OCEANOLOGICA ACTA, 1987, N° SP

Vertical motion in the Eastern equatorial Pacific inferred from drifting buoys

Upwelling
Vertical motion
Equatorial currents
Drifting buoys
Pacific Ocean

Upwelling
Mouvement vertical
Courants équatoriaux
Bouées dérivantes
Océan Pacifique

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Received 12/7/85, in revised form 10/6/86, accepted 19/6/86.

ABSTRACT

Surface current measurements have been obtained from the Eastern tropical Pacific Ocean by means of drifting buoys tracked by the NIMBUS and ARGOS satellite systems since 1977. Near-equatorial divergence of Ekman transport is indicated qualitatively by persistent avoidance of the Equator by drifters. Upwelling velocity and transport were estimated from the horizontal divergence of surface current fields obtained by optimum interpolation of overall and monthly composite data. The estimated divergence is predominantly meridional. Average upwelling velocity and transport in the region 1.5°N-1.5°S, 80°W-130°W are estimated to be 1.5 m d^{-1} and 32×10^6 m^3 s^{-1}. Seasonal variation of the estimated upwelling agrees closely in phase but is larger in magnitude than that implied by the annual and semiannual constituents of 14°C isotherm depth variations in the eastern equatorial Pacific.

Oceanol. Acta, **1987.** Proceedings International Symposium on Equatorial Vertical Motion, Paris, 6-10 May 1985, 27-32.

RÉSUMÉ

Mouvement vertical dans l'est du Pacifique à partir de bouées dérivantes

Des mesures de courants superficiels ont été effectuées dans la partie orientale de l'Océan Pacifique à l'aide de bouées dérivantes suivies par les systèmes de satellites NIMBUS et ARGOS depuis 1977. La divergence proche de l'équateur dans le transport d'Ekman est indiquée qualitativement par l'absence permanente de bouées sur l'équateur. La vitesse verticale et le transport ont été estimés à partir de la divergence horizontale des champs de courants superficiels obtenus par interpolation optimale des données composées mensuelles et totales. La divergence estimée est fortement méridienne. La vitesse moyenne verticale et le transport dans la région 1,5°N-1,5°S, 80°W-130°W sont de l'ordre de 1,5 m j^{-1} et 32×10^6 m^3 s^{-1}. La variation saisonnière de la remontée d'eau est bien en phase, mais est plus importante que celle qui résulte des composantes annuelle et semiannuelle des variations de profondeur de l'isotherme 14°C dans la partie orientale du Pacifique équatorial.

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INTRODUCTION

Equatorial upwelling induced by surface winds has long been a tenet of faith among oceanographers. It is expected from theory, and its prevalence is indicated qualitatively by the distribution of temperature and other water properties, but it is too weak to be directly measured with present methods.

Several investigators have applied the continuity equation,

$$w(z_1) = -\int_0^z \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) dz \,, \tag{1}$$

for indirect determination of the vertical velocity from the divergence of the measured horizontal velocity. Knauss (1966) used this method implicitly to obtain an estimate of 0.5 to 5 m d^{-1} for the near-equatorial upwelling velocity between 118°W and 140°W from measurements of horizontal current made with a current meter lowered from a ship. These results were in contrast to results from a cruise three years earlier (Knauss, 1960) in which no evidence of upwelling was found. Hisard *et al.* (1970) also used current meters lowered from a ship to obtain data that indicated upward movement of as much as 20 m d^{-1}, and downward motion, during two cruises a month apart at 170°E. Taft and Jones (1973) also used data from a current meter lowered from a ship to estimate an upwelling velocity of 7 m d^{-1} at a depth of 28 m in the vicinity of 0° , 110°W.

Halpern (1980 and pers. comm.) used data from 100 km scale triangular arrays of moored current meters to estimate the vertical velocity at 0°, 110°W. From records several months in length he found average upwelling velocities ranging from 2 to 15 m d^{-1} at 50 m depth, and from 50 m d^{-1} (upward) to 2 m d^{-1} (downward) at 250 m depth. He also found variations from 15 m d^{-1} (upward) to 10 m d^{-1} (downward) on a time scale of several days at only 20 m depth.

