Determining the Response. of Sea Level to Atmospheric Pressure Forcing Using TOPEX/POSEIDON Data

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Abstract

The static response of sea level to the forcing of atmospheric pressure, the so-called inverted barometer (IB) effect, is investigated using 3'01'1M/PC)S131)ON data. This response, characterized by the rise and fall of sea level to compensate for the change of atmospheric pressure at a rate of -1 cm/mb, is not associated with any ocean currents and hence is normal] y treated as an error to be removed from sea level observation. Linear regression and spectral transfer function analyses are applied to sea level and pressure to examine the validity of the IB effect, In regions outside the tropics, the regression coefficient is found to be consistently close to the theoretical value except for thc regions of western boundary currents, where the mesoscale variability interferes with the IB effect. The spectral transfer function shows near IB response at periods from 20 to 300 days. The regression coefficient averaged over the regions poleward of 30 degrees is -0.84 +/- 0.29 cm/mb (1 standard deviation). The deviation from -1 cm/mb is shown to be caused primarily by the effect of wind forcing on sea level, based on a multivariate linear regression model involving, both pressure and wind forcing. The regression coefficient for pressure resulting from the multivariate analysis is -0.96 +/- 0.32 cm/mb. In the tropics, the multivariate analysis fails because sea level in the tropics is primarily responding to remote wind forcing. However, after removing from the data the wind-forced sea level estimated by a dynamic model of the tropical Pacific, the pressure regression coefficient improves from -1.22 +/-0.69 cm/mb to -0.99 +/- 0.46 cm/mb, clearly revealing an IB response. The result of the study suggests that, with a proper removal of the effect of wind forcing, the. IB effect is valid in most of the open ocean at periods longer than 10 days and spatial scales larger than 500 km.

1. Introduction

A major issue in applying altimetric observation of sea level to the study of ocean circulation has been the response of sea level to the variation of atmospheric pressure at the sea surface. The magnitude of this response can be sizable, but the current velocity resulting from pressure forcing is generally negligible, when compared to that resulting from wind and/or thermohaline forcing (Ponte, 1993, Philander, 1978). Therefore the effects of pressure on sea level must be removed for studying ocean circulation. To a large extent, the response of sea level to pressure can be approximated by the so-called inverted barometer (IB) effect: Sea level rises (falls) at a rate of 0.995 cm per millibar's decrease (increase) in atmospheric pressure, acting like an IB (e.g. Wunsch, 1972; Gill and Niller, 1973; Chelton and Enfield, 1986). Sometimes this phenomenon is referred to as the atmospheric pressure loading. The pressure gradient resulting from the sea level variation compensates for the atmospheric pressure gradient such that there is no net pressure gradient at the sea surface and hence no surface geostrophic currents are created, However, finite subsurface pressure gradients and velocities are possible in a stratified ocean even when the surface is in full IB equilibrium (Ponte, 1992).

When the IB approximation is valid, the effect of pressure can thus be largely removed via a simple formula with the knowledge of the instantaneous pressure field. However, the validity of the IB approximation is not universal, but a function of the frequency and wavenumber of the pressure forcing and also a function of the geographic location (e.g., Wunsch, 1972; Brink, 1978). Theoretical studies based on a barotropic numerical model (Ponte, 1993) suggested that the dynamic response of sea level could be substantial at periods shorter than about 2 days. Ponte (1993) also reported that the IB approximation could also fail at longer periods over extensive regions (e.g., the tropical Atlantic and Pacific). However, the deviation from the IB response is generally less than

10 percent at periods longer than a week (Ponte et al., 1991; Ponte, 1993). Even in the presence of non-IB response, Tai (1993) showed that the IB correction to sea level was still useful in formulating the dynamic equations governing sea level variation,

Non-IB response of sea level has been documented in many observational studies. Anal yses of sea level data have revealed the dynamic response of sea level to atmospheric pressure in the form of basin modes (Luther, 1982) as well as continental shelf waves (Hammon, 1966; Groves and Hammon, 1968). Based on a global study using the Geosat altimeter data, vanDam and Wahr (1993) reported that the response of sea level to atmospheric pressure change was about -0.6 to -0.7 cm/mb. This result is in contradiction to the belief that the IB approximation is generally valid over most of the open ocean at periods longer than a week, They discussed the apparent departure from the IB effect in terms of errors in the atmospheric pressure data and of the response of sea level to wind that is correlated to pressure in its own way. As discussed in vanDam and Wahr (1993) and Ponte (1994), the effects of atmospheric pressure and wind on sea level are anti-correlated to each other; therefore, the I B effect should be somewhat compensated for by the effect of wind. However, this compensation can only account for less than 10 % deviation from the IB effect. The large discrepancy of the result of vanDam and Wahr from the IB effect is most likely due to the substantial errors in the Geosat data.

In this paper we present the results of using the TOPEX/POSEIDON data in analyzing the IB effect. The high accuracy of the data allows a more robust analysis. In particular, we do not need to apply any orbit error removal procedures to the TOPEX/POSEIDON data because the orbit error is less than 4 cm (Tapley et al., 1994). The orbit error removal procedures, which are necessary for reducing the large orbit errors in the Geosat data (circa 25 cm), may have also removed part of the large-scale IB sea level response. The more accurate corrections for the effects of tropospheric water vapor and ionospheric electrons may also improve the results.

Linear regression and coherence analyses were performed to characterize the relationship between sea level and atmospheric pressure. We also applied the technique of multivariate regression to separating the effects of wind and atmospheric pressure on sea level, in order to test the anti-correlation between pressure-driven and wind-driven sea level at mid-latitudes, as discussed by Ponte (1994) and vanDam and Wahr (1993). In the tropics, sea level is primarily responding to remote wind forcing in terms of equatorial waves. The sea level solution of a model assimilating the TOPEX/POSEIDON data in the tropical Pacific was used to represent the wind-forced response. After subtracting the model solution from the TOPEX/POSEIDON sea level data, the residuals were analyzed to detect any hidden IB effect,

2. Altimeter and Atmospheric Pressure Data

All the sea surface height data from the NASA dual-frequency altimeter collected during the first 470 days of the mission were used for the study. (The CNES altimeter data were not used), Standard corrections and editing procedures suggested in the Geophysical Data Record (GDR) Users handbook (Callahan, 1993) were applied. Also corrected for were the effects of the ocean tides using the model of Cartwright and Ray (1990), as well as the solid earth tides and the pole tide. Because the residual tidal errors were still on the order of 5 cm (Schrama and Ray, 1994), an empirical correction for the residual M2, S2, K1 and 01 tides was applied to the data (Fu et al., 1994, unpublished manuscript). Also available in the GDR are atmospheric pressure data provided by the French Meteorological Office based on the analysis of the European Center for Medium Range Weather Forecast (ECMWF).

Both the altimeter data and the atmospheric pressure data were interpolated to fixed grids 6.2 km apart along each pass for colinear analysis. The time averages of sea level and atmospheric pressure at each grid were calculated and removed. The residuals were then used to evaluate the relationship between sea level and atmospheric pressure.

To create arrays of time series on a regular space-time grid for the analysis, the sea level and pressure data were smoothed and interpolated to a space-time grid of 2° x 2° x 10 da ys. Each interpolated value was an average of all the data in a space-time. window of 5 degrees in latitude, 5/cos(latitude) degrees in longitude (the cos(latitude) factor keeps the two spatial dimensions of the window comparable), and 20 days in time, using the following weighting scheme :

$$q = \frac{\sum_i w_i q_i}{\sum_i w_i} \tag{1}$$

$$w_i = \exp(-r_i^2/D^2) \exp(-t_i^2/T^2) \tag{2}$$

where q is the interpolated value of either sea level or pressure at a given grid location and time, q_i its i*observed value, r_i and t_i the spatial and temporal distances between the i^{th} observation and the location and time of the interpolated value, respectively, D the spatial scale (500 km), and T the temporal scale (10 da ys). These scales were dictated by the sampling of the satellite and the scales of atmospheric pressure variability.

