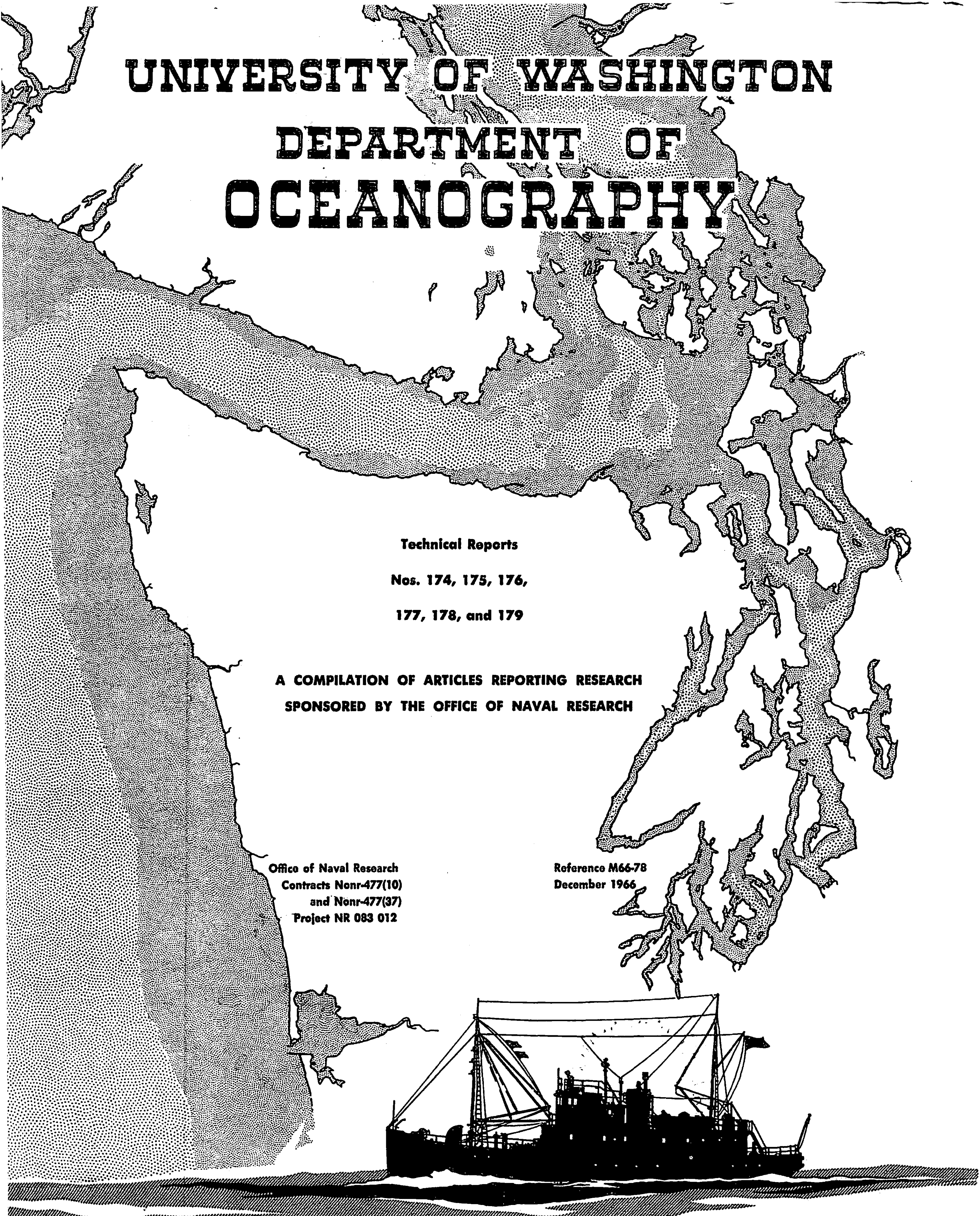
UNIVERSITY OF WASHINGTON
DEPARTMENT OF OCEANOGRAPHY



Technical Reports

Nos. 174, 175, 176,

177, 178, and 179

A COMPILATION OF ARTICLES REPORTING RESEARCH SPONSORED BY THE OFFICE OF NAVAL RESEARCH

Office of Naval Research
Contracts Nonr-477(10)
and Nonr-477(37)
Project NR 083 012

Reference M66-78
December 1966

SEATTLE, WASHINGTON 98105

UNIVERSITY OF WASHINGTON
DEPARTMENT OF OCEANOGRAPHY
Seattle, Washington 98105

Technical Reports

Nos. 174, 175, 176,

177, 178, and 179

A COMPILATION OF ARTICLES REPORTING RESEARCH SPONSORED BY THE OFFICE OF NAVAL RESEARCH

Office of Naval Research
Contracts Nonr-477(10)
and Nonr-477(37)
Project NR 083 012

Reference M66-78
December 1966

RHFleming.
RICHARD H. FLEMING
Chairman

ARTICLES REPORTING RESEARCH SPONSORED BY THE OFFICE OF NAVAL RESEARCH

Technical Report No. 174

IDENTIFICATION AND DETERMINATION OF ORGANIC ACIDS IN SEA WATER BY PARTITION CHROMATOGRAPHY, by Tadashiro Koyama and Thomas G. Thompson. The Journal of the Oceanographical Society of Japan, 20(5):209-220. 1964.

Technical Report No. 175

QUANTITIES OF ZOOPLANKTON AND PROPAGATION OF CALANUS FINMARCHICUS AT PERMANENT STATIONS ON THE NORWEGIAN COAST AND AT SPITSBERGEN, 1959-1962, by Ulf Lie. Fiskeridirektoratets Skrifter, Havundersøkelser, 13(8):5-19. 1965.

Technical Report No. 176

MAGNETIC AND PETROLOGIC STUDIES OF SEDIMENT FOUND ABOVE BASALT IN EXPERIMENTAL MOHOLE CORE EM7, by M. D. Fuller, C. G. A. Harrison, and Y. R. Nayudu. Bulletin, American Association of Petroleum Geologists, 50(3):566-573. 1966.

Technical Report No. 177

AN INSTRUMENT SYSTEM TO MEASURE BOUNDARY-LAYER CONDITIONS AT THE SEA FLOOR, by Richard W. Sternberg and Joe S. Creager. Marine Geology, 3(6): 475-482. 1965.

Technical Report No. 178

SOME CONSEQUENCES OF THE DECOMPOSITION OF ORGANIC MATTER IN LAKE NITINAT, AN ANOXIC FJORD, by Francis A. Richards, Joel D. Cline, William W. Broenkow, and Larry P. Atkinson. Limnology and Oceanography, 10(Suppl.): R185-R201. 1965.

Technical Report No. 179

LOW-FREQUENCY SEA LEVEL OSCILLATIONS ALONG THE PACIFIC COAST OF NORTH AMERICA, by Gunnar I. Roden. Journal of Geophysical Research, 71(20): 4755-4776. 1966.

UNIVERSITY OF WASHINGTON
DEPARTMENT OF OCEANOGRAPHY
TECHNICAL REPORT NO. 179

JOURNAL OF GEOPHYSICAL RESEARCH VOL. 71, NO. 20 OCTOBER 15, 1966

Low-Frequency Sea Level Oscillations along the Pacific Coast of North America[1]

GUNNAR I. RODEN

Department of Oceanography, University of Washington, Seattle

The information obtainable from monthly mean sea level records and the joint variation of sea level, atmospheric pressure, and sea temperature are analyzed. The spectrum of sea level is time invariant and shows significant peaks only at frequencies corresponding to the annual and semiannual oscillations. The bispectrum of sea level indicates a weak interaction of the annual frequency with itself, as well as with other frequencies. Secular sea level changes along the Pacific coast are related to land subsidence and uplift. The mean duration of positive and negative sea level anomalies from long-term monthly means is of the order of 3 months. Extreme durations vary between 10 and 34 months and are closely related to large-scale atmospheric disturbances. The areal coherence of nonperiodic sea level fluctuations is of the order of 1200 km. At temperate and high latitudes there is good coherence between nonperiodic sea level and atmospheric pressure oscillations. The response between these two variables varies from —0.9 cm/mb to —2.3 cm/mb in the frequency range between 1 and 6 cycles per year. The coherence between nonperiodic sea level and sea temperature fluctuations is high in tropical latitudes and low in polar latitudes. The response is largely independent of frequency, varying from 1.0 to 2.7 cm/°C. These findings are in agreement with the coefficient of thermal expansion of seawater and the depth of the mixed layer.

INTRODUCTION

Sea level records from stations along the Pacific coast of the United States date back to the 1850's, when the U. S. Coast Survey established tide stations at San Diego and San Francisco, California, and at Astoria, Oregon. Several stations have since been added, mostly at commercially important ports. The records cover a substantial period of time, and it is interesting to consider what oceanographically useful information can be extracted from them by modern statistical means.

The recording of sea level yields a time series that expresses the displacement of the sea *relative* to that of the land. The processes that take place in the earth's crust are usually slow in comparison with the day-to-day and month-to-month changes in water level but not necessarily with regard to the year-to-year and the secular changes. This has an important bearing upon the interpretation of records. Although short-term sea level oscillations almost certainly reflect fluctuations in water level, long-term trends tend to represent differential motion between the water and the land.

At any fixed location the sea level record is a complicated function of time, involving periodic components as well as continuous random fluctuations. The periodic motion is due to the gravitational effects of the sun-earth-moon system as well as to the effects of solar radiation upon the atmosphere and the ocean. The random fluctuations are principally of meteorological origin and reflect the effect of 'weather' upon the sea surface. If the periodic and random oscillations were truly independent of each other, they could be studied separately and without cross reference. This has been done in the past and has met with some success, notably in the prediction of tides. More recently, however, some evidence that the periodic and random fluctuations are not independent of each other has accumulated, leading to such phenomena as 'tidal cusps' in the spectrum [*Munk, Zetler, and Groves*, 1965]. It is pertinent, therefore, to ask what characteristics of sea level fluctuations can be determined that have a meaningful stability. In particular, the problems considered in this paper deal with (1) the dependence of the sea level spectrum upon time, (2) the stability of trends, (3) the bispectrum of sea level, (4) the mean and extreme durations of nonperiodic sea level fluc-

[1] Contribution 385 from the Department of Oceanography, University of Washington, Seattle.

