# Comparison of Tropical Pacific Temperature And Current Simulations With Two Vertical Mixing Schemes 1 Embedded In An Ocean General Circulation Model And Reference To Observat ions

David 1 Ialpern[1], Yi Chao[2, 3], Chung-Chun Ma [2] and Carlos R. Mechoso [2]

[1] 1 Earth and Space Sciences Division
Jet 1 'repulsion Laboratory
California Institute of Technology
Pasadena, CA 91109

[2] 1 Department of At mospheric Sciences
University of California
Los Angeles, CA 90024

[3] Now **at** Earth and Spare Sciences 1 Division
Jet Propulsion Laboratory
California Institute of Technology
Pasadena, CA 91109

**Abstract.** The Pacanowski-Philander (1'1') and Mellor-Yamada (MY) parameterization models of vertical mixing by turbulent processes were embedded in the Geophysical Fluid Dynamics Laboratory high-resolution ocean general circulation model of the tropical Pacific Ocean. All other facets of the. numeric.al simulations were the same.. Simulations were made for the 1987 - 1988 period. At the *equator,* the M Y simulation produced near surface temperatures more uniform with depth, a deeper thermocline, a deeper core speed of the Equatorial Undercurrent, and a South Equatorial CuJ'rent with greater vertical thickness compared to that computed with the PP method. Along 140°W between 5°N and 10°N, both simulations were the same.

Moored-buoy current and temperature observations had been recorded by the Pacific Marine Environmental Laboratory at three sites (165°E, 140°W, J 10OW) along the equator and at three sites (5°N, 7"N, 9°N) along 140°W. Simulated temperatures were lower than those. observed in the near surface layer, and higher than those observed in the thermocline. Temperature simulations were in better agreement with observations compared to current simulations. At the equator, PP current and temperature simulations were more representative of the. observations than MY simulations.

## J. Introduction

A key component for the success of ocean general circulation model (OGCM) simulations of upper ocean current and temperature fields is the representation or parameterization, in terms of large-scale variables, the turbulent motions and other physical processes occurring at scales smaller than the. mode.1 grid. Vertical mixing of heat, salt and momentum by turbulent processes is a very important phenomenon to parameterize, and many parameterization schemes exist. Some are based on turbulent kinetic energy [*Mellor and Yamada*, 1982; *Gaspar et* al., 1990], Richardson number [*Pacanowski and Philander*, 1981], Richardson number and wind friction velocity [*Reason et al.*, 1993], one-dimensional mixed layer dynamics [*Kraus and Turner*, 1967], observations of mixing [*Peters et al.*, 1988], and combinations of the methods [*Stockdale et al.*, 1993]. The *Pacanowski and Philander* [1981] method, hereafter named "PP", and *Mellor and Yamada* [1982] scheme,

hereafter named "MY", are widely used in uncoupled OGCM [*Stockdale et al.*, 1993] and coupled ocean-atmosphere GCM [*Neelin et al.*, 1992] simulat ions. 1 low MY and PP simulations of temperature and zonal current fields in the tropical Pacific compare with each other, and with observations, is addressed in this paper. The study encompassed the 1987-1988 interval containing El Niño and La Niña. Moored-buoy current and temperature data were recorded by the Pacific Marine Environmental Laboratory along the equator at 165°E, 140°W and 110°W [*McPhaden and Hayes*, 1990; *McPhaden et (il., 1990]* and along 140°W at 5°N, 7°N, and 9°N.

The two OGCM simulations differ only in the. formulation of the parameterization of vertical mixing by t urbulent processes. In the PP parameterization the vertical diffusivity and viscosity are functions of the local Richardson number because of a belief that vertical mixing is strongly influenced by vertical shear of the mean currents. In t he upper 10 m of the model, the coefficient of vertical eddy viscosity has a minimum value 10 compensate for mixing by the high-frequency wind fluctuations which are absent from the monthly mean winds. The KT parameterization is an ad hoc simple solution to include turbulent mixing, in a GCM. The KT scheme employed in our analysis had the same coefficients used by *Philander et al.*, [1987]. In contrast to the simplified representation of mixing portrayed by the PP scheme, t urbulence closure models attempt to explicitly model the turbulent transports of heat, salt and momentum in terms of the larger scale fields. The MY parameterization code dcve.loped by *Rosati and Miyakoda* [ 1988] was used.

The Philander-modified version of the *Bryan* [1969] - *Semtner* [1974] - *Cox* [1984] OGCM [*Philander et al.*, 1987] coven the Pacific Ocean from 30°S to 50°N with realistic coastal geometry. The model ocean has a constant depth of 4149 m. A rigid lid approximation is made at the surface. The longitudinal resolution is 1°. The latitudinal grid size is 1/3° within 10°S and 10°N, and gradually increases poleward to 2.5° at 30°S and 50°N. Coefficients of horizontal eddy viscosity and eddy diffusivit y between 10°S and 10°N are $2 \times 10^3$ m$^2$ s-1. There are 27 levels throughout the water column with 10 uniform layers in the upper 100 m and 8 additional levels between 100 and 317 m. The model time step is 1 hour, and 3-day average temperature, .salinity, and horizontal

velocity components are stored. A model year has *365* clays, or 120 intervals of 3.04 clays, A model month was equal to ten consecutive 3.04-day intervals.

