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SIMULATION OF THE COSTA RICA DOME IN THE EASTERN TROPICAL PACIFIC

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ABSTRACT

General circulation models with fine resolution are used to analyze seasonal and interannual changes of the Costa Rica Dome off Central America. It is found that the Costa Rica Dome with a cyclonic circulation grows rapidly in late boreal spring off the Gulf of Papagayo and matures in summer and early fall in accord with the northward migration of the ITCZ. The Costa Rica Dome is thus maintained by the winds with the cyclonic wind stress curl in boreal summer. In boreal winter strong jet-like northers converging the southernmost ITCZ from three passes in Central America excite warm anticyclonic nonlinear eddies. The Costa Rica Dome is eroded by those warm anticyclones identified as the nonlinear IG eddies (Yamagata, 1982; Williams and Yamagata, 1985).

The model for interannual changes demonstrates that the Costa Rica Dome cannot be generated well in some years like 1981, 1983 and 1988 because of lack of the elements of formation due to a larger scale climate variabilities possibly related to ENSO. The model also shows an ability to capture several interesting synoptic events. For example, the Geosat altimetry data shows that an unusually strong anticyclone was generated in the Gulf of Tehuantepec in October, 1987 and then moved westward keeping its identity. This synoptic event is clearly resolved in the model but the strength of the simulated eddy is less than observed one. This is partly due to adoption of monthly mean winds which smooth out strong gustiness of the northers.

INTRODUCTION

The eastern tropical Pacific is one of the most interesting regions in the world ocean. It is certainly attractive to climate modelers in view of the large-scale ocean-atmosphere interaction which incurs the devastating El Niño along the coast of South America. It is also highly attractive to geophysical fluid dynamicists as well as marine biologists because of formation of mysterious dome-like area of upwelling off the coast of Costa Rica (Cromwell, 1958).

The Costa Rica Dome, which is known as an important area of a tuna fishery, is a persistent nutrient-rich region off Costa Rica where the strong tropical thermocline reaches near the surface due to active oceanic upwelling (Wyrtki, 1964; Broenkow, 1965; Thomas, 1979; Hofmann et al., 1981). The center of the dome is located near 8 - 10° N, 88 - 90° W, of which diameter varies from 200 to 400 km. According to Wyrtki (1964), the upwelling in the dome is maintained by the cyclonic circulation composed of the Equatorial Countercurrent in the south, the Costa Rica Coastal Current in the east, and parts of the North Equatorial Current in the north. That is, the Costa Rica Dome is considered to be one component of the northeastern Pacific circulation system. Hofmann et al. (1981) focussed on the seasonal onset of the phenomenon using a reduced-gravity ocean model driven by mean monthly wind data prepared by the group of Florida State University (Goldenberg and O'Brien, 1981). According to their conclusion the dome is generated locally by a cyclonic wind stress curl associated with the northward migration of the ITCZ in late spring and early summer and weakens in conjunction with southward retreat of the ITCZ in fall. In other words, they assume that the evolution of the Costa Rica Dome is a consequence of a local response to seasonal winds.

The eastern tropical Pacific is noted also as a unique region in meteorology (Hurd, 1929; Roden, 1961; Stumpf, 1975; McCreary et al., 1988; Clarke, 1988). In winter, the atmospheric southward pressure gradient which develops between the Gulf of Mexico and the Pacific region drives strong jet-like northers through three passes (the Isthmus of Tehuantepec, Nicaragua's lake district and the Panama Canal) in Central America (Fig.1). Roden (1961) described the response of the

