Quart J. R. Met. Soc. (1983), **109**, pp. 153–168

551.465.63:551.465.7

Variations of zonal mean sea surface temperature and large-scale air–sea interaction

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(Received 5 May 1982; revised 27 September 1982)

SUMMARY

Twenty-four years data of zonal mean sea surface temperature (SST) were assembled to examine the seasonal and nonseasonal variations of SST. Outside the tropics, annual and semi-annual variations of SST dominate, accounting for over 90% of the variance. In the tropics, the nonseasonal variation of SST is dominated by low-frequency variations with periods longer than 3 years in the Pacific. Variations with periods of 24–32 months account for more than 10% of the variance in the tropical Atlantic from 25 °N to 10 °S and in the equatorial Pacific from 0–10 °S. An empirical orthogonal function (EOF) analysis showed that the major global pattern of zonal mean SST variations is dominated by variations in the equatorial eastern Pacific which precede zonal mean SST variations elsewhere. The time series of this EOF is found to be correlated with an index of the Southern Oscillation. The maximum correlation coefficient is −0·61, with the index leading by 2 months. Finally, the relation between SST and tropical tropospheric temperature (TTT) is examined. SST variations in the Atlantic and in the Indian Ocean are almost contemporary with variations in TTT. Possible explanations for the 1–2 seasons lead time between SST and TTT as found in earlier studies are suggested.

1. INTRODUCTION

In a review paper on the subject of long-range weather forecasting, Nicholls (1980) concluded with the view that forecast techniques which consider air–sea interaction hold most promise for the future of long-range weather forecasting. Such optimism is based on successful searches for predictive relations which generally involve sea surface temperature (SST) as a predictor or as an important factor in the predictive relation. The interested reader is referred to Nicholls for an up-to-date summary of research on this subject.

To characterize objectively the spatial patterns of SST variations, empirical orthogonal function (EOF) analyses have been carried out. Newell and Weare (1976a, b) found that the time series of the first EOF of Pacific SST, which showed an area of maximum weight in the eastern equatorial Pacific, varies one season ahead of the tropical tropospheric temperature (TTT) and two seasons ahead of the northern hemisphere tropospheric temperature; tropical tropospheric temperature is defined here in terms of the geopotential thickness between 300 and 700 mb for seven tropical stations situated between 20 °N and 20 °S. The stations were listed in Table 1 of Navato *et al.* (1981). Similar leads were noted when the actual temperature of the eastern equatorial Pacific was compared with TTT (Newell 1979; Angell 1981). The relationship between equatorial Pacific SST and tropospheric temperature has also been examined by Rowntree (1972, 1979) using a numerical model and observed data.

While higher SST may increase evaporation and hence latent heating of the troposphere, it is known that water molecules have a residence time in the atmosphere of only about 10 days (Peixoto 1973), much shorter than the two seasons noted above. Do regions other than the eastern equatorial Pacific exhibit evaporation changes somewhat later?

To provide the basic data for the study of large-scale air–sea interaction, a data set of zonal mean SST has been assembled (Newell and Hsiung 1979). It is the purpose of this study to present the seasonal cycle of the zonal mean SSTs and the basic statistics of nonseasonal SST variations. It will then be demonstrated by an EOF analysis of zonal mean SSTs that SST variations in the eastern equatorial Pacific dominate nonseasonal

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variations of the global oceans. It will be shown that this mode of SST variation is related to the well-known Southern Oscillation (SO), the major sea level pressure and rainfall variation pattern first pointed out by Walker (1924). The sequence of events in the zonal mean SST variations which lead to the eventual warming of the tropical troposphere is examined by correlation analyses at lags with TTT.

We feel that zonal means can give us a first-order picture of SST variations. Inspection of SST maps shows that surface isotherms are zonally oriented in middle to high latitudes. In a study of the seasonal variation of SSTs, Panfilova (1972) distinguished five types of seasonal variations. The area occupied by each type of variation is basically zonal. Finally, we noted that the principal EOF patterns of nonseasonal SST variations do not show strong east–west variation. The first EOF pattern for the Indian Ocean SST is characterized by weights of the same sign everywhere (Navato *et al.* 1981). For the north and tropical Pacific, it is characterized by opposite changes between the tropical and central north Pacific (Weare *et al.* 1976). For the Atlantic, the first EOF pattern is characterized by opposite changes between lower latitudes (equator to about 30 °N) and middle latitudes and latitudes south of the equator (Weare 1977).

2. THE DATA

The data set used in this study has been described by Newell and Hsiung (1979). It consists of monthly means of SST for every 5 degrees latitude for all three oceans for the period January 1949 to December 1972, making a total of 288 months. Zonal means are computed for the latitude belts 70–65 °N, 65–60 °N, and so on, denoted 67·5 °N, 62·5 °N.

In the equatorial and southern Pacific, data from the western regions were poor. Thus in the computation of zonal means, western limits were 180° from 22·5 °N to the equator and 140 °W south of the equator. A southern limit of 12·5 °S was used because of too few observations to the south. The Atlantic enjoyed a better data coverage and zonal means are computed from 67·5 °N to 32·5 °S. For the Indian Ocean, eastern limits are 120 °E north of the equator and 140 °E south of the equator. Because of gaps in the data, latitudes south of 17·5 °S are not included in this study. Maps of regions included in this data set are given in Navato *et al.* (1981).