Wyrtki and Eldin (1982) estimated vertical velocity from vertical movements of the 26°C isotherm observed from ship and aircraft sampling at 150°W and 158°W during the Hawaii to Tahiti Shuttle Experiment. They observed upwelling and downwelling events with minimum velocities of 3-5 m d^{-1} in response to local wind events on a time scale of 10-20 days. Without specific reference to upwelling, Pullen *et al.* (1987) and Wilson *et al.* (1986) describe observations of vertical displacements of isotherms of as much as 50 m in ten days at the Equator associated with inertia-gravity waves excited by shear instability of the zonal currents (Hansen, Paul, 1984).

The observations of temporal variations make it clear that the variability associated with local wind events and the near-equatorial instability waves is so great that the earlier estimates must be regarded as random samples from a highly variable field. Use of methods that integrate measurements in time and space is required to obtain stable estimates of the equatorial upwelling on climatically important scales. Knauss' (1966) result is the first use of measured currents on large spatial scale, and Halpern's (1980) results are the first derived from use of long time series from current meter arrays. The approach of Wyrtki and Eldin (1982) is useful on short time scales and indicates the nature of the sampling problem, but it cannot be expected to give useful results on longer time scales because temperature is not a conservative property on these time scales. Wyrtki (1981) used climatological information about surface winds and subsurface density distribution to infer equatorial upwelling using dynamical and mass balance constraints. Assuming negligible zonal convergence of the South Equatorial Current, he obtained an estimate of about 1 m d^{-1} for the climatological upwelling velocity at 50 m depth in the region between 100°W and 170°E. Brady and Bryden (1985) also employed this approach as well as other mass balance and kinematic methods to estimate the average vertical velocity between relatively well sampled sections in the central Pacific (110°W and 150°W). They obtained estimates ranging from 1.3 to 2.8 m d^{-1} for the average upwelling velocity at 50 m depth. They favor the method yielding the smallest of these estimates because it is relatively simple and yet provides estimates within the range of uncertainties. The most sophisticated of their methods yields a value of about 2.2 m d^{-1}.

Hansen and Paul (1984) used drifting buoy data from a five-month period of moderately strong southeast tradewinds to obtain an estimate of 2.6 m d^{-1} for the average upwelling velocity at 50 m depth in the region 1.5°N-1.5°S, 100°W-130°W. They assumed that the zonal component of the divergence was negligible on large time and space scales. This paper presents some results from a further attempt to estimate the strength of upwelling on long time and space scales using data from satellite-tracked drifting buoys.

MATERIALS AND METHODS

Satellite-tracked drifting buoys were deployed in the central equatorial Pacific during the Hawaii to Tahiti Test Shuttle Project of 1977-78 (Hansen *et al.*, 1978) and more regularly in the eastern equatorial Pacific since 1979 for the EPOCS Program of NOAA. These buoys were deployed with 2 m × 10 m drogues attached by nylon tethers 30 m long to improve their performance as current following devices. Figure 1 shows the trajectories of all buoys observed from November 1977 through June 1982. Our analysis is confined to this period in order to obtain a result exemplary of non-El Niño conditions.

Some qualitative evidence of equatorial divergence is readily evident in Figure 1. Buoys deployed between 0° and 5°N at 150°W followed a meandering but persistent northward course, moving first westward in the South Equatorial Current, then eastward in the North Equatorial Countercurrent, and finally westward again in the North Equatorial Current. Drifters deployed in this region during 1979-80 for the NORPAX program showed much the same behavior (Wyrtki *et al.*, 1981). Similarly, buoys moving westward in the South Equatorial Current south of the Equator show a persistent southward component, most obviously west of about 100°W, but also east of 100°W within a few degrees of the Equator. The result is a short residence time of buoys near the Equator, expressed as a visible paucity of trajectories there.

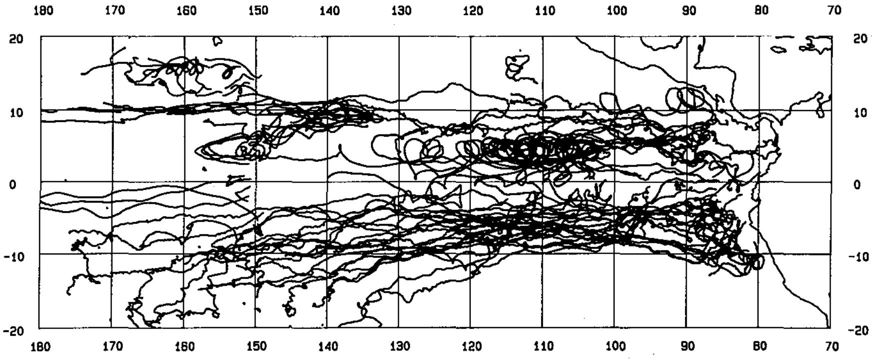

Figure 1
Trajectories of drifting buoys in the Eastern tropical Pacific Ocean, 1977-1982.