Figure 1 shows the geographic distribution of atmospheric pressure variability (in mb) estimated from the data. A major characteristic is a strong latitudinal dependence with the lowest values at tropical latitudes and the highest values at polar latitudes. The magnitude of the 113 effect on sea level in centimeters is also inferred from this map.

3.1 Linear Regression Analysis

To investigate whether sea level is responding to atmospheric pressure in the IB sense, a linear regression analysis was performed on the time. series of sea level and pressure at every grid location. Figure 2 shows the geographic distributions of the regression and correlation coefficients. A value of-1 cm/mb (yellow color) for the regression coefficient represents perfect IB response (the exact value of-0.995 cm/mb has been rounded to -1 cm/mb hereafter). However, the regression coefficient is reliable only when the correlation is significant. Most of the correlation coefficients were estimated based on 44 independent samples (one from each 10-day cycle except for the 3 cycles of the CNES altimeter data not used in the analysis). The resultant 95 % confidence level for non-zero correlation is 0.3 (Bendat and Piersol, 1971). The response is nearly IB in most of the Southern Ocean south of 30° S as well as the northeast Atlantic and the central North Pacific. These are also the regions of high correlation between sea level and pressure.

Mixed results are found in the regions of western boundary currents and the tropics, where the absolute value of the correlation coefficient is relatively low (less than 0.3 in many of these regions), making the regression coefficient relatively unreliable. In the regions of western boundary currents, the effect of strong mesoscale variability is the main reason for the low correlation. In the tropics, the low correlation is mainly caused by the predominant response of sea level to wind. Moreover, the pressure variability is lowest in the tropics, yielding a low signal-to-noise ratio for the IB response.

The globally averaged (area weighted) regression coefficient is -1.11 +/- 0.72 cm/mb. The uncertainty is one standard deviation, of which a good portion is attributed to the tropics. The relatively large standard deviation reflects the spatial variability at the scale of the resolution of our calculation (circa 500 km). Excluding the latitudinal band within 20 degrees from the equator, the average of the coefficient becomes -0.97 +/- 0.40 cm/mb. Displayed in Figure 3 is the zonally averaged regression coefficient as a function of latitude. In the regions equatorward of 30 degrees, the coefficient is not distinguishable from .1 cm/mb to the extent of one standard deviation. Poleward of 30 degrees, however, there is indication that the coefficient is biased away from -1 cm/mb. This bias is a manifestation of the wind effect discussed in Section 1. The direction and magnitude of the bias is similar to the simulation by the model of Ponte (1994).

Both the geographic distribution of the regression coefficient and its global average resulting from the present study are different from the results of vanDam and Wahr (1 993). As noted above, their estimated global regression coefficient is -0.6 to -0.7 cm/mb. One would like to know whether their estimate would be closer to -1 cm/mb if the data in the tropics were excluded. We performed an area-weighted average of the values read off their Figure 3 at latitudes greater than 20 degrees and obtained -0.6 cm/mb. Their Figure 2 shows large areas in a wide range of latitudes where the coefficients are greater than +1 cm/mb, whereas our result shows only limited regions (mostly in the tropics) where the coefficients are positive.

The differences between our result and that of vanDam and Wahr (1993) are probably due to the superiority of the TOPEX/POSEIDON data over the Geosat data. To reduce the relatively large Geosat orbit errors (circa 25 cm), a large-scale orbit error model was fit to the altimeter data and subsequent] y removed from the data before the regression analysis was performed. The same procedure was also applied to the atmospheric pressure data to make the two data sets compatible. However, the procedure removed other large-scale variabilities that had a projection onto the orbit error model as well as the orbit error itself, making the regression calculation somewhat compromised. The larger ionospheric errors and wet tropospheric errors in the Geosat data are undoubtedly also factors causing the differences.

4. Transfer Function and Coherence Analysis

To investigate the IB effect in the frequency domain, a spectral transfer function, $F(\omega)$, was computed as follows:

$$F(\omega) = \langle H(\omega)/P(\omega) \rangle \qquad (3)$$

where ω is frequency, and H and P are the Fourier transforms of sea level and pressure, respectively. The angled bracket denotes an ensemble average performed over 10 degree latitude bands for examining the geographic variation. As revealed in Figure 2, the primary geographic dependence of the regression coefficient is latitudinal, although zonal dependence is quite prominent in the tropics. Under conditions of perfect IB response, the amplitude of F should be unity and the phase should be 180 degrees. The reliability of the estimate of F is dependent on the coherence between sea level and pressure. When the coherence is low, the estimate of F is unreliable.

Shown in Figure 4 are the results of the transfer function and coherence calculations made in 14 latitude bands. The phase shown applies to both transfer function and coherence. Each estimate was based on the average performed according to (3) over all the 2° x 2.° grids within a given latitude band. The 90% confidence level for non-zero coherence was calculated by assuming that each 5° x [5/cos (latitude)]” box, the area from which the data at each grid were derived, provided an independent sample. The degrees of freedom for the coherence estimate are thus proportional to the total number of these boxes (non-overlapping) in each latitude band. The latitudinal variation shown in Figure 4 reveals a degree of symmetry with respect to the equator, Poleward of 30 degrees, the transfer function clearly indicates the IB response at most frequencies, showing amplitude of unity and 180 degree phase, with the coherence being non-zero at a 90% confidence level. At latitudes between 30 and 50 degrees in both hemispheres, however, the coherence shows a decreasing trend at the lowest frequencies, This feature suggests that other processes such as wind-forced fluctuations at those latitudes are interfering with the IB response at low frequencies (see Wunsch, 1972).

At latitudes equatorward of 20 degrees, the transfer function shows significant deviations from the IB response at periods longer than about 30 days. However, the coherence at most frequencies is still non-zero at a 90% confidence level, indicating the existence of a significant linear relationship between sea level and pressure, although not characteristic of the IB effect. Note that the magnitude of the transfer function is generally greater than unity with a phase around 180 degrees. It is well known that the time scales of the ocean's response to wind are shorter in the tropics than mid- and high-latitudes (Philander, 1978). Therefore, the wind-forced response has apparent] y dominated the IB response at periods longer than 30 days in most tropical regions. An attempt to remove the wind-forced response from the sea level data and to examine the IB response in the residual sea level is made in Section 5.

The characteristics of the results in the regions between 20 and 30 degrees are in transition from those in the tropics to those at mid-latitudes; the IB response is valid for periods shorter than 100 days. At lower frequencies, the magnitude of the transfer function is generally greater than unity with a 180 degree phase.

It is worth noting that, without the empirical tidal corrections, the coherence drops below the 90% confidence level at periods close to 60 days at all the tropical latitudes. ‘l’his is due to the large residual M2 and S2 tidal errors in the tropics (e.g., Schrama and Ray, 1994). Both tidal constituents have an aliased period close to 60 days.

5. The Effects of Wind Forcing

As discussed above, the response of sea level to wind at mid-latitudes has a tendency to be anti-correlated to its response to pressure (Ponte, 1994; vanDam and Wahr, 1993), causing the magnitude of the estimated pressure regression coefficient to be less than unity. To analyze this wind effect, we have applied the technique of multivariate linear regression analysis to contemporaneous records of sea level, atmospheric pressure, and wind stress. The wind stress was calculated from the 12.-hourly, 1000 mb wind velocity product obtained from the National Meteorological Center (NMC) by using the bulk formula of Liu et al. (1979). The wind velocit y data were first converted to wind stress. Time averages of the wind stress were calculated and removed. To be used in the calculation described below, the residual wind stress was then interpolated to the same grid as that of sea level and pressure using L@.(1) and (2).

