El Niño and La Niña

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ABSTRACT

El Niño and La Niña are the two complementary phases of the Southern Oscillation. During El Niño, the area of high sea surface temperatures increases, while the atmospheric convection zones of the tropical Pacific expand and merge so that there is a tendency toward spatially homogeneous conditions. La Niña is associated with low sea surface temperatures near the equator, with atmospheric convergence zones that are isolated from each other, and with spatial scales smaller than those of El Niño. It is proposed that both phases of the Southern Oscillation can be attributed to unstable interactions between the tropical ocean and atmosphere. During El Niño, the increased release of latent heat to the atmosphere drives the instability. During La Niña, when the heating of the atmosphere decreases, the compression of the convection into smaller and smaller areas permits an instability that intensifies the trade winds and the oceanic currents. The unstable air–sea interactions are modulated by the seasonal movements of the atmospheric convergence zones, and this determines some of the characteristics of the perturbations that can be amplified. The zonal integral of winds along the equator, rather than winds over a relatively small part of the Pacific such as the region west of the date line, is identified as a useful indicator of subsequent developments in the Pacific.

1. Introduction

Periods during which sea surface temperatures in the central and eastern tropical Pacific Ocean are warmer than normal are referred to as El Niño. Such periods are, of course, separated by intervals during which sea surface temperatures are colder than normal. The term La Niña is apposite for these cold periods because oceanographic and meteorological conditions during El Niño and La Niña are highly complementary. Figure 1, for example, shows how a number of low-pass filtered oceanographic and meteorological variables fluctuate coherently and in phase with the sea surface temperature in the central equatorial Pacific. This interannual fluctuation between complementary El Niño and La Niña conditions, known as the Southern Oscillation, is very irregular. For example, the period between warm El Niño episodes varies enormously, as does the amplitude of an episode: 1963 was a modest episode; 1972 a substantial one. "Normal conditions" can be defined in terms of appropriate averages over a specified period but they depend critically on the chosen period. Figure 1 shows that the tropical Pacific is very seldom in a "normal state," however it is defined, so that El Niño should not be considered as the departure from a "normal state." El Niño is the complement of La Niña, and the two constitute the Southern Oscillation.

The coherence of the fluctuations in the meteorological parameters shown in Fig. 1 can readily be explained by assuming that large-scale motion in the tropical atmosphere corresponds primarily to a direct thermal circulation. In the Pacific Ocean the important regions of heating are the Intertropical Convergence Zone (ITCZ), the South Pacific Convergence Zone (SPCZ) and especially the convective zone over the maritime continent of the western tropical Pacific Ocean. The air that rises in these convective regions, where surface pressure is low, descends in the subtropics and in the high pressure zone of the southeastern tropical Pacific Ocean. The trade winds return the air to the convective regions thus closing the circulation in the lower atmosphere. During El Niño the ITCZ and SPCZ are displaced equatorward, and the large convective zone over the western tropical Pacific Ocean is displaced eastward (Ramage and Hori, 1981; Rasmusson and Carpenter, 1981; Pazan and Meyers, 1982). Because of these movements, rainfall increases in the central but decreases in the western tropical Pacific, surface pressure increases in the west and falls in the east, while the intensity of the trade winds relaxes. During La Niña, when the ITCZ and SPCZ are displaced poleward and the convergence zone in the western tropical Pacific is displaced westward, the trade winds intensify, surface pressure increases over the eastern tropical Pacific but decreases over the western tropical Pacific, while rainfall in the central tropical Pacific decreases. Because of these changes, shown in Fig. 1, zonal gradients are minimized during El Niño when there is a tendency toward spatially homogeneous conditions as convective zones merge. During La Niña, spatial gradients are at a maximum and the various convective regions are separated from each other.