TABLE 1. PERCENTAGE OF VARIANCE EXPLAINED BY ANNUAL AND SEMI-ANNUAL HARMONICS

lat.	Indian Ocean	Pacific	Atlantic
67·5 °N			85·0
62·5 °N			92·4
57·5 °N		92·9	95·3
52·5 °N		96·0	96·6
47·5 °N		97·3	96·1
42·5 °N		97·4	97·1
37·5 °N		98·4	98·0
32·5 °N		99·0	98·0
27·5 °N		98·0	97·1
22·5 °N	94·7	91·7	94·6
17·5 °N	91·0	87·7	92·3
12·5 °N	87·6	76·7	89·9
7·5 °N	82·9	41·1	70·5
2·5 °N	72·6	39·6	84·1
2·5 °S	71·8	71·3	90·6
7·5 °S	83·6	74·8	92·5
12·5 °S	90·0	71·8	93·9
17·5 °S	90·7		93·8
22·5 °S			93·9
27·5 °S			93·6
32·5 °S			93·5

3. SEASONAL VARIATIONS

An harmonic analysis, as set out by Barry and Perry (1973), was performed on the time series of zonal mean SSTs. A computer program which calculates amplitudes and phases of the 144 harmonics was used to perform the analysis. The 24th harmonic, with a period of 288 months per 24 cycles, corresponds to the annual harmonic.

Inspection of all spectra revealed that the annual and semi-annual harmonics are the dominant features. Table 1 shows the percentage of total variance of the time series contained in annual and semi-annual harmonics. Outside the tropics, these two harmonics explain more than 90% of the total variance in most latitudes. A broad region in the tropical Pacific has less than 80% of the total variance explained by the seasonal cycle. At 7·5 °N and 2·5 °N, the seasonal cycle explains only about 40% of the total variance.

The amplitudes and phases of the annual and semi-annual harmonics are presented in Tables 2 and 3. The largest annual and semi-annual amplitude is found at mid-latitudes in the Pacific. The phases are defined relative to December 15. A phase of 8·5 months corresponds to a peak in the harmonic wave at the end of August. A secondary maximum in the semi-annual amplitude appears in the Indian Ocean in early May. In this region, the semi-annual amplitudes are comparable to, and in some latitudes larger than, the annual amplitudes. Panfilova (1972) has used climatological monthly mean grid point values to compute the amplitudes and phases of the annual and semi-annual variations. If the phases of harmonic waves are relatively constant at a latitude, the amplitudes obtained in Panfilova's study can be zonally averaged and compared with our results. There is general agreement. Because Panfilova's study was based on climatology, it was not possible to compare seasonal and nonseasonal variations as is done here.

TABLE 2. AMPLITUDE AND PHASE OF ZONAL MEAN SEA SURFACE TEMPERATURE: ANNUAL COMPONENT

	Indian Ocean		Pacific		Atlantic	
	°C	months	°C	months	°C	months
67·5 °N					2·30	8·2
62·5 °N					1·98	8·1
57·5 °N			4·06	8·2	2·74	8·2
52·5 °N			2·70	8·5	3·16	8·2
47·5 °N			2·92	8·6	3·08	8·4
42·5 °N			4·56	8·7	3·60	8·4
37·5 °N			4·42	8·7	3·52	8·5
32·5 °N			2·94	8·6	3·22	8·6
27·5 °N			3·02	8·5	2·70	8·6
22·5 °N	2·90	7·3	1·86	8·9	1·86	8·9
17·5 °N	1·30	6·6	1·40	8·9	1·68	9·1
12·5 °N	0·88	5·6	0·90	8·2	1·40	9·0
7·5 °N	0·62	5·0	0·42	7·0	0·58	9·0
2·5 °N	0·50	5·0	0·68	4·5	0·94	2·8
2·5 °S	0·60	3·4	1·90	3·6	1·68	2·8
7·5 °S	1·02	2·4	1·74	3·7	1·94	3·2
12·5 °S	1·44	2·2	1·60	3·4	1·92	3·1
17·5 °S	1·94	2·3			2·18	2·9
22·5 °S					2·28	2·6
27·5 °S					2·68	2·5
32·5 °S					3·02	2·1

4. NONSEASONAL VARIATIONS

Time series of nonseasonal variations are obtained by subtracting the long-term monthly means (d_j) from the monthly mean SSTs. At each latitude, the standard deviations (s.d.) of interannual variability for all months are calculated. (d_j and s.d. are defined

TABLE 3. AMPLITUDE AND PHASE OF ZONAL MEAN SEA SURFACE TEMPERATURE: SEMI-ANNUAL COMPONENT

	Indian Ocean		Pacific		Atlantic	
	°C	months	°C	months	°C	months
67·5 °N					0·66	1·9
62·5 °N					0·54	1·7
57·5 °N			0·94	1·8	0·50	1·6
52·5 °N			0·81	2·3	0·42	1·9
47·5 °N			1·02	2·5	0·68	2·1
42·5 °N			1·12	2·5	0·70	2·1
37·5 °N			0·80	2·3	0·54	2·1
32·5 °N			0·42	2·3	0·36	2·3
27·5 °N			0·04	2·1	0·16	2·6
22·5 °N	0·82	4·7	0·16	3·4	0·18	4·0
17·5 °N	1·04	4·7	0·10	3·7	0·24	4·2
12·5 °N	0·84	4·6	0·20	4·4	0·28	4·7
7·5 °N	0·68	4·5	0·16	4·7	0·32	5·0
2·5 °N	0·50	4·4	0·14	4·0	0·42	4·8
2·5 °S	0·42	4·5	0·22	3·5	0·53	4·3
7·5 °S	0·38	4·8	0·18	3·5	0·32	4·3
12·5 °S	0·22	4·9	0·20	3·4	0·22	4·0
17·5 °S	0·06	4·5			0·14	3·0
22·5 °S					0·16	2·4
27·5 °S					0·26	2·0
32·5 °S					0·16	1·9

in the appendix.) The monthly s.d.s for all latitudes are less than 1 K and are generally in the range 0·3 to 0·5 K except at 57·5 °N and 2·5 °S in the Pacific. The maximum monthly s.d. is found in the tropical Pacific in the northern autumn to early winter. At 2·5 °S in the Pacific, the s.d. is a minimum in February (0·5 K) and a maximum in November (1·1 K).