There are many different ways one can write a linear relation between sea level and wind if one includes remote and/or time-lagged wind in addition to local, instantaneous wind. ‘I’he model we used was similar to the one used by Wunsch (1991), i.e.,

$$h_{i,j} = b_1 p_{i,j} + b_2 \tau^x_{i,j+2} + b_3 \tau^x_{i,j-2} + b_4 \tau^y_{i+2,j} + b_5 \tau^y_{i-2,j} \qquad (4)$$

where i,j denote the grid indices in the x and y coordinates, respectively, τ^x and τ^y the wind stress components, h the sea level, p the atmospheric pressure. Note, however, that Wunsch (1991) used wind velocity instead of wind stress in his model. The regression coefficients b_1-b_5 were determined by using a least-squares method, In this form, sea level is responding only to the zonal wind stress two meridional grids (4 degrees in latitude) away and the meridional wind stress two zonal grids away (4 degrees in longitude). The form of the model allows a linear response of sea level to the curl of the wind stress if the data so require. only wind forcing in the vicinity of the sea level is included in the model. The effect of remote wind forcing more than 2 grid spacings (4 degrees) away is not accounted for.

Solutions to (4) were obtained at all grid locations. Our primary interest is the value of b_1: Does it become closer to -1 cm/mb as a result of the multivariate analysis which attempts to separate the effects of wind from pressure? The most prominent results were obtained in the northern hemisphere north of 30° N. Displayed in Figure 5 are the histograms of b_1 and the original regression coefficient based on regression against pressure alone. The values of b_1 are significantly closer to -1 cm/mb than the original result, The average of b_1 is -0.97 +/- 0.43 cm/mb, whereas the average of the original regression coefficient is -0.75 +/- 0.39 cm/mb. Shown in Figure 6 are the histograms for the regions south of 30 S, where the effect of the multivariate regression is less pronounced, but the shift of the centroid of the histogram is in the right direction. The average values are -0.99 +/- 0.27 cm/mb for b, and -0.91 +/- 0.22 cm/mb for the original coefficient,

The regression coefficients b_2-b_5 are quite noisy. The averaged values in the regions north of 30° N and south of 30° S are tabulated in Table 1. The standard deviations are much larger than the results of Wunsch (1991), who used much smoother data resulting from spherical harmonics up to degree and order 10 only. We don't know whether the fact that he used wind velocity instead of wind stress plays a role in causing the difference. Note that b_2 and b_3, the two coefficients for the zonal components of the wind stress, tend to have comparable magnitudes and opposite signs. This feature indicates that the two zonal wind stress variables are correlated with each other in the multivariate regression model. Such feature is more pronounced in the result of Wunsch (1991), probably due to a much smoother database.

The multivariate analysis has not resulted in significant improvement in the solution for b_1 in the tropics. This indicates that the effect of remote wind forcing might be more important in the tropics than the mid- latitude regions. To investigate the effect of remote wind forcing, we obtained the result of a wind-driven model (courtesy of 1. Fukumori). It was a reduced-gravity model covering the tropical Pacific Ocean from 30° S to 30° N. The model was forced by the wind obtained from the National Meteorological Center for the TOPEX/POSEIDON period. 'I"he first year's worth of the TOPEX/POSEIDON sea level data were assimilated into the model with the use of an approximate Kalman smoother (Fukumori, 1994; Fu et al., 1993). The sea level solution was thus a quasi-optimal fit of the wind-driven model to the data. This "smoothed" model sea level accounts for 78 % of the total variance, indicating that the sea level is indeed primarily responding to wind forcing. The IB response, if present, is significantly overridden by the response to wind, which is related to pressure in its own way.

The smoothed model sea level was subtracted from the TOPEX/POSEIDON sea level to remove the effects of wind forcing. The residual sea level was then regressed against atmospheric pressure. The resulting regression coefficient is shown in Figure 7, which also shows the original regression coefficient. The large gradients and complex patterns, as well as the extreme values in the original result have largely disappeared in the new result. The relation of the residual sea level to pressure is apparently much closer to the IB effect. The average regression coefficient has improved from -1.22 +/- 0.69 cm/mb to -0.99 +/- 0.46 cm/mb. Note also that the standard deviation of the new estimate is greatly reduced. This result suggests that the IB effect in the tropical Pacific is basically hidden in a predominantly wind-driven system. It would be interesting to find out whether similar results can be obtained in the tropical Indian and Atlantic Oceans if appropriate wind-driven models are applied there.

The result described in this section has suggested that the IB effect is probably valid over most of the ocean. The apparent disagreement in the simple regression analysis involving pressure alone is caused to a large extent by the missing effects of wind forcing. After the wind effects are removed, the IB effect is more clearly detected,

6. Conclusions

A linear regression of the TOPEX/POSEIDON sea level against the ECMWF atmospheric pressure has resulted in a mean regression coefficient of -0.84 +/- 0.29 cm/mb in the regions poleward of 30 degrees. Most of the uncertainty is attributed to the western boundary current regions, where the intense mesoscale variability interferes with the IB effect.

The deviation of the mean coefficient from the theoretical IB coefficient is shown to be caused primarily by the effect of wind forcing. With the effect of wind forcing accounted for by a multivariate regression model, the mean regression coefficient for pressure is improved to -0.96 +/- 0.32 cm/mb. This result is consistent with the finding of the modeling study of Ponte (1994): Wind-driven sea level at mid-latitudes is anti-correlated to pressure-driven sea level. The effect of wind forcing thus tends to compensate for the IB effect, leading to a pressure regression coefficient with magnitude slightly less than unity if the wind effect is not taken into account.

Analyses of zonally averaged data in the frequency domain have shown that the IB effect is generally valid at periods from 20 (the highest resolvable period) to 300 days in the regions poleward of 30 degrees. There is a slight indication for non-I B response of sea level at periods longer than 300 days.

In the tropics, the IB effect is not directly observable. First, sea level in the tropics is primarily responding to remote wind forcing, Se..ond, the pressure variability is lowest in the tropics, yielding a low signal-to-noise ratio. The multivariate regression fails in the tropics because the remote wind forcing is not properly accounted for. With the wind-driven sea level removed by using a dynamic model in the tropical Pacific, the residual sea level then reveals a clear IB response to pressure.

The result of the study is quite different from the Geosat result of vanDam and Wahr (1993), who reported a global regression coefficient of -0.6 to -0.7 cm/mb. The difference between the two studies is probably due. to the difference in the accuracy of the data. 'I'he relatively large errors in the Geosat orbit as well as the ionospheric and wet tropospheric corrections are among the factors causing the difference.

The result of the study suggests that the IB response of sea level is generally valid in the open ocean at periods longer than 10 days and spatial scales larger than 500 km, consistent with the modeling studies of Ponte (1993, 1994). However, the significance of the apparent spatial variability of the result at the scale of the resolution of our analysis (circa 500 km) is not clear. The 10-day repeat period of the satellite prevents investigation of the shorter period regime where non-113 response is expected to be more significant. The apparent deviation of the result of the pressure-only regression from the theoretical IB effect can mostly be ascribed to the effect of wind forcing on sea level, as demonstrated in both the mid and high latitudes and the tropics.

Acknowledgments

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Table 1. The coefficients of b_2 - b_5 (in dyne/cm^3). The uncertainty is one standard deviation.

	b_2	b_3	b_4	b_5
North of 30° N	**1.25 +/- 3.04**	**-2.57 +/- 4.14**	**-0.40 +/ -4.83**	-0.07 +/ -4.63
South of 30° s	**-0.91 +/- 2.40**	0.45 +/- 1.4-/	-0.23 +/-3.07	-0.41 +/ -2.93

Figure Captions

Figure 1. Therms variability of the atmospheric pressure based on 470 days' worth of data from the ECMWF analysis. The contour interval is 1 mb.

Figure 2. The regression coefficient (upper panel) and the correlation coefficient (lower panel) between sea level and atmospheric pressure.

Figure 3. Zonally-averaged regression coefficient between sea level and pressure as a function of latitude, The dashed lines indicate one standard deviation.

Figure 4a. The spectral transfer function (solid lines) and coherence (dashed lines) bet ween sea level and pressure estimated in 14 latitude bands. The dotted lines indicate the 90 % confidence level for non-zero coherence..