Initial conditions of the OGCM temperature and salinity distributions were the *Levitus* [1982] climatological-mean January distributions. The model ocean was initially at rest. 1 'or the first three years of simulation the model ocean was forced with the *Hellerman and Rosenstein* [1 983] cl i matological -mean monthl y surface wind stress and the *Oort et al.* [1 987] climatological-mean monthly surface air temperature. Surface heat flux was computed according to the method described by *Philander et al.* [1987]. No surface salt flux was i mposed on the OGCM. Si mulated oceanic conditions al the end of the third year became the initial conditions for the 1986-1988 simulations. Simulations for 1986 were used elsewhere and we have not explored the, effect of beginning the simulations in January 1986 instead of a year late.r. 1/or the 1986-1988 simulations, t hc mode.] ocean was forced with surface air temperature from the European Centre for Medium-Range Weather 1 'orecasting and surface wind stress computed from the Florida State Universit y (FSU) monthly mean 2.(' x 2° pseudostress component fields [*Goldenberg and O'Brien*, 1981 ] by using a constant drag coefficient (1.4x10- s). This value is about 15% higher than the 12-b value suggested by *Trenberth et al.* [1989]. In comput ing the wind stress the speed of the water is assured negligible compared with the wind speed, which creates an uncertainty of the monthly wind stress along the equator as large as 17% for several months [*Halpern*, 1988].

## 2. Results

### 2.1. Temperature Along Equator

1 'igure 1 displays the annual mean MY and 1'1' temperat ure differences at t he equator corresponding to 1988, because these differences were larger than in 1987. Differences bet ween 1'1' and MY sea surface temperat ures (SS'1'), i. e., the 0-10 In average temperat ure, were negligible. Temperature differences greater than 1 'C occurred from 25-100 m and from 150°W to 85°W because depths of MY 15-20°C isotherms were 10-20 m deeper than PP isotherms.

Maximum temperature differences (3-4°C) were confined to a thin (10 m) interval near 35 m depth and a narrow longitudinal band ( 15° wide) in the eastern Pacific near 95°W.

At 11 0°W in 1987 (1 'igure 2a), MY temperatures at 10, 25 and 35 m were 0.6°C more representative of the moored-buoy observations, named "M 1 3", than 1'1' temperatures. Also, M Y and MB near surface temperat urcs displayed near- zero vertical gradients. in 1988 at 110° W (1 'igure 2b), MY temperatures at 10 and 25 m again indicated a near-zero vertical gradient; however, 1'1' and M B temperatures indicated that the thermocline reached the. surface.

Simulated temperatures above 50 m were less than that observed, named "cold bias", and simulated temperat ures below 75 m were higher than observat ions, named "warm bias" (Figure 2). Maximum cold and warm biases were 2.2°C (45 m, 110°W, 1987) and 2.7°C (12.0 m, 140°W, 1988), respectively. Biases caused the simulated temperature gradient between the surface and 150 m to bc weaker than that observed at 140°W and 11O°W.

'1'here were 581 monthly mean MB values at 165°E, 140°W and 11 O°W. The M H and MY bias (0.8°C) was significant, according to Student's t-test [*Press et al.*, 1986]. The **95%** significance level is used throughout the paper. The MB and PP bias (0.4°C) was not significant, indicat ing a better correspondence bet wccn M B and PP temperat ures. Correlation coefficients bctwccmh411 and PP and bet ween M B and MY were equal I y high (0.94). Root-mean-square (rms) differences bet ween MB and PI' and bet wccn M B and MY were equal I y large ( 1.9°C).

The amplitude of the annual period fluctuation of the 10-m temperature, $T_{10m}$, during 1987-1988 was determined by harmonic analysis in which wc assume the existence of an annual period fluctuation of the form $(T_{10m})_a$ Sin $(\omega t - \phi)$, where $(T_{10m})_a$ is the amplitude, $\omega$ is the annual angular frequency (= 3600/12), t is the time. in months, and $\phi$ is the. phase lag in degrees. At 11O°'W the M B, MY and 1'1' amplitudes were about the same (1 .6-1.8°C); at 140°W the. amplitudes were approximately the same ( 1 .0-1.1 °C), but 40% smaller than at 11 0°W. The 1 O-m simulated annual cycle of temperature was not sensitive, to the MY or 1'1' parameterizations. Other vertical mixing schemes influence the annual fluctuation of SST in the tropical Pacific [(km *et al.*, 1994].

## 2.2. Temperature Along 140°W

in the 5°N - 10°N band where the North Equatorial Countercurrent (N1 3CC) occurs, the MY ?5, 26 and 27°C isotherms were more uniform with depth than the corresponding 1'1' isotherms, and the MY isotherms extended to greater depths compared to the PP isotherms (Figures 3a and 3b). Throughout the NJ 3CC the PP and M Y differences were less than 1 °C (Figure 3c).

At 5°N (1 'igure 3d), the simulated temperatures at the depth (2.0 m) of the uppermost observation were 0.5-1 .0°C lower than the observed temperature. Whether this situation represents a northward extension of the equatorial cold bias is unknown. I n the center of the N] 3CC at 7°N (Figure 3e), no cold bias occurred near the surface; however, at 9°N (1 'igure 3f), simulated temperatures at 20 and 45 m displayed a cold bias. The warm bias in the thermocline was largest (2°C) at 9°N and smallest at 7°N.

Most of the difference between simulated and observed temperatures within the thermocline at 5°N and 9°N were caused by a 3- to 4-month disturbance (not shown) that was not apparent in either simulation. At 7°N, the simulated and observed mean temperatures were. fortuitously equal (1 'igure 3e) because the disturbance also appeared there, At 5°N, 7°N and 9°N, the departure between a monthly mean observed temperature and the nearly identical simulated temperatures reached 5°C. The correlation coefficient and rms difference between simulated and observed temperatures were 0.94 and 2.2°C, respectively, indicating the inability of either simulat ion to adequately reproduce the magnitude of the fluctuation, The origin of the disturbance, which could have been a norlh-south meandering of the NI3CC [*Wyrtki*, 1978], is unknown.