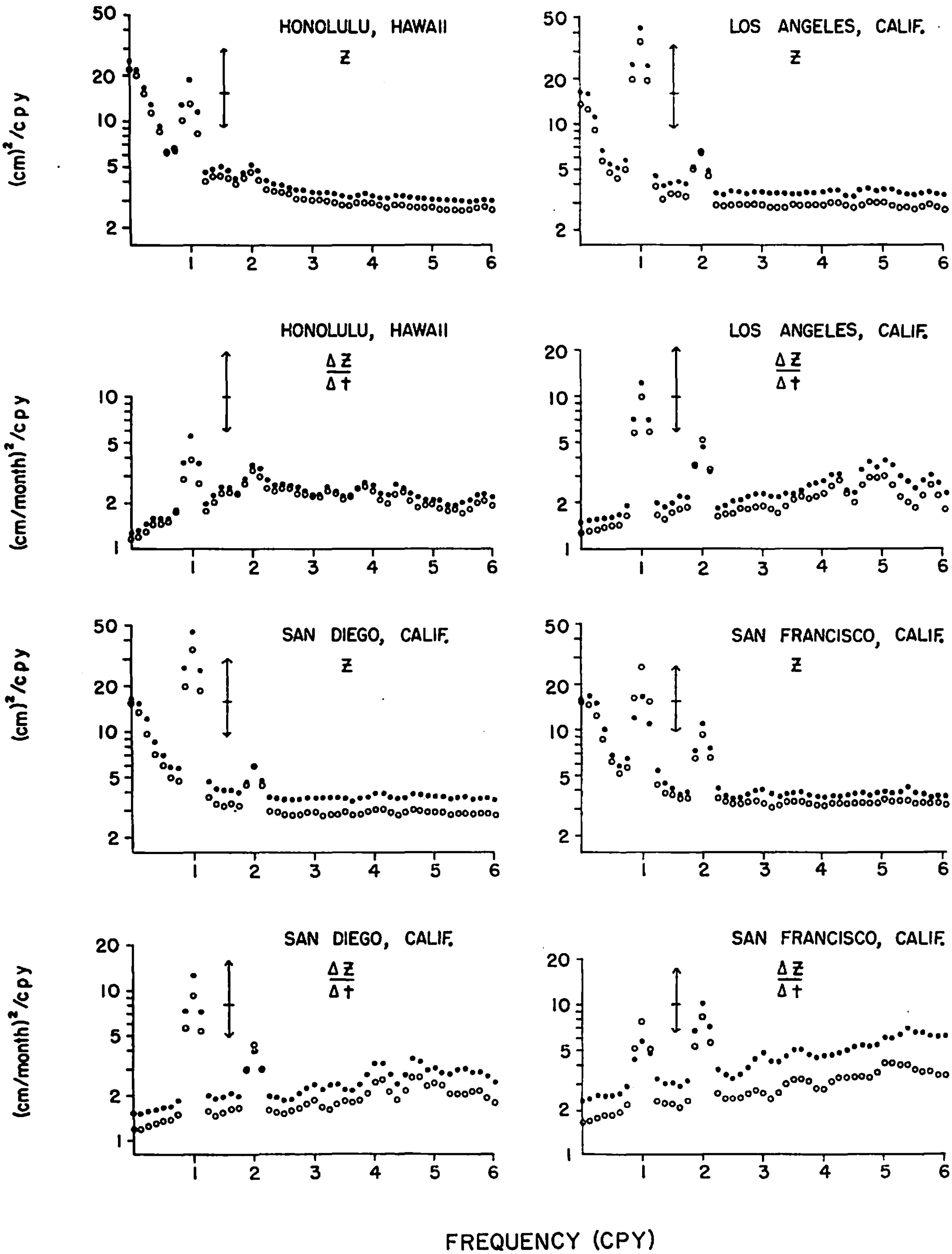
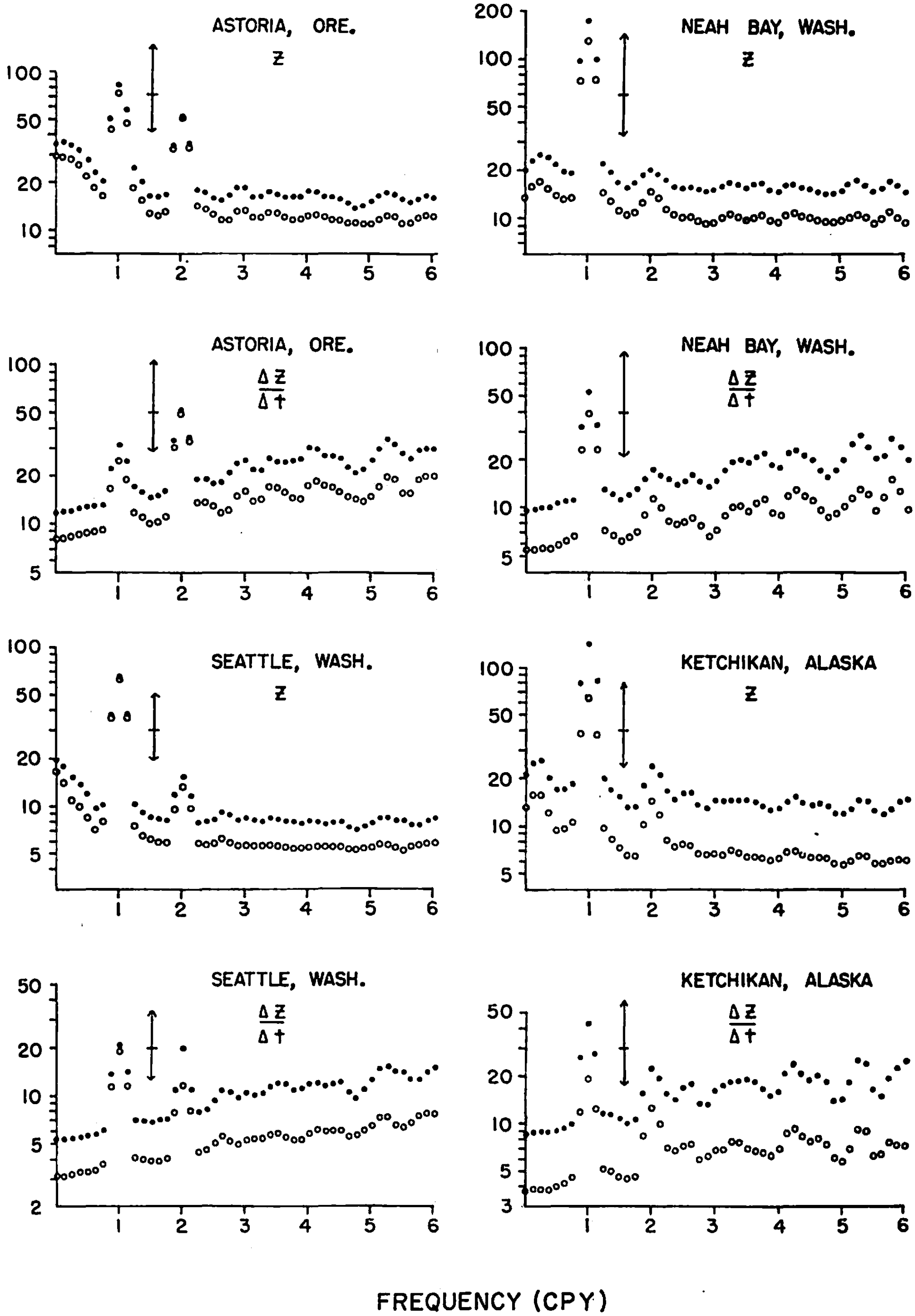


Fig. 1. Spectrums for sea level (z) and its time gradient ($\Delta z/\Delta t$). Dots refer to the observed records, circles to those from which the hydrostatic effect of atmospheric pressure has been eliminated. The arrows indicate the 95% confidence limits.


ASTORIA, ORE.
Z
NEAH BAY, WASH.
Z
ASTORIA, ORE.
ΔZ/Δt
NEAH BAY, WASH.
ΔZ/Δt
SEATTLE, WASH.
Z
KETCHIKAN, ALASKA
Z
SEATTLE, WASH.
ΔZ/Δt
KETCHIKAN, ALASKA
ΔZ/Δt
FREQUENCY (CPY)

tuations and associated probabilities, (5) the areal coherence of nonperiodic sea level fluctuations, and (6) the dependence of nonperiodic sea level fluctuations upon those of atmospheric pressure and sea temperature.

The Spectrum of Sea Level

The spectrum of sea level represents the distribution of 'energy,' or mean-square oscillation, over a range of frequencies. It is based on the assumption of a stationary, Gaussian, and ergodic process, which implies that all relevant information can be obtained from the second moments of a finite piece of record, irrespective of any origin in time [*Lee*, 1960; *Yaglom*, 1962]. If, as often happens with geophysical time series, the above assumptions are only partially fulfilled, the spectrum yields only limited information on the nature of sea level fluctuations. Characteristically, the energy associated with each frequency band in an ordinary spectrum is independent of the energy in any other, which is equivalent to assuming no interaction among different frequencies.

The sea level spectrum is roughly known in the frequency range between 30 cps and 1 cycle in 30 years. It consists of a series of more or less broad peaks superimposed upon a fluctuating level of random noise. The most prominent peaks are due to the diurnal and semidiurnal components of the tide and to sea and swell. Lesser peaks, having typically only 1% of the energy density of the sea-swell band, are due to surges, shelf waves, and surf beat [*Munk, Tucker, and Snodgrass,* 1957]. The peak corresponding to the annual sea level oscillation has about 10% of the energy density of the sea-swell band. On the high-frequency end of the spectrum, there is a very small peak due to capillary waves; its energy density is only 0.001% of that contained in the sea-swell band. There are also some peaks that occur in limited areas of the ocean. A 4-day oscillation has been reported from the equatorial Pacific [*Groves and Grivel,* 1962] and near Australia [*Hamon,* 1962], but it has not been found elsewhere.

Not all of the theoretically expected tidal components have been located in the sea level records. Although the problem of resolution has been solved in some instances [*Munk and Hasselmann,* 1964], there remains the serious problem of interaction between the continuous 'noise' spectrum and the tidal 'line' spectrum. One manifestation of such interaction is the cusplike rise of the continuum toward tidal frequencies [*Munk, Zetler, and Groves,* 1965]. At frequencies much lower than 0.1 cycle per year (cpy), the determination of the sea level spectrum presents considerable difficulties. Part of it is undoubtedly due to the shortness of the available records, but a more serious deficiency is the apparent impossibility of separating the motion of the sea from that of the land. It is therefore hardly surprising that the theoretically predicted 18.6-year period due to the moon's orbital motion has not yet been found in any of the analyzed spectrums.

In the following we shall consider in detail the spectrum in the frequency range between 0 and 6 cpy, with a resolution of 0.125 cpy. The spectrums were calculated from monthly mean sea levels obtained from hourly values by an arithmetic average. This essentially filters out high-frequency fluctuations.[2] In addition, the linear trend was removed from the records to make the time series as stationary as possible.

In Figure 1 the dots refer to the original records and the open circles to those from which the hydrostatic effect of atmospheric pressure has been eliminated by assuming that a pressure increase of 1 mb decreases the sea level by 1 cm. This so-called 'inverted barometer effect,' first described by *Gissler* [1747], follows immediately from the hydrostatic equation. As expected, elimination of the atmospheric pressure effect decreases the total energy contained in the spectrums. The decrease is much more pronounced in high than in low latitudes owing to the greater variability of atmospheric pressure in the former region.

The only significant peaks in the spectrums shown here are those corresponding to the annual and, in some cases, to the semiannual oscillations. These peaks are essentially of meteorological and radiational origin, reflecting the shift of air masses from winter to summer and the annual cycle of solar heating. A characteristic feature of the sea level spectrum is

[2] It would have been better to apply a more effective filter to the hourly data to suppress high-frequency fluctuations. Unfortunately, I had access only to the published monthly means and not to the original hourly values.

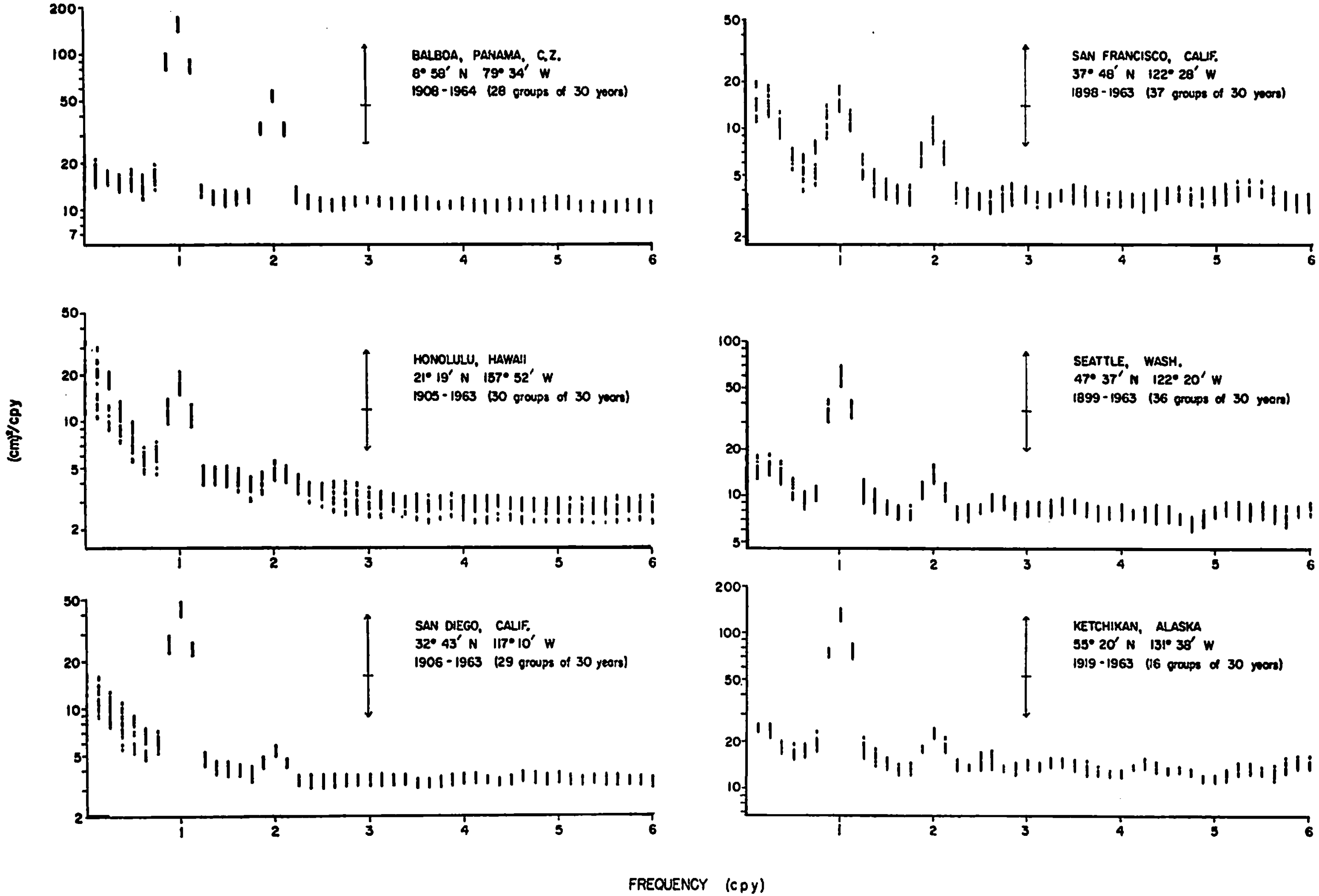


Fig. 2. Time variation of the sea level spectrum for successive 30-year periods. The total number of such periods in a record is referred to as *groups*. The arrows indicate the 95% confidence limits.

the increase in energy density with decreasing frequency. This phenomenon of a 'red spectrum,' a term based on analogy with certain light spectrums, is principally observed in big city ports such as Honolulu, San Diego, San Francisco, and Seattle. It is usually not characteristic of small fishing harbors such as Neah Bay, Washington. There remains the distinct, but yet unexplored, possibility that the growth of the city and the placing of structures on wharves and piers leads to the local settling of the ground on which the reference benchmark is located. If this is so, the 'redness' of the spectrums can be explained by slow changes of the earth's crust due to activities by man.