A composite El Niño starts as a warm sea surface

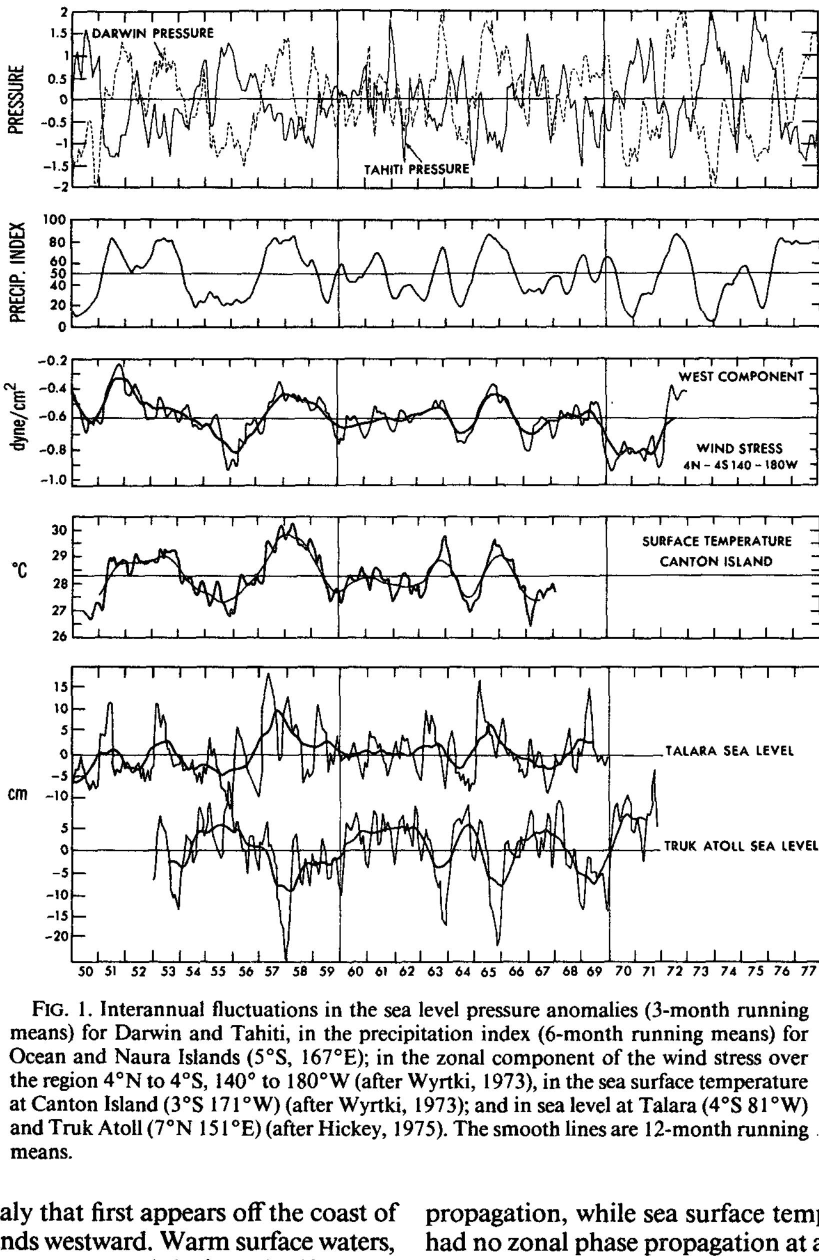

FIG. 1. Interannual fluctuations in the sea level pressure anomalies (3-month running means) for Darwin and Tahiti, in the precipitation index (6-month running means) for Ocean and Naura Islands (5°S, 167°E); in the zonal component of the wind stress over the region 4°N to 4°S, 140° to 180°W (after Wyrtki, 1973), in the sea surface temperature at Canton Island (3°S 171°W) (after Wyrtki, 1973); and in sea level at Talara (4°S 81°W) and Truk Atoll (7°N 151°E) (after Hickey, 1975). The smooth lines are 12-month running means.

temperature anomaly that first appears off the coast of Peru and then expands westward. Warm surface waters, however, do not move westward during El Niño. The westward phase propagation of anomalies is attributable to the reference state, the climatology, which has westward phase propagation of the cold phase of the seasonal cycle (Horel, 1982). During a composite El Niño there is initially an amplification of the warm phase of the seasonal cycle in the east and subsequently a persistence of the high sea surface temperature. The cold season fails to develop fully, and since this low sea surface temperature would have migrated westward, the warm anomalies appear to propagate westward when high sea surface temperatures persist. In Fig. 2 the cold season is seen to fail in 1957, 1963, 1965, 1972 and 1976. El Niño of 1982–83 was different because isotherms on the ocean surface showed eastward phase propagation, while sea surface temperature anomalies had no zonal phase propagation at all (Rasmusson and Wallace, 1983; Gill and Rasmusson, 1983). It thus appears that during El Niño, sea surface temperatures (as opposed to anomalies) have either eastward or no zonal phase propagation. During La Niña, on the other hand, cold surface waters first appear in the east and then expand westward (Rasmusson and Carpenter, 1982). Further analysis of data are necessary to determine what other fields show westward phase propagation and to find out whether Rossby waves are involved. Analysis of the seasonal cycle will be valuable to determine whether the warm phase is associated with eastward (the cold phase with westward) phase propagation. Note that Fourier analysis is inappropriate to determine such an asymmetry.

There is indirect evidence that the Southern Oscil-

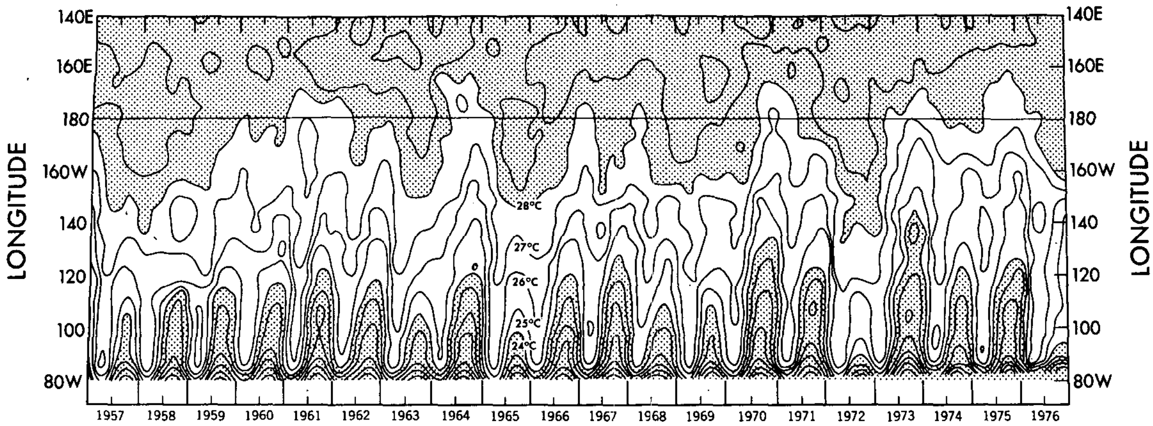

FIG. 2. Sea surface temperature variations along the equator between 100°E and 100°W, and along a line from 0°N 100°W to the South American coast at 10°S. Regions with temperatures less than 24°C and greater than 27°C are shaded. The contour interval is 1°C.

lation involves not only changes in the positions of the convective zones, but also fluctuations in the magnitude of the heating of the tropical atmosphere. Pan and Oort (1983) find that interannual variations in the average temperature of the Northern Hemisphere atmosphere are highly correlated with sea surface temperature in the central equatorial Pacific Ocean. Horel and Wallace (1981) estimate that the interannual variations in the mean temperature of the tropical troposphere are of the order of 1 K. The most probable source of heat for the atmosphere is latent heat from the ocean. (Gill's, 1983, analysis of data for El Niño of 1972 indicates a loss of heat by the ocean to the atmosphere at that time.) Large-scale atmospheric convection over the oceans occurs over the regions of highest sea surface temperature. Figure 3 shows how