## 2.3. Current Along Equator

1 )epths of the PP and MY Equatorial Undercurrent (1 3UC) core speed decreased from west to cast along the equator across the entire width of the. Pacific (Figures 4a and 4b). The 1'1' c.ast-west slope. of the depth of the. core speed was larger and the 1'1' depth of the core speed was less than that simulated by MY, resulting in stronger vertical shear from the surface to the 1 3UC core for PP compared to MY. The PP cast ward current component was greater than MY above the depth of the

1 EUC core speed, except at about 5 m near 105C'W (1 'igure 4c). Below the depth of the core speed, the 1'1' and MY shears were approximatel y equal.

The PP and MY annual mean differences in the zonal current component were larger in J 988 than in 1987, and the maximum difference (0.5-0.6 m s-]), which was greater than one-half the I EUC core speed, occurred at about 65 m in the 145°W- J 30°W longitudinal band (1 'igure 4c). In 1987 the maximum difference (0.3-0.4 $m s^{-1}$) was also at about 65111, but the position shifted cast ward to 120°W- J 05°W.

Wc investigated whether the large point-to-point differences between PP and MY zonal currents (Figure 4 c) appeared in integral properties of the EUC and South EquatorialCurrent (S1 EC) at J 65°E, J 40°W and 110°W. The 2-year mean1'1'1 EUC transports per unit width were larger compared to MY by about J 6% (13 $m^2 s^{-1}$). The opposite situation was found for the SEC because the MY transport per unit width was greater than 1'1' by about 27% (4 $m^2 s^{-1}$). The 1'1' and MY vertical thicknesses of the. EUC were almost the same (235 m), in marked contrast to the 50% difference between the PI' and MY vertical thicknesses of the S1 EC. The PP EUC transport (32x $10^6$ $m^3 s^{-1}$, or 32 Sv) was nearly 4 Sv, or about 13%, greater than the MY value, and therms difference (5.8 Sv) was not much greater than the bias. The correlation coefficient between 1'1' and MY 1 EUC total transports was 0.95.

At J J O°W during both years the. 1'1' annual mean zonal current profiles were more representat ive of the observat ions with regard to current speed and shear compared to t he MY simulation (Figure 5), At 140°W, the }'1' and MY currents were similar, except at 80 m where the M Y eastward current was too small by 0.3 m $s^{-1}$ in J 987 and ().5 m $s^{-1}$ in 1988 compared to the observations. At 165°E, the PP and MY differences were negligible in J 987, and in 1988 the MY currents at 50 and 100 m were more rcprc.sent at ive of the observat ions.

On many occasions the monthly mean observed and simulate.d zonal currents differed by more than 0.5 m $s^{-1}$ (not shown). At 11 0°W, the 10-m simulated annual mean west ward current was 2-3 times greater than that observe.cl (Figure 5) because the simulat ions did not adequatel y reveal the observed eastward flow in March-May that is associated with the annual Kelvin wave.

I'here were 341 monthly mean MB values at 165°E, J 40°W and 11 O°W. The MB and 1'1' zonal current bias (0.02 m $s^{-1}$) was not significant, and the MB and MY bias ((). 13 m $s^{-1}$) was significant. The 1'1' monthly mean zonal current at 165°E, 140°W and 11O°W was more representative of the observations compared to MY,

### 2.4. Current Along 140°W

The 1988 mean PP and h4Y zonal current components throughout the. N] 3CC had differences that were less than 0.05 m $s^{-1}$ (not shown). At 7°N near the middle of the N] 3CC, simulated currents at 20 and 40 m were 0.15 m $s^{-1}$ larger than observations (Figure 6), and this feature was persistent each month. The observed current was almost uniform bet ween 20 and 80 m, in contrast to the large simulated shear (2x $10^{-3}$ $s^{-1}$). Differences between simulated and observed current decreased with depth and were nonexistent at the 20(-111 bottom of the NJ 3CC.

## 3. Discussion

### 3.1. Simulations

Vertical profiles of upper ocean zonal current and temperature at the equator are sensitive to the vertical mixing of momentum and heat. 1 'or example, the 1 3UC thickness is too small and the maximum speed is too large if the vertical turbulent momentum flux is too small [*McCreary*, 1981 ]. The situation is expected to be different in the NI3CC which is in geostrophic balance, although important nongeostrophic components exist [*Johnson et al.*, 19881.

That the MY model created increased mixing compared to the PP technique was illustrated by the smaller shear above the depth of the 1 3UC core speed (Figures 4a and 4b), deeper penetrat ion of the westward-flowing SI 3C (Figures 4a and 4b), and a more verticall y homogeneous temperature profile. (Figures 2 (110° W), 3a, anti 3b). Differences bet ween 1'1' and M Y renal currents were much larger within the 3UC and S1 3C compared to the NI 3CC (1 igures 5 and 6). The PP and MY current bias was significant at the equator but not at 7°N, and the rms difference between 1'1' and

MY currents was significantly larger at the equatorial sites compared to 7°N. Point-to-point differences between PP and MY EUC simulations were substantial (Figure 4c); however, integral characteristics of the EUC, such as vertical thickness and total transport, agreed within 15%.

## 3.2. Observations and Simulations

**3.2.1. Longitudinal Variations.** That the 1'1' zonal current and temperature along the equator were more representative of observations compared to MY was not a surprise because the PP scheme had been tuned for the equatorial Pacific [*Pacanowski and Philander*, 1981]. The 1'1' scheme favors regions where the current has large vertical shear. The 1987 mean maximum MB vertical shears, which were computed in the interval defined by current-meter depths, were. 0.5, 1.3 and 1.3x 10-2 $s^{-1}$ at 165°E, 140°W and 110°W, respectively. At 165°E, where the maximum vertical shear was nearly 60% smaller than at 140°W and 110°W, the 1'1' and MY zonal currents were approximately equal, in contrast to the situation at 140@W and 110°W where PP currents were more representative of MB (Figure 5). Also, at 7°N, 140°W where the maximum MB shear ($0.2x\ 10^{-2}\ s^{-1}$) was less than that at 165°E, the PP and MY zonal currents were the same (Figure **6).**

Comparison of observed and simulated near surface temperatures is sensitive to the longitude along the equator of the intercomparison test. The rms differences between 10-m observed and simulated temperatures in 1987 and 1988 were 1.5 and 2.4°C at 140°W and 110°W, respectively; the difference was significant. in addition, the correlation coefficient between observed and simulated 10-m temperatures was significantly higher at 140°W than at 110°W.