The redness of the sea level spectrum can be eliminated by studying its time derivative. Typical features of the time-derivative spectrum are that its energy density is low at low frequencies and that it increases with increasing frequency. To a certain extent, these results are expected. The time-derivative spectrum can be obtained from the original spectrum by multiplying it by frequency squared [*Lee,* 1960]. If the energy density of the original spectrum decreased with increasing frequency, the energy density of the time-derivative spectrum should be exactly zero at zero frequency and should approach zero at high frequencies. The actual time-derivative spectrums shown in Figure 1 differ from this for two reasons. First, with a finite record length, the first spectral estimate is not made at exactly zero but at a frequency centered at ¼ times the resolution [*Munk, Snodgrass, and Tucker,* 1959]; hence the energy density is not zero at the low-frequency end of the spectrum. Second, because of environmental high-frequency noise and measurement errors, the energy density of the sea level spectrum does not decrease with increasing frequency but approaches a rather constant 'background noise' level. Thus there is a tendency for the observed time-derivative spectrum to increase as frequency squared, at high frequencies.

A problem of considerable practical importance concerns the time variation of the sea level spectrum. In a truly stationary time series the spectrum should be independent of any translation in time. This means that spectrums computed over different time intervals should be identical. To find out whether this actually happens, spectrums were computed for successive 30-year intervals for all stations having 45 or more years of continuous record. The results are shown in Figure 2. Each ordinate consists of a series of points (sometimes merged into a line) equal to the total number of 30-year periods in the record. The frequency resolution is 0.125 cpy and the arrows indicate the 95% confidence limits [*Blackman and Tukey,* 1959]. The outstanding feature is that at most stations the energy density fluctuations are statistically insignificant. The sea level spectrums for these stations are therefore time invariant. This is a noteworthy result because it permits us to regard the sea level fluctuations as stationary.

The Stability of Trends

Sea level is recorded *relative* to a fixed benchmark. Movements of both the water and the land may contribute to its change. As far as long-term, or secular, changes are concerned, only this differential motion between water and land can be determined accurately. The causes for the secular sea level change cannot be found from a study of tide-gage records alone; instead, comparable records of atmospheric and oceanic variables and repeated geodetic surveys must be studied. At present, only the former method is feasible. If the climatological records contain no long-term trends, we can assume that land motion caused the observed sea level change. If trends do exist in the climatological records, the relative contribution to the sea level trend may be determined from a knowledge of the response factors between sea level and the climatic variables. On very long time scales the effect of glacial melting must be considered. If the melting occurs on a global scale, the sea level changes caused by this process should be the same along all coasts. There is, however, no conclusive evidence of such an occurrence during the last 100 years.

Along the west coast of the United States, climatic fluctuations over the past century have been stationary and are noted for the absence of long-term trends [*Roden,* 1965, 1966]. The rather pronounced sea level trends observed at some stations are therefore due to land motion. Evidence of topographic movements of the land in many coastal regions has been given [*Marmer,* 1951]. In southeast Alaska, for example, the land has risen from 1 to 4 m during the past century [*Hicks and Shofnos,* 1965], a rate of rise which exceeds even the 0.5- to 1-m per century rate observed in Fennoscandia [*Heiskanen and Vening Meinesz,* 1958]. The uplift in these regions is thought to be the result of postglacial isostatic adjustment of the earth's crust.

In contrast to these slow changes, earthquakes may lead to sudden displacements of large pieces of land. The great Alaska earthquake of March 27, 1964, affected an area in excess of 10^5 km^2 and changed land elevations by 2.5 m in places [*Grantz et al.,* 1964]. Nor are endeavors by man always negligible in bringing about local movements of the land on which benchmarks are located. The possible effect of city growth on land settlement was mentioned above. There is evidence in the Panama Canal Zone that the excavation and subsequent filling with water of Gatun Lake decreased the local benchmark elevations by several centimeters between 1908 and 1938 [*Sheldon,* 1938; *Roden,* 1963*b*]. The exploitation of oil reserves off Terminal Island, California, resulted in a rather constant land subsidence rate of 5 cm/year between 1942 and 1959, the harbor facilities (where the benchmark is located) literally sinking into the sea. Fortunately, this trend has now been arrested by pumping water into the partially depleted oil reservoirs. All this serves to emphasize that, along the west coast of North America, tectonic forces and activities by man have to a considerable degree affected the stability of benchmarks to which sea levels are referred.

It is of interest to inquire into the stability of sea level trends. Obviously, the trend cannot be a constant for all time; otherwise the sea level might increase or decrease to infinity. A particular piece of record may be fitted with a trend, but its value will frequently differ from that computed for any other comparable record length. To see what the time variations are, I computed the *linear* trend for successive 30-year periods for all stations having 45 or more years of continuous record. The mean month-to-month variation was eliminated by subtraction before starting the computations in order to avoid interference between periodic and random components. A noteworthy feature of the resulting nonperiodic sea level fluctuations is that they are autocorrelated. The autocorrelations are of an exponential type and may, to the first approximation, be fitted by a simple exponential curve [*Roden,* 1960]. Denoting the sea level fluctuations by $z(t)$, the decay constant of the autocorrelation by c, the total record length by T, the variance by m_0, and time by t yields the linear trend

$$b = \frac{6c^2}{c^2T^2 + 6cT + 12} \sum_{t=1}^{T} \left(\frac{2t}{T} - 1\right) z(t) \quad (1)$$

and the mean-square error of the estimate

$$\epsilon^2 = \frac{24cm_0}{T(c^2T^2 + 6cT + 12)} \quad (2)$$

according to *Zadeh and Ragazzini* [1950].

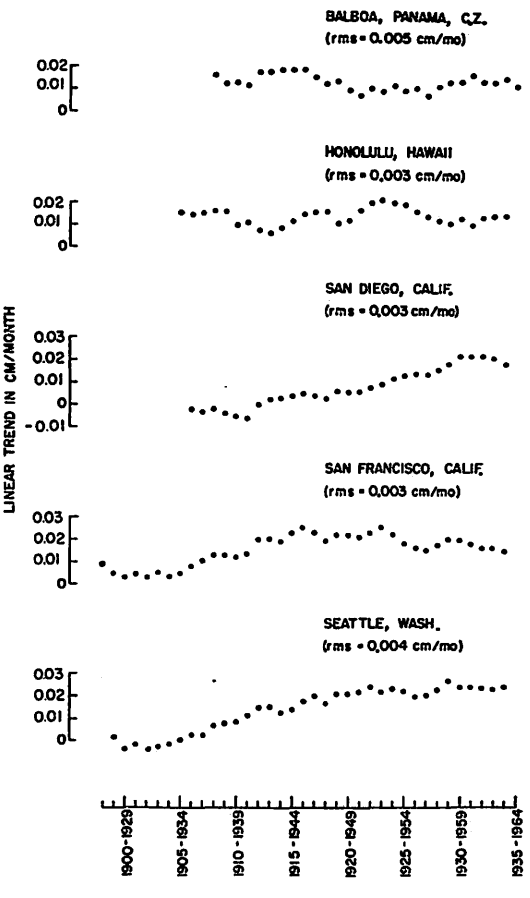


Fig. 3. Time variation of the linear sea level trend for successive 30-year periods. The rms error is a constant, to the decimal points given here.

The results of the computations are shown in Figure 3. The linear trend computed for successive 30-year periods is not constant but varies more or less irregularly with time. The rms error of the estimate also changes with time, but its variation is so small that it is a constant to the decimal places given in the figure. At Balboa, Panama, the irregular fluctuations are small, falling within the limits of the rms error; hence the trend may be regarded as constant. At Honolulu, Hawaii, the trend fluctuates widely and shows no apparent pattern. At San Diego, California, the time change of the linear trend is almost constant, suggesting a parabolic increase in sea level. At San Francisco, California, and at Seattle, Washington, the sea level increase appears to have been parabolic in the early part of the record and linear later. The time change of the trend thus appears to be different for each station, which largely suggests local land motion rather than climatic variation as the cause of the observed secular sea level change.

The Bispectrum of Sea Level

To a first approximation the sea level fluctuations may be regarded as a linear superposition of statistically independent waves. This leads to the concept of the (ordinary) spectrum, which is fully described by the second moments of distribution. If the process under consideration is nonstationary, or if there is interaction among different frequencies, the ordinary spectrum no longer suffices. The simplest extension is to study interaction between *pairs* of frequencies, which leads to the bispectrum [*Hasselmann et al.*, 1963; *Haubrich*, 1965].

The bispectrum is completely described by the third moments of distribution. A fundamental measure of the third moment of distribution is the bicorrelation, defined by

$$\psi(\tau_1, \tau_2) = \langle z(t)z(t + \tau_1)z(t + \tau_2)\rangle \qquad (3)$$

where $z(t)$ denotes sea level as a function of time t, τ_1 and τ_2 are lags, and the angle brackets denote the average with respect to time. Because of the various symmetries involved, the bicorrelation is fully determined in the octant $\tau_1 > \tau_2 > 0$. Note that $\psi(0, 0)$ denotes the mean cube power associated with the sea level fluctuations, which is related to the coefficient of skewness by

$$\gamma = \frac{\psi(0, 0)}{[\psi(0)]^{3/2}} \qquad (4)$$

where $\psi(0)$ is the variance of the record, obtained from $\langle z(t)^2\rangle$.

The bispectrum can be obtained from a two-dimensional Fourier transform of the bicorrelation as follows:

$$S(\omega_1, \omega_2) = \frac{1}{4\pi^2} \iint_{-\infty}^{\infty} \psi(\tau_1, \tau_2)F(\tau_1, \tau_2) \cdot e^{-i\omega_1\tau_1 - i\omega_2\tau_2}\, d\tau_1, d\tau_2 \qquad (5)$$

where $\omega_1 = 2\pi f_1$ and $\omega_2 = 2\pi f_2$ denote frequency and $F(\tau_1, \tau_2)$ is a suitable two-dimensional fading function which is used to avoid serious side bands. No two-dimensional fading function appears to have been worked out, though to a first approximation one might try a generalized 'Tukey fader' of the type

$$F(\tau_1, \tau_2) = \frac{1}{4}\left(1 + \cos\frac{\pi\tau_1}{\tau_{\max}}\right)\left(1 + \cos\frac{\pi\tau_2}{\tau_{\max}}\right) \qquad (6)$$

where $\tau_{\max}$ denotes the maximum lag used in the bicorrelation.

Thus the calculation of the bispectrum closely parallels that of the ordinary spectrum. The chief drawback to this straightforward method is that it is wasteful of computer time because of the number of multiplications involved. A different approach to obtain the bispectrum is sometimes used [*Haubrich*, 1965]. The original time series, $z(t)$, is divided into n groups of length N each. For each of the groups the Fourier transform is computed from

$$Z_j(\omega) = \frac{1}{(2\pi N)^{1/2}} \sum_{t'=1}^{N} e^{-i\omega t'} z_j(t')W_N(t') \quad \text{for} \quad \begin{array}{l} j = 1, 2, \cdots, n \\ t' = 1, 2, \cdots, N \end{array} \qquad (7)$$

where $W_N(t')$ is a weighting function given by

$$W_N(t') = 3^{1/2}(2t' - 1)/N \quad \text{for} \quad t' = 1, 2, \cdots, N/2 \qquad (8)$$

The bispectrum is then obtained from the *average* triple product

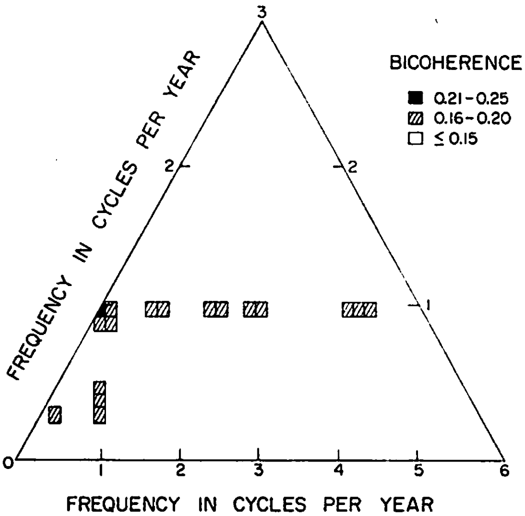


Fig. 4. Bicoherence of sea level. For clarity, only values 0.15 or larger have been entered. Note the ridges extending from the annual frequency pair (1, 1).

$$S(\omega_1, \omega_2) = \frac{\sum_{j=1}^{n} Z_j(\omega_1) Z_j(\omega_2) \bar{Z}_j(\omega_1 + \omega_2)}{n} \tag{9}$$

where the bar denotes the complex conjugate.

Since the bispectrum is complex, it is of advantage to express it in polar form. In analogy with the ordinary spectrum, the *bicoherence* is obtained from

$$R^2(\omega_1, \omega_2) = \frac{n^3 \, |S(\omega_1, \omega_2)|^2}{\sum_{j=1}^{n} |Z_j(\omega_1)|^2 \sum_{j=1}^{n} |Z_j(\omega_2)|^2 \sum_{j=1}^{n} |Z_j(\omega_1 + \omega_2)|^2} \tag{10}$$

and the *biphase* from

$$\pi(\omega_1, \omega_2) = \tan^{-1} \frac{\mathrm{Im}\,[S(\omega_1, \omega_2)]}{\mathrm{Re}\,[S(\omega_1, \omega_2)]} \tag{11}$$

The bicoherence and the biphase express the degree of interaction between the frequencies ω_1, ω_2, and $\omega_1 + \omega_2$. If the bicoherence equals 1, the interaction is perfect; if it equals 0, there is no interaction. For most geophysical time series, the values will lie somewhere in between.