Paul and Hansen (1984) used an optimum interpolation method to derive an average surface current vector field from the drifting buoy data shown in Figure 1. The subsurface drogues were lost from many of these buoys while they remained otherwise fully operational, but as only surface floats. While a difference in behavior is rarely apparent when a drogue is lost from a buoy, it is expected that there are differences of performance of importance on the longer time scales relevant to our analysis. Comparison of averaged velocity data from common times and regions shows that undrogued buoys tend to have larger westward velocity components, reflecting buoy windage and waveage, and westward vertical current shear in the near surface waters, all due to the persistent easterly winds in the region. Drogued buoys on the other hand show stronger poleward movement, which seems to be consistent with theoretical ideas about the vertical structure of the Ekman transport. We conclude that data from the undrogued buoys have much of the same information content as that from drogued buoys, but that there are systematic differences between them. To make the data from undrogued buoys more representative of the mean velocity of the mixed layer for inclusion in an otherwise quite sparse data set, the undrogued buoy data were transformed by an operation of the form $\bar{v}' = \mathbb{A}\bar{v} + \mathbb{B}$, in which $\bar{v}$ denotes undrogued buoy velocity data and $\mathbb{A}$ and $\mathbb{B}$ are coefficient matrices determined by linear regression with drogued buoy data within each of several regions for which it was thought that behavior might be distinct for dynamical or empirical reasons. Zonal and meridional decorrelation scales of 2 400 and 500 km were derived empirically from the composite data set for use in the optimum interpolation. The result of their analysis is shown as Figure 2. The major familiar features of the surface currents in the eastern tropical Pacific are readily

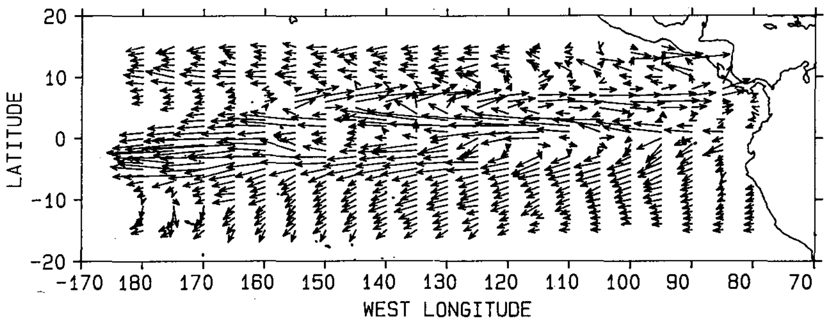

Figure 2
Average surface currents in the eastern tropical Pacific Ocean derived from optimum interpolation of buoy drift data. Vector length scale is 50 cm s^{-1}/8° latitude.

apparent. In regions of sparse data the method tends toward the average value of the entire field, and indicates larger uncertainties in an error map that is produced in parallel with the analysis. Thus, the North Equatorial Countercurrent does not appear west of about 155°W in the analysis because the drifting buoys did not sample that region but the method fills the gap with average values from more distant points. Paul and Hansen also produced average surface current maps for each of the twelve months of the year. Each of the monthly ensemble data sets is of course smaller and therefore yields maps of greater uncertainty. Decorrelation scales were also found to be smaller, about 1 600 and 300 km.