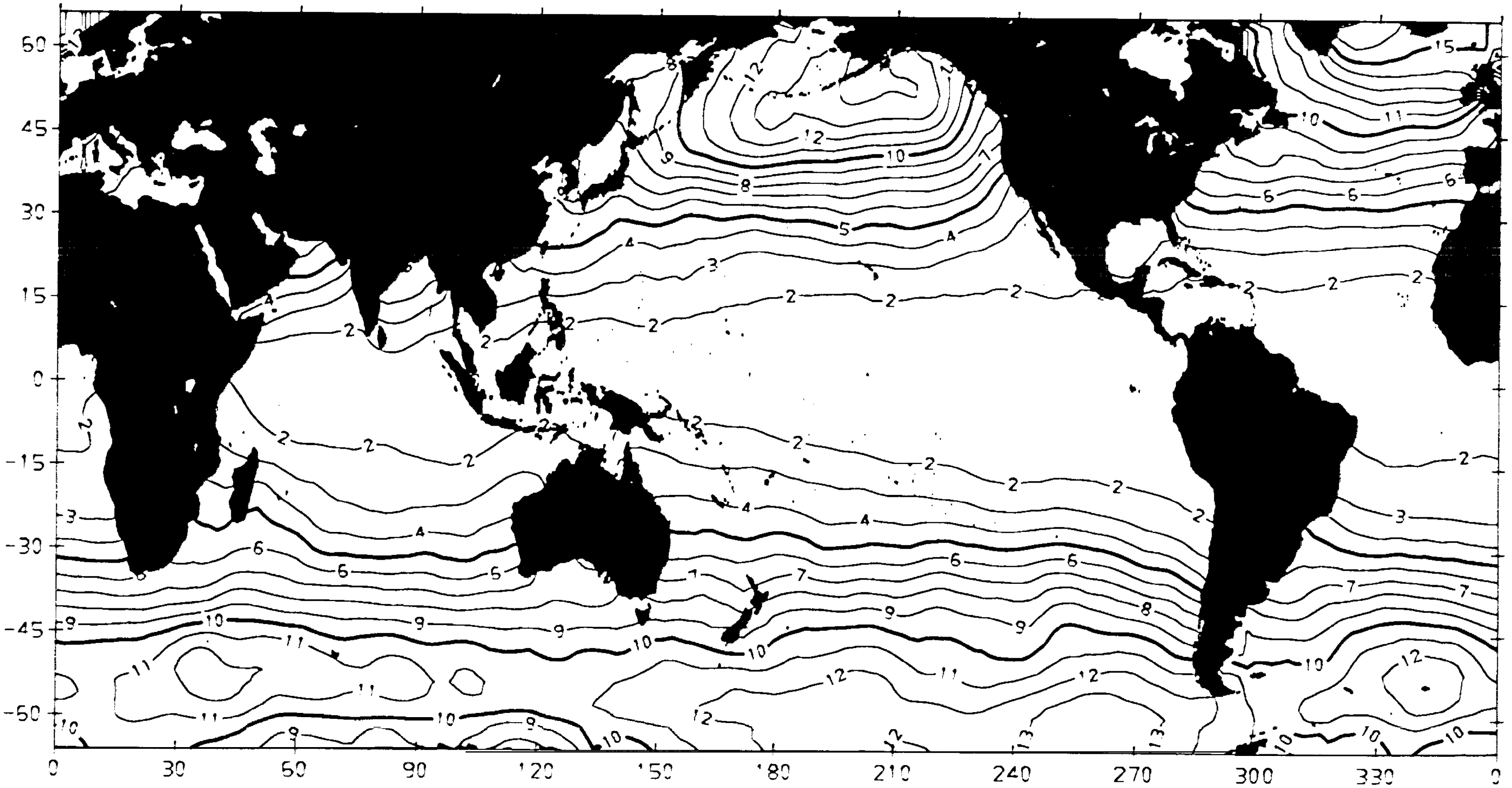
Figure 4b. The phase for the transfer function and coherence shown in 4a,

Figure 5. Histograms (expressed in area] population in arbitrary unit) of the regression coefficient of sea level versus pressure at all locations north of 30° N from the regression against pressure alone (solid line) and from the multivariate regression (dashed line).

Figure 6. Same as in Figure 5 except for all locations south of 30° S.

Figure 7. Regression coefficient of sea level versus pressure based on the original TOPEX/POSEIDON sea level (upper panel) and the residual sea level after the wind-driven model sea level was removed (lower panel).