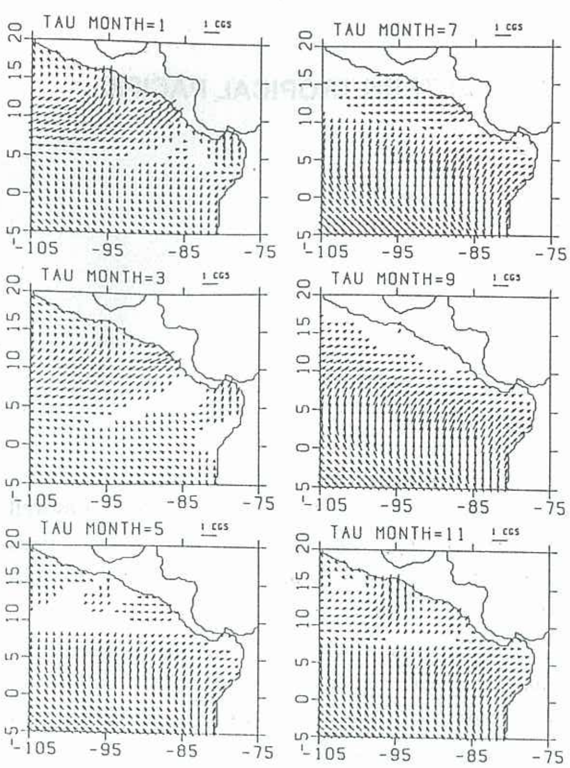

Figure 1: Revised Hellerman-Rosenstein monthly mean wind stresses used in the present work. Three winter northers through three passes in Central America are well resolved.

ocean to a Tehuantepecer: one of the above northers. The winds move the water to southward causing considerable mixing along the wind axis due to entrainment of the water from the sides and below. This process cools the SST by several degrees not only in the Gulf of Tehuantepec but also farther offshore.

Recent satellite infrared imagery of SST has made up the lack of systematic oceanographic observations both in space and time. Stumpf (1975), Stumpf and Legeckis (1977) and Clarke (1988) reported the evolution of SST associated with wind-induced upwelling in the winter near the Gulf of Tehuantepec and the Gulf of Papagayo, which is consistent with Roden's cruise observations. In particular, they all noticed the active anticyclonic eddy formation along the western edge of the jet-like northers. McCreary et al. (1988) tried to explain this preference of the anticyclonic eddy in terms of a mixed-layer physics. Stumpf and Legeckis (1977) concluded that the anticyclonic eddies propagating westward are generated by local winds from November through March. Matsuura and Yamagata (1982) introduced the IG dynamics to explain the longevity of those anticyclonic eddies.

The lack of unified understanding on those phenomena off Central America may be partly attributable to sparse hydrographic observations, which only provide low resolution images in both time and space scales, as contrasted with the rather detailed satellite images of SST and altimetry. Also, it may be partly attributable to synoptic, seasonal and interannual changes of the oceanic condition itself. As Hofmann et al.(1981) pointed out, the seasonal migration of the Intertropical Convergence Zone (ITCZ) is at least responsible for those oceanic changes.

In the present study, as an extension of the recent work of Umatani and Yamagata (1991), we try to obtain a comprehensive picture of synoptic, seasonal and interannual variations in the eastern tropical Pacific by use of high resolution ocean circulation models (originally supplied by GFDL) forced by realistic winds.

MODELS

Here we summarize our model parameters. The model resolution is 0.25 degree in both longitude and latitude for the limited area model and 0.5 degree for the Indo-Pacific basin model. There are 20 levels in the vertical in both models and the actual topographic and geometrical data supplied by GFDL are fitted. The coefficient of lateral eddy viscosity is 2×10^7 cm^2s^{-1}, whereas the coefficient of lateral eddy diffusivity is 1×10^7 cm^2s^{-1}. The vertical eddy viscosity and diffusivity are based on the formulae given by Pacanowski and Philander (1981). The initial condition is the climatological annual mean of temperature field (Levitus, 1982). Monthly mean climatological winds of Hellerman and Rosenstein (1983) are used in the limited area model, whereas the ECMWF winds are used in the Indo-Pacific basin model.

RESULTS FROM THE LIMITED AREA MODEL

Seasonal Cycle

As is well known, there are three major components of the lateral surface circulation in the eastern tropical Pacific (Fig.2). Those are the westward South Equatorial Current (SEC) near the equator, the eastward North Equatorial Countercurrent (NEC) near 5 °N and the westward North Equatorial Current (NEC) north of 10 °N. The currents show high seasonal variations due to the seasonal changes of the surface winds which are associated with the meridional migrations of the ITCZ.