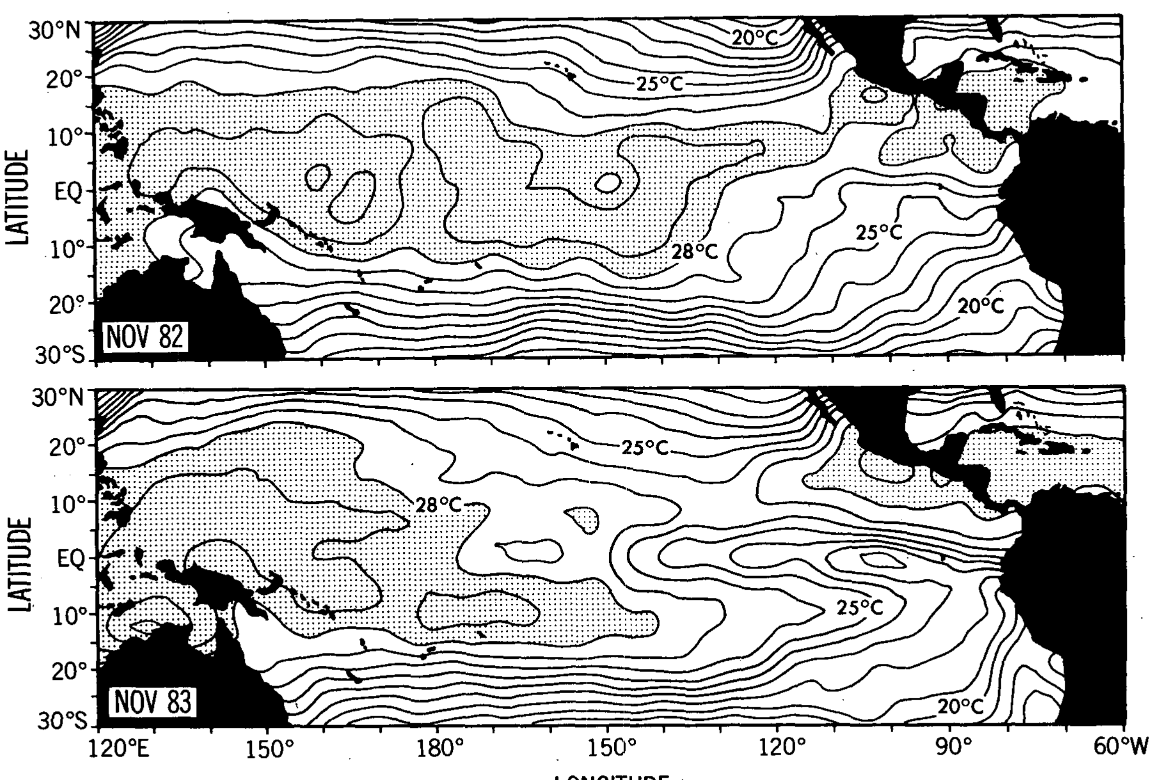

FIG. 3. Sea surface temperatures in November 1982 during El Niño and in November 1983 during La Niña.

the area of the pool of surface waters with temperatures in excess of 28°C can change between El Niño (November 1982 in this case) and La Niña (November 1983 in this case). Confirmation that the changes in the sea surface temperature of the tropical Pacific Ocean result in changes in the heating of the atmosphere and hence in the Southern Oscillation come from experiments with a general circulation model of the atmosphere (Lau, 1985; Manabe and Hahn, 1981; Lau and Oort, 1985). Variability of the atmosphere was simulated for a 15-year period in two experiments. In the first experiment climatological (monthly mean) sea surface temperature variations were specified. The statistics of variability in midlatitudes were realistic, but there was practically no variability in the tropics other than the seasonal cycle. In the second experiment, climatological sea surface temperatures were again specified except in the tropical Pacific Ocean where the observed sea surface temperatures for the period 1962–76 were specified. This second calculation simulated the observed Southern Oscillation during that period remarkably well. The principal conclusion from this calculation is that changes in the heating of the atmosphere cause the Southern Oscillation. As mentioned previously, there is evidence that the heat gained from the ocean by the atmosphere increases during El Niño. There has as yet been no attempt to determine whether this heat gain decreases during La Niña.

An explanation for the sea surface temperature (and other oceanic) changes associated with the Southern Oscillation follows readily from the assumption that they are part of the response of the ocean to the variations in the surface winds. During La Niña, sea surface temperatures are below normal because the exceptionally intense trade winds drive the warm waters of the upper ocean westward and expose cold water to the surface in the central and eastern tropical Pacific. Sea level, which measures the heat content of a water column, is therefore unusually high in the west and low in the east (Fig. 1). The westward transfer of warm surface waters is effected by an intensification of the South Equatorial Current, to the south of 3°N, and a weakening of the eastward North Equatorial Countercurrent, approximately between 3 and 10°N (Wyrtki, 1973). During El Niño, complementary conditions result in the trade winds relaxing, warm surface waters returning eastward to the central and eastern Pacific because of an intensification of the eastward Countercurrent and a weakening, sometimes a reversal, of the westward South Equatorial Current. Confirmation that the oceanic changes associated with the Southern Oscillation are primarily in response to fluctuations in the surface winds come from the success with which models of the ocean simulate the observed changes when forced with the observed winds. Busalacchi and O'Brien (1981) have used a shallow water model to reproduce observed sea level changes in the tropical Pacific Ocean during the 1960s. Philander and Seigel (1985) simulated oceanic changes (including those in sea surface temperature) during the 1982–83 El Niño with a multilevel general circulation model of the ocean.