Fig 1



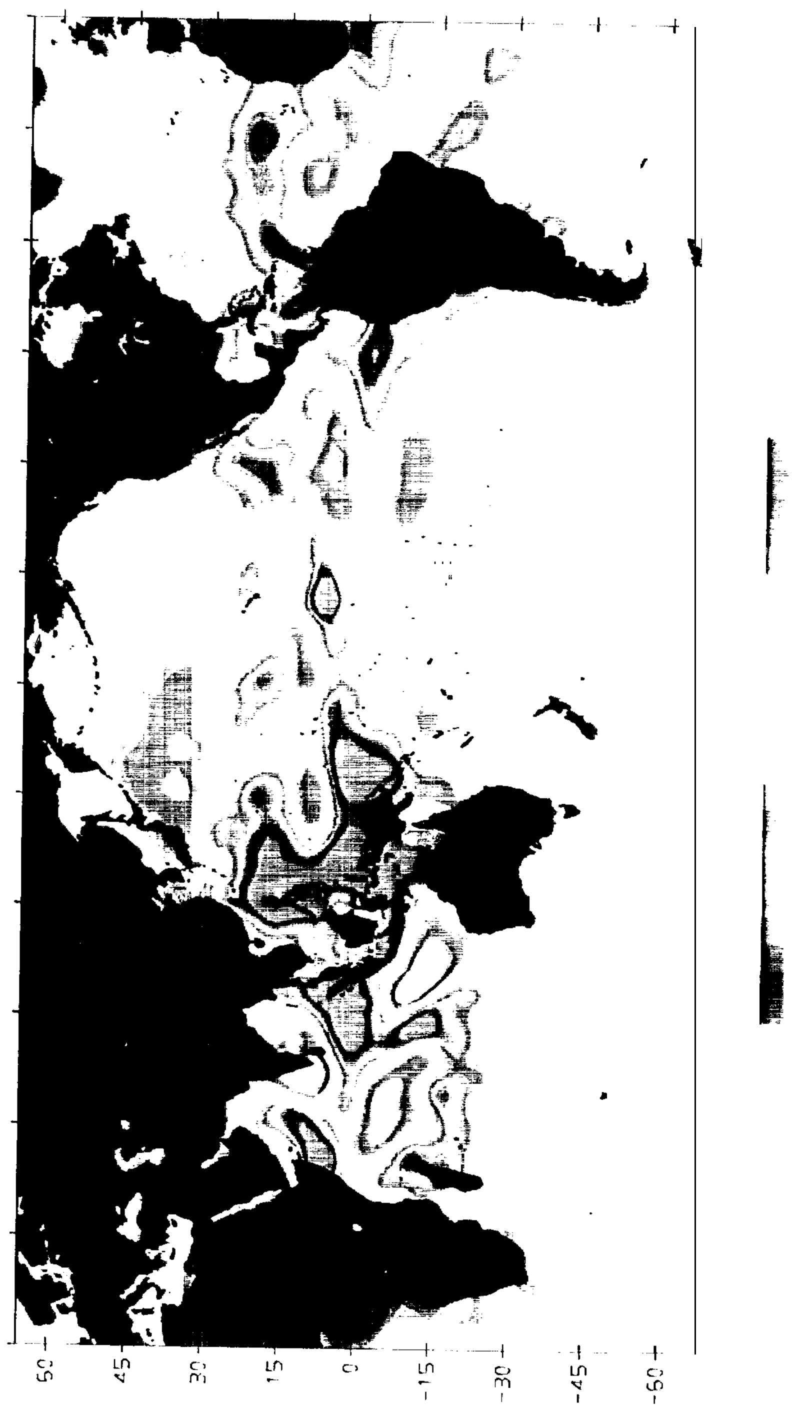
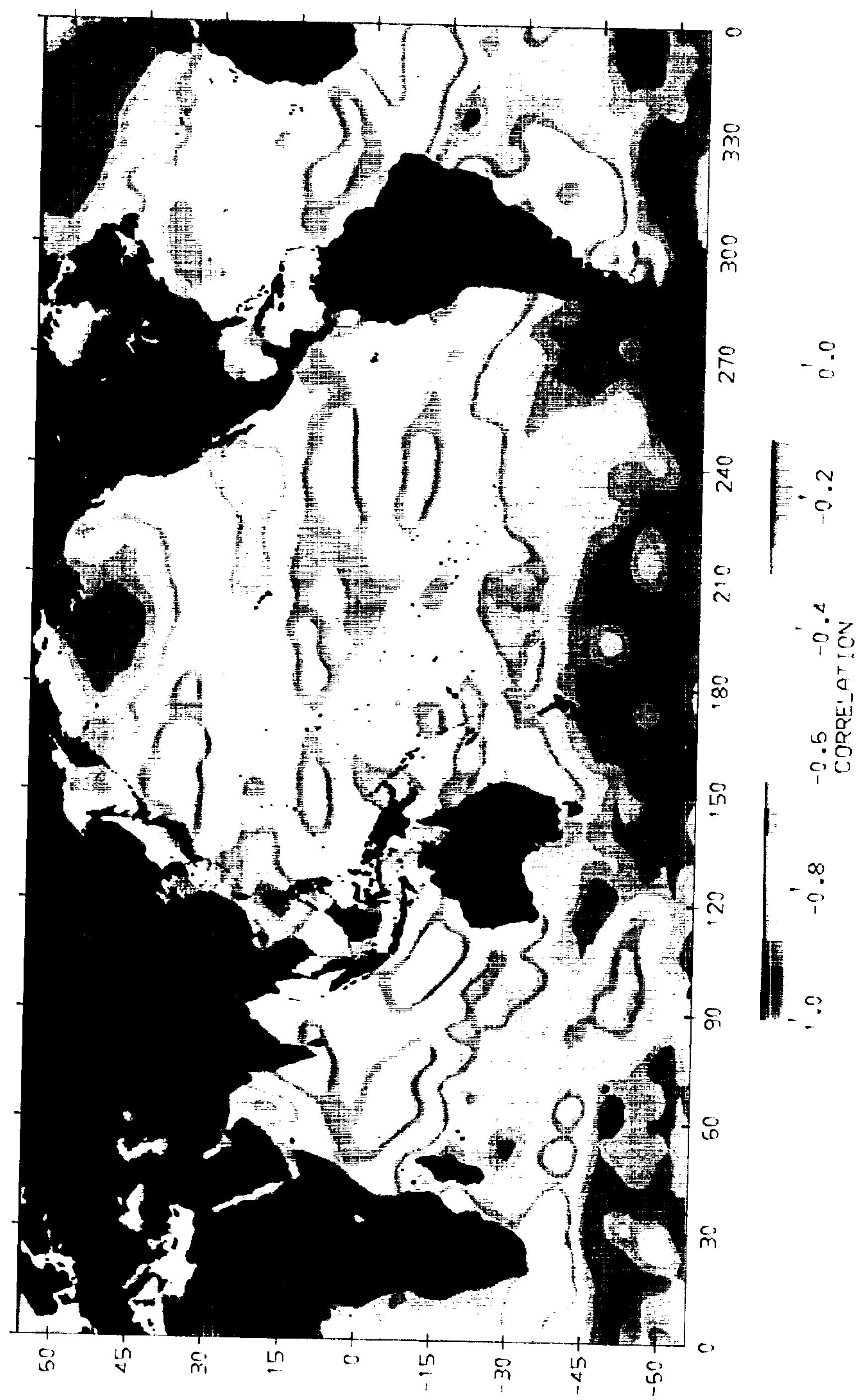


Fig. 2 (upper)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-60

Fig 2 (lower)

