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Seasonal variation of the equatorial zonal geopotential gradient in the eastern Pacific Ocean

by Mizuki Tsuchiya

ABSTRACT

Evidence is presented to show that the zonal westward geopotential gradient along the equator in the eastern Pacific between 95W and 120W exhibits a seasonal variation with vanishingly low values in March-May and high values ($5 \times 10^{-7}$ m s$^{-2}$) in June-July followed by a gradual decrease toward the end of the year. This variation is found to be in phase with the seasonal variation of the westward component of the wind stress near the equator, which also shows low values in northern spring and high values in northern summer and fall.

1. Introduction

The sea surface of the equatorial ocean slopes up toward the west as a consequence of the prevailing easterly trade winds. Montgomery and Palmén (1940) have estimated the zonal geopotential gradient resulting from this slope to be of the order of $4 \times 10^{-7}$ m s$^{-2}$ both for the Pacific and Atlantic Oceans. This gradient and its relationship to the wind field have been studied by various investigators (Austin, 1958; Knauss, 1963; Lemasson and Piton, 1968; Hickey, 1975).

The circulation in the eastern intertropical Pacific has been known to vary in response to the seasonal variation of the wind field over the ocean (e.g., Wyrtki, 1966; Stroup, 1969; Tsuchiya, 1974). The seasonal variation of the wind near the equator in the eastern Pacific is associated with the annual meridional migration of the intertropical convergence zone (ITCZ). During northern summer and fall, when...
the ITCZ moves farthest north (11-15N) and the southeast trades extend well into the northern hemisphere, the region of the equator is dominated by the strong southeast trades. During northern spring, when ITCZ moves closest to the equator (2-5N) and the southeast trades retreat to the south, the wind near the equator is generally weak (e.g., see Meteorological Office, 1956).

Various monthly maps of the sea-surface meteorology (e.g., Meteorological Office, 1959) indicate that this seasonal variation of the wind field in the eastern equatorial Pacific is very much like that in the equatorial Atlantic. Recently, Katz et al. (1977) have presented evidence that the zonal pressure gradient along the Atlantic equator varies seasonally in correlation with the variation of the zonal wind stress near the equator: when the wind is weak (northern spring), the gradient is small; when the wind is strong (northern summer to fall), the gradient is large. It is interesting to study how the zonal pressure gradient varies with season in the eastern Pacific. Theoretical studies (Hurlbert et al., 1976; McCreary, 1976) suggest that variations in the equatorial oceans can be interpreted in terms of equatorially trapped Rossby and Kelvin waves excited by changing winds. The zonal scale of the ocean is an important factor in this regard. This paper describes findings based on the data from the Eastropac expedition (February 1967-April 1968).

2. Geopotential anomaly along the equator

Eastropac was designed primarily for the study of the monthly or seasonal variations in the distributions of oceanographic properties in the eastern intertropical Pacific Ocean. It consisted of seven bimonthly cruise periods; during each period extensive oceanographic observations were made along meridional tracks usually separated by 7° of longitude. A detailed description of the expedition, including the method of data processing, has been published in the introduction to the Eastropac Atlas (Love, 1972).

All Eastropac STD stations that lie no farther than 1° of latitude from the equator and that have observations at least down to 500 m were used to compute geopotential anomaly (\( \phi_a \)) along the equator with respect to 500 db. The procedures used here to estimate the zonal pressure gradient essentially follow those of Katz et al. (1977). For each longitude, the computed geopotential anomalies were averaged over all stations within 1° of the equator. The average geopotential anomalies at 0, 50, and 100 db are plotted against longitude for each of the seven bimonthly periods (Figs. 1-7). The number of stations averaged is usually three to seven and is indicated at the bottom for each longitude of the observations. The standard deviations from the averages are mostly less than 0.2 m²s⁻² (1 m²s⁻² = 10 dyn. cm) for 0 and 50 db and less than 0.1 m²s⁻² for 100 db (the largest standard deviations are 0.44 m²s⁻² for 0 db, 0.25 m²s⁻² for 50 db, and 0.22 m²s⁻² for 100 db).

To estimate the large-scale zonal gradient of geopotential along the equator, linear regression lines were fitted to the data points in Figures 1-7. East of 92W (the
Figure 1. Geopotential anomaly \((\phi_a)\) at 0, 50, and 100 db with respect to 500 db along the equator in February-March 1967 (\(1 \text{ m}^2 \text{s}^{-2} = 10 \text{ dyn. cm}\)). Each point represents the arithmetic mean over stations within 1° of the equator. The number of stations averaged at each longitude is given at the bottom, and the dates of observations are given at the top. The straight lines are linear regressions of geopotential anomaly on longitude.

longitude of the Galápagos Islands) the distribution of geopotential anomaly appears to be more irregular (Figs. 1, 3 and 7), and there might be a reversal in the gradient (Lemasson and Piton, 1968). For this reason, this area was excluded from the regression analysis. The zonal gradients of geopotential anomaly thus estimated by fitting straight lines are listed in Table 1. The unit of the geopotential gradient \(10^{-7} \text{ m s}^{-2}\) in Table 1 is equivalent to \(10^{-5} \text{ dyne \cdot g}^{-1}\) used by Katz et al. (1977). The standard errors of the regression coefficients (estimated geopotential gradients) are also shown in the table.²