From a meteorological point of view, the interannual variations in the tropical atmosphere are caused by changes in the sea surface temperature. From an oceanographic point of view, the SST changes are in turn caused by the atmospheric variability. The circularity of these arguments implies that unstable interactions between the ocean and atmosphere can cause modest perturbations to amplify (Philander, 1983). Consider, for example, a modest initial relaxation of the trade winds that results in a modest increase in sea surface temperatures in the central and eastern equatorial Pacific Ocean. If these changes succeed in moving the convective zone of the western tropical Pacific eastward then there is a further relaxation of the trades that leads to a further warming of the central equatorial Pacific, a further displacement of the convective zone, and so on. A simple model in which an increase in thermocline depth causes an increase in sea surface temperature, which in turn causes increased local heating of the atmosphere, shows that this instability occurs when convergent winds in the atmosphere drive convergent oceanic currents. This is possible in a nonrotating system, a case to which Lau's (1981) calculations are relevant. (Lau assumed that meridional motion is negligible in both the ocean and atmosphere. This assumption, that zonal winds do not cause any equatorial upwelling in the ocean, is indefensible on an equatorial β-plane but is justifiable in a nonrotating system.) If the rotation of the earth is taken into account, but the Coriolis parameter is assumed to be a constant, then these air–sea interactions are stable because convergent winds in the atmosphere drive divergent Ekman flow in the ocean (Philander *et al.*, 1984). On a rotating sphere the interactions are unstable near the equator where the Coriolis parameter vanishes, but the difference between east and west now becomes important. Consider a pool of unusually warm surface waters, on the equator, that heats the atmosphere locally. The low-level winds will converge on this pool. To the west, the westerly winds drive convergent oceanic currents so that the thermocline deepens and the pool expands. To the east, matters are more complex because two competing mechanisms affect the depth of the thermocline (and hence the SST) in that region. One is equatorial upwelling induced by the local easterlies that converge on the heated region. This reduces the SST. The other is an eastward travelling Kelvin wave, excited by the westerly winds to the west, which deepens the thermocline and increases the SST. Whether this warming caused by the Kelvin wave dominates over the local upwelling depends not only on the relative intensities of the easterlies and westerlies but also on their zonal extent so that the zonal integral of the winds is the critical factor (McCreary, 1976). In a case where the winds are symmetrical about the heat source, warm conditions amplify to the west while cold conditions amplify to the east (Philander *et al.*, 1984). On an equatorial β-plane the westerly winds are more intense (and have a larger zonal integral) than the easterly winds so that the warming caused by the Kelvin wave dominates over the local upwelling to the east of the heat source. The warm surface waters and the associated atmospheric convection therefore expand eastward, as during the El Niño event of 1982–83. If the initial perturbation that evolves into El Niño is along the eastern boundary of the ocean basin, then the model shows that there is no zonal phase propagation associated with the amplification of El Niño conditions.

This explanation for the development of El Niño conditions can, to a limited degree, also explain the development of La Niña conditions. If there is convection over a large part of the tropical Pacific Ocean, as is the case towards the end of El Niño, for example, and if a perturbation diminishes this convection in a certain region, then the instability described above can work in reverse to amplify the perturbation. This mechanism can not operate very long, however, because negative convection is impossible. Furthermore, it can explain the eastward expansion of warm conditions but not the westward expansion of cold conditions. Another problem is the intensification of the tradewinds during La Niña. Is this possible even when the heating of the atmosphere decreases?

This paper explores the possibility that El Niño conditions, when spatial gradients are at a minimum and when there is convection over much of the anomalously warm tropical Pacific Ocean, are unstable to certain perturbations that can amplify and thus cause La Niña. The proposed instability depends critically on the difference between the easterly and westerly winds that converge on an equatorial heat source. It is necessary for the eastward expansion of warm conditions, in the model of El Niño discussed earlier, that the effects of the westerly winds to the west of a heat source prevail over the effects of the easterly winds to the east of the heat source. When El Niño reaches its mature phase, westerly wind anomalies cover most of the western and central tropical Pacific Ocean. Suppose that at this stage an external factor reduces the intensity of the westerly winds. (This factor could be a new heat source in the far western Pacific.) The reduction in the intensity of the westerly winds over the central Pacific will cause a change in sea surface temperature in the eastern Pacific because upwelling caused by the local easterly winds in the east can now be dominant over downwelling associated with the remotely excited Kelvin wave. The lower sea surface temperatures in the east reduce the area of the convective zone. If this reduction in the scale of the convective regions results in more intense easterlies and, hence, in more upwelling, then cold conditions will expand westward. To demonstrate that a decrease in the spatial scale L of a heat source Q can cause more intense winds, even though the heat released to the atmosphere decreases, consider the response of a shallow water model of the atmosphere to the heating.

$$Q = \begin{cases} Q_0 \cos\left(\dfrac{\pi x}{2L}\right) \exp(-y^2/4\lambda^2), & |x| < L \\ 0, & |x| > L. \end{cases} \quad (1)$$

The zonal winds u on the equator ($y = 0$), to the east ($x > L$) of the heated region is (Gill, 1980)

$$U = -\frac{Q_0 L}{a^2c^2 + \epsilon^2 L^2}(1 + e^{(-2\epsilon L)/c})e^{(\epsilon/c)/(L-x)}, \quad (2)$$

where c is the long gravity or Kelvin wave speed, λ the radius of deformation, a a constant and ϵ is both the coefficient of Rayleigh damping and Newton cooling. From Eq. (2) it is clear that as L decreases from a value much larger than L^*, the easterly winds intensify. [It can be shown that $L^* = O(c/\epsilon)$. This is the distance an atmospheric Kelvin wave travels in the dissipative time $1/\epsilon$ and is a few thousand kilometers.] In an initial value problem where the heat source appears suddenly at time $t = 0$, the Kelvin wave introduces atmospheric motion to the east of the heat source $x > L$. Since the Kelvin wave dissipates significantly in a distance L^*, motion to the east of $x = L$ is primarily in response to the heating in the region $(L - L^*) < x < L$. As the value of L decreases and approaches L^*, the region of maximum heating near $x = 0$ affects the winds east of $x = L$ more and more. It is for this reason that a decrease in the dimensions L of the heat source Q in Eq. (1) intensifies the winds to the east of the heat source. The more intense winds cause more upwelling, lower sea surface temperatures and a further reduction in the area of the convective zone. The instability that results can therefore cause the westward expansion of La Niña conditions.

The instability described above can be initiated by the appearance of easterly winds over the far western tropical Pacific Ocean. This can lead to the termination of El Niño and the onset of La Niña. The easterly winds appear in the early calendar months of the year following the onset of El Niño, and they appear before sea surface temperatures in the tropical Pacific start to fall (Rasmusson and Carpenter, 1982). The appearance of easterly winds over the western Pacific is therefore not attributable to changes in SST but must be related to the other factors that affect the movements of the atmospheric convergence zones. Changes in the heating of northern Australia and the eastern Indian Ocean are possible factors. This paper, however, is not concerned with the reason for the appearance of easterly winds over the western tropical Pacific towards the end of El Niño, but is concerned with the consequences of these winds for developments over the central and eastern Pacific. We shall demonstrate how cold La Niña conditions can expand westward after a mature El Niño is suitably perturbed. The model for the air–sea interactions is described in section 2. Results are presented in section 3 and discussed in section 4.