The horizontal divergence was computed from the analyzed surface current fields. The result for the ensemble average is shown in Figure 3. The spatial variability in this computed divergence field is considerable, which is not surprising in consideration of the temporal variability of the surface currents, and the still modest amount of drifting buoy data distributed over the large region shown in Figure 1. Some prominent features do emerge from the analysis however. Most notable is the pattern of divergence along the Equator, almost continuous from the coast of South America to 165°W, beyond which the data are insufficient for its estimation. There is also an indication of a zone of convergence along approximately

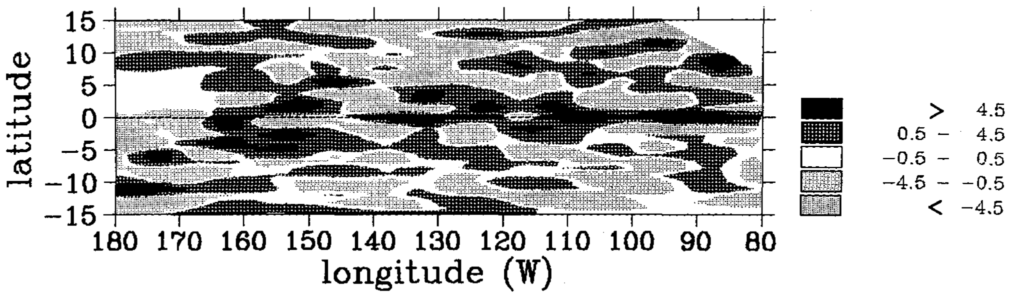

Figure 3
Horizontal divergence of surface currents. Units are 10^{-7} s^{-1}.

5°N, corresponding to the Countercurrent ridge, in the eastern half of the region, and a region of locally strong divergence at the approximate location of the Costa Rica Dome. Most of the strong and organized features in Figure 3 are obtained from the divergence of the meridional component of velocity. The divergence of the zonal component is generally weaker and more random but slightly positive overall.

RESULTS

Through the logic of the continuity equation, Figure 3 can be interpreted as a map of vertical motion. If it is supposed that the currents and divergence described by the drifting buoys apply to a 50 m deep mixed layer, a horizontal divergence of $1 \times 10^{-7}\ s^{-1}$ corresponds via equation (1) to an upwelling velocity across the bottom of the mixed layer of 5×10^{-4} cm s^{-1}, or 0.43 m d^{-1}.

The data base for the individual months is insufficient to show even the equatorial divergence clearly in all cases. To obtain a more definite if less detailed result, the data shown on Figure 3 and those from each of the monthly analyses were averaged zonally between 80°W and 130°W and between latitudes 1.5°N and 1.5°S. Results of the monthly analyses are shown in Figure 4. The ensemble average horizontal velocity

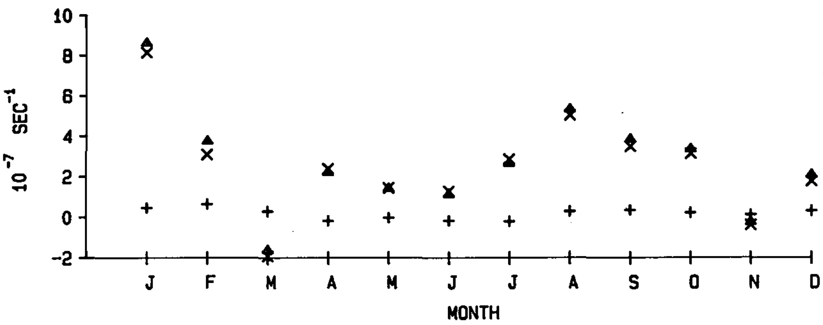

Figure 4
Monthly variation of average surface current divergence in the region 1.5°N-1.5°S, 80°W-130°W. Symbols are : Δ, divergence ; +, δu/δx ; x, δv/δy.

divergence over this region was determined to be $3.5 \times 10^{-7}\ s^{-1}$, which implies an average upwelling of 1.5 m d^{-1} at 50 m depth, or an upwelling transport into the surface mixed layer of $32 \times 10^6\ m^3\ s^{-1}$. These results are very comparable to those of Wyrtki (1981) and of Brady and Bryden (1987). The nature of the data that they used precluded a systematic analysis of uncertainty. Our data lend themselves to a more rigorous analysis.