Lemasson and Piton (1968) summarize previous measurements of geopotential anomaly along the entire length of the equator in the Pacific and list values 5 to \(7 \times 10^{-7} \text{ m s}^{-2}\) for the 0-db geopotential gradient.³ These values are significantly larger

<table>
<thead>
<tr>
<th>Period</th>
<th>Westward zonal geopotential gradient, (10^{-7} \text{ m s}^{-2})</th>
<th>Westward zonal wind-stress comp., (10^{-2} \text{ N m}^{-2})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Feb.-Mar. 1967</td>
<td>2.6 (1.5)</td>
<td>0.3 (0.1)</td>
</tr>
<tr>
<td>Apr.-May 1967</td>
<td>-0.8 (2.1)</td>
<td>1.2 (0.4)</td>
</tr>
<tr>
<td>June-July 1967</td>
<td>5.1 (0.5)</td>
<td>1.5 (1.1)</td>
</tr>
<tr>
<td>Aug.-Sept. 1967</td>
<td>2.9 (0.9)</td>
<td>0.6 (0.1)</td>
</tr>
<tr>
<td>Oct.-Nov. 1967</td>
<td>2.8 (0.8)</td>
<td>0.4 (0.2)</td>
</tr>
<tr>
<td>Dec. 1967-Jan. 1968</td>
<td>4.3 (2.2)</td>
<td>-0.3 (0.7)</td>
</tr>
<tr>
<td>Feb.-Apr. 1968</td>
<td>-0.1 (1.8)</td>
<td>0.4 (0.2)</td>
</tr>
</tbody>
</table>

2. The standard errors of the regression coefficients in Table 1 are of roughly the same order of magnitude as the uncertainties in the estimates of the geopotential gradient due to the scatter of geopotential anomaly about the average at each longitude of the observations.

3. Lemasson and Piton (1968) use a reference pressure of 700 db or 1000 db. This choice does not affect the estimates of the surface geopotential gradient because the 500-db surface is practically horizontal with respect to the 1000-db surface in the region of the equator (see their Fig. 1 and Wyrtki, 1974, Fig. 16).
than those shown in Table 1 for the eastern Pacific between 95W and 120W. This difference is due to a greater gradient of geopotential anomaly in the central Pacific (see Lemasson and Piton, 1968, Fig. 3).

As can be seen from Table 1 and Figures 1-7, the standard errors are large in some periods for which the data points show a large deviation from a linear distribution (April-May 1967; December 1967-January 1968; February-April 1968). Because of these large standard errors, it is difficult to discuss the temporal variation of the gradient with certainty. Nevertheless, the general trend is that the gradient at 0 db is small in March-May, increases rather quickly in June-July, and then decreases gradually. It is not immediately obvious that this variation indicated by the data represents a seasonal variation, because the data cover a time period of only slightly more than a year. However, as will be seen below, this variation is found to be in phase with the known seasonal variation of the wind near the equator and appears to be seasonal. (The data suggest the presence of a semi-annual cycle with maxima in June-July and December-January. However, the large gradients in December 1967-January 1968 are a result of the high geopotential-anomaly values at 119W in comparison with those at the other three longitudes. If the data at 119W
are excluded, the zonal geopotential gradients decrease to 0.7, 1.0, and $-0.3 \times 10^{-7} \text{ m s}^{-2}$ for 0, 50, and 100 db, respectively. The standard errors of the regression coefficients also decrease to 1.3, 1.7, and $1.2 \times 10^{-7} \text{ m s}^{-2}$ for 0, 50, and 100 db).

The geopotential gradient at 50 db shows the same trend except for a sharp drop in August-September 1967. The gradient at 100 db is too small to discuss its variation with any confidence.

This variation in the zonal geopotential gradient is similar to that found by Katz et al. (1977) for the equatorial Atlantic and is consistent with the seasonal variation of the zonal wind stress near the equator indicated by the long-term monthly mean charts of the wind field (Meteorological Office, 1956; Wyrtki and Meyers, 1975).

The variation of the zonal geopotential gradient can also be compared with the winds observed during the period of Eastropac. Two sets of data are presented here. The first set is based on the weather observations made on Eastropac STD stations. All wind measurements taken on STD stations between 2N and 2S (this choice again follows Katz et al., 1977) and west of 95W were used to compute the average zonal component of wind stress for each bimonthly period of Eastropac. The calculations were made according to the same formula with the same values of the drag coefficient and air density as those assumed by Wyrtki and Meyers (1975). The results are shown in the last column of Table 1.
The second set is based on the wind-stress data compiled from the Marine Deck of the Environmental Data Service by Wyrtki and Meyers (1975). They have computed the average wind stress over quadrangles of 1° latitude by 10° longitude for individual months from 1950 to 1972. From these data the monthly average zonal wind stress over the equatorial belt between 2N and 2S and between 90W and 120W was estimated for each month of the Eastropac period (Table 2).