2. The model

a. The atmosphere

Steady atmospheric motion in response to a heat source Q is described by the equations

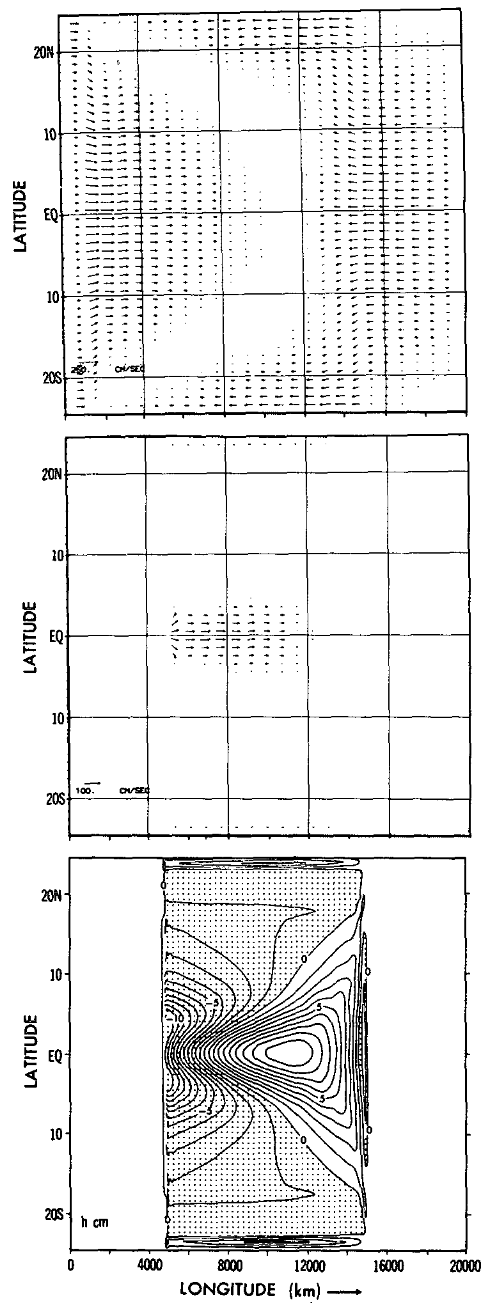

FIG. 4. (a) The low level winds in response to the heat source of Eq. (4); (b) the oceanic currents and (c) departure of the thermocline (equivalent) depth from a depth of 20 cm in response to the steady winds in Fig. 4a.

$$-fV + gH_x = -AU, \tag{3a}$$

$$fU + gH_y = -AV, \tag{3b}$$

$$D(U_x + V_y) = -BH - Q. \tag{3c}$$

The coordinates (x, y) measure distance in an eastward and northward direction; the corresponding velocity components are U and V. The one-layer atmosphere has an equivalent depth D. Perturbations to this depth are measured by H. The coefficient for Rayleigh friction is A and the coefficient for Newtonian cooling is B. The Coriolis parameter $f = \beta y$ vanishes at the equator ($y = 0$) and $C = (gD)^{1/2}$ is the long gravity wave speed.

Figure 4 shows the atmospheric response to a top-hat heat source

$$\begin{aligned} Q &= Q_0 \quad 2500 \text{ km} < x < L \\ &= 0 \quad x < 2500 \text{ km} \quad \text{and} \quad x > L \end{aligned} \tag{4}$$

where L = 15 000 km and Q_0 = 10 cm s^{-1}. The longitudinal extent of the model atmosphere is $0 < x < 20\,000$ km. Its meridional extent is $|y| < 2500$ km. Highly diffusive layers, 1000 km wide, along $x = 0$ and $x = 20\,000$ km damp the motion to such an extent that the atmosphere effectively is longitudinally unbounded. (This was established in experiments that varied the width of the diffusive layers where the values of A and B are increased by a factor of 10.) The results in Fig. 5 are for the case C = 66 m s^{-1} so that the radius of deformation is 2000 km approximately. Reduced gravity has the value 40 cm s^{-2} and the dissipation rates are

$$A = 1/5 \text{ days}, \quad B = 1/15 \text{ days}. \tag{5}$$

Philander *et al.* (1984) describe the manner in which Eqs. (3) are solved numerically and discuss the values assigned to the parameters.

b. The ocean

Oceanic motion in response to a body force (τ^x, τ^y) is described by the equations

$$u_t - fv + gh_x = -au + \tau^x, \tag{6a}$$

$$v_t + fu + gh_y = -av + \tau^y, \tag{6b}$$

$$h_t + d(u_x + v_y) = -bh. \tag{6c}$$

The zonal u and meridional v velocity components in an ocean with an equivalent depth d are associated with depth perturbations h. Motion is damped by Rayleigh friction (coefficient a) and Newtonian cooling (coefficient b). Reduced gravity has the value 2 cm s^{-2}, the equivalent depth d = 20 cm, a = 1/100 days, b = 1/100 days and the radius of deformation is 250 km. The model ocean extends from 5000 km < x < 15 000 km as shown in Fig. 4. Highly diffusive layers 600 km wide along the northern and southern coasts dissipate coastal Kelvin waves. The equilibrium oceanic currents and the associated thermocline displacements in re-