A first-order question in use of drifting buoys is their accuracy in measuring currents. Kinder *et al.* (1980) summarize evidence indicating that the buoy/drogue configuration used in EPOCS and FGGE can be expected to have a downwind bias ranging from 0.3 to 3 % of the wind speed, depending upon whether the buoy is drogued or not. Because of the average westward flow, drifters tend to sample the eastern part of the region early, and the western part later, in their operating lifetimes. This imposes an east-west trend in the proportion of data from undrogued drifters, ranging from about 10 % in the eastern part of the region to about 70 % in the west. If uncorrected, this effect would introduce a spurious zonal divergence of less than $0.3 \times 10^{-7}\ s^{-1}$ into the data. Although unimportant for the present application, this bias was largely removed by the statistical correction applied to the data from undrogued drifters by Paul and Hansen (1984). In the eastern part of the region where we have focussed our attention because of the generally better data coverage, the wind and wave induced bias is believed not to exceed one-half percent of the wind speed. We therefore expect the bias in estimation of the mixed layer current divergence to mimic the divergence of the surface winds but in the ratio 0.005. Wyrtki and Meyers (1975) show surface wind convergence of $10^{-5}\ s^{-1}$ near 5°N and divergence of $0.5 \times 10^{-5}\ s^{-1}$ near the Equator over the eastern Pacific. The magnitude of this systematic error in estimation of divergence in the oceanic mixed layer is therefore $0.5 \times 10^{-7}\ s^{-1}$, or less. It is thus a significant, but not critical source of error.

Hansen and Paul (1984) showed that in common with other methods of measuring ocean currents, the principal source of error in determination of average values is the geophysical variability. The random errors associated with geophysical variability can be estimated according to the usual rules for estimation and propagation of errors. The uncertainty of the divergence is estimated as

$$\varepsilon_D = \sqrt{2}\left[\frac{\varepsilon_u^2 - \varepsilon_{uu}}{\delta x^2} + \frac{\varepsilon_v^2 + \varepsilon_{vv}}{\delta y^2}\right]^{1/2}, \quad (2)$$

in which ε_u, ε_v are the uncertainty of the estimates of the mean velocity components, computed from the variance and degrees of freedom of the data. The error covariances on the scales, δx, δy, at which the divergence was computed are denoted by ε_{uu}, ε_{vv}. Hansen and Paul (1984) reported average variance of surface currents between ± 5° latitude in the eastern Pacific of about 500 $cm^2\ s^{-2}$ for the zonal component, and about 800 $cm^2\ s^{-2}$ for the meridional component. Their estimates were made from data collected during a period of particularly energetic mesoscale wave activity, and are therefore probably a conservative overestimate for the longer term average. Five to six drifter locations, quasi-random in time, are available daily from tropical drifters using the ARGOS system. After being edited for infrequent large errors, these data were smoothed and interpolated to uniform six-hour intervals (δt). To construct the maps exemplified by Figure 2, the data were averaged in 1° latitude by 10° longitude rectangles. In the overall data set portrayed by Figures 1 to 3, there were about 400 six hour velocity vectors per 1° × 10° cell in the region between 6°N and 8°S, and from 140°W to the coast of South America. The decorrelation scales used by Paul and Hansen (1984) imply an effective number (N) of about 2 500 observations for each average velocity determination in the composite data set (Fig. 2).

The fact of the variance of the meridional current component being greater than or equal to that of the zonal component while the meridional scales of the flow are less than the zonal scales makes the meridional term predominant in estimation of error of the divergence. Based upon other work (Hansen, Hwang, 1987) we use a Lagrangian integral time scale (τ) of five days to estimate the number of degrees of freedom in our data (*see* also Lukas, 1986), and ε_{vv} $(1°\varphi) \approx 0.7\ \varepsilon_v^2$. Also, in as much as one of the effects of the correction applied for drogue loss is partial replacement of data from undrogued drifters with the regional averages from drogued drifters, the number of degrees of freedom is further reduced by about 10 %.

The estimate of the meridional velocity component error therefore is $\varepsilon_v = (\tau\sigma_v^2/N\delta t)^{1/2} = 2.5$ cm/s, and that for the divergence computation is

$$\varepsilon_D \approx 0.77\frac{\varepsilon_v}{|\delta y|} = 1.8 \times 10^{-7}\ s^{-1}. \quad (3)$$

By averaging over the region ± 1.5° latitude, 130°W to the coast of South America, the error estimate for the divergence in the overall data set is reduced to $0.5 \times 10^{-7}\ s^{-1}$, and the corresponding uncertainties of mean upwelling velocity and transport for this region are 0.2 m d^{-1}, and $5 \times 10^6\ m^3\ s^{-1}$.

