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A numerical study of the response of tropical Pacific SST to atmospheric forcing

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सार — भूमंडलीय मियामी आइसोपिकनिक समन्वय महासागर मॉडल के प्रचालन के लिए एन. सी. ई. पी/एन. सी. ए. आर. के चालीस वर्षों की पुनः विश्लेषित परियोजना से वायुमंडलीय संवर्धनों के आँकड़े दो सैटों में लिए गए हैं, जिनमें से एक सैट का आधार मासिक औसत जलवायविक आँकड़े हैं तथा दूसरे का आधार 1982 से 1983 तक के मासिक औसत आँकड़े हैं । इन्ही आँकड़ों का उपयेाग यहाँ किया गया है। इन देानों का जलवायविक प्रयोगों तथा 1982-83 के इलनिनो प्रयोगों से संबंध है। पवन-समुद्र ऊष्मा और संवेगी अभिवाहों के विभिन्न प्राचलीकरण समूहों पर उष्णकटिबंधीय प्रशान्त महासागरीय एस.एस.टी. के संवेदी परीक्षण दो प्रयोगों में किए गए हैं। प्रारम्भिक परिणाम यह दर्शाते हैं कि ऊष्मा अभिवाहों के निरन्तर स्थानान्तरित गुणांक (1.2 x 10^{-3}) उष्णकटिबंधीय प्रशान्त महासागरीय एस.एस.टी. की अति गणना करते हैं, जबकि लियु-कैट सरोसस-बुजिंगर (ल्यू आदि 1979) पद्धति के अनुरूपण द्वारा इसमें विशेष रूप से अतिन्यून वायु गति की स्थिति में उल्लेखनीय सुधार हेा सकता है। दूसरी ओर लार्ज और पॉण्ड (1982) कर्ष गुणांक सूत्रीकरण से उष्णकटिबंधीय प्रशान्त महासागरीय एस.एस.टी. अनुरूपण में बहुत कम अंतर आता है। यद्यपि इससे महासागरीय सतह परिसंचरण में परिवर्तन आने की संभावना है। एस.एस.टी. का ऋतु चक्र और उष्णकटिबंधीय प्रशान्त महासागर एस.एस.टी. की अंतःवार्षिक विभिन्नता का अध्ययन भी इस शोध-पत्र में किया गया है। चूँकि एस.एस.टी. वायुमंडल और महासागर के बीच सम्पर्क कायम करने वाला सबसे अधिक महत्वपूर्ण महासागरीय पैरामीटर है, अतः विभिन्न प्राचलीकरण स्कीमों के इस मूल्यांकन से भविष्य में महासागरीय वायुमंडलीय युग्म संख्यात्मक मॉडलों के अध्ययन में सहायता मिल सकती है।

ABSTRACT. Two sets of atmospheric forcing from NCEP/NCAR 40-year reanalysis project, one based on monthly averaged climatological data and the other on 1982-83 monthly averaged data, are used to derive the global Miami Isopycnic Coordinate Ocean Model (MICOM). These two runs are referred to as the climatological experiments and 1982-83 El Nino experiments. Sensitivity tests of tropical Pacific SST to different bulk parameterizations of air-sea heat and momentum fluxes are carried out in the two experiments. Primary results show that constant transfer coefficients (1.2×10^{-3}) for heat flux greatly overestimate the tropical Pacific SST, whereas the Liu-Katsaros-Businger (Liu *et al.* 1979) method can significantly improve the SST simulation especially under very low-wind speed conditions. On the other hand, Large and Pond (1982) formulation of the drag coefficient made little difference on the tropical Pacific SST simulation although it might modify the surface ocean circulation. The SST seasonal cycle and interannual variability of tropical Pacific SST are also examined in this study. Since SST is the most important oceanic parameter that provides the link between the atmosphere and the ocean, this evaluation of different parameterization schemes may facilitate future studies on coupling ocean-atmospheric numeric models.

Key words — Tropical Pacific, Sea surface temperature (SST), Simulation, El Nino, Parameterization, Numerical model.

1. Introduction

Progress in understanding the earth as a complete system is increasingly dependent on improvement in understanding the interactions between the atmosphere and the underlying oceans. The atmosphere provides primary driving forces for ocean circulation and exerts great effects on oceanic thermal structure. On the other hand, the ocean also plays a significant role in the atmospheric circulation as an important moisture and heat source. In this tightly-coupled system, the most important oceanic parameter that provides the link between the atmosphere and the ocean is the sea surface temperature (SST). The tropical Pacific experiences the strongest air-sea interaction in the world ocean, and accurate simulation of tropical pacific SST is critical to climate prediction in many regions around the world.

In the tropical Pacific Ocean, SST variation can be influenced by solar (shortwave) and planetary (longwave) radiation, heat flux across the ocean surface, advection, upwelling, and mixing processes. The last three terms depend on the past history of the surface wind stress, *i.e.*, the momentum flux, across the ocean surface. In an ocean numerical model, the air sea heat and momentum fluxes are often derived from SST field and atmospheric wind, temperature and humidity fields. However, because of the sparsity of observational data, inaccurate atmospheric forcing has always been a limitation of large-scale ocean models (Chen *et al.* 1994, Kirtman and Schneider 1996, Mestas-Nunez *et al.* 1994). For example, estimates of the uncertainty in monthly mean surface winds are on the order of $2ms^{-1}$ (Halpern 1988, Harrison *et al* 1984). Uncertainties in atmospheric forcing are sometimes so large that the differences between model results and observations can be explained by errors in the atmospheric forcing. Recently, NCEP/NCAR 40-year Reanalysis Project (Kalnay *et al.* 1996) which uses a frozen state-of- the-art global data assimilation system and an up-to-date database has become available. This new dataset should enhanced our ability to simulate SST.