From May through December, the northward trade winds are intense. In particular, they are most intense during the boreal fall because the ITCZ is located farthest north at that time. From January through April, however,

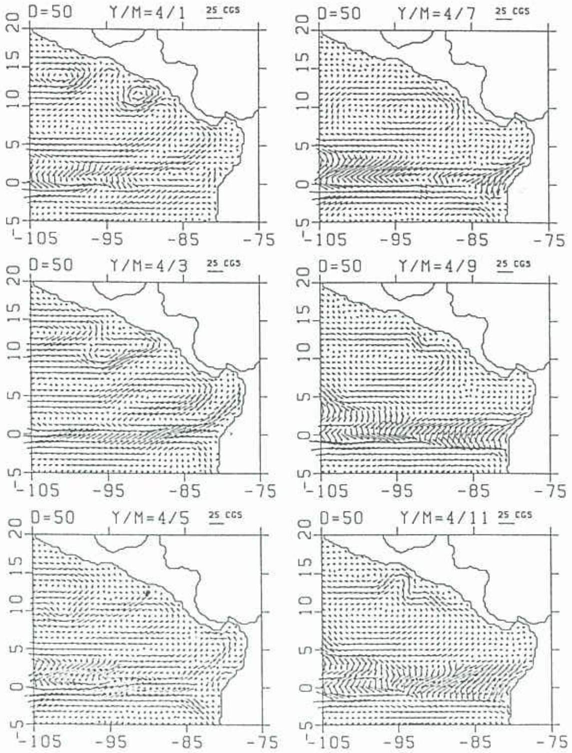

Figure 2: Simulated annual cycle of horizontal velocity vectors at a depth of 50 m

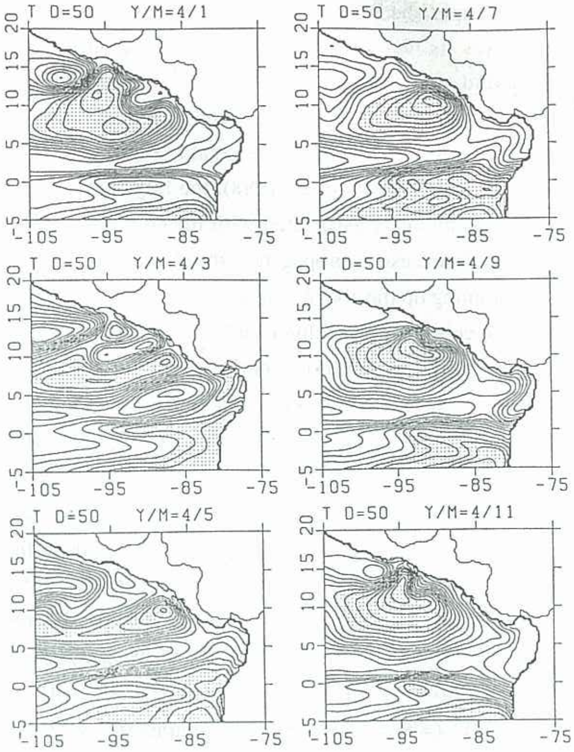

Figure 3: Simulated annual cycle of subsurface temperatures at a depth of 50 m. Contour interval is 1 °C. The temperature less than 20 °C is shaded.

these trade winds relax with the southward retreat of the ITCZ. In accordance with those changes of winds, the NECC intensifies in summer and fall, reaching the eastern end of the Pacific basin (see also Cane, 1979; Philander et al., 1987). During winter and spring, however, the NECC relaxes and shows high variability.

From December through March, winds from the Gulf of Mexico are persistent and intense north of 5 °N. In particular, three passes in the Central America are clearly resolved in the Hellerman-Rosenstein winds (1°×1° grid resolution) as shown in Fig. 1. Those strong jet-like northers are associated with three dipole-like structures of the wind stress curl and excite three noticeable anticyclonic circulations associated with comparatively weak cyclonic circulations off the coast of Central America. Thus, in winter the westward current near 10 °N shows a complicated meandering pattern.