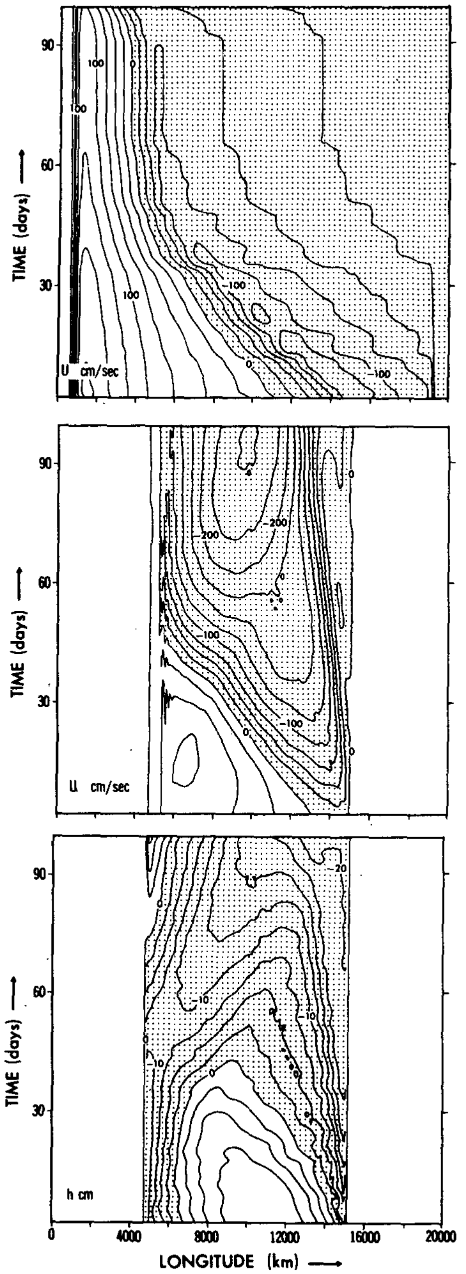

FIG. 5. Evolution of La Niña conditions along the equator: (a) surface winds; (b) zonal currents and (c) thermocline depth departures from a depth of 20 cm. Shaded regions indicate westward motion or an elevation of the thermocline.

sponse to the winds of Fig. 4a are shown in Fig. 4b, c. It is assumed that

$$(\tau^x, \tau^y) = \gamma(U, V), \tag{7}$$

where $\gamma = 10^{-6}$ s^{-1}. For a discussion of the values assigned to the various parameters see Philander *et al.* (1984).

c. The perturbation

The equilibrium conditions in Fig. 4, which will persist indefinitely provided the heat source in Eq. (4) is maintained, are not unlike oceanic and atmospheric conditions in the tropical Pacific Ocean towards the end of El Niño. The westerly winds over the western and central part of the ocean basin drive eastward currents, elevate the thermocline in the west, and depress it in the central Pacific. At this stage in the development of El Niño, which is usually in January, easterly winds appear west of the date line (Rasmusson and Carpenter, 1982), presumably because of seasonal movements of the convergence zone in the western Pacific towards Australia and the southeastern Indian Ocean. In the model, this can be simulated by introducing a heat source to the west of the ocean,

$$Q = \begin{cases} Q_0/10, & 0 < x < 5000 \text{ km} \\ 0, & x > 5000 \text{ km}. \end{cases} \tag{8}$$

d. Coupling between the ocean and atmosphere

The easterly wind anomalies that converge on the new heat source will elevate the oceanic thermocline in the eastern side of the ocean basin, lower sea surface temperatures there and reduce the areal extent of the original heat source in Eq. (4). The effect of the wind on the ocean is calculated from (7). The effect of the thermocline changes on the heating function is taken into account by making the eastern boundary ($x = L$) of the heated region dependent on the depth h_e of the thermocline at the eastern coast (x = 15 000 km) of the ocean basin:

$$L = \left(1 + \frac{h_e - \bar{h}_e}{\alpha \bar{h}_e}\right) 15\,000 \text{ km}, \tag{9}$$

where $\alpha = 1/25$.

Here $\bar{h}_e$ is the original depth of the thermocline in the east. As the thermocline in the east rises (as h_e decreases), the value of L decreases and the net heating of the atmosphere decreases. Formulations other than that in (9) were tried and were found to give results quantitatively different but qualitatively similar to those described next. (In some experiments the value of α was changed; in others the heating of the atmosphere was set equal to zero wherever the depth of the thermocline was less than a specified value.)

These models for the ocean and atmosphere, though simple, are extremely crude so that the interactions between the ocean and atmosphere are also parameterized in a crude manner. The sea surface temperature, the only oceanic parameter that affects the atmosphere, is related to the thermocline depth in a complex manner. In the study of how El Niño develops (Philander *et al.*, 1984) we simply assumed that thermocline depth is proportional to the sea surface temperature and to the heat released to the atmosphere. Here we make a different assumption: an increase in thermocline depth does not affect the heat released to the atmosphere, but a decrease in this depth in the eastern equatorial Pacific decreases the areal extent of the region in which the atmosphere is heated. The two different parameterizations for El Niño and La Niña are both justified, but in different regions and under different conditions. In the eastern equatorial Pacific where the thermocline is usually shallow, an increase in its depth is correlated with an increase in sea surface temperature. However, in regions where the SST is high and the thermocline is relatively deep, most of the time in the western and the eastern tropical Pacific during El Niño, further increases in the depth of the thermocline barely affect the SST. The interactions described by Eqs. (4) and (9) take this into account because they do not permit an increase in heating when the depth of the thermocline increases. To simulate both El Niño and La Niña with the same model, a combination of the assumptions made in this and the earlier study of El Niño is necessary. Far better will be an improved model for the ocean in which SST is an explicit variable. This will be done at a later date. Further comments on another unrealistic feature of the model, the assumed relation between SST and the heating of the atmosphere, appear in section 4.

3. Results

Figure 5 shows the evolution of the oceanic currents, the thermocline depth and the surface winds along the equator after the introduction of the perturbation heat source to the west of the ocean basin. The westerly winds over the ocean immediately start to decelerate and in due course become easterly. These changes, which have westward phase propagation, are attributable to a westward contraction of the original heat source [Eq. (4)] that shrinks according to Eq. (9). The easterly winds are seen to drive accelerating westward ocean currents, and to elevate the thermocline along the equator. From Fig. 6, which shows the fields after 40 days, it is evident that the elevation of the thermocline at the equator is associated with a deepening off the equator. This figure suggests that the oceanic motion is in direct response to the easterly winds and includes an equatorial Rossby wave excited at the eastern coast of the basin.

Initially, there is little change in the winds over the western side of the ocean basin. The changes are predominantly in the east where the eastern extreme of the heated zone moves westward. After approximately thirty days, the winds in the west change significantly and after forty days, by which time the heating of the atmosphere is predominantly to the west of the ocean basin, the winds over the ocean are essentially steady. The ocean then comes into equilibrium with these winds. An oceanic Kelvin wave excited at the western boundary is evident in the thermocline depth variations of Fig. 5 and plays an important role in the adjustment of the ocean to the steady winds. Note how this wave is associated with an ultimate deepening of the thermocline in the west. Figure 6 shows the winds and the oceanic fields 100 days after the initial perturbation.