Uncertainty in the bulk atmospheric parameterization schemes is another source of error for the atmospheric forcings. However, this uncertainty is often neglected by ocean modelers. The variation of fluxes between the ocean and the atmosphere is very sensitive to the choice of parameterization, especially in low wind regimes (Webster and Lukas 1992). This has been verified by Miller *et al.* (1992), who found dramatic improvements in simulated tropical phenomena by strengthening the air-sea coupling in the light-wind regime. The conventional coefficients used in the bulk formulae were derived and verified with observations taken in moderate wind condition (4-5m/s), however, over large areas in the tropical Pacific, the mean wind is weak (< 3 m/s). In low wind speed regimes it is necessary to account for buoyancy effects on turbulent transport. Standard stability- dependent bulk schemes (*e.g.* Liu *et al.* 1979) show good performance in the tropics. This scheme, termed LKB method hereafter, was further modified by Fairall *et al.* (1996) to include a different specification of the roughness/stress relationship; a gustiness velocity to account for the additional flux induced by boundary layer scale variability, and profile functions obeying the convective limit. The transfer coefficient for momentum, also known as the drag coefficient, was also studied extensively. For deep water with large fetch, it has been expressed as a function of wind speed or assumed to be constant over a range of moderate wind speeds. Summaries of these studies have been given by Businger (1975), Garratt (1977) and Smith (1988).

The motivations for this study are twofold. First, we wish to test the sensitivity of the tropical Pacific SST to different atmospheric forcing parameterizations in a global isopycnic ocean numerical model. Knowledge gained from sensitivity tests on these parameterizations might be used to evaluate the various schemes and facilitate future studies on coupling ocean-atmospheric numerical models. The second motivation is to examine both the seasonal cycle and the interannual variability of SST in tropical Pacific. The performance of coupled ocean-atmospheric models used for ENSO prediction critically depends on the ability of the ocean model to accurately simulate the SST variation. In this study we attempt to simulate the SST seasonal cycle and interannual variation in the tropical Pacific Ocean and try to identify the important processes that are responsible for the SST variability.

The remainder of this paper is outlined as follows: Section 2 briefly describes the ocean numerical model and atmospheric forcing used in this study. Section 3 briefly discuses the bulk parameterization schemes that are tested by the ocean model and the experimental design. Section 4 presents the results of simulations, followed by some concluding remarks and discussions in section 5.

2. Ocean model and atmospheric forcing

(a) *Ocean model*

The ocean model used in this study is the Miami Isopycnic Coordinate Ocean Model (MICOM), which has been described in detail by Bleck *et al.* (1992). It is a primitive equation model containing prognostic equations that have a coordinate of density in the vertical direction. The split-explicit, time-differencing scheme is used to extend the time step in the baroclinic calculations. The domain is that of the World Ocean between 69°S and 66°N. The coastline and bottom topography are realistic and the major islands within the domain are also treated realistically. The zonal resolution for our simulations was chosen to be 2 degrees. The meridional grid spacing is 2cosφ degrees at latitude φ since the horizontal mesh is superimposed on a Mercator projection of the earth's surface. The vertical stratification is reproduced by choosing 16 isopycnic layers with specific densities, in which the uppermost layer represents a Kraus and Turner 1967 surface mixed layer.

The model is forced by time- and space-dependent zonal and meridional component of wind speed, air temperature, specific humidity, radiation and precipitation. The atmospheric forcing datasets will be described in the following section. The air-sea heat and momentum fluxes in the model are calculated by bulk parameterizations:

$$H_s = \rho_a C_{pa} C_h V(T_s - \theta) \tag{1}$$

$$H_e = \rho_a L_e C_e V(q_s - q) \tag{2}$$

$$\tau_i = \rho_a C_d V(u_{si} - u_i) \tag{3}$$

Where,

H_s - sensible heat flux

H_e - latent heat flux

τ_i - ith component of wind stress

ρ_a - air density

V - magnitude of wind speed

C_{pa} - specific heat

L_e - latent heat

C_h - transfer cofficient of sensible heat flux

C_e - transfer coefficient of latent heat flux

C_d - drag coefficient

T_s - air-temperature at the surface

θ - air-temperature at 10 meter height

q_s - specific humidity at the surface

q - specific humidity at 10 meter height

u_{si} - ith component of wind speed at the surface

u_i - ith component of wind speed at 10 meter hight above the surface

In the original ocean model, the transfer coefficients C_h,C_e and C_d are all constants (1.2×10^{-3}).

The initial conditions for our simulations are derived from climatological data (Levitus 1982). The model is spun up for 100 years forced with the monthly averaged climatological forcing derived from Comprehensive Ocean Atmospheric Data Set (COADS).The subsequent simulations use the results at 100 years as the initial conditions.

(b) *Atmospheric forcing*

The monthly-averaged atmosphric forcings are from the NCEP/NCAR 40-year Reanalysis Project (Kalnay *et al.* 1996). The original datasets have a resolution of 2.5°, and are linearly interpolated onto the MICOM grid. There are two monthly averaged datasets used in the study: climatological data and 1982-83 data. The climatological data (13-year averaged monthly fields) are used for the study of SST variation at normal atmospheric conditions, whereas the 1982-83 data, which cover one of the strongest El Nino events, are for the study of SST variation at strong El Nino conditions. The radiation at the air-sea surface is calculated as the sum of the following four terms in the dataset: downward shortwave radiation, upward shortwave radiation, downward longwave radiation and upward longwave radiation. In addition, air temperature and specific humidity at 2-meter height are used.