The cold Costa Rica Dome evolves in boreal spring off the Gulf of Papagayo and matures in summer and early fall (Fig. 3). In late fall and winter the dome weakens and the more variable eddies forced by the northers take the place. The center of the simulated dome is located near 10 °N, 90 °W and the temperature in the core is less than 26 °C. The diameter is about 500 km in early summer although in fall the dome elongates in the zonal direction. These values are remarkably consistent with those observed. It is noteworthy that the Costa Rica Dome becomes prominent as the ITCZ reaches farthest north in the boreal fall .

The present limited area model demonstrates that the global northward trade winds in boreal summer which converge into the ITCZ are most responsible for the maintenance of the Costa Rica Dome. Those winds in summer are associated with rather weak curl. In addition, the center of the dome is located about 500 km west of that of the weak wind stress curl. Those results apparently show a contrast to the conclusion of Hofmann et al. (1981), in which the localized cyclonic curl of the FSU wind stress data in summer was identified as the principal mechanism responsible for upwelling in the Costa Rica Dome.

Warm IG Eddies in Boreal Winter

The annual cycle of the subsurface currents and temperatures at a depth of 50 m shows clearly that three warm anticyclonic eddies exist off the coast of Central America in winter although the curl of the wind stress has a dipole structure above the three regions (Figs. 2 and 3) Those anticyclonic eddies are confined above the sharp thermocline which exists at a depth of about 100 m. In addition, they propagate westward faster than the long Rossby wave speed as estimated for the same region.

For example, the Papagayo eddy generated in December preserves its identity for a few months and propagates westward at a speed of about 0.1 m s^{-1} which is almost twice as large as the long Rossby wave speed. As suggested in Matsuura and Yamagata (1982) and more recently in McCreary et al.(1988), the large upper layer thickness anomaly associated with the eddy may partly explain the excessive propagation speed.

Summing up the above, it is quite reasonable to expect that the coherent anticyclonic eddies generated in winter off the coast of Central America are in a dynamical balance between the planetary wave dispersion and the nonlinear planetary geostrophic divergence associated with the order-one layer thickness change. In other words, they may be governed by the same intermediate geostrophic (IG) dynamics as applied to the ovals in the Jupiter's atmosphere (Williams and Yamagata, 1985). Actually, an approximate estimate for the set of three nondimensional parameters β^*, $\varepsilon^*$and s^* using a β-plane approximation gives ($\beta^*=0.1$, $\varepsilon^* = 0.02$ and $s^* = 0.1$), where β^* ($=\beta L f_0^{-1}$) is the beta parameter with the Coriolis parameter f_0 evaluated at a reference latitude, ε^* ($=U f_0^{-1} L^{-1}$) is Rossby number and s^* ($= L_R^2 L^{-2}$ where L_R is the deformation radius $\sqrt{g^* H} f_0^{-1}$ with the reduced gravity g^* and the mean depth H of the upper layer) is the stratification parameter. The above values satisfy the unique conditions:

$$s^* \sim S\,\beta^*,\ \varepsilon^* \sim E\,\beta^{*2}, \qquad (1)$$

where S, E are numbers of O(1). The above conditions lead to the IG dynamics which is governed by the IG equation:

$$\eta_T - \eta_X - \beta^*(ES^{-1}\eta\,\eta_X + S\Delta\,\eta_X - 2\,y\,\eta_X - E\,J(\Delta\,\eta,\eta)) = 0, \qquad (2)$$

where J denotes the Jacobian operator, η (scaled with $LUf_0 g^{*-1}$) measures the interface displacement from the mean depth H of the upper layer, time T is scaled by $f_0^{-1} s^{*-1} \beta^{*-1}$ and X (scaled with L) denotes a coordinate in the zonal direction. A remarkable property of the above equation is that only warm, anticyclonic eddies are long lived (if dissipation is negligible) due to the balance between the scalar nonlinearity due to finite thickness changes of interface (the first term in the bracket) and the planetary wave dispersion (the second term in the bracket).

Heat Budgets off the Coast of Costa Rica

Let us consider two artificial boxes A and B in the upper 60 m off Central America. The box A has a rectangular domain bounded by 86.5 °W, 88.5 °W, 8 °N and 10 °N, which covers the cyclonic circulation localized off the Gulf of Papagayo. The other box B is also a rectangular domain bounded by 89 °W, 91 °W, 9 °N and 11 °N, which covers the central part of the Costa Rica Dome during its mature stage. Since the dome is a shallow phenomena, the adopted depth of the two boxes is enough to calculate the budgets. The results for the heat budget are shown in Fig. 4.

