: Geology, v. 14, p. 141-145.

Kerr, R.A., 1981, Assessing the risk of eastern U.S. earthquakes: Science, v. 214, p. 169-171.

King, P.B., 1969, Tectonic map of North America: U.S. Geol. Surv. map, scale 1:5,000,000.

King, P.B., 1970, Tectonics and geophysics of North America, in H. Johnson and B.L. Smith, eds., The megatectonics of continents and oceans, New Brunswick, Rutgers University Press, 282 pp.

Klitgord, K.D., Popenoe, P., and Schouten, H., 1984, Florida: A Jurassic transform plate boundary: J. Geophys. Res., v. 89, p. 7753-7772.

Klitgord, K.D., and Hutchinson, D.R., 1985, Distribution and geophysical signatures of early Mesozoic rift basins beneath the U.S. Atlantic continental margin, in Robinson, G.R., Jr., and Froelich, A.J., eds., Proceedings of the Second U.S. Geol. Survey Workshop on the early Mesozoic basins of the eastern United States: U.S. Geol. Survey Circular 946, Ch. 9.

Kondratyev, K. Ya., 1984, Book Review of Carbon Dioxide Review: 1982: EOS, v. 65, 382 p.

Kuenen, Ph.H., 1950, Marine Geology: New York, John Wiley & Sons, Inc., 568 pp.

Lambert, A., and Vanicek, P., 1979, Contemporary crustal movements in Canada: Can. J. Earth Sci., v. 16, p. 647-668.

Lepage, C.A., and Johnson, R.A., 1983, Earthquakes in Maine: October, 1975 to December, 1982: Maine Geol. Survey Open-File Map 83-2.

Lisitzin, E., 1958, Le niveau moyen de la mer: Bull. Inf. C.O.E.C., v. 10, p. 254-262.

Lisitzin, E., and Pattullo, J.G., 1961, The principal factors influencing the seasonal oscillation of sea level: J. Geophys. Res., v. 66, p. 845-852.

Lundgren, L., and Ebblin, C., 1972, Honey Hill fault in eastern Connecticut regional relations: Geol. Soc. Amer. Bull., v. 83, p. 2773-2794.

Meade, R.H., and Emery, K.O., 1971, Sea level as affected by river runoff, eastern United States: Science, v. 173, p. 425-428.

Mercer, J.H., 1978, West Antarctic ice sheet and CO_2 greenhouse effect: a threat of disaster: Nature, v. 271, p. 321-325.

Minard, J.P., 1969, Geology of the Sandy Hook quadrangle in Monmouth County, New Jersey: U.S. Geol. Surv. Bull., v. 1276, 43 p.

Murray, G. E., 1961, Geology of the Atlantic and Gulf Coastal Province of North America: New York, Harper & Brothers, 692 pp.
National Research Council (NRC), 1979, Carbon Dioxide and Climate: A Scientific Assessment: Washington, DC, National Academy Press.
Newman, W.S., Cinquemani, L.J., Pardi, R.R., and Marcus, L.F., 1980, Holocene deleveling of the United States east coast, in Mörner, N.-A., ed., Earth Rheology, Isostasy, and Eustasy: New York, John Wiley and Sons, Ltd., p. 449-463.
Niiler, P.P., and Richardson, W.S., 1973, Seasonal variability of the Florida Current: J. Mar. Res., v. 31, p. 144-167.
Pattullo, J.G., 1960, Seasonal variation in sea level in the Pacific Ocean during the International Geophysical Year, 1957-1958: J. Mar. Sci., v. 18, p. 168-184.
Peltier, W.R., 1984, The thickness of the continental lithosphere: J. Geophys. Res., v. 89, p. 11,303-11,316.
Peltier, W.R., 1986, Deglaciation induced vertical motion of the North American continent: J. Geophys. Res., in press.
Peltier, W.R., and Andrews, J.T., 1976, Glacial-isostatic adjustment - I.The forward problem: Geophys. J. R. Astr. Soc., v. 46, p. 605-646.
Peltier, W.R., Farrell, W.E., and Clark, J.A., 1978, Glacial isostasy and relative sea level: A global finite element model: Tectonophysics, v. 50, p. 81-110.
Peltier, W.R., and Wu, P., 1982, Mantle phase transitions and the free air gravity anomalies over Fennoscandia and Laurentia: Geophys. Res. Lett., v. 9, p. 731-734.
Perry, A.M., 1983, Estimating the greenhouse effect: EOS, v. 222, 1072p.
Pocklington, R., 1972, Secular changes in the ocean off Bermuda: J. Geophys. Res., v. 77, p. 6604-6607.
Quinlan, G. and C. Beaumont, 1981, A comparison of observed and theoretical postglacial relative sea level in Atlantic Canada: Can. J. Earth Sci., v. 18, p. 1146-1163.
Revelle, R.R., 1983, Probable future changes in sea level resulting from increased atmospheric carbon dioxide, in Changing Climate, Report of the Carbon Dioxide Assessment Committee, National Academy Press, Washington, DC., p. 433-448.
Roemmich, D., 1985, Sea level and the thermal variability of the ocean, in Glaciers, Ice Sheets, and Sea Level: Effect of a CO_2-Induced Climatic Change, National Academy Press, Washington, D.C., p. 104-115.
Roemmich, D., and Wunsch, C., 1984, Apparent changes in the climatic state of the deep North Atlantic Ocean: Nature, v. 307, p. 447-450.
Sbar, M.L., and Sykes, L.R., 1973, Contemporary compressive stress and seismicity in eastern North America: An example of intra-plate tectonics: Geol. Soc. Amer. Bull., v. 84, p. 1861-1882.
Scholz, C.H., 1972, Crustal movements in tectonic areas: Tectonophysics, v. 14, p. 201-217.
Seeber, L., Armbruster, J.G., and Bollinger, G.A., 1982, Large-scale patterns of seismicity before and after the 1886 South Carolina earthquake: Geology, v. 10, p. 382-386.
Shakal, A.F., and Toksoz, M.N., 1977, Earthquake hazard in New England: Science, v. 195, p. 171-173.