3. Bulk parameterization schemes and experimental design

(a) *Bulk parameterization schemes*

The practical way of determining largescale air-sea momentum and heat exchanges in ocean numerical

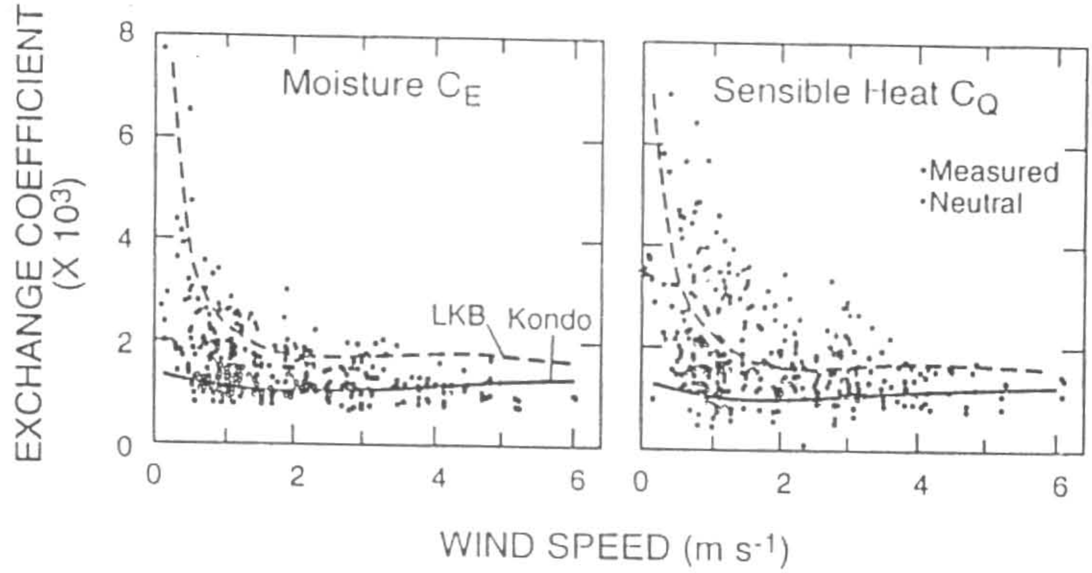


Fig. 1. Eddy correlation measurements of exchange coefficients for: (a) water vapour C_E; (b) Sensible heat C_Q. Models are from Liu *et al.* (1979) (dashed curve) and Kondo (1975) (solid line). After Bradley *et al.* (1991)

model is through bulk parameterization. The transfer coefficients of heat and momentum fluxes are empirically derived. Constant transfer coefficients C_d, C_h and C_e are usually used in moderate wind conditions (4-15 m/s) under near neutral conditons. Over large areas in the tropical Pacific, however, the mean wind is weak (< 3 m/s). In the low wind speed regimes it is necessary to account for buoyancy effects on turbulent transport, and a stability-dependent bulk scheme is needed. One example of this kind of scheme is LKB method developed by Liu *et al.* (1979). In LKB method, three non-dimensional equations based on similarity theory are solved, and the profiles of C_h and C_e are shown in Fig. 1 in which it is evident that the coefficients are much higher in low-wind speed condition than in moderate wind conditions.

In 1982, Large and Pond (LP hereafter) developed a wind speed-dependent two-branch drag coefficient C_d to calculate the wind stress:

$$10^3\, C_d = \begin{cases} 0.49 + 0.065\, u, & u > 10\ m/s \\ 1.14 & 3 \le u \le 10\ m/s \\ 0.62 + 1.56\, u - 1 & 1 \le u \le 3\ m/s \\ 2.18 & u \le 1\ m/s \end{cases}$$