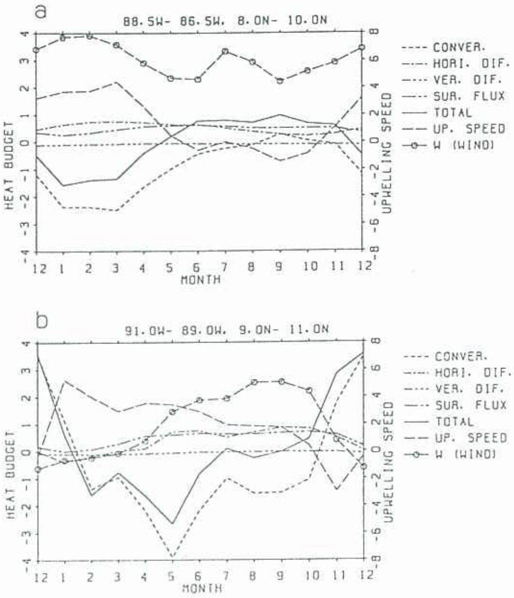

Figure 4: (a) Monthly mean values of the heat budget for the Box A (in 10^{12} cal/sec). The rate of change of heat storage is determined by the convergence of heat transport, flux across the surface, horizontal and vertical diffusion. The modelled upwelling speed (in 10^{-4} cm/sec) at a depth of 60 m and the contribution of the wind stress curl are also shown. (b) Same as in (a) but for the Box B.

The temperature averaged within the nearshore box A drops most sharply in January at a rate of -1.4 °/month mainly due to the active divergence of the heat transport. It is clearly seen that the divergence is related to active upwelling in boreal winter. The contribution of the wind stress curl (due to local forcing) is about three times larger than the model upwelling velocity. This means that another contribution of the planetary geostrophic convergence (due to remote forcing) cancels two thirds of the local wind forcing. This planetary effect is associated with westward movement of a tilted thermocline shoaling westward. In other words, the westward propagation of long Rossby waves cannot be negligible. The active cooling in the box A lasts from December through April while strong northers prevail off Costa Rica. From late spring through late summer, however, the situation changes totally because of a seasonal stop of local upwelling. The surface temperature begins to increase due to surface heating and lateral diffusion and completes the annual cycle.

The temperature averaged within the offshore box B begins to drop in early spring and the rate reaches the maximum in May at -2.4 °/month due to the strong divergence of the heat transport . In contrast to the box A, the contribution of the local wind stress curl is negligible in winter and spring. In addition, the local wind stress curl does not show any seasonal maximum in May. Therefore it is reasonable to expect that a contribution of the planetary geostrophic divergence (due to remote forcing) plays a major role in the upwelling process. This planetary effect is associated with westward movement of a tilted thermocline shoaling eastward. The local upwelling due to the positive wind stress curl revives within the box B from summer through fall in accord with the northward migration of the ITCZ. However, the heat storage within the box B remains almost the same. This is because the cooling due to the upwelling balances with surface heating and lateral diffusion. The Costa Rica Dome is now in a steady state. From November through January, a warm water accumulates in the box B and thus reduces the cold Costa Rica Dome. In particular, the dome is warmed at a rate of 3.2 °/month in December and reaches the maximum temperature of 25.5 ° C in January. This warming is mainly associated with the lateral convergence of the heat transport due to propagating warm anticyclonic eddies originally excited near the coast of Costa Rica..

Comparison with Climatological Data

Figure 5 compares the subsurface temperature field at a depth of 50 m produced in the standard run and the climatological field at the same depth prepared by Levitus (1982). The correspondence of the model results with the data is rather good except for the period of February - April. In particular the cold Costa Rica Dome in summer and fall is simulated well with regard to the lateral scale and location. The minimum temperature at the center of the simulated dome, however, is lower than the climatology by two or three degrees. This is mainly because the model surface heat flux is too small compared to the available estimates. For example the mean surface heat flux calculated by Reed (1983, 1985) ranges from over 75 W m^{-2} to around 100 W m^{-2}, almost twice as large as the model value. In order to simulate more reasonable values of the sea temperature, the flux formula, as well as the mixed-layer parameterization, needs to be refined.