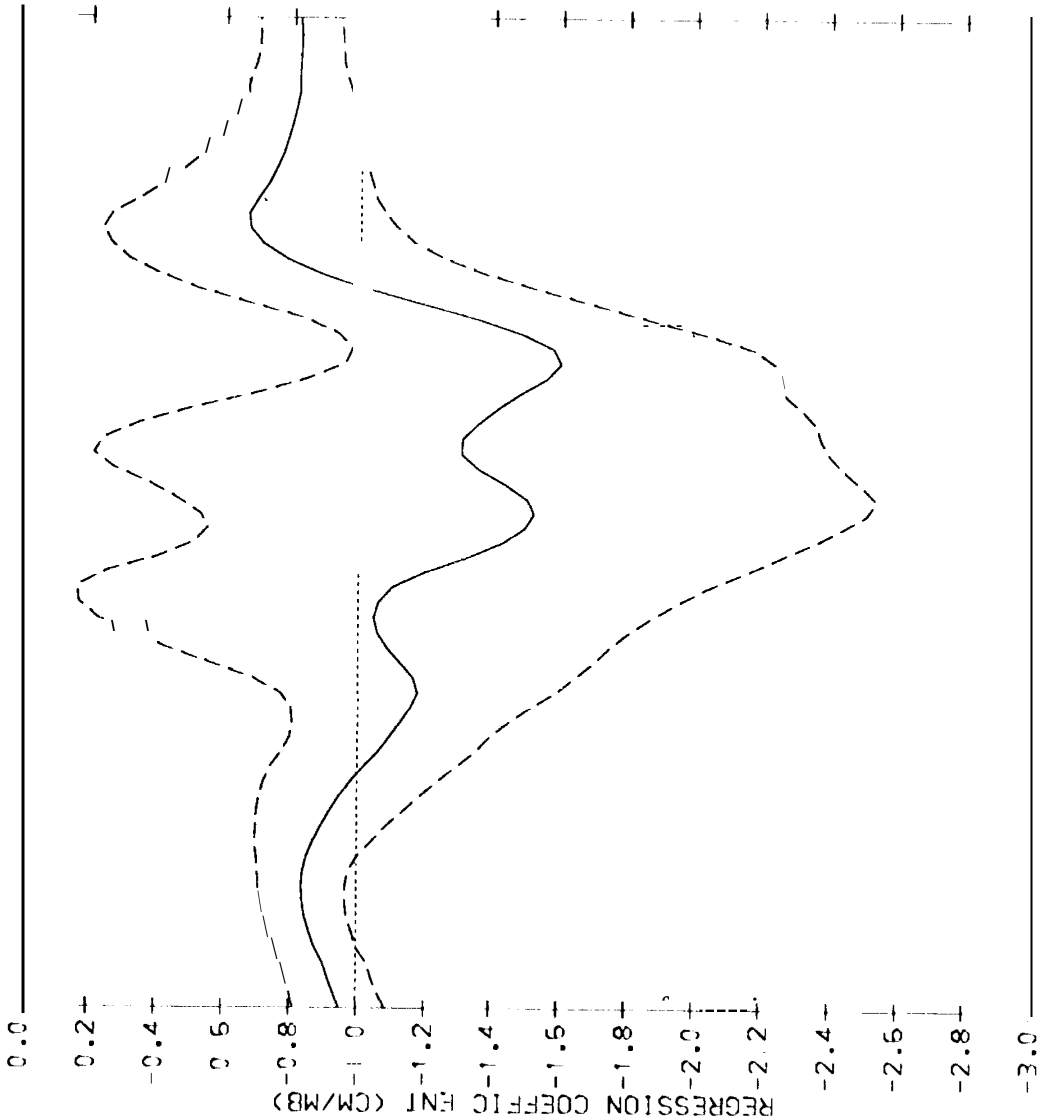
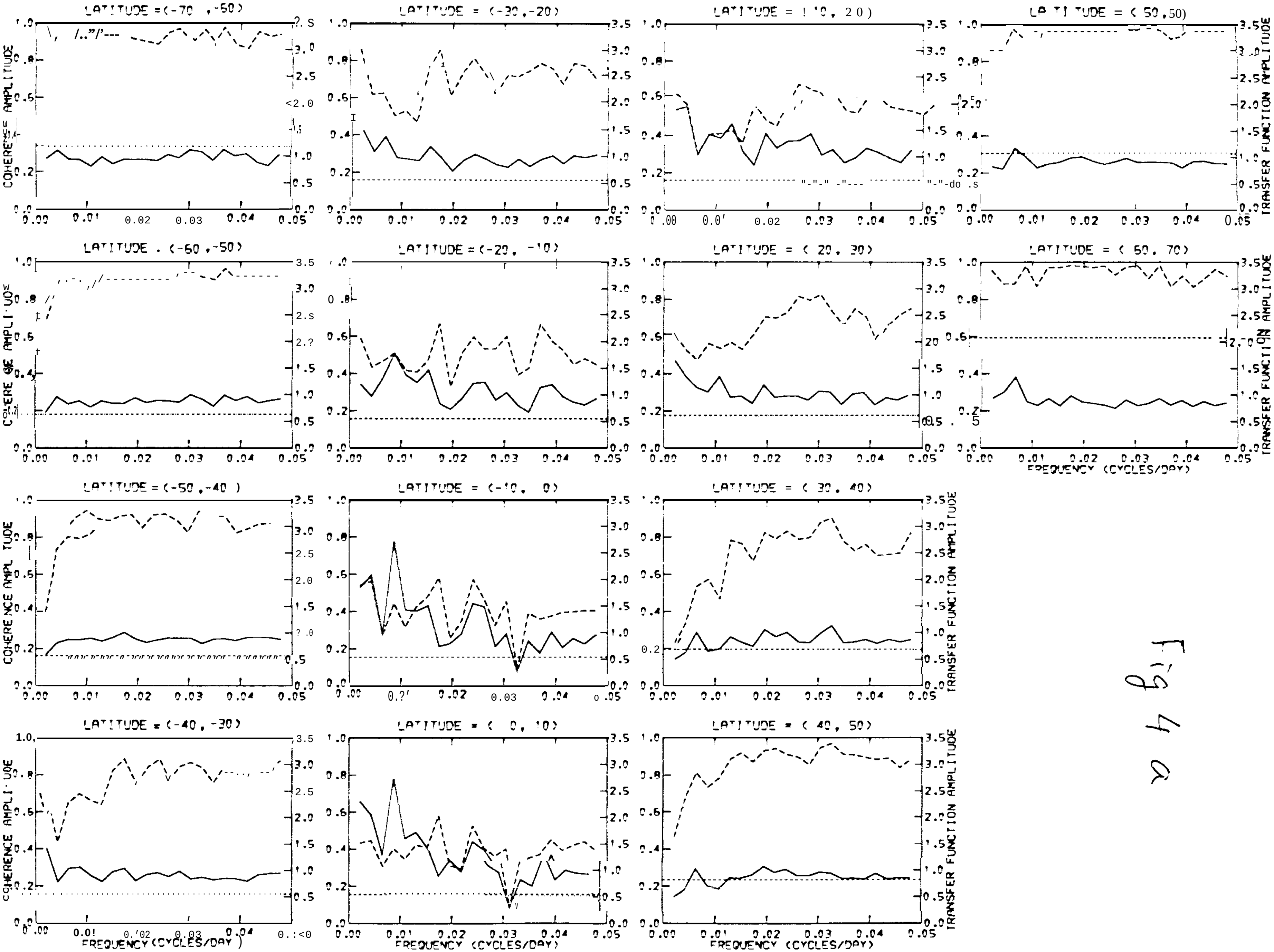


Fig 3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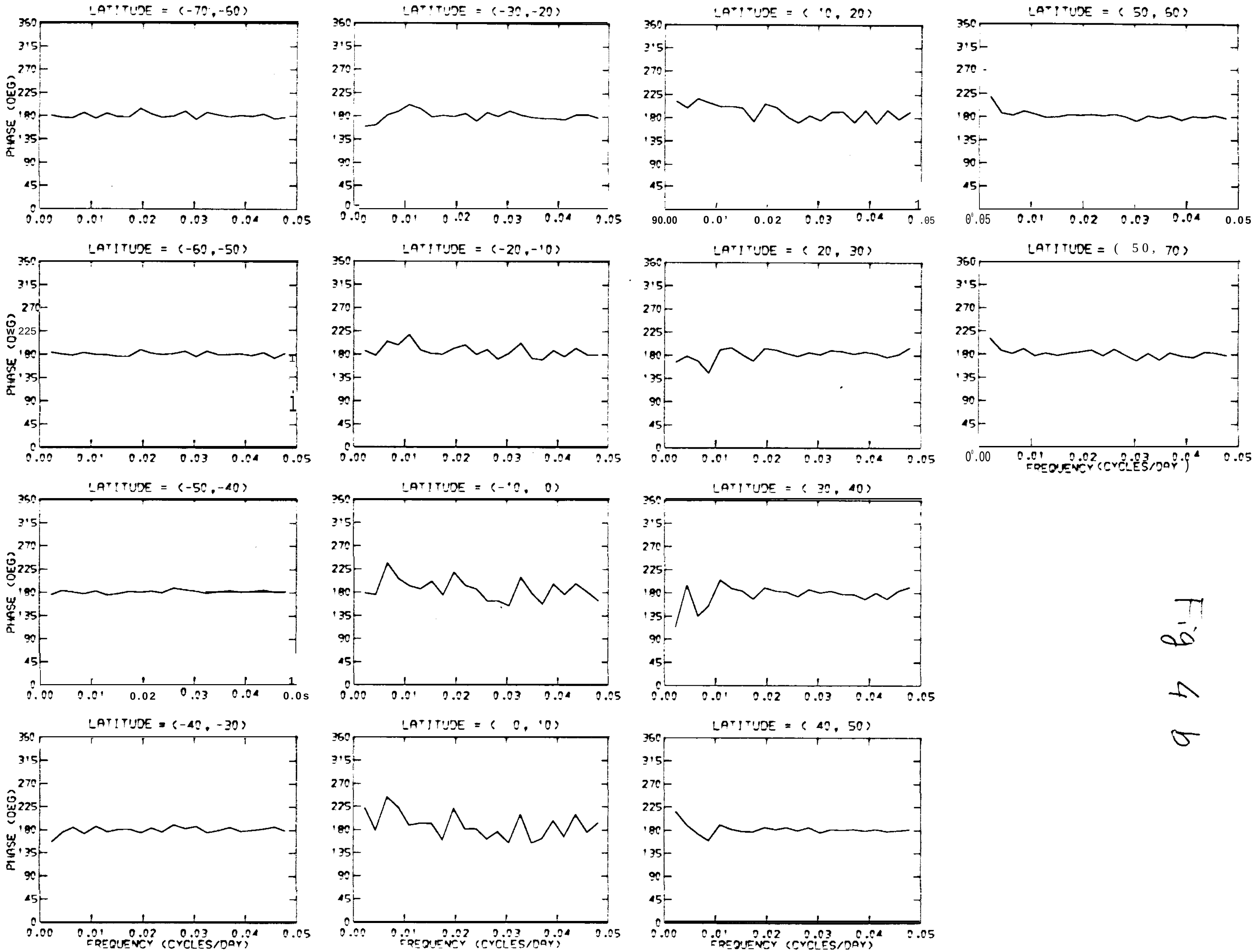
LATITUDE = (-30, -20)
LATITUDE = (10, 20)
LATITUDE = (50, 50)
LATITUDE = (-60, -50)
LATITUDE = (-20, -10)
LATITUDE = (20, 30)
LATITUDE = (50, 70)
LATITUDE = (-50, -40)
LATITUDE = (-10, 0)
LATITUDE = (30, 40)
LATITUDE = (-40, -30)
LATITUDE = (0, 10)
LATITUDE = (40, 50)
COHERENCE AMPLITUDE
TRANSFER FUNCTION AMPLITUDE
FREQUENCY (CYCLES/DAY)