(a) *Stationarity*

The nonseasonal time series are tested for stationarity using the run test as outlined by Bendat and Piersol (1971). Each time series is divided into 12 time segments and the means and s.d.s of each segment as well as the median of these quantities are calculated. Runs above and below the median are counted. Too many runs indicate oscillations and too few trends. For twelve segments the 5% level of significance for the run test is $3 < r < 10$, where r is the number of runs. From the run test, it was found that most zonal means are stationary except for 57·5 °N and 52·5 °N in the Atlantic where the number of runs is two. Examination of the segment means at these latitudes showed a cooling trend. These trends have been noted in Newell and Hsiung (1979).

(b) *Normality*

The frequency distribution of each time series is tested for normality by the chi-square test (Bendat and Piersol). The monthly means are normalized by the respective monthly s.d.s to downweight the seasonal dependence of the monthly variance. Chi-square statistics, defined as

$$\chi^2 = \sum_{i=1}^{16} (f_i - F)^2/F$$

where f_i is the observed frequency distribution in a class interval and F is the frequency in the same class interval as would be expected from a normal distribution, are calculated for 16 class intervals. For 13 degrees of freedom, the chi-square value, significant at the 95% and 99·5% levels, are 22·4 and 29·8, respectively. The chi-square values at latitudes 22·5 °N, 7·5 °N in the Indian Ocean, 47·5 °N, 27·5 °N and 2·5 °N in the Pacific and 42·5 °N, 12·5 °N and 7·5 °N in the Atlantic are significant at above the 99·5% level. Latitudes with chi-square values significant at the 95% level are 2·5 °N in the Indian Ocean, 12·5 °N, 7·5 °N and 7·5 °S in the Pacific and 27·5 °N, 17·5 °N, 2·5 °N and 22·5 °S in the Atlantic. The frequency distributions of zonal mean SST at these latitudes therefore deviate significantly from that of a normal distribution. Inspection of the frequency distribution at 12·5 °N in the Atlantic showed a bimodal distribution.

(c) *Persistence*

Long-range forecasts are sometimes based on persistence (Nicholls 1980). To examine the persistence of the time series, the autocorrelation coefficients for lags up to 15 months for all latitudes are calculated. The autocorrelation functions vary smoothly with lags. Shown in Table 4 are the autocorrelation coefficients at a lag of 6 months and the lag in months when the autocorrelation coefficient drops below 0·5. A striking feature is the strong persistence in the tropical and subtropical Pacific from 22·5 °N to 7·5 °S, where the autocorrelation coefficients stay above 0·5 even after 4 months. Wright (1979) and Fleer (1981) noted a strong persistence in rainfall in the central equatorial Pacific. Wright suggested a positive feedback between the Walker circulation and SST which produces the strong persistence. The most persistent time series is the one at 22·5 °N, the autocorrelation coefficient at a lag of 15 months is 0·32.

TABLE 4. AUTOCORRELATION COEFFICIENTS OF NONSEASONAL ZONAL MEAN SST AT A LAG OF 6 MONTHS (a) AND LAG IN MONTHS WHEN AUTOCORRELATION COEFFICIENT DROPS BELOW 0·5 (b)

Latitude	Indian Ocean		Pacific		Atlantic	
	(a)	(b)	(a)	(b)	(a)	(b)
67·5 °N					23	1
62·5 °N					27	2
57·5 °N			33	4	35	2
52·5 °N			25	3	21	3
47·5 °N			23	2	21	2
42·5 °N			25	3	25	3
37·5 °N			31	3	40	3
32·5 °N			12	2	35	2
27·5 °N			13	2	33	3
22·5 °N	10	1	61	10	25	3
17·5 °N	17	2	53	7	28	4
12·5 °N	17	2	51	7	26	4
7·5 °N	30	3	45	6	21	2
2·5 °N	12	1	34	5	16	3
2·5 °S	12	1	27	5	15	3
7·5 °S	15	1	36	5	28	3
12·5 °S	19	1	26	3	18	2
17·5 °S	16	2			08	2
22·5 °S					20	3
27·5 °S					24	2
32·5 °S					09	1

Autocorrelation coefficients expressed in hundredths

The Indian Ocean showed the least persistent SST variations among the oceans, with autocorrelation coefficients dropping below 0·5 in less than a month in most latitudes. This comparatively low persistence was also noted in the time series of the first EOF pattern of Indian Ocean SSTs (Navato *et al.* 1981) and nonseasonal rainfall in this region (Fleer 1981).

(d) *Low-frequency variations*

In this section, we examine low-frequency variations of SST with respect to large-scale atmospheric phenomena. Walker (1923) first pointed out that the Southern Oscillation phenomenon is the dominant feature in global weather. Eigenvector analysis of global sea level pressure substantiated Walker's claim (Kidson 1975). A comprehensive review of the phenomenon has been given by Julian and Chervin (1978). Trenberth (1976) showed that the Southern Oscillation has a period between three and six years. Next to the Southern Oscillation is the Quasi-biennial Oscillation (QBO). This signal is strongest in the tropical lower stratosphere (Newell *et al.* 1974), but a QBO signal with a period of 24–31 months is also evident in extratropical temperatures (Landsberg 1962; Rasmussen *et al.* 1981). Spectral analysis of southern hemisphere ultra-long waves showed a peak at about 26 months but this QBO signal seems to be unrelated to the tropical QBO (Trenberth 1980).