TABLE 1

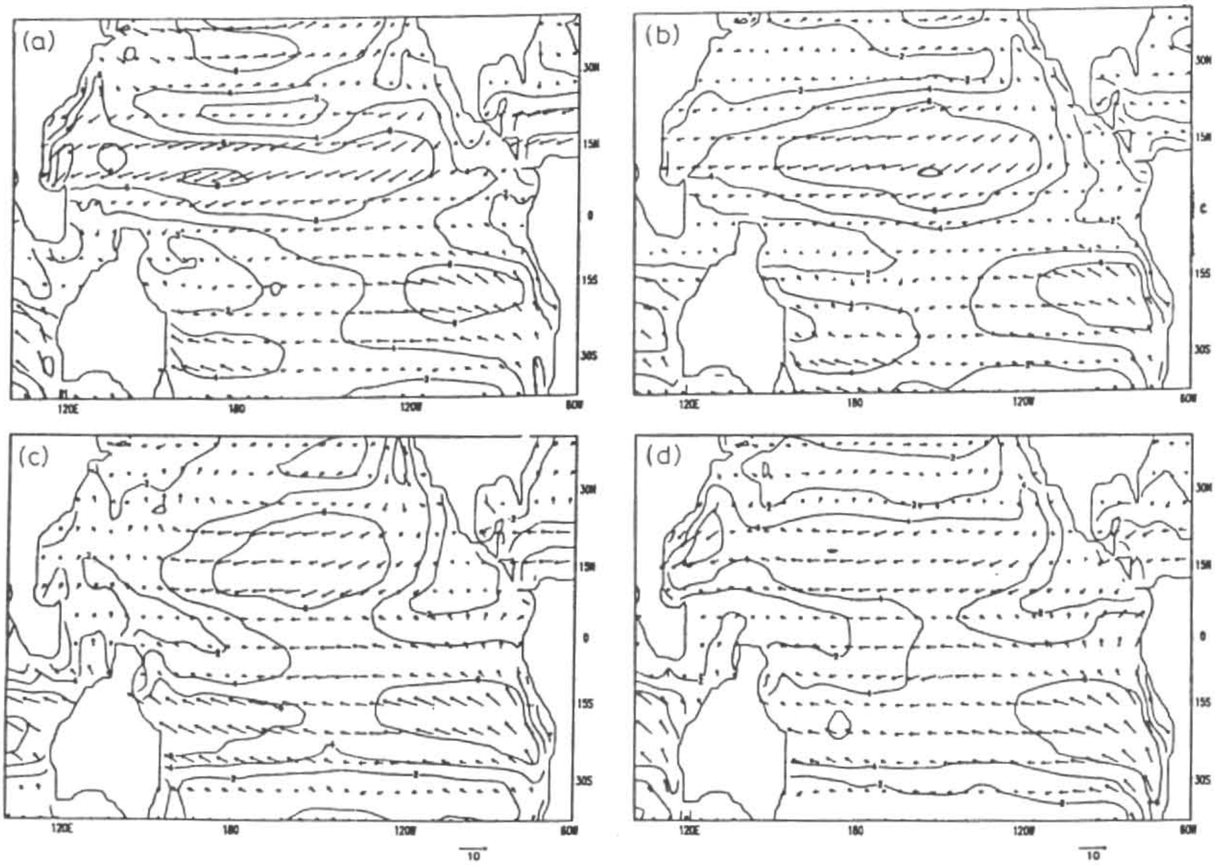
Climatological Experiments

Experiment •	Transfer coefficient of momentum flux	Transfer coefficient of heat flux
1.1	C_d = constant	C_h, C_e = constant
1.2	LP	C_h, C_e = constant
1.3	C_d = constant	LKB
1.4	LP	LKB

The form of the 1-3 m/s wind speed branch was based on the notion that the drag coefficient increases as the wind speed approaches zero.

(b) *Experimental design*

Two sets of experiments which use the two different atmospheric forcings described in the last section were carried out. In one set of experiments, monthly mean climatological data were used to drive the ocean model. In the second set of experiments monthly mean 1982-83 data were used. In each set of experiments, there are four cases with combinations of different air-sea heat and momentum flux parameterization schemes as listed in Tables 1 and 2.



Figs. 2(a-d). Monthly averaged climatological wind field (m/s) in (a) January, (b) April, (c) July and (d) October

TABLE 2

1982-83 El Nino case experiments

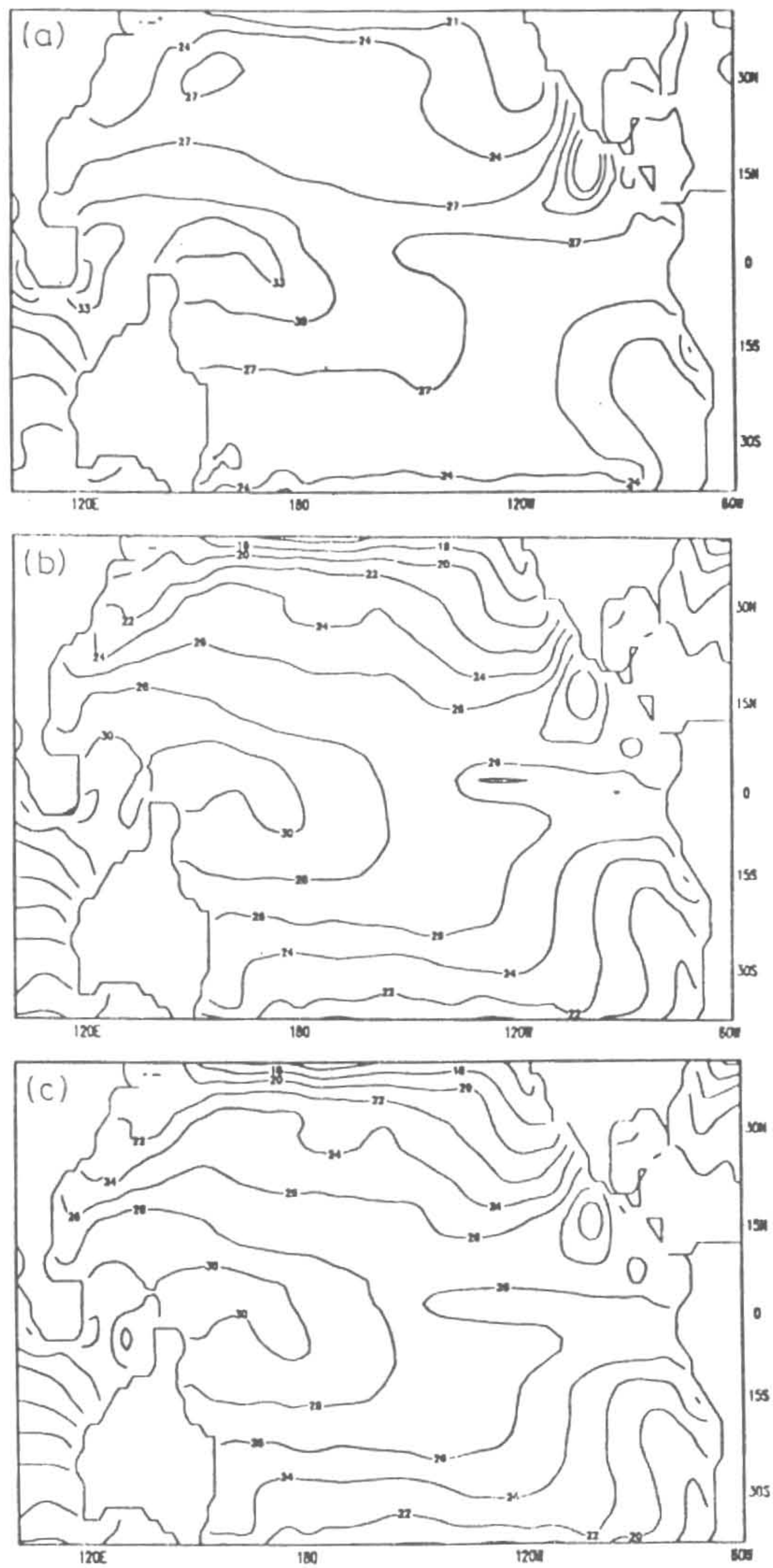
Experiment	Transfer coefficient of momentum flux	Transfer coefficient of heat flux
2.1	C_d = constant	C_h, C_e = constant
2.2	LP	C_h, C_e = constant
2.3	C_d = constant	LKB
2.4	LP	LKB