Fig 4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LATITUDE = (-60, -50)
LATITUDE = (-50, -40)
LATITUDE = (-40, -30)
LATITUDE = (-30, -20)
LATITUDE = (-20, -10)
LATITUDE = (-10, 0)
LATITUDE = (0, 10)
LATITUDE = (10, 20)
LATITUDE = (20, 30)
LATITUDE = (30, 40)
LATITUDE = (40, 50)
LATITUDE = (50, 60)
LATITUDE = (60, 70)
PHASE (DEG)
FREQUENCY (CYCLES/DAY)

Fig 4b

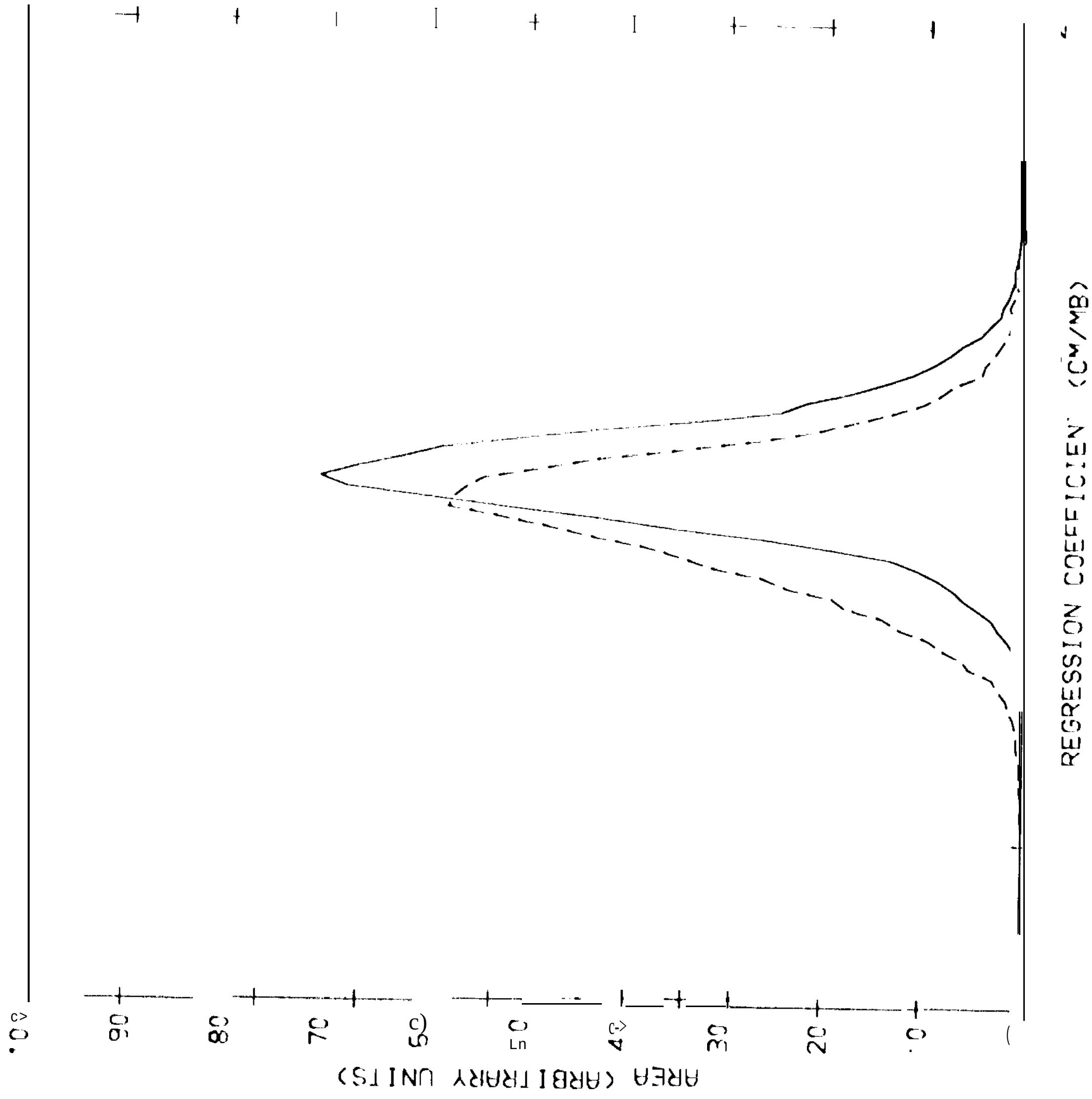

Fig 5

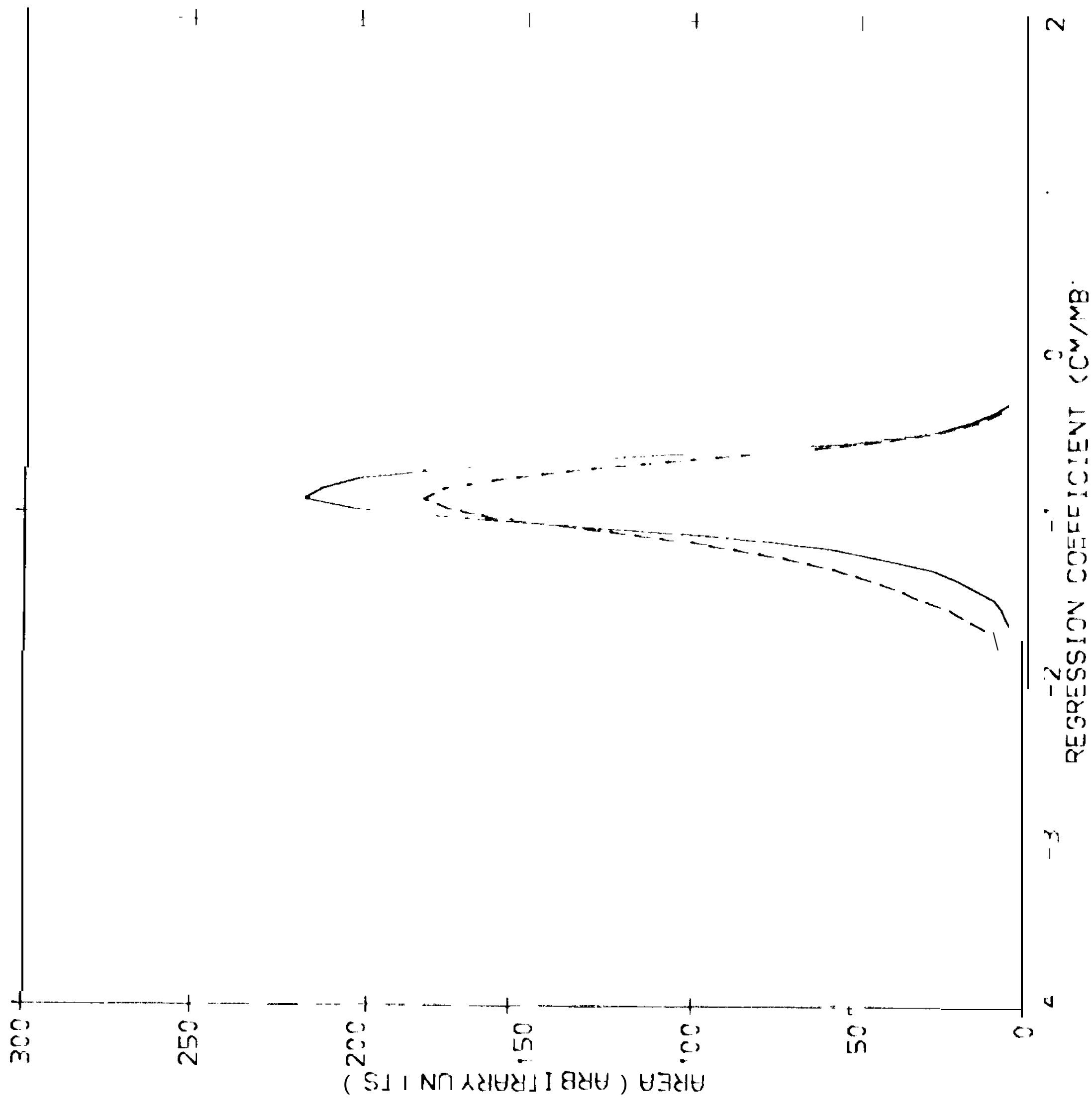
REGRESSION COEFFICIENT (CM/MB)
AREA (ARBITRARY UNITS)
300
250
200
150
100
50
0
-2
-1
0
2