3.2.2. **Cold Bias.** In the approximate 50-m near surface layer on the equator at 140°W and 110°W, the simulated temperatures were lower than that observed during the 1987 El Niño but not during the 1988 La Niña (Figure 2). Why the cold bias had a year-to-year variation is beyond the scope of this report.

Possible causes of the cold bias are excessive upward advection of cold water, too-low heat flux into the ocean, excessive westward advection of cold water, and incorrect OGCM physics of the upper ocean heat budget. To reconcile some features of the temperature profile, the vertical

velocity was computed from the simulated cas!-west ancl north-soul}} components of velocity using the equation of continuity with zero vertical motion at the surface. The simulated maximum upwelling velocities at 140°W and 11 O°W (1 'igure 7) were about 1.5 times greater than that observed several years earlier [*Halpern and Freitag, 1987; Halpern et* **al., 1989],** which are referenced because vertical velocity was not measured during 1987 and 1988. Uplifting the simulated thermocline by a greater amount would lead to near surface temperatures that would be lower than observed, although we can not separate the relative contributions produced by upwelling and mixing of the uplifted thermocline.

Another factor that likely influenced the simulation of temperature, especiall y at 10 m, is the neglect of the surface renal current in calculating surface zonal wind stress, which is appl ied as a body force to the uppermost 1 O-m layer of the OGCM. '1 'he 17% variat ion in surface wind stress that is associated with the surface current [*Halpern*, 1988] is approximately the same as the uncertainty of the drag coefficient, and is sufficient to alter an OGCM simulation of SST by about 2°C [*Harrison et al.*, 1990]. The 1987-1988 average MY cold bias at 10 m at 11 O°W was 17% large.r in March-May than in September-November, which is consistent with the idea that the sulfate zonal current was not included in the computation of the surface wind stress.

We investigated whether salinity could be a factor contributing to the cold bias. Vertical profiles Of PP and MY annual mean salinities from the surface to 300 m for 1987 and ] 988 al 11O°W were similar. Salinity measurements [*Hayes et al.*, 1991] indicated stronger near surface salinity gradient and less saline near surface watt.r compared to simulations. This inhibits vertical mixing and contributes towards elevated near surface temperatures compared to those simulated.

3.2.3. **Warm Bias.** Within the equatorial thermocline at 1 65°E, 140°W and 11 O°W, simulated temperatures were higher than observed (1 'igure 2). Onc reason for the warm bias is that observed upwelling speeds at 150 m al 140°W [*Halpern et al.*, 1989] and 110°W [*Halpern and Freitag*, 1 987], which were not made during 1 987-1988, were greater than simulated values (1 'igure 7). '1'bus, the simulated thermocline would be deeper than observe.d. Another possible contribution to the warm bias is mixing created by the breaking of short-period internal gravity

**wave.s** [*Moum et al.*, 1992]. The OGCM employed in this analysis did not contain a parameterization of short-period internal waves.

### 3.3. Comparison With Other Studies

**Wc** are aware of results from one similar kind of investigation. The 1'1' and MY zonal current differences displayed in 1 'igure 4c are different than those computed by *Delecluse et al.* (Figure 3.3.2.c *in Stockdale et al. [ 1993])* from OGCM simulations with the *Gaspar et al.* [ 1990] turbulent kinetic energy, named "TKE", and 1'}' parameterizations for vertical mixing. in their result, the isoline of zero difference between 1'1' and TKE zonal currents was approximately linear from 300 m near 150°E to **25 m** near 120°W. This yielded a 0.7 $m\,s^{-1}$ difference at 60 m near 90°W, in contrast to the near zero difference between PP and MY currents at 60 m near 90°W (Figure 4c). In their result (Figures 3.3.2.a and 3.3.2.b in *Stockdale et al. [* 1993]), the TKE EUC core speed was *20%* greater than the PP value, in contrast to our investigation in which the MY EUC core speed was 10% less than the )'1' value (1 'igures 4a and 4b). The apparent disagreement between our results and those of *Delecluse et al. could* bc influenced by experimental design. Our OGCM had six additional layers above. 160 m. Also, different levels of complexity of turbulent kinetic energy models were employed. The MY 2..5 -level scheme contains three-dimensional diffusion and advection, and the 1.5-level TKE model dots not include the effects of horizontal shear and horizontal turbulent diffusion. *Smith and Hess [ 1993]* have shown that a change of the *Mellor and Yamada [1982]* model from level 2, in which advection and diffusion of turbulent kinetic energy arc ignored, to level 2.5 significant] y alters the mixing regime. Differences between our analyses and those by *Delecluse et al.* need further investigation.

## 4* conclusion

At the equator, 1'}' and MY simulated current and temperat urc had three generic differences: (i) MY mixing in the near surface layer was stronger, (ii) MY 1 EUC core speed was dccpcr, and

(iii) MY SEC had a larger vertical thickness. No substantial difference was found between PP and MY simulations of integral properties of the current and temperature. Between 5°N and 9°N, PP and MY simulations of temperature and current were very similar.

Three generic features distinguish our simulations and observations: (i) warm bias occurred in the thermocline and a cold bias occurred in the near surface layer, (ii) temperature was better simulated than current, and (iii) agreement between PP and MB currents and temperatures at the equator was slightly better than that between MY and MB.