Because the SO and QBO phenomena are not oscillations at a unique frequency, the spectra of the nonseasonal time series have been divided into 4 frequency bands and the percentage of nonseasonal variance contained within each band computed. The frequency bands are: (a) 114–24th harmonics, i.e. sub-annual frequencies; (b) 13–24th harmonics, corresponding to oscillations with periods 12 to 22·1 months; (c) 9–12th harmonics, i.e. periods 24–32 months, corresponding to quasi-biennial frequencies; and (d) 1–8th harmonics, corresponding to periods of 36 months and higher, the SO frequency band. The percentage of nonseasonal variance contained in each frequency band is depicted in Table 5. For a random time series, the percentages of variance contained in these bands are 79, 10·5, 3·5 and 7%, respectively. Because of persistence, there is inflation of variance at low frequencies and deflation at high frequencies. As can be anticipated from the results of the autocorrelation analysis, there are substantial low-frequency variations in the tropical and subtropical Pacific. More than 50% of the nonseasonal variance is contained in frequency band (d) in this region.

TABLE 5. TABLE OF PERCENTAGE OF NONSEASONAL VARIANCE CONTAINED IN VARIOUS FREQUENCY BANDS

Latitude	Indian Ocean				Pacific				Atlantic			
	a	b	c	d	a	b	c	d	a	b	c	d
67·5 °N									61	8	4	27
62·5 °N									44	13	6	37
57·5 °N					30	13	5	50	35	15	8	42
52·5 °N					34	12	11	42	38	16	11	35
47·5 °N					41	11	11	36	44	11	8	36
42·5 °N					38	14	2	46	40	10	9	40
37·5 °N					34	15	4	48	35	5	14	46
32·5 °N					55	12	4	28	42	13	9	36
27·5 °N					43	18	5	34	34	15	9	42
22·5 °N	66	11	3	20	18	7	4	70	31	14	14	42
17·5 °N	50	15	2	33	25	7	5	63	28	9	18	44
12·5 °N	37	15	7	41	26	7	3	64	24	10	21	44
7·5 °N	38	8	13	41	24	9	5	62	37	12	11	40
2·5 °N	49	12	7	31	19	16	8	57	32	18	19	30
2·5 °S	56	15	5	24	19	18	10	53	34	20	16	31
7·5 °S	59	9	6	26	19	15	11	56	32	15	11	42
12·5 °S	56	10	7	26	42	10	4	43	42	17	5	36
17·5 °S	44	15	7	32					45	15	6	34
22·5 °S									37	16	6	41
27·5 °S									41	12	5	42
32·5 °S									61	16	3	20

Nonseasonal variance contained in the frequency band (a) from wavenumber 114–24 (i.e. sub-annual frequencies); (b) wavenumber 13–24 (period 12–22·1 months); (c) wavenumber 9–12 (period 24–32 months) and (d) wavenumber 8 and lower (period 36 months and longer)

The QBO signal showed a maximum in the tropical north Atlantic, explaining more than 20% of the variance at 12·5 °N. But if the harmonic amplitudes of the 12th harmonic (period 24 months) are examined, the largest amplitude (0·28 K) is found in the equatorial Pacific region from 2·5 °S to 7·5 °S.

Nicholls (1979) proposed an air–sea interaction model for the tropospheric QBO in the Indonesian–north Australian regions. It is interesting to note that Hastenrath and Kaczmarczyk (1981) also found a marked preference for frequencies around 2·5 years in equatorial Atlantic sea level pressure. The frequency distribution of nonseasonal SST at 12·5 °N in the Atlantic as found here resembles a bimodal distribution, a feature implicit in Nicholls' model.

(*e*) EOF *analysis and lag correlations*

EOF analyses have been performed separately for the north Pacific by Davis (1976, 1978), for the north and tropical Pacific by Weare *et al.* (1976), for the Atlantic by Weare (1977) and Haworth (1978), for the Indian ocean by Weare (1979) and Navato *et al.* (1981) and for the tropical Atlantic and Pacific by Hastenrath (1978). The main features of the first EOF for each ocean are described in section 1. The general pattern of the first EOF for a conglomerate set of global SST can be anticipated from results of EOF analyses performed on each ocean or oceans. Hastenrath (1978) showed that the equatorial eastern Pacific varies in phase with the equatorial Atlantic in the first EOF pattern. Weare (1979) showed that the time series associated with the first Pacific EOF pattern is positively correlated with the time series of the first Indian Ocean EOF. The first EOF of global SST is thus expected to show the pattern in which SST variations in the eastern equatorial Pacific is negatively correlated with SST variations in mid latitudes in the north Pacific and north Atlantic, and positively correlated with SST variations elsewhere.

An EOF analysis of the global zonal mean SST confirms our synthesis. The first EOF pattern is shown in Table 6 and the associated time series in Fig. 1. The procedure for the EOF analysis follows that which is described in Essenwanger (1976). The first EOF explains 26% of the variance whereas the second and third explain 11 and 8%, respectively. When logarithms of the eigenvalues are plotted against eigenvalue numbers (not shown), the first eigenvalue deviates significantly from a straight line fit of the linear portion of the smaller eigenvalues and is therefore significant according to the test suggested by Craddock and Flood (1969).