Fig 6

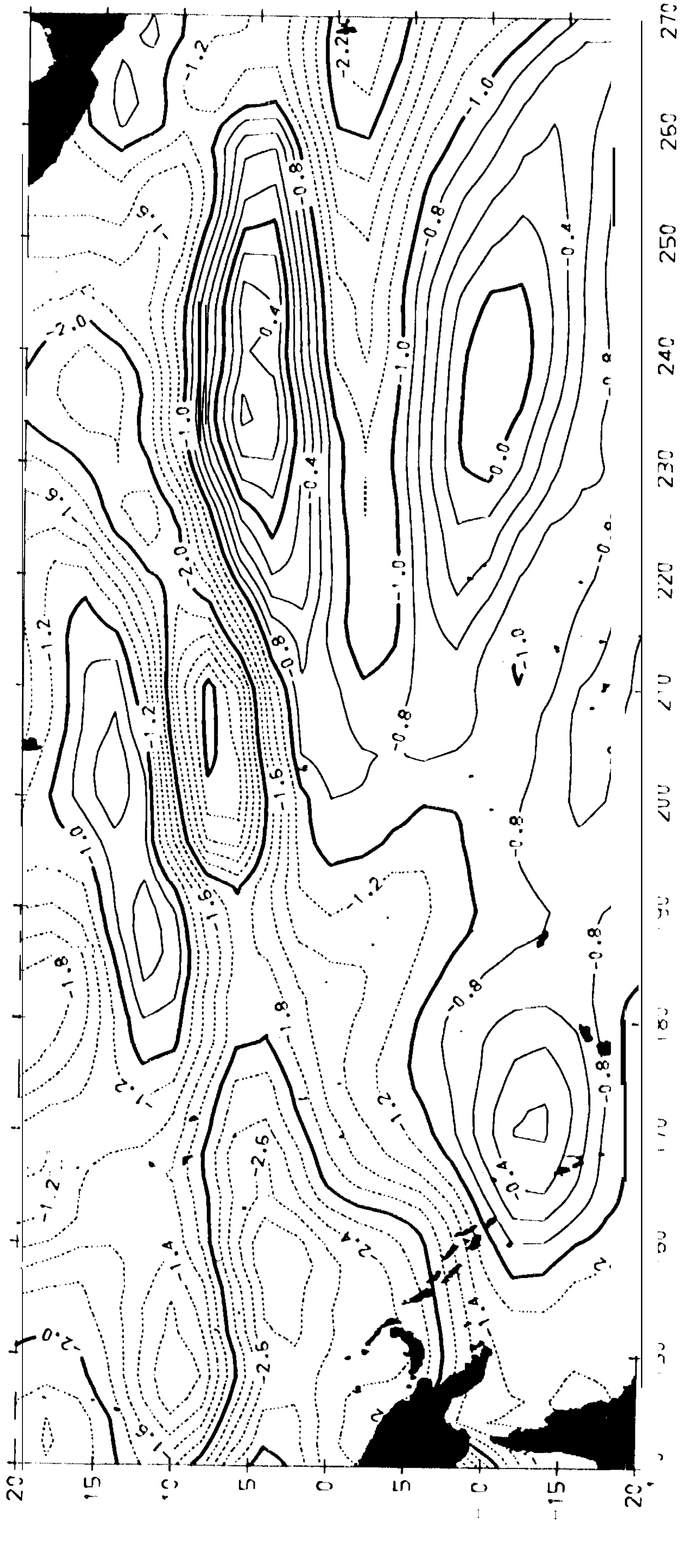

Fig 7 (upper)

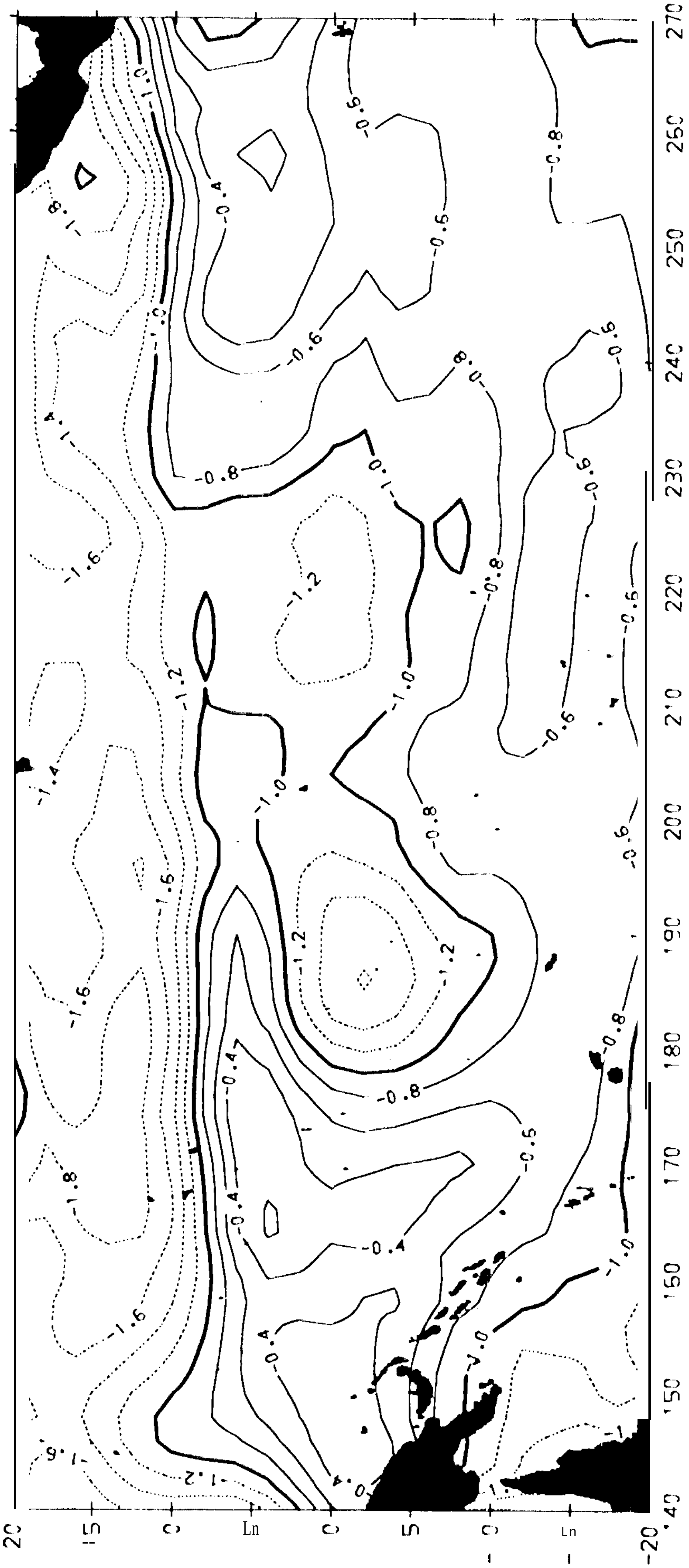

Fig 7 (lower)