4. Results and discussion

(a) *Climatological experiment*

The monthly averaged climatological wind field in January, April, July and October over the tropical Pacific are shown in Fig. 2. The most pronounced feature of tropical Pacific wind is the northeast and southeast trade winds that lie north and south of the inter-tropical convergence zone (ITCZ). Both trade wind systems undergo an annual cycle of equatorward migration and zonal expansion into the western ocean in winter and spring seasons of its hemisphere, followed by poleward migration and contraction into the central and eastern ocean during summer and fall. During this cycle, it can be seen clearly that the southeast trades in the eastern Pacific Ocean intensified in summer and fall and relaxed in winter and spring. Another characteristic of the wind field is that over large areas in the ITCZ, the annual mean wind speed is very weak (< 4 m/s), and in some areas the wind speed is even weaker (< 2m/s).

Using the climatological wind forcing described above, we tested the sensitivity of tropical Pacific SST to different parameterization schemes of air-sea heat and momentum fluxes listed in Table 1. In these simulations, the factors that affect the SST variations are the wind-stress-induced oceanic advection and mixing and the air-sea sensible and latent heat fluxes. The solar and longwave radiations are identical among the four cases of the experiments although they are also very important factors that can affect the SST



Figs. 3(a-c). Simulated sea surface temperature (SST) (^{o}C) of October in (a) case 1.1; (b) case 1.3 and (c) case 1.4

variation in the ocean. Furthermore, in the calculation of sensible and latent heat fluxes, the air temperature and specific humidity at 2-meter height are also identical in all of the four cases, *i.e.*, the differences of heat fluxes in different cases are caused only by the differences of the transfer coefficients C_h and C_e.

To show the results of sensitivity tests, we take the simulated SST Fig. 3, surface current speed and mixed layer depth Fig. 4 in October as examples. In case 1.1 in which all the transfer coefficients are constants (1.2×10^{-3}), it is evident that the SST is overestimated in the tropical Pacific although the general SST distribution pattern is reasonable Fig. 3a. The highest SST in warm pool of western Pacific is 33°C, and in the warm spot off the coast of Mexico the SST is even higher. When we compare this SST field with the corresponding wind field of October (Fig. 2d), it is found that there are two low-wind-speed centers (< 2m/s) over the warm pool and off the coast the Mexico which are exactly the same locations of highly overestimated SST. Therefore, constant transfer coefficients overestimate tropical SST especially under low-wind-speed conditions.