Smith, W.E.T., 1966, Earthquakes of eastern Canada and adjacent areas 1928-1959: Dominion Observ. Publ. 32, Ottawa, Ont., p. 87-121.

Sykes, L.R., 1978, Intraplate seismicity, reactivation of preexisting zones of weakness, alkaline magmatism, and other tectonism postdating continental fragmentation: Rev. Geophys. Space Phys., v. 16, p. 621-688.

Thompson, K.R., and Hazen, M.G., 1983, Interseasonal changes of wind stress and Ekman upwelling: North Atlantic, 1950-1980: Can. Tech. Rep. Fish. Aquat. Sci., No. 1214, 175 p.

Thorarinsson, S., 1940, Present glacier shrinkage and eustatic changes in sea level: Geogr. Ann., v. 12, p. 131-159.

Tyler, D.A., and Ladd, J.W., 1980, Vertical crustal movement in Maine, in Thompson, W.B., ed., New England seismotectonic study activities in Maine during fiscal year 1980, Maine Geol. Survey Report to USNRC, p. 99-153.

Vanicek, P., 1975, Vertical crustal movements in Nova Scotia as determined from scattered geodetic relevelings: Tectonophysics, v. 29, 183-189.

Vanicek, P., 1976, Pattern of recent vertical crustal movements in Maritime Canada: Can. J. Earth Sci., v. 13, p. 661-667.

Vanicek, P., 1978, To the problem of noise reduction in sea-level records used in vertical crustal movement detection: Phys. Earth Planet. Interiors, v. 17, p. 265-280.

Walcott, R.I., 1970, Isostatic response to loading of the crust in Canada: Can. J. Earth Sci., v. 7, p 716-727.

Walcott, R.I., 1972a, Past sea levels, eustasy and deformation of the earth: Quat. Res., v. 2, p. 1-14.

Walcott, R.I., 1972b, Late Quaternary vertical movements in eastern North America: quantitative evidence of glacio-isostatic rebound: Revs. Geophys. Space Phys., v. 10, p. 849-884.

Winkler, C.D., and Howard, J.D., 1977, Correlation of tectonically deformed shorelines on the southern Atlantic coastal plain: Geology, v. 5, p. 123-127.

Woollard, G.P., 1958, Areas of tectonic activity in the United States as indicated by earthquake epicenters: AGU Trans., v. 39, p. 1135-1150.

Wu, P., and Peltier, W.R., 1983, Glacial isostatic adjustment and the free air gravity anomaly as a constraint on deep mantle viscosity: Geophys. J. R. Astr. Soc., v. 74, p. 377-449.

Wu, P., and Peltier, W.R., 1984, Pleistocene deglaciation and the earth's rotation: A new analysis: Geophys. J. R. Astr. Soc., v. 76, p. 753-791.

Wunsch, C., 1972, Bermuda sea level in relation to tides, weather, and baroclinic fluctuations: Rev. Geophys. Space Phys., v. 10, p. 1-49.

Yang, J.-P., and Aggarwal Y.P., 1981, Seismotectonics of northeastern United States and adjacent Canada: J. Geophys. Res., v. 86, p. 4981-4998.

York, J.E., and Oliver, J.E, 1976, Cretaceous and Cenozoic faulting in eastern North America: Geol. Soc. Amer. Bull., v. 87, p. 1105-1114.