In practical analysis, difficulties arise because of discrete sampling and a finite record length. If samples are taken Δt time units apart, the Nyquist frequency is given by $f_N = (2\Delta t)^{-1}$. If there is any nonzero energy density associated with frequencies above the Nyquist frequency, it will be folded back toward lower frequencies. This phenomenon is called 'aliasing' and results from undersampling a function [*Hamming*, 1962]. To avoid aliasing in the bispectrum, the sampling must be frequent enough to ensure that the bispectrum is small at all frequencies outside the region $|f_1| \le f_N$, $|f_2| \le f_N$, $|f_1 + f_2| \le f_N$. On the other hand, because of the various symmetries involved, the bispectrum (of a continuous and infinitely long record) is determined by its values in the octant $0 \le f_1 < \infty$, $0 \le f_2 \le f_1$. Multiplying these two regions together yields a triangular region for which the bispectrum of a discrete and finite process is determined [*Hasselmann et al.*, 1963]. The triangular region is defined by the frequency pairs (0, 0), $(f_N, 0)$, and $(\frac{1}{2}f_N, \frac{1}{2}f_N)$.

The bicoherence and the biphase were computed for all stations with more than 45 years of record. A typical example is shown in Figure 4. The station selected is Seattle, Washington, but the essential features are repeated at all other stations. Over the whole triangular region, the bicoherence is remarkably low, indicating that no strong interaction takes place between pairs of frequencies. The highest value of the bicoherence (0.25) occurs at the frequency pair (1, 1). This suggests that there is some weak interaction of the annual peak with itself to produce a semiannual peak. The ridges extending from the frequency pair (1, 1) might indicate that there is also interaction of the annual frequency with most other frequencies, but, as the confidence limits for the bicoherence have not yet been worked out, the significance of this low ridge cannot be determined. The biphase (not shown) at the frequency pair (1, 1) is 5°, and along the ridges it fluctuates not more than 15° from zero. This means that the interactions are essentially in phase. Where the bicoherence is less than 0.15, the biphase fluctuates randomly, as expected.

Duration of Sea Level Oscillations Above and Below Arbitrarily Selected Levels

In many applications it is important to know the mean and extreme durations of sea level fluctuations. Suppose we have a record of monthly mean sea level from which the mean annual oscillation and the long-term trend have been eliminated and we wish to determine the duration of the remaining fluctuations above or below some level of interest. We may draw a horizontal line at that level and divide the record into two separate parts. The upper part will consist of a sequence of random pulses of varying duration and the lower part will furnish another such sequence. For each, we count the duration of the individual pulses and determine the probability function. From it we may compute the mean, the variance, and other relevant statistical parameters.

If the record under consideration is stationary and Gaussian, the *mean* duration can be determined from a knowledge of the variances of the original variable and its time derivative [*Rice*, 1958; *Longuet-Higgins*, 1962]. Let $z(t)$ denote sea level, and $z'(t)$ its first derivative with respect to time. The mean duration above an arbitrarily fixed level I (or below $-I$) is then given by

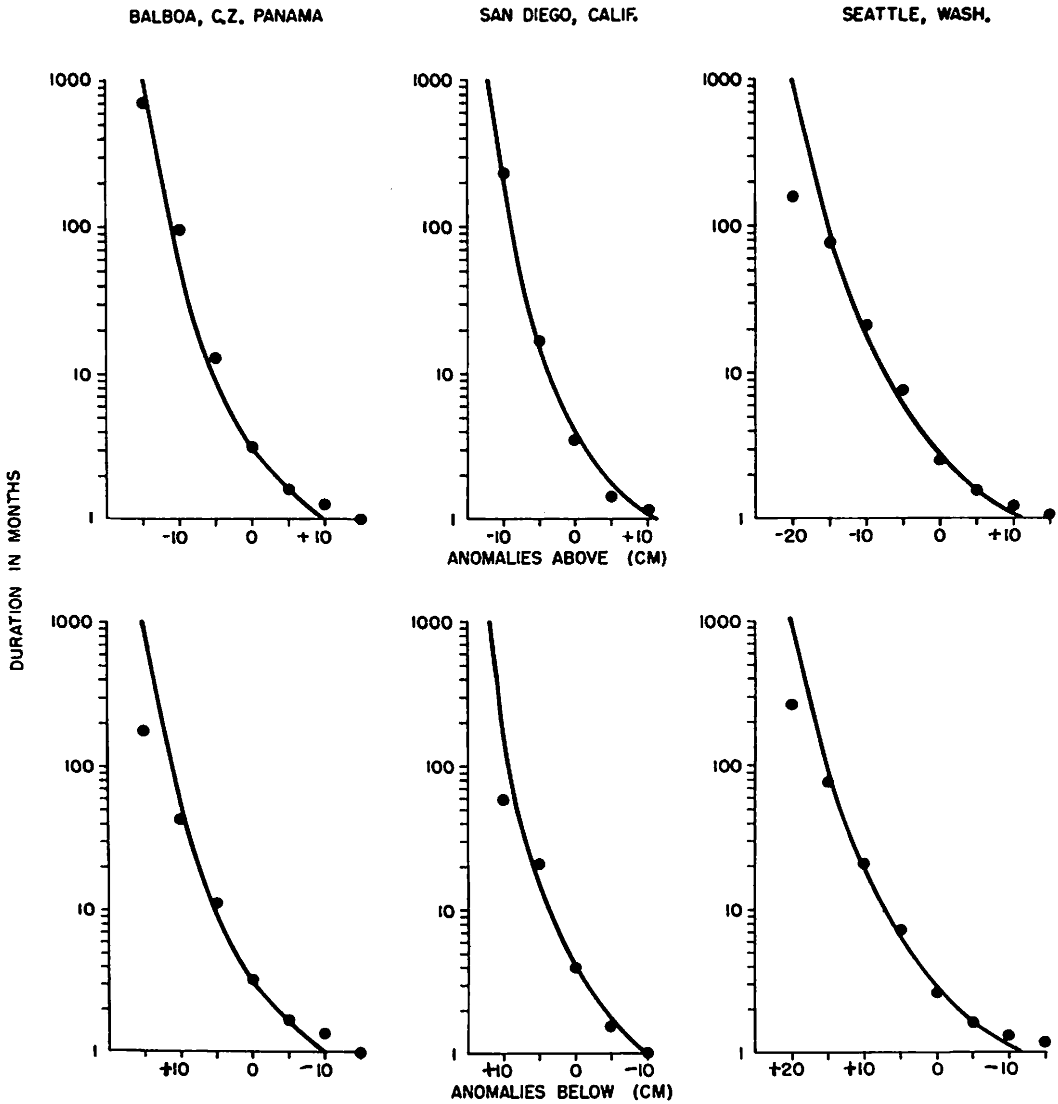


Fig. 5. Observed (dots) and theoretical (solid curve) mean durations of sea level anomalies (from long-term monthly means).

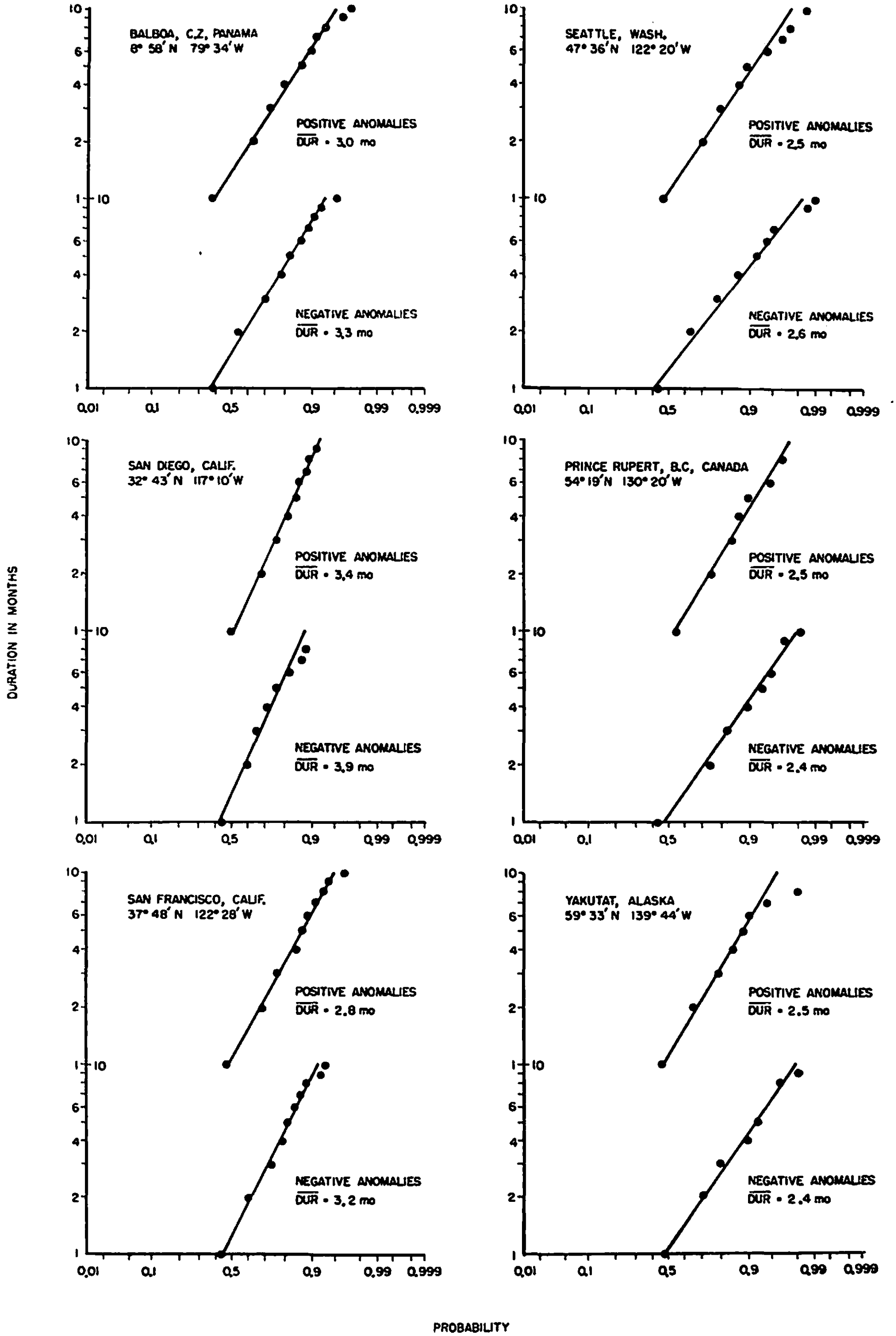


Fig. 6. Cumulative probability distributions for the duration of positive and negative sea level anomalies (from long-term monthly means). Dots refer to the observed probabilities, to which the straight line has been fitted by inspection.

$$\mathrm{DUR}\,(z(t)) = \pi\left(\frac{m_0}{m_2}\right)^{1/2} \exp\,(I^2/2m_0) \cdot\left(1 - \frac{2}{(2\pi)^{1/2}}\int_0^{I/m_0^{1/2}} e^{-t^2/2}\,dt\right) \qquad (12)$$

where $m_0 = \langle(z(t)^2)\rangle$ and $m_2 = \langle(z'(t)^2\rangle$ are the variances of $z(t)$ and $z'(t)$, respectively.

The agreement between calculated and observed mean durations is shown in Figure 5. The dots refer to the observations and the curves were derived from (12). The agreement is reasonably good. The mean duration above a fixed level decreases rapidly as that level increases, confirming the empirically known fact that large anomalies do not last long. The mean duration below a fixed level shows similar features. The average duration above or below the zero level is of the order of 3 months, and it decreases slightly with increasing latitude.

The cumulative probability distributions for the duration of positive and negative sea level anomalies (from long-term monthly means) are shown in Figure 6. The ordinate gives the duration on a logarithmic scale and the abscissa denotes the probability on a normal scale. The dots refer to the observed probabilities and the straight line was fitted to them by inspection. The outstanding feature is that, for durations not exceeding two or three times the mean duration, the cumulative probability distributions are log-normal. This is a characteristic feature not only of sea level but also of atmospheric pressure and temperature fluctuations [*Roden,* 1964*a,* 1965].

Having determined the mean and the probability distribution of the duration, it is of interest to investigate other possible useful measures that characterize the process. Suppose we want to know whether the departures from the mean duration are persistent, i.e., whether one long duration is likely to be followed by another. We can solve the problem by computing the autocorrelation for the random sequence of duration pulses. This was done for several stations. At all of them the autocorrelation consisted of a large spike at the origin, followed by very small random fluctuations. The physical meaning of this is that the duration pulses are *not* persistent; on the average, therefore, one long duration is not likely to be followed by another. This makes it difficult if not impossible to predict the future duration from a knowledge of the past durations.