Temperature was not as good a discriminator as current because simulated and observed temperatures were more highly correlated compared to currents. The correlation coefficients between simulated and observed currents and temperatures at the three equatorial sites were 0.65 and 0.94, respectively, and the difference was significant. Also, in the NECC the correlation coefficients between simulated and observed currents and temperatures were 0.81 and 0.94, respectively, which were significantly different.

The PP and MY schemes have a number of adjustable parameters, and the simulation sensitivity to a range of values should be tested. That MY temperatures were more uniform with depth in the near surface layer compared to those of PP indicates the PP minimum level of mixing was too small; a similar conclusion was reached by *Richardson and Philander* [1987].

In addition to further attention devoted to vertical mixing, parameterization, increased accuracy of simulated current and temperature require high frequency and high wavenumber components of the surface wind field, parameterization of internal gravity wave motion, and representation of salinity flux. Studies of the sensitivity of the EUC to the parameterization of horizontal mixing by turbulent processes are warranted because the maximum speed of the EUC is 25% greater for a doubling of the horizontal eddy viscosity coefficient [*Pacanowski and Philander*, 1981]. However, it is a tenet of faith that assimilation of observations will mitigate imperfect surface wind stress and OGCM physics and will yield improved simulations of subsurface temperature and current.

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1 'igure 1. Annual mean PP and MY temperature difference along the equator during 1988. Contour interval is 0.5°C, and negative values (dashed line) mean that the PP temperature was smaller than that of MY. Arrows show locations of moored-bmy observations along the equator.

1 'igure 2. Vertical distributions of annual mean 1'1' and MY simulated and MB observed temperatures along the equator at 165°E, 140°W and 11 0°W for two (a) 1987 and (b) 1988.

1 'igure 3. Simulations of annual mean temperature along 140°W during 1988: (a) PP and (b) MY. (c) PP and MY temperature difference computed from (a) and (b), and negative values mean that the 1'1' temperature was smaller than MY. Contour interval of (a), (b) and (c) is 1 °C. (d) to (f) Vertical dist ributions of June-l)cccmbcr 19881'1' and MY simulated and MB observed temperatures along 140°W at (d) 5°N, (c) 7°N, and (f) 9"N.

1 'igure *4.* Simulations of annual mean renal current (eastward is positive, solid line; westward is negative, dashed line) along the equator during 1988: (a) 1'1' and (b) **MY.** (c) PP and **MY** difference computed from (a) and (b), anti positive values (solid lines) mean that the PP current was greater than that of MY. Contour interval of (a), (b) and (c) is 10 cm $s^{-1}$.

**1** 'igure **5.** Vertical distributions of annual mean PP and MY simulated and M B observed zonal current component along the equator at 165°E, 140°W and 11 O°W for (a) 1987 and (b) 1988. Posit ive values arc eastward and negat ive value.s arc westward.

Figure 6. Vertical distributions of June.-] )ecember 19881'1' and MY simulated and M B observed current at 7°N, 140°W. Positive values are eastward and negative values are westward.

1 'igure **7.** Vertical distributions of 1987-1988 mean 1'1' and MY vertical velocities along the equator at 165°E, 140°W and 11O°W. Positive values are upward.

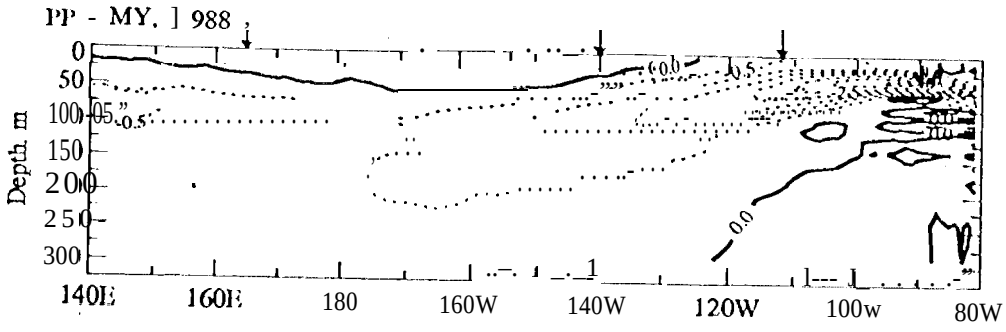

PP - MY. ] 988
Depth m
0
50
100
150
200
250
300
140E
160E
180
160W
140W
120W
100w
80W
0.0
0.5
0.5
0.0

Figure ]