Errors in estimation of the monthly values are larger due to both the order of magnitude reduction in the number of observations available and the shorter decorrelation scales used for the monthly analyses. The effective number of observations for each interpolated map value is thereby reduced to about one hundred. The average uncertainty of the meridional velocity component estimations is therefore about 12 cm s^{-1}, and that for the divergence and upwelling velocity are $2.2 \times 10^{-7}\ s^{-1}$ and 0.9 m d^{-1}, respectively in the eastern equatorial region for which results are shown in Figure 4.

Uncertainties associated with geophysical variability are thus seen to be satisfactory small (~ 15 %) for the overall average, but are exceeded by only about half of the estimates for the monthly composites. These estimates of error are believed to be conservative (large) because they are based on estimates of surface current variances obtained during a period of intense mesoscale eddy activity, and because we have treated all of the individual velocity estimates as independent in time. Many of these estimates are obtained sequentially from the same drifter, in which case the error of the means decreases in part like N^{-1} rather than $N^{-1/2}$.

DISCUSSION

Although the data set from drifting buoys is still somewhat sparse for the purpose, they seem to provide a basis for quite reasonable estimates of upwelling into the mixed layer. The paucity of near equatorial data revealed in Figure 1 may suggest some doubt as to the relevance of the drifter data to equatorial upwelling. However, the meridional decorrelation scales used in the optimum interpolation introduces strong influences from latitudes at which the data are more plentiful into the field from which the divergence in the zonal band 1.5°N-1.5°S is computed. The decorrelation scales were derived from data for the complete fields to be mapped. The meridional decorrelation scales so obtained, particularly for the overall average, are somewhat longer than those usually ascribed to equatorial wave processes (Philander, 1978), and so may tend to suppress the estimated strength of upwelling very near to the Equator. In regions where the data are particularly sparse, such as that between 1°S and 9°N west of 160°W, and at the margins of the sampled area, the interpolation method assigns greater importance of data at distant locations, resulting in larger scale smoothing that yields a locally uniform and featureless field. The average and monthly calculation of upwelling was for this reason limited to the region east to 130°W. Some of the low monthly values shown in Figure 4 may be affected by this characteristic of the analysis. The months shown with the largest values of divergence in Figure 4 are also generally better sampled.

A salient feature of Figure 4 is the predominance of the meridional component of divergence over the zonal component. This is not particularly surprising, the same result was reported by Taft and Jones (1973) for instance, but it provides support for the results of Hansen and Paul (1984) and others who have tried to estimate the divergence on the basis of the meridional component alone. Figure 4 also shows major annual and semiannual variations. With the exception of the anomalous convergence found for March, all of the major maxima and minima occur within a month of those obtained from the results of Meyers' (1979) analysis of the vertical movements of the 14°C isotherm in the eastern equatorial Pacific. The maximum value of horizontal divergence and upwelling in August is presumably due to local upwelling driven by the maximum in the southeast tradewinds during the northern summer. Kindle (1979) and Busalacchi and O'Brien (1980) have shown that semiannual variation of thermocline depth at the eastern boundary is remotely forced from the central Pacific where the wind stress has a substantial semiannual component and maximum westward (upwelling favorable) stress occurs during December-January. The region over which we have averaged the data for Figure 4 extends far enough to the west to be directly influenced by the semiannual forcing. The amplitude of the seasonal cycle in our result is about three times as large as the sum of Meyers (1979) annual and semiannual constituents. A considerable part of this difference is probably due to the fact that the 14°C isotherm is two to three times as deep on the average as the representative 50 m mixed layer depth that we have used, and that his results are based upon data spanning many years, including El Niño events.

The upwelling estimate by Hansen and Paul (1984) for the period June through October was nearly twice as large as the average of the corresponding period in Figure 4. Their data were from only the region 100°W to 130°W, and were from a single year in which winds and circulation were relatively strong.

The drifting buoy data analyzed were limited to the period prior to the El Niño event of 1982-83 to provide a basis for diagnosis of that event. Additional drifter data are being collected for the EPOCS and TOGA programs. These data and those collected for NORPAX during FGGE will be used to improve the analyses.

Acknowledgments

Preparation of this paper was aided by discussions with Drs. A. Leetmaa, R. L. Molinari, and J. J. O'Brien. The computation and computer graphics were done by Ms. M. Pazos, and the copy was typed by Ms. G. Derr.

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