In these calculations, the perturbation heat source intensifies the easterly winds that directly cause upwelling in the eastern side of the basin. This triggers the instability. This instability develops in essentially the same way if the perturbation easterly winds are confined to the western side of the ocean basin. The oceanic response to such winds include an eastward travelling Kelvin wave that elevates the thermocline and which, upon its arrival at the eastern boundary, triggers the instability.

4. Discussion

Both the warm El Niño phase and the complementary cold La Niña phase of the Southern Oscillation can be explained in terms of unstable interactions between the tropical Pacific Ocean and the atmosphere. During El Niño, these interactions cause modest initial warm perturbations in the western Pacific to expand eastward, as happened in 1982. If the initial perturbation is in the eastern Pacific, then there is no zonal phase propagation and the amplification is essentially independent of longitude (Philander *et al.*, 1984). For the development of El Niño, it is crucial that the effect of the westerly wind anomalies to the west of a heat source, on the region to the east, exceeds the local effect of the easterly winds to the east of the heat source. This permits the deepening of the thermocline by the Kelvin wave to overwhelm the equatorial upwelling induced by the easterlies. Should the effect of the westerlies be weaker than that of the easterlies—the introduction of a heat source to the west of the original heat source in section 3 accomplishes this—then the equatorial upwelling and low sea surface temperatures induced by the easterlies will expand westward as demonstrated in section 3. These results imply that developments over the tropical Pacific Ocean depend not on the wind stress over one relatively small part of the Pacific, but on the zonal integral of the windstress along the equator. It has been suggested (Barnett, 1984, for example) that the zonal wind over the region to the west of the date line is a good indicator of subsequent developments in the tropical Pacific Ocean. In June 1982 westerly winds appeared to the west of the date line, and El Niño developed. In December 1982 these westerlies become easterlies, but now they failed to be a good predictor because El Niño continued for several more months. Furthermore, exceptionally intense westerlies appeared west of the date line in 1979 when there was no El Niño. Analysis of data is needed to determine whether the zonal integral of the winds along the equator would have been a good indicator of developments in 1979 and 1982.

The discussions thus far have concerned the manner in which wind variations affect the ocean, and the sea surface temperature in particular, and the manner in which heat sources affect the atmosphere. The relation between the sea surface temperature and the heating of the atmosphere is, however, a complex one. An increase in sea surface temperature in a certain region

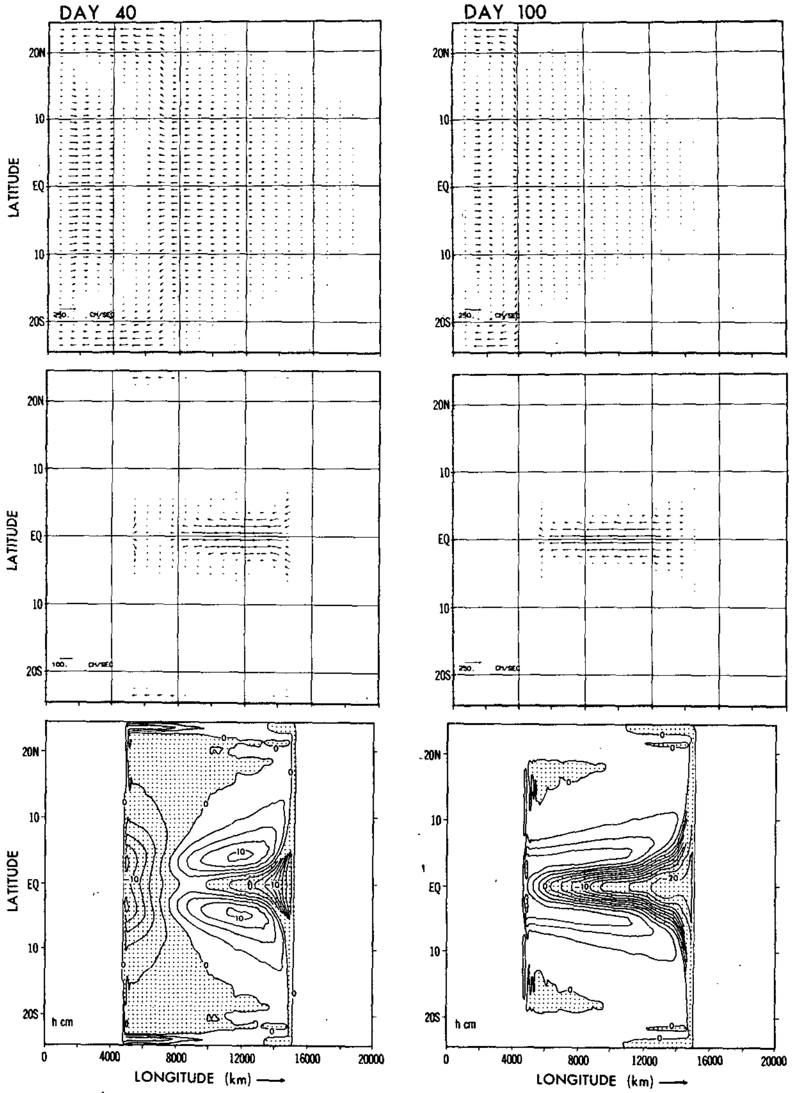

FIG. 6. As in Fig. 4, but conditions 40 days and 100 days after the perturbation heat source appears.

will cause a local heating of the atmosphere only under favorable large-scale atmospheric conditions. Specifically, it is necessary that there be large-scale rising atmospheric motion for a warm SST anomaly to cause local heating of the atmosphere and to initiate unstable air–sea interactions that amplify the anomaly. A warm anomaly that appears to the south of the equator in the eastern tropical Pacific during the boreal summer or autumn is unlikely to initiate an instability because that is a region of large-scale descending atmospheric motion at that time. Water vapor will be released to the atmosphere locally, but the condensation and heating of the atmosphere will occur where the air rises, probably over the western Pacific. The most favorable time for a warm anomaly in the eastern equatorial Pacific to cause local heating and to amplify is early in the calendar year when the seasonal migration of the ITCZ takes it to its lowest latitude and when the SST is at a seasonal maximum. This explains why, in Figs. 2 and 4, El Niño in the eastern tropical Pacific starts during the early months of the year when it amounts to an amplification of the warm phase of the local seasonal cycle. During the subsequent months the growth of anomalous conditions continues to be modulated by the seasonal movements of the ITCZ (Ramage and Hori, 1981).