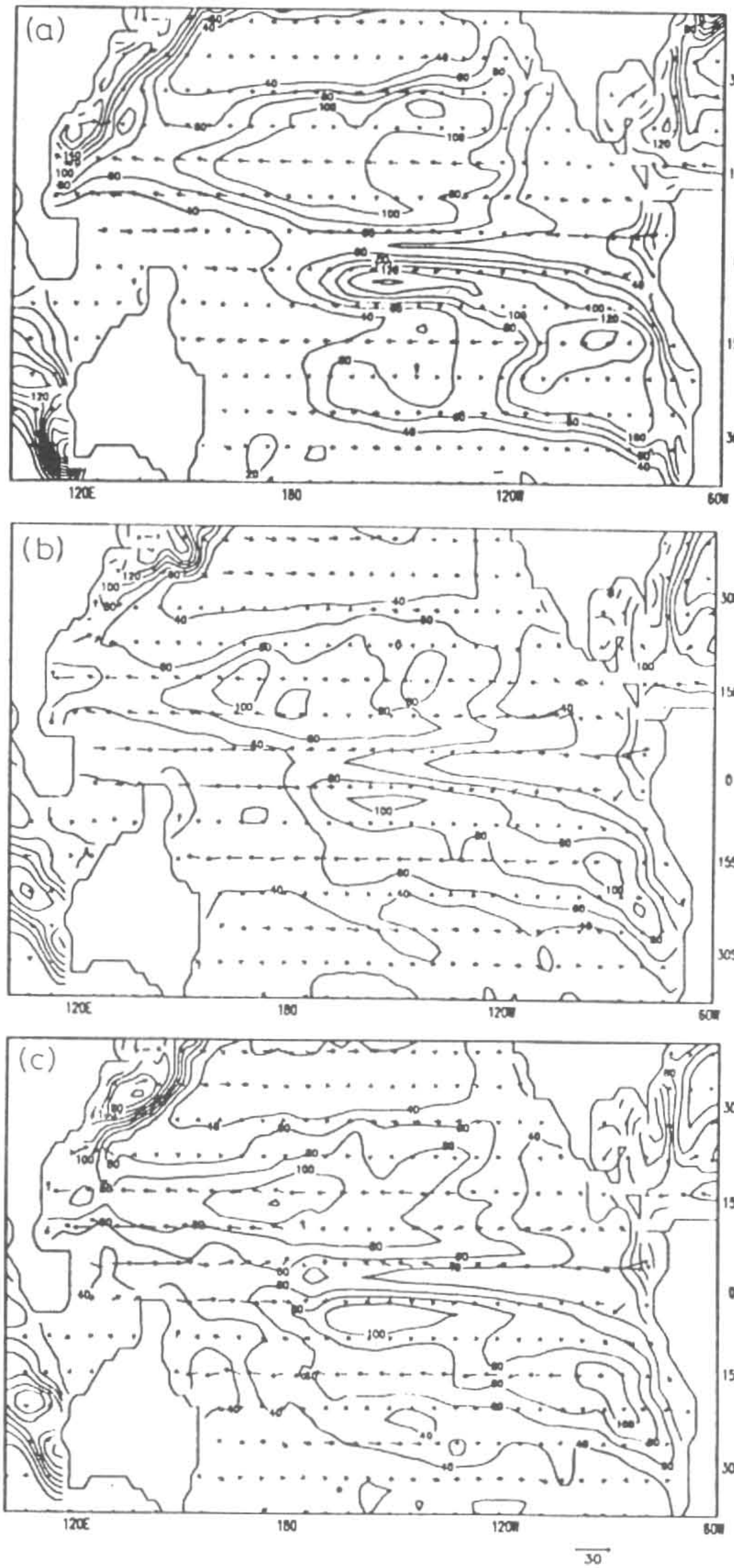
In case 1.3, the transfer coefficients of sensible and latent heat fluxes are calculated by LKB method, but the drag coefficient is the same as in case 1.1. In this simulation, the SST become realistic and the highest temperature falls within realistic range (Fig. 3b). The warm surface water with temperatures higher than 28°C occupies the western tropical Pacific and extends eastward both north and south of the equator. Colder water appears in the southeast and equatorial ocean, forming a cold tongue extending from the Peru coast to the central equatorial ocean. In the warm pool of western Pacific and the warm spot off the coast of Mexico the highest SST is 30°C. This great improvement in tropical Pacific SST simulation is mainly due to LKB method. Since the drag coefficient C_d in case 1.3 is the same as in case 1.1, the wind stress is the same and the simulated ocean circulation are similar in both of the simulations as we can see in Figs. 4 (a&b). The only factor that makes the difference in SST between case 1.3 and case 1.1 is heat flux across the ocean surface.

In case 1.4, the transfer coefficients of heat flux are calculated by LKB method and the drag coefficient is calculated by LP method. The contribution of LP method is that it increases the estimation of wind stress at low-wind speed conditions, so that the wind speed-induced ocean circulation is also affected Fig. 4c. However, the modified pattern of the ocean circulation does not seem to significantly change the simulated SST. By comparing Figs. 4 (b & c) it is obvious that the SST distribution are almost the same in the two figures except that the cold tongue in Fig. 4(c) is slightly stronger than in Fig. 4(b). This can be explained by the upwelling effect. In the eastern Pacific, off the coast of the South America, increased wind stress can cause cooling of surface water due to the enhanced upwelling. However, in the western and central Pacific, where the upwelling is very weak and the SST horizontal gradient is small, increased wind stress does not cause significantly SST cooling although the ocean circulation system is modified.

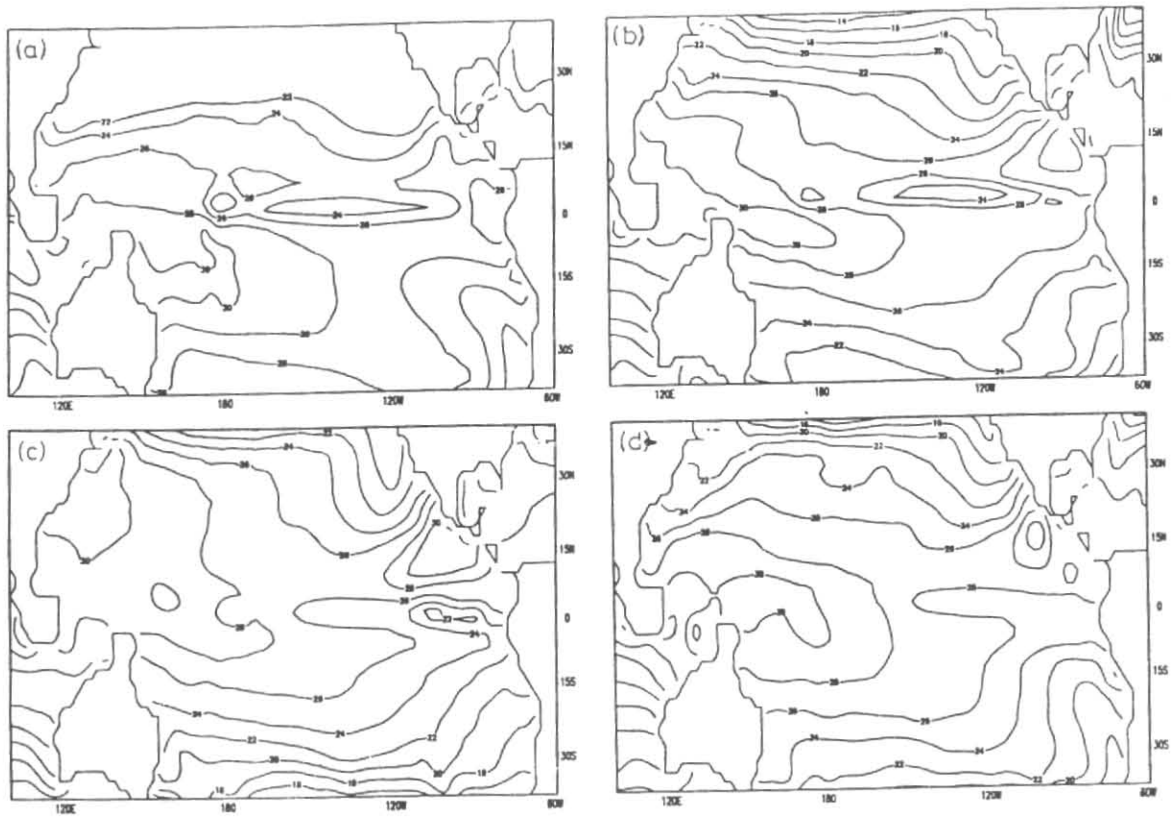
The annual SST cycle of the eastern Pacific Ocean is closely related to the variations of southeast trades. Simulated tropical Pacific SSTs in January, April, July and October are shown in Fig. 5. During the period of relaxed southeast trades (north winter and spring), the extent of the cold tongue is at its minimum, and so is the horizontal thermal gradient (Figs. 5 a & b). In north summer and fall when the southeast trades are intensified, the extent of the old tongue is maximum (Figs. 5c & d). On the other hand, the annual SST change in warm pool in the western Pacific is not induced by the trade-wind system, but by the solar radiation: in northern spring the warmest surface water is south of the equator (Fig. 5a), whereas in northern fall it is in the north (Fig. 5d).