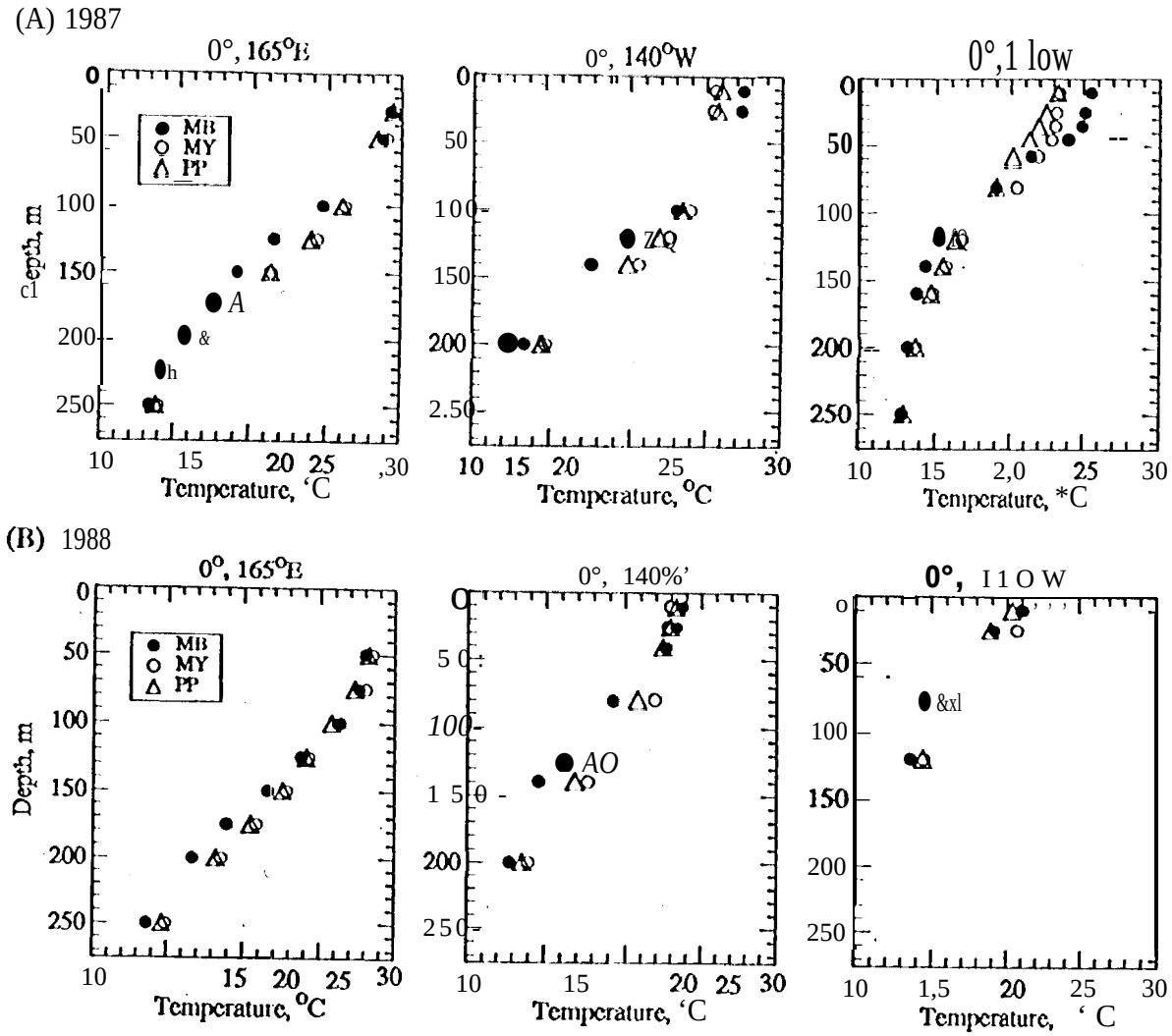

(A) 1987
0°, 165°E
0°, 140°W
0°,1 low
MB
MY
PP
Depth, m
Temperature, °C
(B) 1988
0°, 165°E
0°, 140%'
0°, I1OW
MB
MY
PP
Depth, m