In the western tropical Pacific the onset of El Niño is usually in April and May (Fig. 7). Those are the months when the seasonal movement of the convergence zone in the west is from the Southern to the Northern Hemisphere. During El Niño, this northward movement is modified in an eastward direction. The seasonal movement of the convergence zone in the western tropical Pacific is back towards the Southern

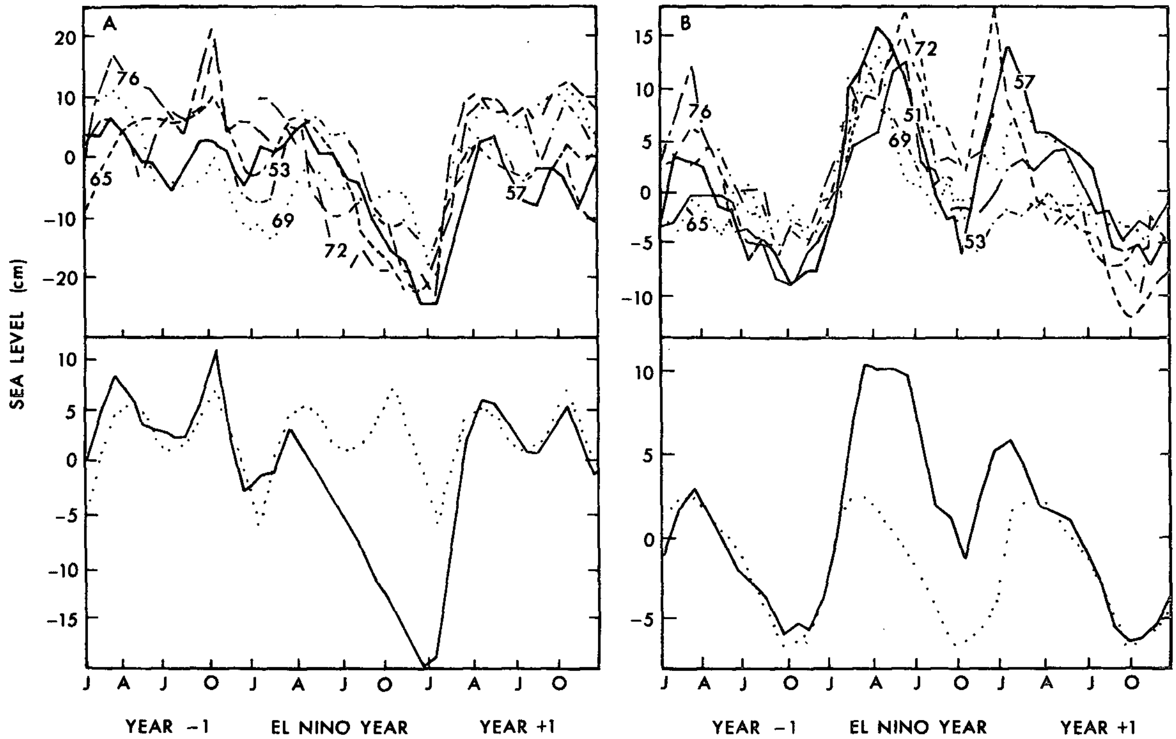

FIG. 7. (a) Sea level at Truk (152°E 7°N during indicated El Niño years (top panel), for the composite El Niño (bottom panel: continuous line) and for the annual mean in non-El Niño years (bottom panel: dotted line) (after Meyers, 1982). (b) Curves as in (a) but for sea level at Callao (79°W, 12°S). In the eastern Pacific, El Niño events typically appear as an enhancement of the annual cycle. In the western Pacific, a fall in sea level is correlated with westerly wind anomalies and a rise in sea level is correlated with easterly wind anomalies.

Hemisphere in December and January. This southward migration is remarkably regular and happens even during El Niño when sea surface temperatures in the central and eastern tropical Pacific are exceptionally high. The rise of the sea level in the western Pacific in Fig. 7 is associated with easterly winds that in turn are associated with the southward movement of the atmospheric convergence zone. This movement, which terminates El Niño and initiates La Ninã, clearly demonstrates that sea surface temperature is but one of the factors that influences the movements of the atmospheric convergence zones. The factors other than SST that affect the convergence zones must be responsible for the westerly wind anomalies west of the dateline at the beginning of El Niño and the easterly winds in that region at the beginning of La Niña.

The seasonal movements of the convergence zones modulate the unstable interactions between the tropical ocean and atmosphere. This seasonal cycle has interannual variability, which is caused by factors other than SST in the tropical Pacific, but that can be amplified by the unstable interactions thus causing the Southern Oscillation. The degree to which perturbations are amplified depends on their character and this may be the reason for the considerable interannual variation in the amplitude of the Southern Oscillation. Certain perturbations—those that amount to an amplification of the warm phase of the seasonal cycle in the eastern tropical Pacific, and those that correspond to an eastward movement, in April and May, of the convergence zone in the western tropical Pacific—often grow to enormous amplitudes. Others, which also merit attention, fail to amplify significantly: anomalous conditions with a large amplitude developed in neither 1979, when westerly wind anomalies appeared in the western tropical Pacific, nor in 1975 when the eastern tropical Pacific was unusually warm early in the year (Wyrtki *et al.*, 1976).

The period of the Southern Oscillation is approximately three years (Rasmusson and Carpenter, 1981). This is probably determined by the time it takes an unstable interaction to run its course through El Niño, a failure of the cold season in the east, and La Ninã, an enhancement of the subsequent cold season in the east. There is evidence that during El Niño the ocean loses an unusually large amount of heat to the atmosphere. Presumably the ocean recovers this heat during La Niña. It is unclear to what extent heat budget constraints affect the frequency of warm and cold events.

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