(b) *1982-83 experiment*

Tropical Pacific monthly averaged wind field in 1982 and 1983 are shown in Fig. 6. The most important feature in the 1982-83 wind data is the gradually eastward movement of the convergence zone across the tropical Pacific. The convergence zone was around 160°E in April 1982 (Fig. 6b), 180°E in July 1982 (Fig. 6c) and 160°W in October 1982 (Fig. 6d). The eastward movement of convergence zone coincided with a collapse of the trade winds as westerly wind anomalies propagated eastward. These conditions reached the coast of South America in late 1982, by which time westerly winds prevailed over much of the western and central equatorial Pacific. During the El Nino of 1982-83, the wind speed in the tropical Pacific is extremely weak, especially in the convergence zone



Figs. 4(a-c). Simulated surface current-speed (cm/s) and mixed layer depth (m) of October in (a) case 1.1; (b) case 1.3 and (c) case 1.4



Figs. 5(a-d). Simulated sea surface temperature (SST) (°C) in case 1.4 for (a) January, (b) April (c) July and (d) October

in June-October of 1982 and all over the equatorial Pacific in November 1982-March 1983. Under this kind of low wind speed condition, the air-sea heat and momentum flux estimation by the constant transfer coefficients becomes even worse than seen in the climatological experiment as we can see below.

The sensitivity of tropical Pacific SST to different air-sea momentum and heat flux parameterization schemes is primarily tested in the four cases listed in Table 2. The results are similar to that in climatological experiment except that the SST is even more over-estimated under the extremely low-wind speed conditions by using constant transfer coefficients. Take the SST field in October 1982 (Fig. 7) as an example. In case 2.4, in which LKB and LP method are used, the SST is realistic with the highest SST of 30°C (Fig. 7a). However, in case 2.1, in which all the transfer coefficients are treated as constants, there is a patch of extremely high SST (36°C) (Fig. 7b). This patch of high SST is exactly the same location of convergence zone in Fig. 6 where the wind speed is less than 1m/s. As discussed in the climatological experiment, the improvement in SST simulation is mainly made by the LKB method. LP method makes little difference to the SST simulaton in tropical Pacific.

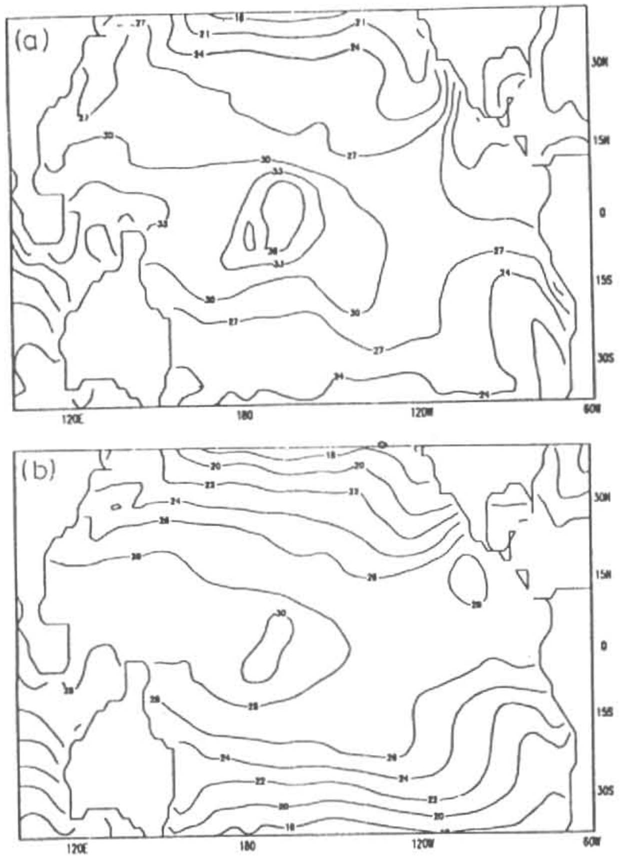
Another result worth noting in the 1982-83 experiment is the SST variation. El Nino of 1982-83 was one of the strongest ones in the past decades and had an unusual evolution. In the atmosphere, the 1982-83 El Nino started with an eastward movement of convergence zone in western Pacific in May 1982. In the ocean, this coincided with an eastward propagation of warm surface waters caused by a local weakening of the winds and hence reduced evaporation. This eastward propagation of warm SST can be seen clearly in both simulated and observed time-longitude cross sections of SST (Fig. 8). These conditions reached the coast of south America in late 1982 when the westerly winds prevailed over much of the western and central equatorial Pacific. By then the eastern tropical Pacific had become exceptionally warm because