Temperature, °C
Temperature, 'C
Temperature, ' C

Figure 2

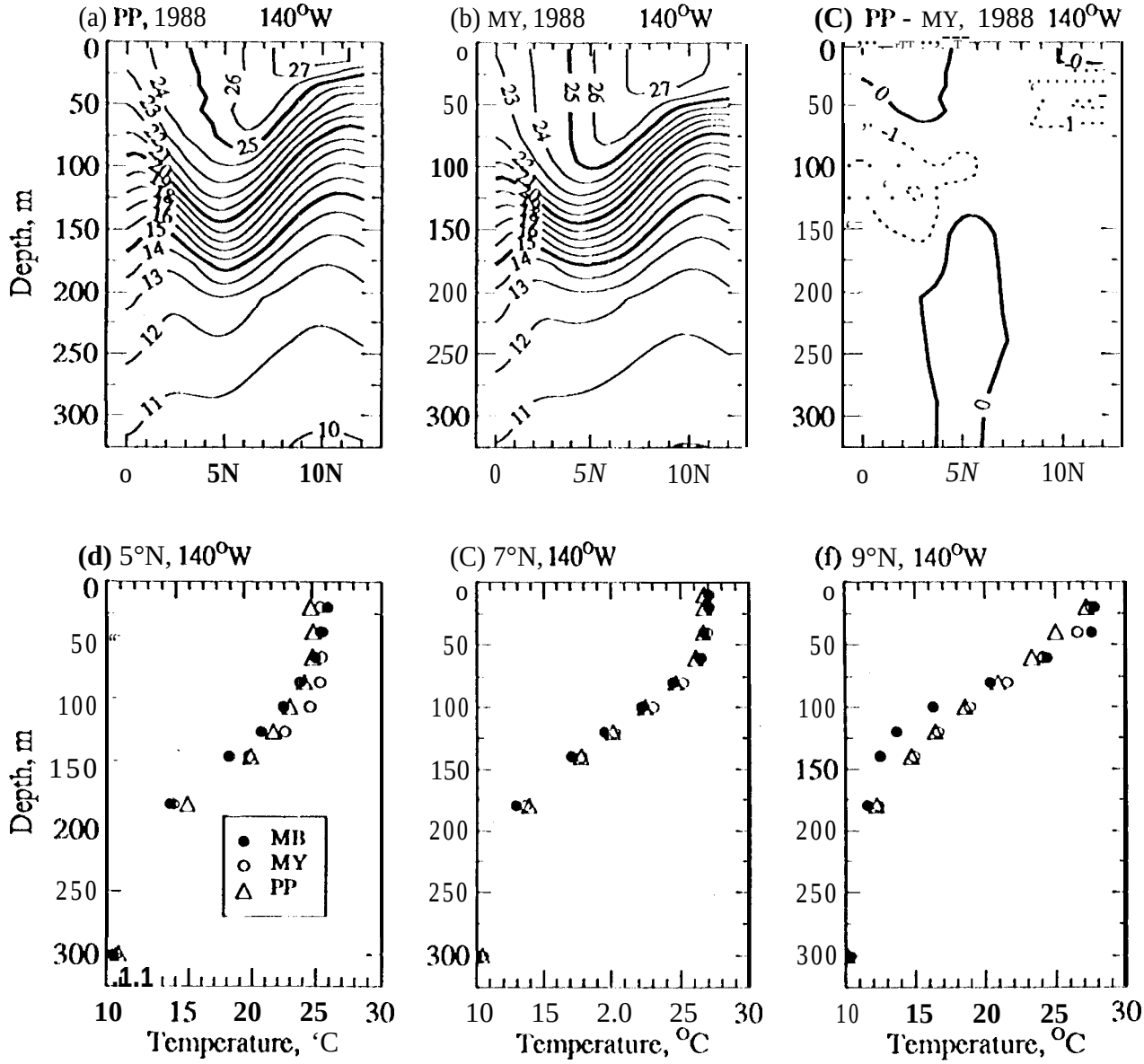
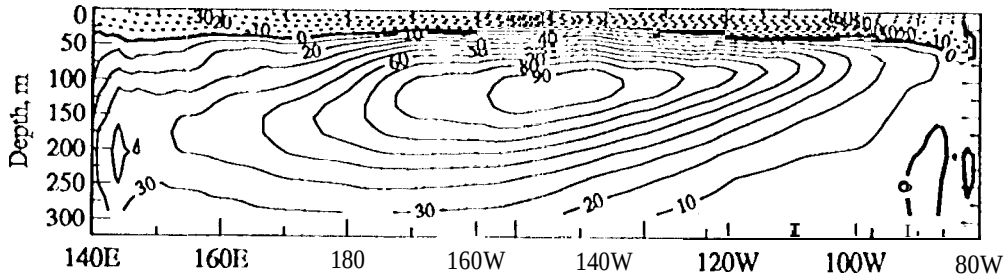
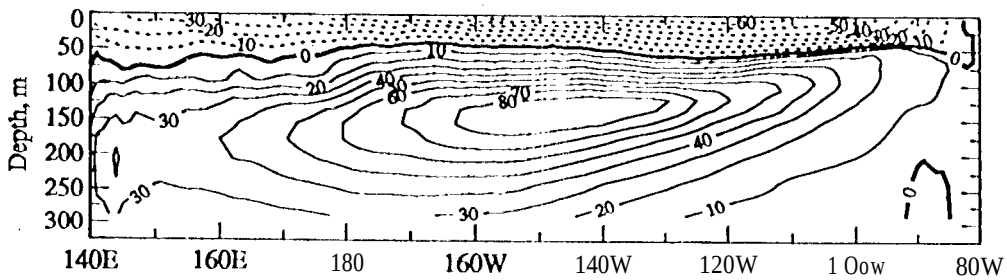