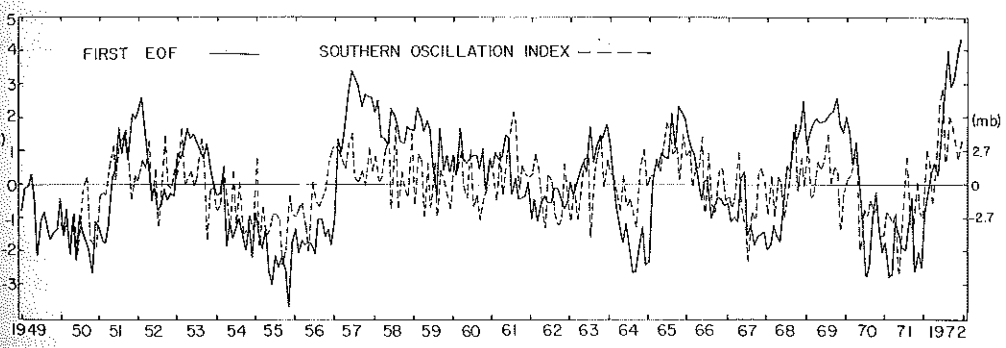

Figure 1. Time series of the first empirical orthogonal function of zonal mean sea surface temperature and an index of the Southern Oscillation. The Southern Oscillation index used here is the nonseasonal pressure difference between Easter and Darwin. The time series of the Southern Oscillation index has been inverted so that the time series can be compared.

The contemporaneous correlations between zonal mean SSTs at different latitudes are given elsewhere (Newell and Chiu 1981). The correlation coefficients between the

TABLE 6. FIRST EIGENVECTOR OF NONSEASONAL ZONAL MEAN SEA SURFACE TEMPERATURE

	Indian Ocean	Pacific	Atlantic
67·5 °N			0·10
62·5 °N			0·01
57·5 °N		0·18	0·01
52·5 °N		0·12	−0·02
47·5 °N		0·02	−0·07
42·5 °N		−0·09	−0·08
37·5 °N		−0·10	−0·06
32·5 °N		−0·04	−0·04
27·5 °N		0·02	0·03
22·5 °N	0·11	0·12	0·06
17·5 °N	0·11	0·12	0·05
12·5 °N	0·11	0·15	0·06
7·5 °N	0·10	0·23	0·05
2·5 °N	0·10	0·36	0·05
2·5 °S	0·10	0·48	0·07
7·5 °S	0·08	0·42	0·06
12·5 °S	0·07	0·33	0·06
17·5 °S	0·08		0·09
22·5 °S			0·13
27·5 °S			0·14
32·5 °S			0·13

zonal mean SSTs and the zonal mean at 2·5 °S in the Pacific for lags up to ±15 months have been calculated here. To test the significance of these correlation coefficients, we followed an approach adopted by Angell (1981) which involved estimating the number of independent samples, n^*, for autocorrelated time series with N observations through the relation $n^* = N/f$ where

$$f = 1 + 2\left\{\sum_{i=1}^{\infty} P_i Q_i\right\}$$

and P_i and Q_i are autocorrelation coefficients at a lag of i months of the variables, i.e. zonal mean SST and zonal mean SST at 2·5 °S in the Pacific. To get a conservative estimate of n^*, which will be adopted for all latitudes, autocorrelation coefficients are calculated using the most persistent time series, namely 22·5 °N in the Pacific (see subsection 4(*c*)), and the summation of the series in the brackets is terminated whenever either P_i or Q_i becomes negative. The computed value for f is 7·5. The corresponding n^* is 38. This procedure of estimating significance is used in subsequent tests of the significance of correlation coefficients. We noted that N is different from 288 for calculating correlation coefficients at lags. At a lag of 15 months, N equals 273. The difference is small (about 6·5%) and is neglected.

Table 7 shows the maximum correlation coefficient between SST variations at 2·5 °S in the Pacific and SST variations elsewhere at various lags. Only coefficients significant at the 95% level are included. The number in parenthesis is the lag in months of the maximum correlation coefficient. The lag is positive if zonal mean SST leads. It is interesting to note that variations at 2·5 °S lead zonal mean SST variations elsewhere. They are contemporary with SST variations in the equatorial Pacific from 7·5 °N to 12·5 °S, and lead variations in the Indian Ocean by 2–4 months and in the equatorial Atlantic by 6–7 months.

TABLE 7. TABLE OF MAXIMUM CORRELATION COEFFICIENTS AT VARIOUS LAGS BETWEEN NONSEASONAL ZONAL MEAN SST AND ZONAL MEAN SST AT 2·5 °S IN THE PACIFIC

Latitude	Indian Ocean	Pacific	Atlantic
22·5 °N	39(−4)		
17·5 °N	44(−2)	34(−4)	40(−6)
12·5 °N	50(−3)	47(−4)	41(−7)
7·5 °N	48(−3)	65(0)	39(−6)
2·5 °N	42(−4)	85(0)	
2·5 °S	36(−3)	100(0)	
7·5 °S		84(0)	
12·5 °S		56(0)	
17·5 °S			
22·5 °S			34(−1)

Correlation coefficients are expressed in hundredths. Only coefficients significant at the 95% level are included. The effective number of independent observations is 38. The values of r at the 95% and 99·5% levels are 33 and 47. The number in parenthesis is the lag in months of the maximum correlation coefficient, and is positive if zonal mean SST leads zonal mean SST at 2·5 °S.