In studies of historical records of nonperiodic sea level fluctuations, the question arises as to the *extreme duration* of such fluctuations above and below specified levels. At present, this problem can be solved only empirically. In Table 1 are listed the observed extreme durations of positive and negative anomalies. Positive anomalies last from 10 to 34 months, and negative ones from 12 to 19 months. There is a tendency for the extreme durations to be slightly less at high than at low latitudes. An examination of the dates of occurrence of the extremes brings out the interesting fact that they occurred more or less simultaneously over long distances. This strongly suggests that

TABLE 1. Observed Extreme Durations, in Months, of Sea Level Anomalies from Long-Term Monthly Means

Dates of occurrence are also given. The reference period is from 1926 to 1963 except for Neah Bay (1935-1963) and Sitka (1938-1963).

Station	N Latitude	W Longitude	Positive Anomalies: Months	Positive Anomalies: Dates	Negative Anomalies: Months	Negative Anomalies: Dates
Balboa, Canal Zone	8°58′	79°34′	28	Jan 1957–Apr 1959	15	Mar 1949–May 1950
San Diego, Calif.	32°43′	117°10′	34	Apr 1957–Jan 1960	19	Oct 1932–Apr 1934
La Jolla, Calif.	32°42′	117°15′	34	Apr 1957–Jan 1960	18	Sep 1932–Feb 1934
Los Angeles, Calif.	33°43′	118°16′	28	Oct 1957–Jan 1960	18	Sep 1932–Feb 1934
San Francisco, Calif.	37°48′	122°28′	23	Apr 1957–Feb 1959	18	Jun 1954–Nov 1955
Seattle, Wash.	47°36′	122°20′	15	Dec 1957–Feb 1959	16	Apr 1938–Jul 1939
Neah Bay, Wash.	48°22′	124°37′	17	Sep 1940–May 1942	16	Apr 1938–Jul 1939
Ketchikan, Alaska	55°20′	131°38′	18	Dec 1940–May 1942	13	Jul 1961–Jul 1962
Sitka, Alaska	57°03′	135°20′	10	Dec 1940–Sep 1941	12	Nov 1950–Oct 1951

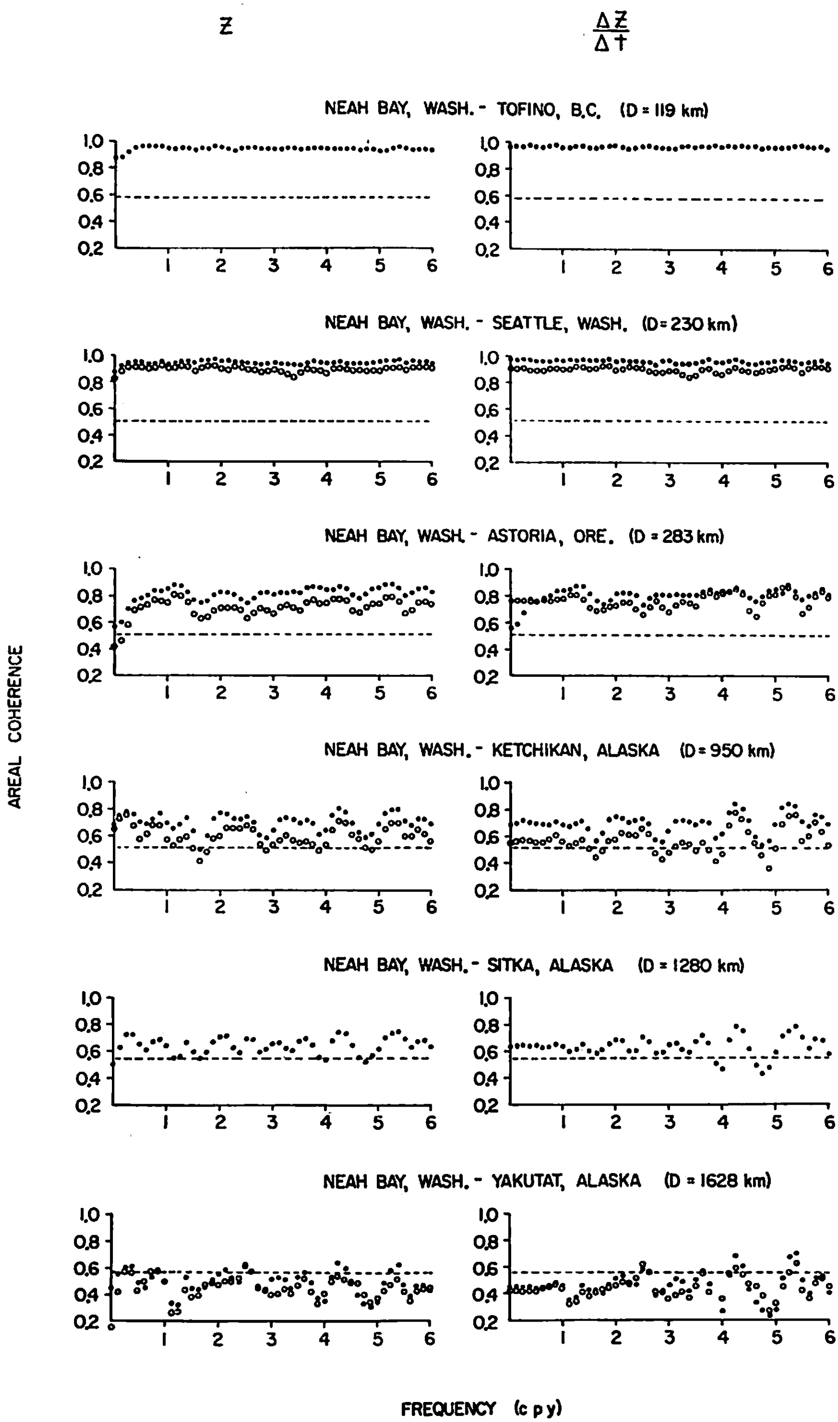


Fig. 7. Areal coherence of nonperiodic fluctuations of sea level (z) and its time gradient ($\Delta z/\Delta t$). D denotes the shortest distance between station pairs. Dots refer to the observed records, circles to those from which the hydrostatic effect of atmospheric pressure has been eliminated (where possible). The dashed line refers to the 95% confidence limit.

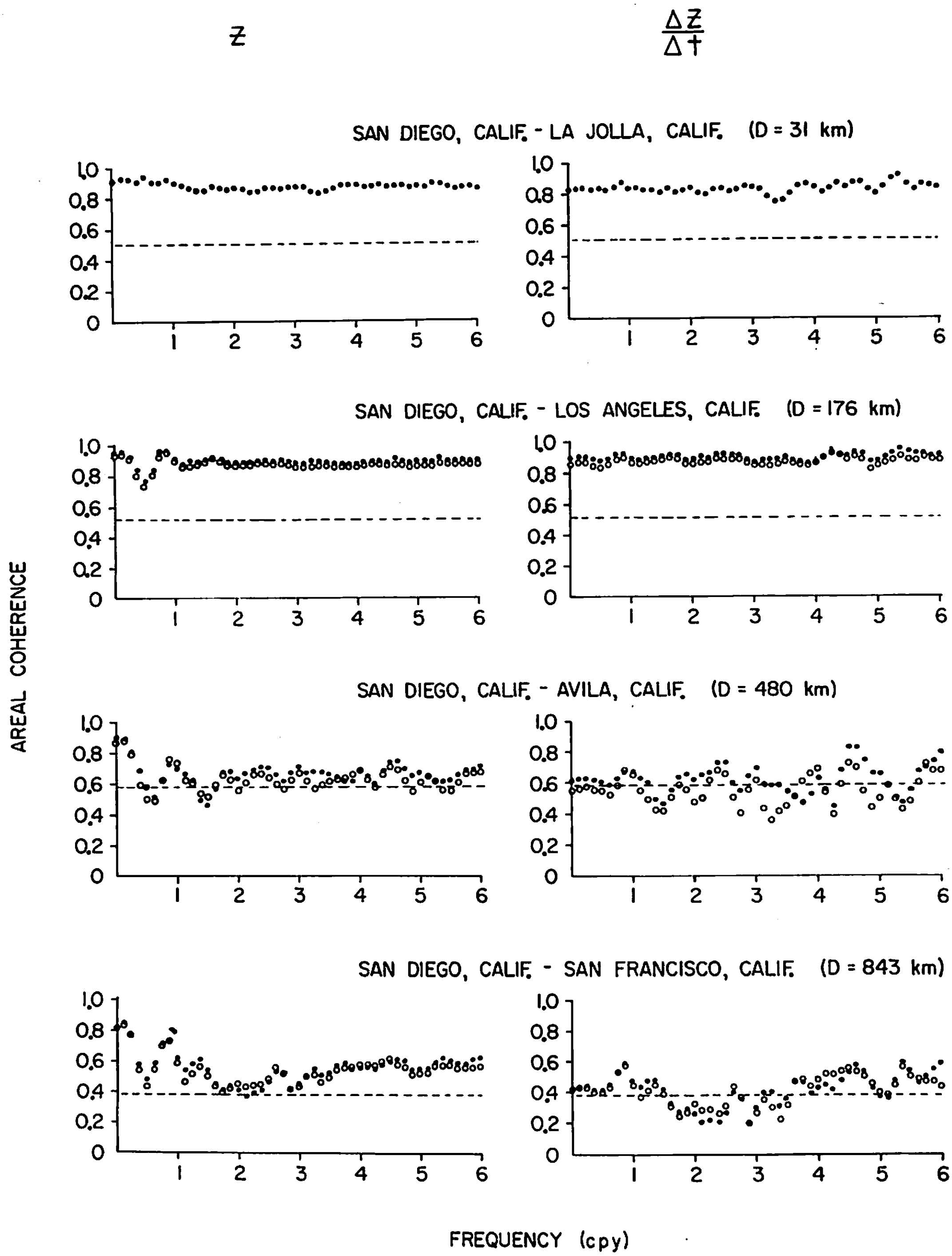


Fig. 8. Areal coherence of nonperiodic fluctuations of sea level (z) and its time gradient ($\Delta z/\Delta t$). D denotes the shortest distance between station pairs. Dots refer to the observed records, circles to those from which the hydrostatic effect of atmospheric pressure has been eliminated (where possible). The dashed line refers to the 95% confidence limit.

there is some relation to large-scale atmospheric pressure disturbances. The winters of 1940-1941, 1941-1942 and 1957-1958, 1958-1959 are a case in point. During these winters, the Aleutian low pressure was unusually strong and considerably south and east of its normal position [*Namias*, 1959]. In coastal regions, atmospheric pressures were below normal, southerly winds extended to lower than usual latitudes, and temperature and precipitation were above normal. Because the sea surface responds inversely to atmospheric pressure and directly to the heat content of the water, and because the effect of the earth's rotation leads southerly winds to pile up water along the coast in the northern hemisphere, the abnormally high sea levels observed during this period are not surprising. More interesting is the fact that throughout the summer the sea levels remained above average throughout the summer and the atmospheric circulation was normal. This fact suggests that disturbances in the ocean take longer to decay than in the atmosphere. If the ratio of the mean duration of sea level to the mean duration of atmospheric pressure anomalies [*Roden*, 1965] is any indication of this relationship, the disturbances in the ocean should last 1.5 times as long as those in the atmosphere.

Prolonged periods of below average sea levels along the west coast of the United States might be expected when the atmospheric pressures are unusually high, when the northerly winds are stronger than normal, and when the sea temperatures are below average. Such meteorological conditions prevailed, for example, during 1932 and 1933 along the southern California coast [*Roden and Reid*, 1961], with the result that sea levels remained below normal for about a year and a half. The extreme durations of sea level anomalies thus seem to be intimately linked to large-scale atmospheric disturbances.

The Areal Coherence of Sea Level Fluctuations

It is frequently important to know how coherent the sea level fluctuations are in area. Knowledge of this may shed light on such practical problems as the representativeness of a single tide-gage station and the optimum and most economical spacing of water-level recorders in surveys. If the data from great numbers of densely and equally spaced recorders were available, the answer could be obtained from a detailed study of the wave number spectrum [*Longuet-Higgins*, 1957]. Such records have not been made, however. There are only about 20 unequally spaced tide-gage stations along the Pacific coast of the United States. The approach used here, therefore, is to take sea level records from which the mean month-to-month variation and the linear trend have been eliminated and to study the coherence of the remaining nonperiodic fluctuations between *pairs* of stations, keeping the variable distance as a parameter.

The change of coastwise coherence with distance from a given reference station is shown in Figures 7 and 8 for both sea level and its time gradient. Over distances of a few hundred kilometers the coherence is obviously high and is not much affected by atmospheric pressure. Over distances of the order of 1000 km, the coherence is significant at most frequencies between 0 and 6 cpy, though it is somewhat decreased when the atmospheric pressure effect is taken into account. When the distance between station pairs increases to about 1500 km, the coherence becomes statistically insignificant. The phase between sea level oscillations (not shown) did not differ significantly from 0° at any of the station pairs. This indicates that nonperiodic sea level fluctuations are in phase up to 1200 km from the reference station. Similar results were also obtained when investigating the areal coherence of sea level in the tropical Pacific [*Roden*, 1963*a*] and around the Japanese islands [*Roden*, 1964*b*]. Thus the dimensions of areal coherence of sea level are comparable to those of large-scale atmospheric disturbances.