(a) PP, 1988 140°W
(b) MY, 1988 140°W
(C) PP - MY, 1988 140°W
Depth, m
0
50
100
150
200
250
300
o
5N
10N
(d) 5°N, 140°W
(C) 7°N, 140°W
(f) 9°N, 140°W
MB
MY
PP
10
15
20
25
30
Temperature, °C

Figure 3

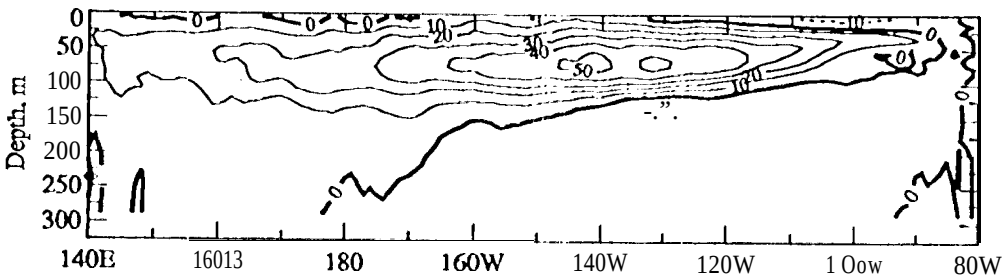
(a) P}', 1988




(b) MY, 1988




(c) PP - MY, 1988




F'i gin-e 4

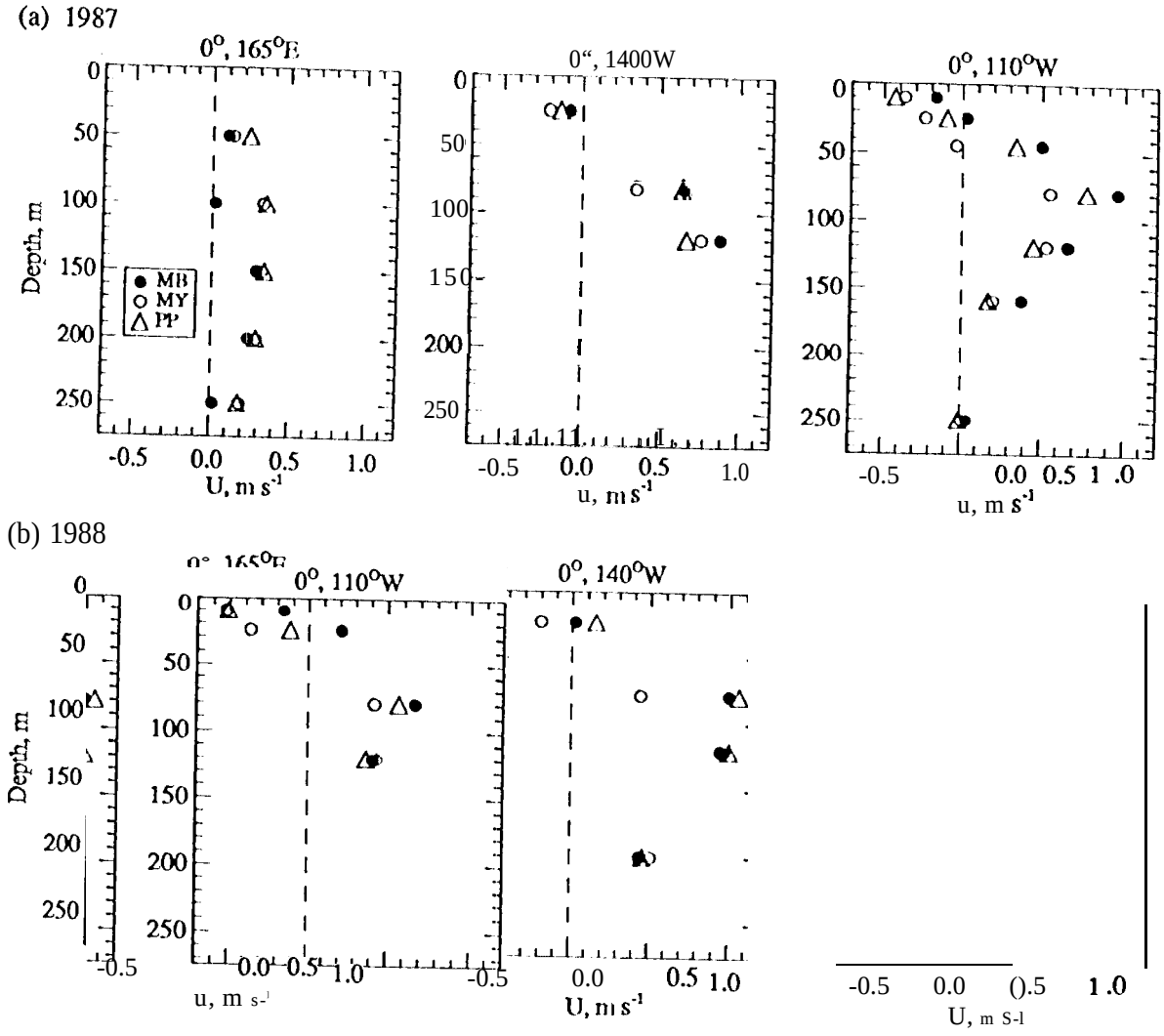

(a) 1987
0°, 165°E
0°, 140°W
0°, 110°W
Depth, m
MB
MY
PP
U, m s-1
u, m s-1
(b) 1988
0°, 110°W
0°, 140°W
u, m s-1
U, m s-1
U, m S-1

Figure 5

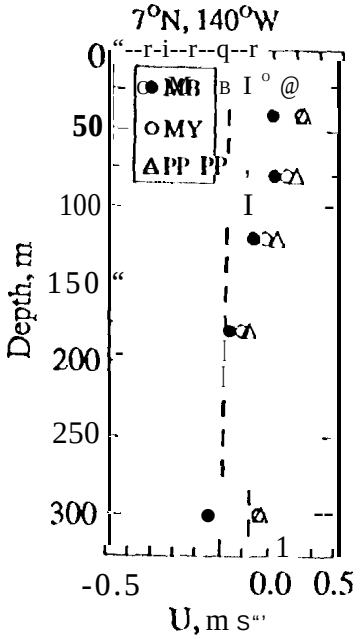



Figure 6

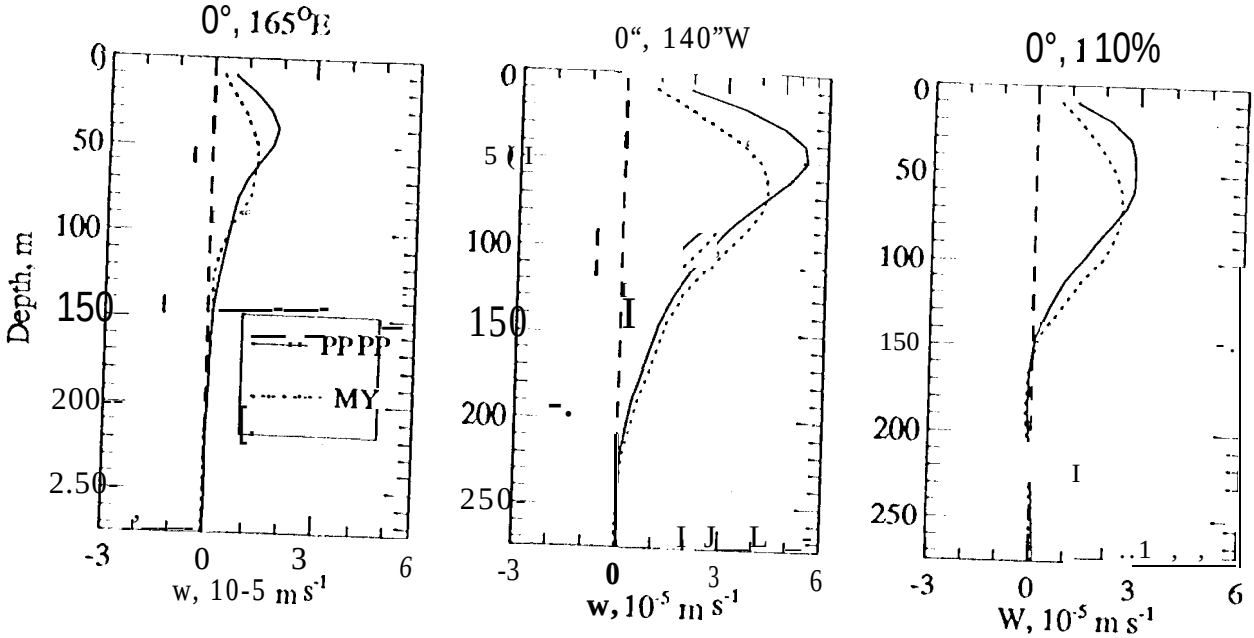

0°, 165°E
0°, 140"W
0°, 110%
Depth, m
0
50
100
150
200
250
PPPP
MY
-3
0
3
6
w, 10-5 m s-1
W, 10-5 m s-1

Figure 7