5. SOI–SST RELATION

To examine the interaction between the major modes of SST and sea level pressure variations, an index of the SO is superposed on the time series of the first EOF in Fig. 1. Trenberth (1976) examined the sea level pressure variations in the south Pacific and suggested as an index of the SO the pressure difference between Darwin and some weighted mean of pressure in Tahiti and Easter Island. The SO index (SOI) used here is the nonseasonal pressure difference between Easter Island and Darwin (Quinn and Burt 1972). Figure 2 shows the graph of correlation coefficients at lags between SOI and the

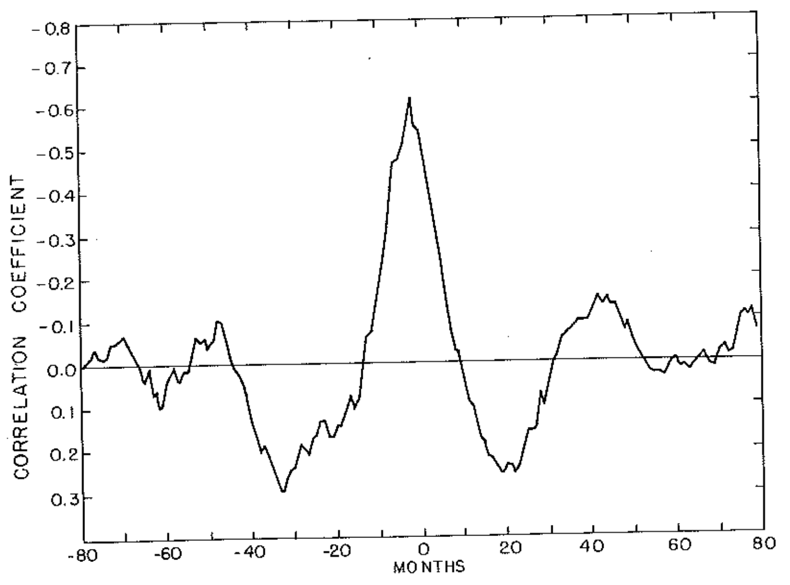

Figure 2. Graph of correlation coefficients at lags (in months) between SOI and the time series of the first EOF. The lags are defined negative if SOI leads.

time series of the first EOF. It can be seen that the peak correlation is fairly well defined. The best correlation between these time series is −0·61, with SOI variations leading by two months. This correlation is significant at the 99·9% level. The effective number of independent samples, n^*, is 190 (see sub-section 4(*e*)).

Correlation coefficients are computed using seasonally stratified data to examine the seasonal dependence of the correlations; the results are shown in Table 8. Only coefficients greater than 0·49, significant at the 95% level, are included. Most of the coefficients appear in the upper tridiagonal, confirming the lead relationship between SOI and zonal SST. A correlation coefficient as large as 0·83 is found between SOI variations in May and SST variations in July.

TABLE 8. CORRELATION COEFFICIENTS BETWEEN SOI AND EOF1 FOR DIFFERENT MONTHS

EOF1 SOI	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Jan		54		58		50						
Feb		55	53	63	55	64			52	63	61	
Mar	52	70	66	56	53							
Apr				57		63			55	61	61	
May			53	51	59	77	83	71	63	73	79	78
Jun						56	57	73	61	66	68	63
Jul						51		52	51			
Aug							55	62	58	59	62	62
Sep						53	67	68	76	72	79	77
Oct											57	
Nov												
Dec						58	58	66	57	59	65	63

All coefficients are negative and are expressed in hundredths.
For 24 years of data, the 95% and 99·9% levels of significance are $r = 49$ and 62, respectively. Only values larger than 49 are included.

It had been argued that the occurrence of warm water off the coast of Peru, often known as the El Niño phenomenon, is due to the slackening of the tropical easterlies associated with a decrease of east–west pressure gradient, hence causing less upwelling along the coastal regions and warming of the sea surface by surface radiative, sensible and latent heat fluxes (e.g. Bjerknes 1969). Hickey (1975) reported that minima in zonal easterlies in the central Pacific are associated with El Niño. Wyrtki (1975) proposed a different explanation for the occurrence of warm water. It entailed a gradual rise in sea level in the western tropical Pacific under the influence of stronger wind in the central Pacific followed by a relaxation of the wind and a suppression of upwelling in the east due to dynamic effects. The exact cause of this warming has not been established (Barnett 1977) but is no doubt related to the changing wind pattern.

The first eigenvector of sea level pressure from Kidson's (1975) analysis also showed a tendency of the pressures at high northern latitudes to be opposed to those in the Indonesian–north Australian region. When the SOI is negative (lower pressure over the southeast Pacific), there is lower pressure in high latitudes. Associated with an increase of meridional pressure gradient, there are stronger westerlies at mid-latitudes (Van Loon and Madden 1981, Van Loon and Rogers 1981). The stronger westerlies at mid-latitudes may enhance evaporative cooling and mixing, thus giving rise to lower SST there.

Markham and McLain (1977) and Markham (1979) found that increased February–April California rainfall is associated with negative SST anomalies in the central Pacific and positive anomalies off the coast of North America in the previous winter. Li *et al.* (1979) found that monsoon rainfall in China (June–July rain) is preceded by positive SST anomalies south of 35 °N off the coast of Japan and negative anomalies north of 35 °N in the preceding winter. It is interesting to note that the pattern of correlations between SST and California rainfall and Chinese monsoon rainfall can be interpreted in terms of increased westerlies: stronger westerlies are associated with an intensification of the oceanic gyres, providing warm (cold) advection in the Kuroshio (Oyashio) region, more evaporative cooling and more mixing in the central Pacific.