The first set of the data, which represents small portions of the second set, shows substantially smaller wind-stress values than the second set. Both sets, however, exhibit roughly similar seasonal variations with low values in February-April and high values in June-July followed by a gradual decrease toward the end of the year. Thus, the temporal variation in the zonal wind stress during Eastropac does not deviate qualitatively from the seasonal variation indicated by the long-term monthly average charts (Meteorological Office, 1956).

Table 2. Monthly mean westward wind-stress component averaged over the area between 2N and 2S and between 90W and 120W during the Eastropac expedition. Data source Wyrtki and Meyers (1975).

<table>
<thead>
<tr>
<th>Month</th>
<th>Year</th>
<th>Westward wind-stress comp., 10^{-2} N m^{-2}</th>
<th>Month</th>
<th>Year</th>
<th>Westward wind-stress comp., 10^{-5} N m^{-2}</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mar.</td>
<td></td>
<td>0.9</td>
<td>Nov.</td>
<td></td>
<td>2.0</td>
</tr>
<tr>
<td>Apr.</td>
<td></td>
<td>0.9</td>
<td>Dec.</td>
<td></td>
<td>1.6</td>
</tr>
<tr>
<td>May</td>
<td></td>
<td>2.5</td>
<td>Jan.</td>
<td>1968</td>
<td>2.2</td>
</tr>
<tr>
<td>June</td>
<td></td>
<td>3.8</td>
<td>Feb.</td>
<td></td>
<td>0.8</td>
</tr>
<tr>
<td>July</td>
<td></td>
<td>2.9</td>
<td>Mar.</td>
<td></td>
<td>1.0</td>
</tr>
<tr>
<td>Aug.</td>
<td></td>
<td>1.9</td>
<td>Apr.</td>
<td></td>
<td>1.6</td>
</tr>
<tr>
<td>Sept.</td>
<td></td>
<td>3.1</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

4. The meridional component is toward the north and is of about the same magnitude as the zonal component. Its seasonal variation is in phase with that of the zonal component.
The ratio of the zonal wind stress (divided by the density of sea water) to the zonal geopotential gradient is a measure of the depth below which the effect of the wind stress vanishes and above which the horizontal zonal shear stress is in balance with the zonal pressure gradient. For the Atlantic, this depth is found to be nearly uniform at about 100 m (Katz et al., 1977). For the Pacific, however, this ratio is not as uniform as in the Atlantic (Tables 1 and 2).

Nevertheless, in view of the paucity of the Pacific data (see next section), the correlation between the geopotential gradient and the wind stress is considered to be qualitatively good (when the wind is strong, the geopotential gradient is large, and when the wind is weak, the geopotential gradient is small). In fact, if the December 1967 data at 119W noted before are disregarded, the only notable offset from the correlation is a small (negative) geopotential gradient relative to a fairly large (positive) zonal wind stress in April-May 1967.

3. Discussion and summary

The data set used in this study is better than the Atlantic data set used by Katz et al. (1977) in that it covers a continuous time period of more than a year at regular bimonthly intervals. However, it has much fewer observations in each longitude than the Atlantic data; there are no time series of measurements except one short series taken at 112W in August 1967. This data coverage was apparently insufficient to eliminate short-period fluctuations and resulted in large values of the standard errors in the estimated zonal geopotential gradients. The short-period variability in geopotential anomaly can indeed be large; Taft and Jones (1973) report that the range of the geopotential anomaly at 0/500 db computed from nine STD lowerings at 0°, 115W over a period of 6 days in July-August 1969 was as large as 1.2 m² s⁻².

Despite this difficulty, the Eastropac data provide evidence that the zonal geopotential gradient along the equator exhibits a seasonal variation with low values in March-May and high values in June-July followed by a gradual decrease toward the end of the year. This variation is in phase with the seasonal variation of the zonal wind-stress component near the equator.
Independent evidence for the seasonal variation in the zonal geopotential gradient at the sea surface near the equator can be found in tide-gauge data at Talara (4°35'S, 81°17'W), Peru and Canton Island (2°48'S, 171°43'W). From these data, Wyrtki (1975) finds that the difference in monthly mean sea level between Canton Island and Talara (Canton Island minus Talara) varies from 33 cm in May to 47 cm in October. This corresponds to a variation of the zonal geopotential gradient from $3.3 \times 10^{-7}$ m s$^{-2}$. He remarks that this variation is largely in response to the variation of the zonal wind-stress component in the southeast trades between 4S and 10S and between 100W and 180W. From June to October, when the trades are strong, the difference in sea level increases rapidly; from December to February, when the trades become weak, the difference decreases rapidly and remains small until the wind speed increases again in June. In these data, however, the variation of the sea-level difference lags that of the zonal wind-stress component by 2 to 3 months (Wyrtki, 1975, Fig. 6). The Eastropac data are too coarse to resolve a phase difference of this magnitude.

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REFERENCES


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