How representative are observations from a single tide gage? The very high coherence (0.9) at nearby stations suggests that sea level observations made at a point are representative of conditions over at least 200 km. Even stations with dissimilar exposure, such as San Diego, California (enclosed bay), and La Jolla, California (open coast), yield almost identical results. For the frequency range and type of nonperiodic fluctuations considered here, the relevant information could be obtained by spacing the tide gages as much as 200 km apart.

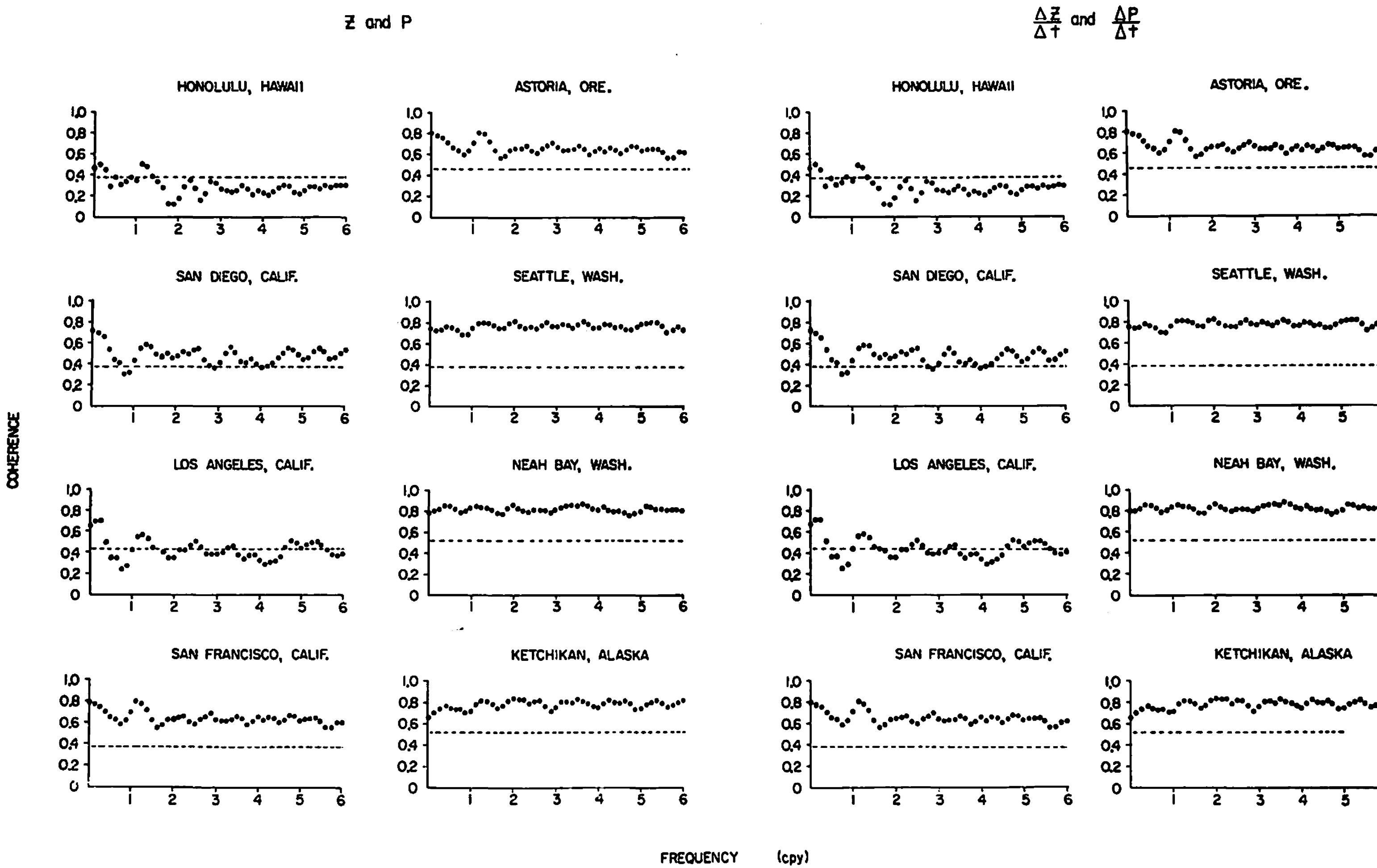


Fig. 9. Coherence between nonperiodic fluctuations of sea level (z) and atmospheric pressure (p), and for the corresponding time gradients. The dashed line indicates the 95% confidence limit.

The above is true as long as the macroenvironment in which the stations are located remains the same. In Figure 7, for example, all stations from Neah Bay, Washington, to Yakutat, Alaska, are affected by the northward-flowing Alaska current, and in Figure 8 the stations between San Diego and San Francisco, California, are influenced by the southward-setting California current. For a homogeneous macroenvironment, differences in local exposure seem to be unimportant.

In the vicinity of current boundaries, however, conditions are more complicated. Sea level fluctuations on one side of the boundary are not necessarily the same as or even similar to those on the other. The distances over which sea level fluctuations are coherent in current boundary regions are difficult to obtain, owing to the absence of suitably placed tide gages. Investigations of the Kuroshio-Oyashio boundary have shown a distance of less than 100 km [*Shoji*, 1961; *Roden*, 1964*b*].

Coherence and Response Between Sea Level and Atmospheric Pressure Oscillations

The knowledge that sea level responds inversely to atmospheric pressure and that the relationship can be derived from the hydrostatic equation is more than 200 years old [*Gissler*, 1747]. The use of the hydrostatic equation to determine the response factor is somewhat artificial because absence of motion in the sea and in the atmosphere is assumed. Under these conditions and the further assumption that the total volume of water in the world ocean remains the same, the response is given by [*Proudman*, 1953; *Fel'zenbaum*, 1960]

$$\frac{\partial z}{\partial t} = -\frac{1}{\rho g}\left[\frac{\partial p_a}{\partial t} - \frac{1}{\sigma}\frac{\partial}{\partial t}\int_\sigma p_a \, d\sigma\right] \qquad (13)$$

where z is sea level, ρ is (constant) water density, g is gravity, t is time, p_a is atmospheric pressure, and σ is the total area covered by the world ocean. The important feature in this equation is that the response depends upon the *difference* between the local pressure change and the change of the mean pressure over all the oceans. If the latter were zero, a pressure increase by 1 mb would decrease the sea surface elevation by approximately 1 cm. Accurate estimates of the mean pressure change over the world ocean are difficult to obtain, however, owing to the paucity of observations in the southern hemisphere, and the hydrostatic response of sea level to atmospheric pressure can be determined only within the limits of a constant.

The coherence between the nonseasonal fluctuations of sea level and atmospheric pressure is shown in Figure 9. The coherence is low in tropical latitudes, moderate in subtropical latitudes, and high in temperate and high latitudes. This is not surprising because atmospheric pressure fluctuations are much larger in polar than in tropical regions. At most stations the coherence fluctuates around a constant level in the frequency range between 0 and 6 cpy. This, and the equally constant phase of 180° (not shown), indicate an inverse response between sea level and atmospheric pressure, as expected from the hydrostatic equation. The magnitude of the response for different frequencies could be determined formally from

$$\mathrm{resp}_{zp}(\omega) = \left[\frac{E_z(\omega)}{E_p(\omega)}\right]^{1/2} C_{zp}(\omega) \qquad (14)$$

where $E_z(\omega)$ and $E_p(\omega)$ are the energy density spectrums of the nonperiodic sea level and atmospheric pressure fluctuations, respectively, and where $C_{zp}(\omega)$ denotes the coherence between these two variables. The straightforward use of (14) presents some difficulties. We have seen that the coherence is almost independent of frequency. The same can be said for the power spectrums of nonperiodic pressure fluctuations [*Roden*, 1965]. The spectrums of nonperiodic sea level fluctuations, on the other hand, are decidedly red, showing a concentration of spectral energy at low frequencies

TABLE 2. Frequency Response between Nonperiodic Sea Level and Atmospheric Pressure Fluctuations, in cm/mb

Frequency, cpy	San Diego, Calif.	San Francisco, Calif.	Neah Bay, Wash.	Ketchikan, Alaska
1	−1.3	−2.0	−2.3	−1.9
2	−1.5	−1.5	−1.7	−1.5
3	−0.9	−1.2	−1.8	−1.3
4	−0.9	−1.2	−2.0	−1.4
5	−1.2	−1.3	−1.7	−1.3
6	−1.2	−1.3	−1.9	−1.4

TABLE 3. Change of the Specific Volume of Seawater with Temperature for Selected Oceanic Salinities at Sea Level Pressure [after *Fofonoff and Froese*, 1958]. All units are in 10^4 cm^3 g^{-1} °C^{-1}.

Temperature, °C	Salinity, ‰ 32	33	34	35	36	37
−2	0.134	0.165	0.195	0.225	0.255	0.285
0	0.416	0.445	0.473	0.501	0.529	0.557
2	0.678	0.704	0.731	0.757	0.783	0.808
4	0.922	0.946	0.971	0.995	1.019	1.043
6	1.151	1.174	1.196	1.218	1.241	1.263
8	1.367	1.388	1.408	1.429	1.449	1.469
10	1.572	1.591	1.609	1.628	1.647	1.665
12	1.767	1.784	1.801	1.818	1.835	1.852
14	1.953	1.969	1.984	2.000	2.015	2.030
16	2.133	2.147	2.161	2.174	2.188	2.202
18	2.306	2.319	2.331	2.343	2.355	2.367
20	2.475	2.486	2.496	2.507	2.518	2.528
22	2.633	2.649	2.658	2.667	2.676	2.685
24	2.800	2.808	2.816	2.823	2.831	2.838
26	2.958	2.965	2.971	2.977	2.983	2.989
28	3.114	3.119	3.124	3.129	3.133	3.138
30	3.269	3.272	3.276	3.279	3.282	3.285

[*Roden*, 1960]. From the above-said and the form of (14) it follows that the frequency response between the two variables will be much higher at low than at high frequencies. Is this a valid conclusion? The answer depends upon how the low-frequency sea level fluctuations are interpreted. If they are assumed to reflect entirely the motion of the water level, the conclusion is certainly correct. If, however, they represent differential motion between the land and the water, or of the land alone, the conclusion is obviously false. In the discussion of secular sea level trends it was shown that along the tectonically active west coast of the United States land motion was important in producing such long-term changes. Therefore, the magnitude of the response between sea level and atmospheric pressure cannot be determined accurately for frequencies much lower than 1 cpy. For the frequency range between 1 and 6 cpy the response factors are listed in Table 2. Most of the values fall between 1 and 2 cm/mb, a result also obtained by *Saur* [1962] on a seasonal basis.

Coherence and Response Between Sea Level and Sea Temperature Fluctuations

The specific volume of seawater is a complicated function of temperature, salinity, and pressure [*Fofonoff and Froese*, 1958]. For a constant salinity and pressure, the temperature change of the specific volume is a function of the temperature itself. As is shown in Table 3, this measure of the thermal expansion of seawater increases rapidly with increasing temperature, being 6 to 7 times as large as 30°C than at 0°C. The response between sea level and sea temperature fluctuations should therefore be larger in tropical and temperate latitudes than in polar regions.

As a test of this hypothesis the coherence between nonperiodic sea level and sea surface temperature fluctuations was computed for several stations along the west coast of North America, from Balboa, Panama, to Yakutat, Alaska. Some of the results are shown in Figure 10. As expected, the coherence is somewhat higher at low than at temperate latitudes. It is noteworthy that at none of the Alaska tide-gage stations was there any coherence between sea level and sea temperature fluctuations. Elimination of the hydrostatic effect of atmospheric pressure from the records does not change these findings.

The magnitude of the response can be estimated from an expression similar to (14):

$$\mathrm{resp}_{z\theta}(\omega) = \left[\frac{E_z(\omega)}{E_\theta(\omega)}\right]^{1/2} C_{z\theta}(\omega) \qquad (15)$$

where $E_\theta(\omega)$ denotes the energy-density spectrum of nonperiodic sea temperature fluctuations, θ is temperature, and the other symbols retain their previous meanings. The results are listed in Table 4 together with the relevant thermal expansion coefficients. The response between sea level and sea temperature varies between 1 and 2.7 cm/°C and tends to decrease slightly with increasing frequency. As expected, it is larger in low than in high latitudes. At Balboa, Panama, the response is about 2.6 times stronger than at Neah Bay, Washington. This agrees favorably with the coefficient of thermal expansion, which is about 2.1 times larger at Balboa than at Neah Bay.

Conclusions

The following conclusions can be derived from a statistical analysis of monthly mean sea levels:

1. The sea level spectrums are time invariant within the 95% confidence limits; hence the sea level fluctuations may be regarded as stationary (after trend elimination, if necessary).