The contemporaneous correlation between SST and SOI was computed by Wright (1977) who found large contemporaneous negative correlations between SOI and SST in the equatorial eastern Pacific and Indian Ocean. To examine the lead and lag relationship, the correlation coefficients between SOI and zonal mean SSTs have been computed for lags up to ±15 months here. The results are summarized in Table 9. It can be seen that for all latitudes the correlation coefficient is less than 0·61, which is the correlation between SOI and the time series of the first EOF. It is this pattern of SST variations that best correlates with the SOI. The SOI correlates negatively with SST in all latitudes and leads SST variations in the equatorial Pacific by about 1–2 months, leads SST variations in the Indian Ocean by 5–6 months and those in the tropical Atlantic by 9–10 months.

TABLE 9. AS TABLE 7 EXCEPT FOR THE SOUTHERN OSCILLATION INDEX

Latitude	Indian Ocean	Pacific	Atlantic
22·5 °N		−27(−2)	
17·5 °N	−28(−4)	−32(−2)	−26(−9)
12·5 °N	−40(−5)	−42(−4)	−27(−10)
7·5 °N	−38(−5)	−49(−3)	−25(−9)
2·5 °N	−34(−5)	−54(−2)	
2·5 °S	−29(−5)	−52(−2)	
7·5 °S	−27(−7)	−55(−2)	
12·5 °S		−47(−2)	
17·5 °S			−35(−2)
22·5 °S			−41(−2)
27·5 °S			−34(−2)
32·5 °S			−27(0)

As Table 7 except the effective number of independent samples is 62. The values of r at the 95% and 99·5% levels are 25 and 37.

From a hundred years (1861–1960) of Pacific seasonal SST data, Newell *et al.* (1981) found that the SOI is positively correlated with SST variations in the western north Pacific (160 °W to 160 °E) 3 seasons ahead and negatively correlated with SST variations along the coast of North and South America 1–2 seasons later. We found no significant correlation between SOI and zonal mean SST north of 25 °N in the Pacific. South of 25 °N, zonal means of SST which only include the eastern Pacific (see section 2) lag behind SOI by 2–4 months.

We noted that there is maximum contemporary correlation between SOI and SST at 32·5 °S in the Atlantic. This may suggest an antarctic origin of the Southern Oscillation first proposed by Walker (1923). Chiu (1981) showed that recent variations in antarctic sea ice area are related to the SOI. The correlation found here, −0·27, is, however, only marginally significant by our test.

The role of Indian Ocean SST on the Indian monsoon has been debated. Shukla and Misra (1977) found that increased monsoon rainfall is associated with positive SST anomalies in an area to the south-west of the Indian continent. Contrary to this, Weare (1979) found that the time series of the first EOF of Indian Ocean SST, which has positive weights everywhere in the northern Indian Ocean, is negatively correlated with monsoon rainfall over India. Pant and Parthasarathy (1981) found that Indian monsoon rainfall is positively correlated with an index of the SO. Our study showed that there is high negative correlation between SOI and SST in the Indian Ocean, with correlation coefficients in latitudes 12·5 °N and 7·5 °N exceeding the 99·9% level of confidence. Our result is thus consistent with those of Weare and of Pant and Parthasarathy. When the SOI is negative there is higher SST in the northern Indian Ocean, higher sea level pressure over the Arabian Sea, higher SST in the eastern equatorial Pacific and less monsoon rainfall in India. The decrease in monsoon rainfall may be a consequence of decreased convergence over India. The higher sea level pressure is presumably associated with enhanced subsidence, less cloudiness and therefore higher SST in the Arabian Sea region (Newell 1979; Weare 1979).

6. Relation with tropical tropospheric temperature (TTT)

In section 1, the question of why TTT lags behind the time series of the first EOF of the Pacific SST by one season and the tropical eastern Pacific zonal mean SST by two seasons is raised. From the results in sub-section 4(*e*), it is seen that variations in 2·5 °S in the Pacific are precursors of SST changes elsewhere. One therefore expects that the actual heating of the troposphere takes place in regions other than the equatorial Pacific. Correlation analyses between TTT and zonal means of SST were performed at lags up to ±15 months. The maximum correlation coefficients are given in Table 10. It can be seen that SST variations at 2·5 °S and 7·5 °S lead TTT by 5–6 months. This agrees with the results of Newell (1979) and Angell (1981). The best correlation appears north of the equator in the Pacific, with correlation coefficients exceeding 0·6 at 7·5 °N–17·5 °N at a lead time of about two months. In the Atlantic and Indian Ocean, the maximum correlation appears at a lead or lag of about one month. This almost contemporary correlation is suggestive of the simple heating mechanism of the troposphere by increased SST through air–sea surface flux exchange, with a time scale of less than one month.

TABLE 10. As Table 7 except for tropical tropospheric temperature (TTT)

Latitude	Indian Ocean	Pacific	Atlantic
22·5 °N		55(3)	51(−1)
17·5 °N		60(3)	54(−1)
12·5 °N	51(1)	64(2)	60(−1)
7·5 °N	55(1)	64(2)	52(−1)
2·5 N	52(1)	49(2)	
2·5 °S	55(1)	47(5)	
7·5 °S		48(6)	
12·5 °S			

As Table 7 except the effective number of independent samples is 24. The period used for this calculation is from January 1958 to December 1972. The values of *r* at the 95% and 99·5% levels are 43 and 57.