Figs. 6(a-f). Monthly averaged wind field (m/s) in (a) January 1982, (b) April 1982, (c) July 1982, (d) October 1982, (e) January 1983 and (f) April 1983

of the reduced southeast trade wind and hence reduced upwelling. The difference between the simulation and observation of SST in Fig. 8 is the onset of eastward propagation of warm SST in the western Pacific. The warm SST began to move eastward in June 1982 in observation, but in August 1982 in the simulation which has 2-month lag compared with observations.

5. Conclusions

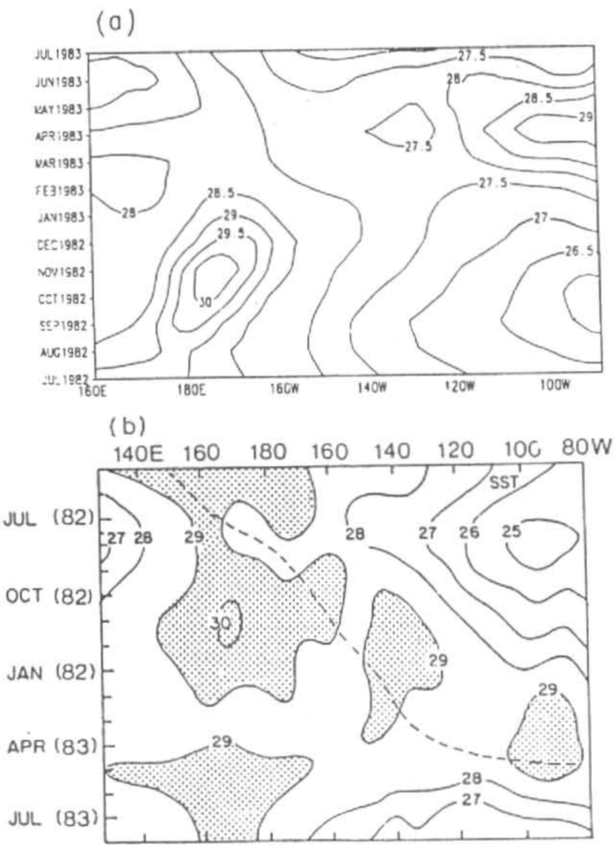
Two sets of monthly averaged atmospheric forcings from NCEP/NCAR 40-year Reanalysis Project are used to drive the global Miami Isopycnic Coordinate Ocean Model. One set of forcings is monthly averaged climatological data (13-year mean), and the other is 1982-83 monthly data. Sensitivity tests of tropical



Figs. 7(a&b). Simulated sea surface temperature (SST) (°C) in October 1982 for (a) case 2.1 and (b) case 2.4

Pacific SST to different combinations of heat and momentum transfer coefficients under both the normal condition and the strong El Nino condition are carried out. The tropical Pacific SST annual cycle and interannual variation (1982-83) are also examined in this study. Our findings are the following:

(*i*) Constant transfer coefficients (1.2×10^{-3}) for heat fluxes greatly under-estimate the air-sea heat fluxes, thus, over-estimate the SST in tropical Pacific especially under very low-wind speed conditions. Liu *et al.* (1979) method can significantly improve the tropical Pacific SST simulation, since it has a wind-speed-dependent parameterization scheme that increases the air-sea heat fluxes under low-wind speed condition. This improvement is evident in both the climatological studies and the 1982-83 El Nino studies. Large and Pond (1982) formulation of the drag coefficient, on the other hand, made little difference on the tropical Pacific SST



Figs. 8(a&b). Time-longitude cross-section of tropical Pacific SST (5°N-5°S): (a) simulated and (b) observed (from Gill and Rasmusson 1983)

simulation although it might modify the surface ocean circulation under low-wind speed conditions.

(*ii*) A qualitatively realistic tropical annual SST cycle is reproduced by the model. The SST in the eastern Pacific is warm in northern winter and spring and cold in summer and fall. In the western Pacific the warm pool has a southward movement in north winter and spring and a northward movement in summer and fall. In the 1982-83 El Nino studies, the eastward warm SST propagation across the tropical Pacific is clearly simulated by the model. The highest SST in the eastern tropical Pacific increased from 26°C in August 1982 to 29°C in March 1983. However, the onset of eastward propagation of warm SST in the western Pacific has 2- month lag compared with observations.

(*iii*) A possible reason for the 2-months lag of the onset of eastward warm SST propagation is that monthly-averaged data lacks necessary high frequency information. The onset of El Nino in the western tropical Pacific can be triggered by a westerly wind burst. However, in the monthly-averaged data the westerly wind burst could be decreased and even disappear, so the triggering machanism for the onset of warning could similarly be decreased and even disappear. To test this hypothesis, we plan to use European Center for Medium-Range Weather Forecasts (ECMWF) 1982-83 twice-daily data to drive the ocean model. Furthermore, by analyzing the model results produced by the monthly averaged data and by high-frequency twice-daily data, we may explore potential atmospheric forcing parameterization schemes which may compensate the lost of high frequency information in the monthly averaged forcings.

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