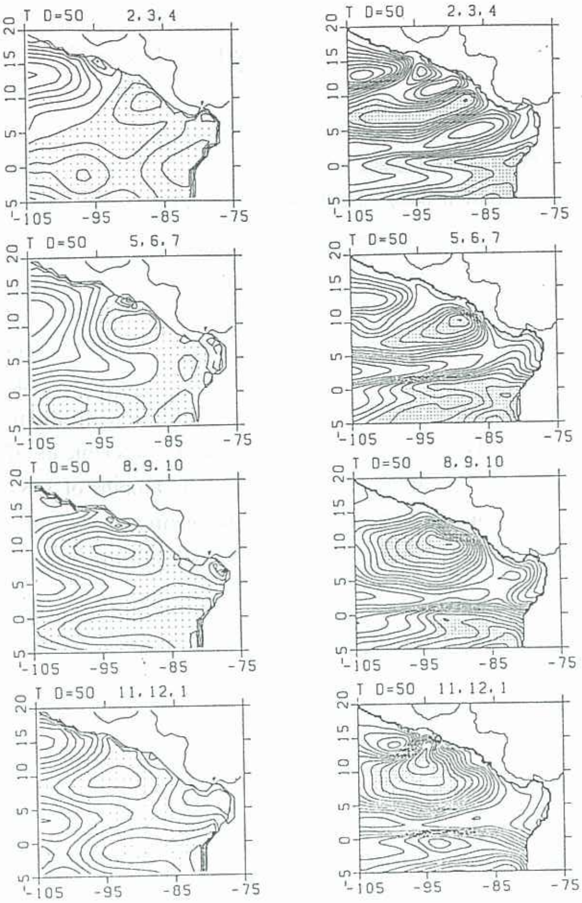

Figure 5: Subsurface temperatures averaged for three months at a depth of 50 m in four seasons. Left panel: the present model simulation. Right panel: Levitus climatology. Contour interval is 1 °C. Temperatures less than 20 °C are shaded.

The correspondence of the model results with the Levitus climatology is rather poor for February - April. In particular it looks terrible off the Gulfs of Papagayo and Panama. It has turned out, however, that the discrepancy is not due to a model fault but rather due to the low resolution of the climatology data. Actually, using the global area coverage (GAC) data of the NOAA polar orbiting satellite, Legeckis (1988) recently summarized the positions of the fronts off Central America from March 7 to 20, 1985. It is remarkable how well the location of the fronts corresponds with our model product for February - April.

RESULTS FROM THE BASIN MODEL

So far we have analyzed the seasonal cycle of the ocean circulation off the Costa Rica using a fine resolution limited area model. It is also interesting to analyze synoptic and interannual changes of the oceanic conditions off the Costa Rica. The model used here is the Indo-Pacific basin model with 0.5 degree resolution, which was driven by the ECMWF winds with 2 degree resolution.

Interannual Changes

Figure 6 shows the subsurface temperatures at a depth of 57 m on September 15 in each year from 1981 through 1988. The model shows strong interannual changes in the eastern tropical Pacific. In particular, the Costa Rica Dome develops well in 1982, 1986 and 1987. The years 1981, 1983 and 1988 are bad years for the dome since the dome cannot be identified in the subsurface temperature field. This seems to be due to lack of the elements of formation as follows. During the early months of 1981 the oceanic condition was unusually warm and upwelling of cold water was not so active. In 1983, the oceanic condition was also warm in winter in the eastern Pacific because of the big El Niño. In addition, the winter northers could not last long. In 1988, both oceanic and atmospheric conditions in winter were favorable for the evolution of the dome but unusual winds which favor downwelling persisted off the coast of Costa Rica during the summer.