Using the time series of the first nonseasonal EOF of the Atlantic and Pacific SST, Newell and Weare (1976a) found that the Atlantic does not contribute significantly to the variance of TTT. However, Navato *et al.* (1981) showed that the Atlantic does contribute if the time series of the normalized EOF is used. The difference between using normalized and unnormalized data is discussed in Navato *et al.* Inspection of the first EOF pattern of the Atlantic (Figures 4 and 5 in Navato *et al.*) showed that the unnormalized pattern is dominated by variations at mid-latitudes (with a maximum around 40 °N), whereas the normalized first EOF is dominated, to a much lesser extent, by variations in the tropics, with a maximum located around 20 °N, 70 °W. From Table 10, it can be seen that there is high correlation between TTT and zonal mean SST in the Atlantic from 7·5 °N to 22·5 °N. Since the first EOF of the unnormalized Atlantic SST is characterized by high and middle latitude SST changes, its contribution to TTT is thus expected to be minimal.

Correlation coefficients between SOI and TTT were also calculated for lags up to 15 months. The maximum coefficient is −0·42, with SOI leading by eight months. This is consistent with results in previous sections and the results of Navato *et al.*

7. Summary and discussion

Using 24 years of zonal mean SST data, it is shown that the annual and semi-annual variations account for more than 90% of the variance outside the tropics. In the tropics, there is high persistence and substantial low-frequency variations. The low-frequency variations are examined with respect to large-scale features of the atmosphere, the Southern Oscillation and the Quasi-Biennial Oscillation. Spectra of nonseasonal variations of zonal mean SST are divided into frequency bands representative of the SO and QBO. The SO signal is strongest in the equatorial Pacific, with variations of period three years and longer accounting for 70% of the nonseasonal variance. The QBO signal, with period 24–32 months, is strong in the equatorial Atlantic, accounting for about 20% of the variance at 0–5 °N and 10–20 °N. The SST variations at 10–15 °N also showed a bimodal distribution.

From an EOF analysis, it is found that the major pattern of zonal mean SST variations is dominated by variations in the equatorial Pacific, with opposite changes at mid-latitudes in the north Pacific and Atlantic and changes of the same sign elsewhere. Variations in the equatorial Pacific are precursors of SST changes elsewhere, leading changes in the Indian Ocean by 2–4 months and changes in the tropical Atlantic by 6–7 months.

This pattern of SST variations is closely related to the SO. The SOI leads zonal mean SST changes in the tropical Pacific by 1–4 months, the Indian Ocean by 4–7 months and the northern tropical Atlantic by 9–10 months. In the south Atlantic from 15–35 °S, the SOI leads SST changes by 0–2 months, hence is suggestive of a potential source region of the SO.

Earlier studies (Newell and Weare 1976a; Newell 1979; Angell 1981) showed that SST variations in the equatorial eastern Pacific preceded changes in the tropical tropospheric temperatures by 1–2 seasons. From a correlation analysis between SST and tropical tropospheric temperature, it is found that there is maximum contemporary correlation between temperature and SST in the Indian Ocean and Atlantic. Newell *et al.* (1981) found that seasonal SOI variations are correlated contemporaneously with SST variations in the eastern equatorial Pacific but lead SST variations in the equatorial central Pacific by about one season. While this may suggest a propagation of SST anomalies westward, it also points to another possible source region of tropospheric heating, i.e. the central equatorial Pacific, which is not included in our data set. This region also dominates in the first EOF pattern of nonseasonal rainfall (Kidson 1975). Since precipitation is an indicator of the amount of latent heating in the troposphere (Newell *et al.* 1974), we are led to speculate that in addition to the Indian Ocean and the tropical Atlantic, the central Pacific (rather than the eastern equatorial Pacific) contributes to the variations in the heating of the troposphere.

The question of whether the ocean forces the atmosphere or vice versa has often been raised. In this study, we showed that on a global scale, there is evidence that the major pattern of sea level pressure change, the SO, precedes changes in the major mode of SST variations. The zonal mean SST anomalies in turn affect atmospheric temperature. This question may perhaps be better answered within the context of specified spatial and temporal scales defined with reference to air–sea interaction mechanisms.

APPENDIX

Let a_{ij} be the zonal mean SST of a particular latitude belt at the ith year and jth month, $i = 1, \ldots, N$; $j = 1, \ldots, 12$, where N is the number of years. The long-term monthly means (d_j) and standard deviations (S_j) are defined as

$$d_j = \frac{1}{N} \sum_{i=1}^{N} a_{ij}; \qquad S_j = \frac{1}{N} \left\{ \sum_{i=1}^{N} (a_{ij} - d_j)^2 \right\}^{1/2}, \qquad j = 1, \ldots, 12.$$

Let the normalized zonal mean SST be defined as $a'_{ij} = (a_{ij} - d_j)N^{-1/2}/S_j$ so that

$$\sum_{i=1}^{N} (a'_{ij})^2 = 1.$$

To compute the correlation coefficients between two variables using seasonally stratified data, let b'_{ij} be the normalized matrix of variable b. The correlation coefficients (C_{kl}) between variable b at month k and variable a at month l is given by the matrix multiplication

$$C_{kl} = \sum_{i=1}^{N} b'_{ik} a'_{il}.$$

ACKNOWLEDGMENTS

Thanks are due to Dr A. Navato for valuable comments and discussions. Sea surface temperature data for the Atlantic and Indian Ocean were kindly supplied by the British Meteorological Office. This work is supported by the Climate Dynamic branch of the U.S. National Science Foundation under Grant ATM-7918016 and by the U.S. Department of Energy under contract number DE-AC02-76EV12195.

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