2. The only significant periodicities are those pertaining to the annual and, sometimes, to the semiannual oscillations. Both are principally of meteorological origin. The total energy contained in the spectrums increases with

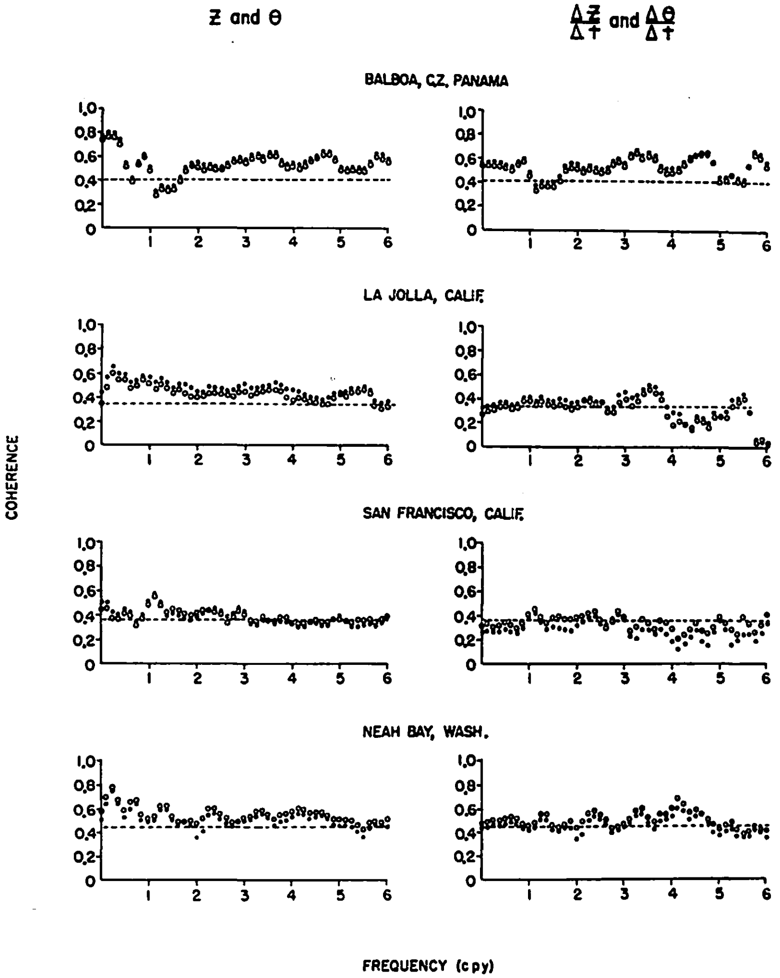


Fig. 10. Coherence between nonperiodic fluctuations of sea level (z) and sea surface temperature (θ), and for the corresponding time gradients. Dots refer to the original records, circles to those from which the hydrostatic effect of atmospheric pressure has been eliminated. The dashed line indicates the 95% confidence limit.

TABLE 4. Frequency Response between Nonperiodic Sea Level and Sea Temperature Fluctuations, in cm/°C

θ refers to the mean annual temperature, S to mean annual salinity, and α_t is the thermal expansion coefficient.

Frequency, cpy	Balboa, Panama	La Jolla, Calif.	San Francisco, Calif.	Neah Bay, Wash.
1	2.6	1.7	1.4	1.3
2	2.6	1.4	1.4	1.0
3	2.7	1.4	1.2	1.0
4	2.6	1.4	1.2	1.0
5	2.6	1.4	1.2	1.0
6	2.6	1.2	1.2	1.0
θ (°C)	28.1	16.9	13.2	9.8
S (‰)	30.7	33.6	29.5*	31.3
$\alpha_t(10^4\ g^{-1}\ cm^3\ °C^{-1})$	3.1	2.3	1.8	1.5

* Measured at Fort Point.

latitude owing to a similar increase in atmospheric activity.

3. The bispectrum of sea level suggests weak interaction of the annual frequency with itself as well as with some other frequencies. Except for the weak ridges extending from the annual frequency, the bispectrum is essentially zero.

4. Along the Pacific coast of the United States secular sea level trends are due to land uplift or subsidence rather than to long-term climatological changes. The trends at nearby stations are frequently different, which also points toward this conclusion.

5. The mean duration of nonperiodic sea level fluctuations above and below arbitrarily selected reference levels can be obtained from a knowledge of the variances of sea level and its first time derivative. There is good agreement between observed and calculated mean durations.

6. The observed mean durations of positive and negative sea level anomalies (from long-term monthly means) are of the order of 3 months and decrease slightly with increasing latitude. The extreme durations of these anomalies vary between 10 and 34 months and are closely related to large-scale atmospheric disturbances such as occurred during the winters of 1940-1941 and 1957-1958. The probability distributions for the duration are log-normal, for durations not exceeding 2 or 3 times the mean duration.

7. The areal coherence of nonperiodic sea level fluctuations is of the order of 1200 km for stations located in the same macroenvironment. Differences in exposure within the same macroenvironment are relatively unimportant. The areal coherence is not much decreased by eliminating the hydrostatic effect of atmospheric pressure from the records.

8. There is a strong and inverse relationship between nonperiodic fluctuations of sea level and atmospheric pressure at temperate and high latitudes. At frequencies between 1 and 6 cpy the response factor varies mostly between 1 and 2 cm/mb. At very low frequencies the response cannot be determined accurately because of the influence of land motion upon secular sea level changes.

9. There is a moderate and direct relationship between nonperiodic fluctuations of sea level and sea surface temperature at low and temperate latitudes. No such relationship is evident in high latitudes. At frequencies between 1 and 6 cpy the response factor varies mostly between 1 and 2.6 cm/°C and decreases with increasing latitude. This agrees well with the results expected from the temperature dependence of the coefficient of thermal expansion.

Acknowledgments. I am indebted to C. A. Barnes, L. K. Coachman, R. H. Fleming, G. W. Groves, W. H. Munk, and M. Rattray for discussion and advice. The sea level and meteorological data were kindly furnished by the U. S. Coast and Geodetic Survey and the U. S. Weather Bureau of the Environmental Science Services Administration, Washington, D. C.

The research reported herein was supported by the Office of Naval Research under contract 477(37), project 083012.

References

Blackman, R. B., and J. W. Tukey, *The Measurement of Power Spectra,* 190 pp., Dover Publications, New York, 1959.

Fel'zenbaum, A. I., *Theoreticheskie osnovy i metody racheta ustanovivshikhsia morskikh techenii (Theoretical basis and methods of calculating stationary ocean currents),* 126 pp., Academy of Sciences, USSR, Moscow, 1960.

Fofonoff, N. P., and C. Froese, Tables of physical properties of sea water, *Fisheries Res. Board Can., Manuscript Rept. Ser. 24,* 35 pp., 1958.

Gissler, N., Anledning at finna hafvets affall för vissa år (Reason for finding a decreased sea level in certain years), p. 142, *Kgl. Svenska Vetenskapsakad. Handl.,* Stockholm, 1747.

Grantz, A., G. Plafker, and R. Kachadoorian, Alaska's Good Friday earthquake, March 27, 1964, *U. S. Geol. Surv. Circ. 491,* 35 pp., 1964.

Groves, G. W., and F. Grivel, Some relationships between sea level and wind in the equatorial Pacific, *Geofis. Intern., Mexico, 2,* 1962.

Hamming, R. W., *Numerical Methods for Scientists and Engineers,* 411 pp., McGraw-Hill Book Company, New York, 1962.

Hamon, B. V., The spectrums of mean sea level at Sydney, Coff's Harbour, and Lord Howe Island, *J. Geophys. Res., 67*(13), 5147–5155, 1962.

Hasselmann, K., W. H. Munk, and G. MacDonald, Bispectra of ocean waves, in *Time Series Analysis,* edited by M. Rosenblatt, pp. 125–139, John Wiley & Sons, New York, 1963.

Haubrich, R. A., Earth noise, 5 to 500 millicycles per second, *J. Geophys. Res., 70*(6), 1415–1440, 1965.

Heiskanen, W. A., and F. A. Vening Meinesz, *The Earth and Its Gravity Field,* 470 pp., McGraw-Hill Book Company, New York., 1958.

Hicks, S. D., and W. Shofnos, The determination of land emergence from sea level observations in southeast Alaska, *J. Geophys. Res., 70*(14), 3315–3320, 1965.

Lee, Y. W., *Statistical Theory of Communication,* 509 pp., John Wiley & Sons, New York, 1960.

Longuet-Higgins, M. S., The statistical analysis of a random moving surface, *Phil. Trans. Roy. Soc. London, A, 249,* 321–387, 1957.

Longuet-Higgins, M. S., Intervals between zeros of a random function, *Phil. Trans. Roy. Soc. London, A, 254,* 557–559, 1962.

Marmer, H. A., Tidal datum planes, *U. S. Coast and Geodetic Survey Spec. Publ. 135,* 142 pp., 1951.

Munk, W. H., and K. Hasselmann, Super-resolution of tides in *Studies on Oceanography Dedicated to Professor Hidaka in Commemoration of his Sixtieth Birthday,* pp. 339–344, University of Tokyo Press, 1964.

Munk, W. H., F. E. Snodgrass, and M. J. Tucker, Spectra of low frequency ocean waves, *Bull. Scripps Inst. Oceanog., 7*(4), 282–362, 1959.

Munk, W. H., M. J. Tucker, and F. E. Snodgrass, Remarks on the ocean wave spectrum, in *Naval Hydrodynamics, Natl. Acad. Sci.—Natl. Res. Council Publ. 515,* pp. 45–60, 1957.

Munk, W. H., B. Zetler, and G. W. Groves, Tidal cusps, *Geophys. J., 10,* 211–219, 1965.

Namias, J., Recent seasonal interactions between North Pacific waters and the overlying atmospheric circulation, *J. Geophys. Res., 64*(6), 631–646, 1959.

Proudman, J., *Dynamical Oceanography,* 409 pp., Methuen & Company, London, 1953.

Rice, S. O., Distribution of the duration of fades in radio transmission, *Bell System Tech. J., 37*(3), 581–635, 1958.

Roden, G. I., On the nonseasonal variations in sea level along the west coast of North America, *J. Geophys. Res., 65*(9), 2809–2826, 1960.

Roden, G. I., On sea level, temperature, and salinity variations in the central tropical Pacific and on Pacific Ocean islands, *J. Geophys. Res., 68*(2), 455–472, 1963*a*.

Roden, G. I., Sea level variations at Panama, *J. Geophys. Res., 68*(20), 5701–5710, 1963*b*.

Roden, G. I., On the duration of nonseasonal temperature oscillations, *J. Atmospheric Sci., 21*(5), 520–528, 1964*a*.

Roden, G. I., Spectral analysis of Japanese sea level records, in *Studies on Oceanography Dedicated to Professor Hidaka in Commemoration of his Sixtieth Birthday,* pp. 166–180, University of Tokyo Press, 1964*b*.

Roden, G. I., On atmospheric pressure oscillations along the Pacific coast of North America, 1873–1963, *J. Atmospheric Sci., 22*(3), 280–295, 1965.

Roden, G. I., A modern statistical analysis and documentation of historical temperature records in California, Oregon and Washington, 1821–1964, *J. Appl. Meteorol., 5*(1), 3–24, 1966.

Roden, G. I., and J. L. Reid, Jr., Sea surface temperature, radiation and wind anomalies in the North Pacific Ocean, *Records Oceanog. Works Japan, 6*(1), 36–52, 1961.

Saur, J. F. T., The variability of monthly mean sea level at six stations in the eastern North Pacific, *J. Geophys. Res., 67*(7), 2781–2790, 1962.

Sheldon, R. C., Precise level checks on permanent bench marks, *Panama Canal Co. Rept.,* May 1938. (Available from the Chief Hydrographer, Panama Canal Company, Balboa Heights, Panama.)

Shoji, D., On the variations of the daily mean sea levels along Japanese islands, *J. Oceanog. Soc. Japan, 17*(3), 141–152, 1961.

Yaglom, A. M., *An Introduction into the Theory of Stationary Random Functions,* 235 pp., Prentice-Hall, Englewood Cliffs, N. J., 1962.

Zadeh, L. A., and J. R. Ragazzini, An extension of Wiener's theory of prediction, *J. Appl. Phys., 21,* 645–655, 1950.

Data Sources

Association d' Océanographique Physique, Union

Géodésique et Géophysique Internationale, Monthly and annual mean heights of sea level up to and including the year 1936, *Publ. Sci.*, *5*, 1940.

———, Monthly and annual mean heights of sea level, 1937–1946, and unpublished data for earlier years, *Publ. Sci.*, *10*, 1950.

———, Monthly and annual mean heights of sea level, 1947–1951, and unpublished data for earlier years, *Publ. Sci.*, *12*, 1953.

———, Monthly and annual mean heights of sea level, 1952–1956, and unpublished data for earlier years, *Publ. Sci.*, *19*, 1958.

———, Monthly and annual mean heights of sea level for the period of the International Geophysical Year 1957 to 1958, and unpublished data for earlier years, *Publ. Sci.*, *20*, 1959.

———, Monthly and annual mean heights of sea level 1959 to 1961, and unpublished data for earlier years, *Publ. Sci.*, *24*, 1963.

Panama Canal Company, Sea level records for Balboa, Panama, 1908–1965, *unpublished.*

U. S. Coast and Geodetic Survey, Environmental Science Services Administration *A.* Nineteenth century sea levels: Astoria, Oregon, 1870–1876 (hourly); San Diego, California, 1854–1871 (hourly); San Francisco, California, 1854–1876 (hourly); Sausalito, California, 1877–1896 (hourly and half-hourly); Honolulu, Hawaii, 1878–1891 (hourly), all *unpublished;* also original marigram records rating back to 1854, all *in storage. B.* Twentieth century sea levels: records of monthly mean sea level over the period of observation for Pacific coast and Pacific island tide gages, *unpublished.*

(Manuscript received May 2, 1966.)

UNCLASSIFIED TECHNICAL REPORTS DISTRIBUTION LIST
for OCEANOGRAPHIC CONTRACTORS
of the OCEAN SCIENCE AND TECHNOLOGY GROUP
of the OFFICE OF NAVAL RESEARCH
(Revised November 1966)

DEPARTMENT OF DEFENSE

Director of Defense Research
and Engineering
Office of the Secretary of Defense
Washington, D.C. 20301
1 Attn: Office, Assistant Director
(Research)

Navy

2 Office of Naval Research
Ocean Science and Technology Group
(Code 408-P)
Department of the Navy
Washington, D.C. 20360
1 Attn: Surface Branch (Code 463)
1 Attn: Undersea Programs (Code 446)
1 Attn: Field Projects (Code 418)
1 Attn: Geography Branch (Code 414)

1 Commanding Officer
Office of Naval Research Branch Office
495 Summer Street
Boston, Massachusetts 02110

1 Commanding Officer
Office of Naval Research Area Office
207 West 24th Street
New York, New York 10011

1 Commanding Officer
Office of Naval Research Branch Office
219 South Dearborn Street
Chicago, Illinois 60604

1 Commanding Officer
Office of Naval Research Branch Office
1030 East Green Street
Pasadena, California 91101

1 Commanding Officer
Office of Naval Research Area Office
1076 Mission Street
San Francisco, California 94103

5 Commanding Officer
Office of Naval Research Branch Office
Box 39, Fleet Post Office
New York, New York 09510

6 Director
Naval Research Laboratory
Washington, D.C. 20390
Attn: Code 5500

(Note: 3 copies are forwarded by the above addressee to the British Joint Services Staff for further distribution in England and Canada).

2 Commander
U.S. Naval Oceanographic Office
Washington, D.C. 20390
Attn: Library (Code 1640)

Chief, Bureau of Naval Weapons
Department of the Navy
Washington, D.C. 20360
1 Attn: FASS
1 Attn: RU-222

1 Office of the U.S. Naval Weather
Service, Washington Navy Yard
Washington, D.C. 20390

1 Chief, Bureau of Yards and Docks
Office of Research
Department of the Navy
Washington, D.C. 20390
Attn: Code 70

U.S. Navy Electronics Laboratory
San Diego, California 92152
1 Attn: Code 3102
1 Attn: Code 3060C

1 Commanding Officer and Director
U.S. Naval Civil Engineering Laboratory
Port Hueneme, California 93401

1 Commanding Officer
Pacific Missile Range
Pt. Mugu, California 93041

1 Commander, Naval Ordnance Laboratory
White Oak
Silver Spring, Maryland 20910

1 Commanding Officer
Naval Ordnance Test Station
China Lake, California 93557

1 Commanding Officer
Naval Radiological Defense Laboratory
San Francisco, California 94135

1 Commanding Officer
U.S. Naval Underwater Ordnance Station
Newport, Rhode Island 02844

Chief, Bureau of Ships
Department of the Navy
Washington, D.C. 20360
1 Attn: Code 1622B

1 Officer-in-Charge
U.S. Navy Weather Research Facility
Naval Air Station, Bldg. R-48
Norfolk, Virginia 23511

1 Commanding Officer
U.S. Navy Air Development Center
Johnstown, Pennsylvania 159
Attn: NADC Library

1 U.S. Fleet Weather Central
Joint Typhoon Warning Center
COMNAVMARIANAS Box 12
San Francisco, California 942

1 Pacific Support Group
Naval Oceanographic Office
San Diego, California 92152

1 Superintendent
U.S. Naval Academy
Annapolis, Maryland 21402

2 Department of Meteorology & Oceanography
U.S. Naval Postgraduate School
Monterey, California 93940

1 Commanding Officer
U.S. Naval Underwater Sound Laboratory
New London, Connecticut 06321

1 Commanding Officer
U.S. Navy Mine Defense Laboratory
Panama City, Florida 32402

1 Officer-in-Charge
U.S. Fleet Numerical Weather Facility
U.S. Naval Postgraduate School
Monterey, California 93940

Air Force

1 Headquarters, Air Weather Service
(AWSS/TIPD)
U.S. Air Force
Scott Air Force Base, Illinois 62225

1 ARCRL (CRZF)
L. G. Hanscom Field
Bedford, Massachusetts 01730

Army

1 Coastal Engineering Research Center
Corps of Engineers
Department of the Army
Washington, D.C. 20310

1 Army Research Office
Office of the Chief of R&D
Department of the Army
Washington, D.C. 20310

1 U.S. Army Beach Erosion Board
5201 Little Falls Road, N.W.
Washington, D.C. 20016

OTHER GOVERNMENT AGENCIES

20 Defense Documentation Center
Cameron Station
Alexandria, Virginia 20305

2 National Research Council
2101 Constitution Avenue, N.W.
Washington, D.C. 20418
Attn: Committee on Undersea Warfare
Attn: Committee on Oceanography

1 Laboratory Director
California Current Resources Laboratory
Bureau of Commercial Fisheries
P.O. Box 271
La Jolla, California 92037

1 Commanding Officer
Coast Guard Oceanographic Unit
Bldg. 159, Navy Yard Annex
Washington, D.C. 20390

1 Director
Coast and Geodetic Survey - U.S. ESSA
Attn: Office of Hydrography and Oceanography
Washington Science Center
Rockville, Maryland 20852

1 Chief, Publications Branch (AD 72)
Office of Scientific Information and Documentation Division
U.S. ESSA
Washington Science Center
Rockville, Maryland 20852

1 Director
Atlantic Marine Center
Coast and Geodetic Survey - U.S. ESSA
439 West York Street
Norfolk, Virginia 23510

1 Director
Institute for Oceanography
U.S. ESSA
Gramax Building
Silver Spring, Maryland 20910

1 U.S. ESSA
Geophysical Sciences Library (AD 712)
Washington Science Center
Rockville, Maryland 20852

1 Chief, Office of Mar. Geol. & Hydrol.
U.S. Geological Survey
Menlo Park, California 94025

1 Director
Pacific Marine Center
Coast and Geodetic Survey - U.S. ESSA
1801 Fairview Avenue, East
Seattle, Washington 98102

1 Geological Division
Marine Geology Unit
U.S. Geological Survey
Washington, D.C. 20240

1 Director
U.S. Army Engineers Waterways Experiment Station
Vicksburg, Mississippi 49097
Attn: Research Center Library

1 Laboratory Director
Bureau of Commercial Fisheries
Biological Laboratory
450-B Jordon Hall
Stanford, California 94035

1 Bureau of Commercial Fisheries
U.S. Fish and Wildlife Service
P.O. Box 3830
Honolulu, Hawaii 96812

1 Laboratory Director
Biological Laboratory
Bureau of Commercial Fisheries
P.O. Box 3098, Fort Crockett
Galveston, Texas 77552

1 Laboratory Director
Biological Laboratory
Bureau of Commercial Fisheries
P.O. Box 1155
Juneau, Alaska 99801

1 Laboratory Director
Biological Laboratory
Bureau of Commercial Fisheries
P.O. Box 6
Woods Hole, Massachusetts 02543

1 Laboratory Director
Biological Laboratory
Bureau of Commercial Fisheries
P.O. Box 280
Brunswick, Georgia 31521

1 Laboratory Director
Tuna Resources Laboratory
Bureau of Commercial Fisheries
P.O. Box 271
La Jolla, California 92037

1 Bureau of Commercial Fisheries & Wildlife
U.S. Fish and Wildlife Service
Librarian
Sandy Hook Marine Laboratory
P.O. Box 428
Highlands, New Jersey 07732

1 Laboratory Director
Biological Laboratory
Bureau of Commercial Fisheries
#75 Virginia Beach Drive
Miami, Floria 33149

1 Director, Bureau of Commercial Fisheries
U.S. Fish and Wildlife Service
Department of Interior
Washington, D.C. 20240

1 Bureau of Commercial Fisheries
Biological Laboratory, Oceanography
2725 Montlake Boulevard, East
Seattle, Washington 98102

1 Director
National Oceanographic Data Center
Washington, D.C. 20390

1 Dr. Gene A. Rusnak
U.S. Geological Survey
Marine Geology and Hydrology
345 Middlefield Road
Menlo Park, California 94025

1 Assistant Director
Oceanography Museum of Natural History
Smithsonian Institution
Washington, D.C. 20560

1 Bingham Oceanographic Laboratories
Yale University
New Haven, Connecticut 06520

1 Advanced Research Projects Agency
The Pentagon
Washington, D.C. 20310
Attn: Nuclear Test Detection Office

RESEARCH LABORATORIES

2 Director
Woods Hole Oceanographic Institution
Woods Hole, Massachusetts 02543

1 Director
Narragansett Marine Laboratory
University of Rhode Island
Kingston, Rhode Island 02881

1 Gulf Coast Research Laboratory
Ocean Springs, Mississippi 39564
Attn: Librarian

1 Chairman, Department of Meteorology and Oceanography
New York University
New York, New York 10453

1 Director
Lamont Geological Observatory
Columbia University
Palisades, New York 10964

1 Director
Hudson Laboratories
145 Palisade Street
Dobbs Ferry, New York 10522

1 Great Lakes Research Division
Institute of Science & Technology
University of Michigan
Ann Arbor, Michigan 48105

1 Director
Chesapeake Bay Institute
Johns Hopkins University
Baltimore, Maryland 21218

1 Director, Marine Laboratory
University of Miami
#1 Rickenbacker Causeway
Miami, Florida 33149

2 Head, Department of Oceanography
and Meteorology
Texas A&M University
College Station, Texas 77843

1 Director
Scripps Institution of Oceanography
La Jolla, California 92037

1 Allan Hancock Foundation
University Park
Los Angeles, California 90007

1 Head,Department of Oceanography
Oregon State University
Corvallis, Oregon 97331

1 Director, Arctic Research Laboratory
P. Barrow, Alaska 99723

1 Head, Department of Oceanography
University of Washington
Seattle, Washington 98105

1 Geophysical Institute of the
University of Alaska
College, Alaska 99735

1 Director
Bermuda Biological Station
for Research
St. Georges, Bermuda

1 Director, Hawaiian Marine Laboratory
University of Hawaii
Honolulu, Hawaii 96825

1 President
Osservatorio Geofisico Sperimentale
Trieste, Italy

1 Department of Engineering
University of California
Berkeley, California 94720

1 Applied Physics Laboratory
University of Washington
1013 N.E. Fortieth Street
Seattle, Washington 98105

1 Physical Oceanographic Laboratory
Nova University
1786 S.E. Fifteenth Avenue
Fort Lauderdale, Florida 33316

1 Director
Ocean Research Institute
University of Tokyo
Tokyo, Japan

1 Marine Biological Association
of the United Kingdom
The Laboratory
Citadel Hill
Plymouth, England

1 New Zealand Oceanographic Institute
Department of Scientific and
Industrial Research
P.O. Box 8009
Wellington, New Zealand
Attn: Librarian

1 Serials Department
University of Illinois Library
Urbana, Illinois 61801

1 Mr. J. A. Gast
Wildlife Building
Humboldt State College
Arcata, California 95521

1 Department of Geodesy & Geophysics
Cambridge University
Cambridge, England

1 Institute of Geophysics
University of Hawaii
Honolulu, Hawaii 96825

1 Division of Engineering and
Applied Physics
Harvard University
Cambridge, Massachusetts 02138

1 Underwater Warfare Division
of the Norwegian Defense Research
Establishment
Karljohansvern, Horten, Norway

1 Department of Geology and Geophysics
Massachusetts Institute of Technology
Cambridge, Massachusetts 02139

1 Marine Science Center
Lehigh University
Bethlehem, Pennsylvania 18015

1 Lieutenant Nestor C. L. Granelli
Montevideo 459
Buenos Aires, Argentina

1 Oceanographische Forschungsantalt
der Bundeswehr
Lornsenstrasse 7
Kiel, Federal Republic of Germany

ADDENDUM

To The

UNCLASSIFIED TECHNICAL REPORTS DISTRIBUTION LIST
for OCEANOGRAPHIC CONTRACTORS
of the OCEAN SCIENCE AND TECHNOLOGY GROUP
of the OFFICE OF NAVAL RESEARCH
(Revised November 1966)

DEPARTMENT OF DEFENSE

Navy

1 Commander
U.S. Naval Oceanographic Office
Washington, D.C. 20390
1 Attn: Code 031
1 Attn: Code 70
1 Attn: Code 90

1 West Coast Support Group
U.S. Naval Oceanographic Office
c/o U.S. Navy Electronics Laboratory
San Diego, California 92152

1 U.S. Naval Oceanographic Office
Liaison Officer (Code 332)
Anti-Submarine Warfare Force
U.S. Atlantic Fleet
Norfolk, Virginia 23511

1 U.S. Naval Oceanographic Office
Liaison Officer
Anti-Submarine Warfare Force Pacific
Fleet Post Office
San Francisco, California 96610

1 Commander-in-Chief
Submarine Force Pacific Fleet
Fleet Post Office
San Francisco, California 96610

1 Commander-in-Chief
Pacific Fleet
Fleet Post Office
San Francisco, California 96610