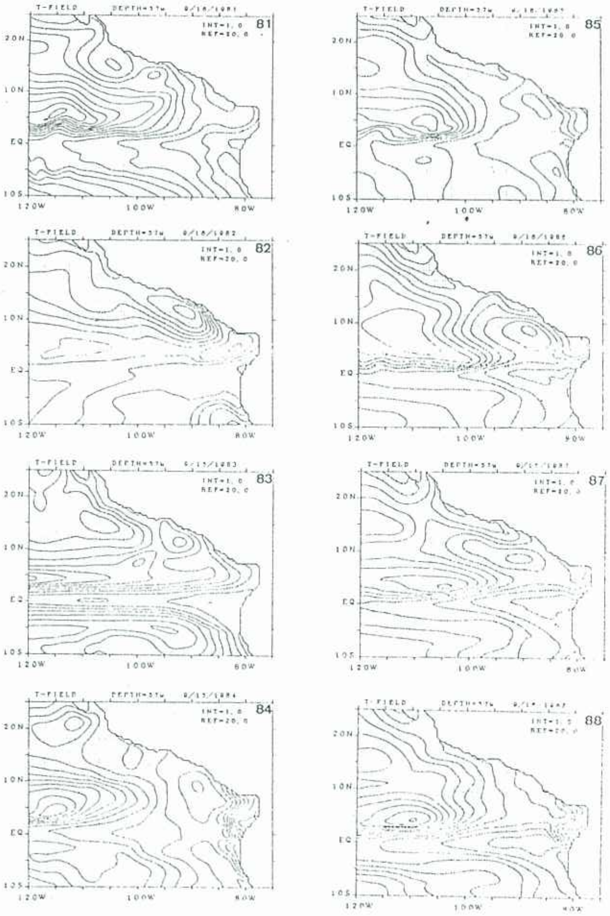

Figure 6: Simulated subsurface temperatures at a depth of 57 m on September 15 in each year from 1981 through 1988. Contour interval is 1 °C. Temperatures less than 20 °C are shaded.

Some Synoptic Events

Although the present basin model for the interannual changes is less fine than the limited area model, it is still fine enough to capture prominent synoptic events of eddy formation off the coast of Costa Rica. As an example we focus on a particular event which occurred in the Gulf of Tehuantepec in 1987. A sequence of the dynamic height fields calculated from the model clearly shows that a circular dynamic height anomaly (which suggests an anticyclonic eddy) was generated by unusual northers in the Gulf of Tehuantepec in October, 1987. The anomaly then evolved and expanded westwards as a linear long Rossby wave (Fig. 7). A sequence of GEOSAT altimetry data processed with the method of collinear analyses (Shibata and Kitamura, 1990) also shows an evolution of the same anticyclone with surface elevation of almost double strength (not shown). The anticyclone, however, kept the circular shape and moved westward with a phase speed of about 15 km/day which is faster than the long Rossby wave speed (Matsuura and Yamagata, 1982; McCreary et al., 1988; Umatani and Yamagata, 1991). The anticyclone was also tracked by the EPOCS drifting buoy 6858 for a certain period (Hansen and Maul, 1991).

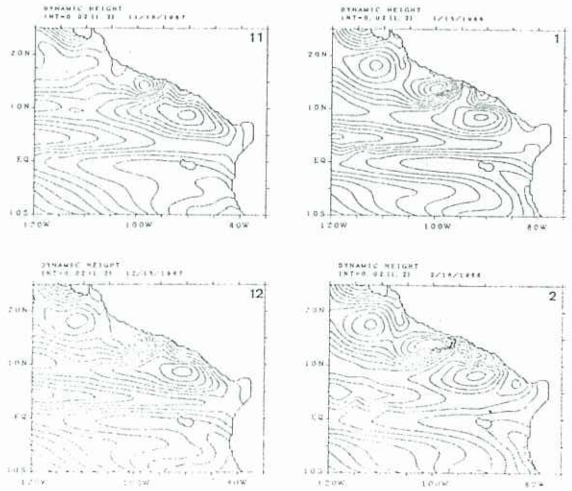

Figure 7: Simulated dynamic height fields (reference level: 854 m) from November 1987 through February 1988. Contour interval is 0.02 dynamic meter. Values less than 1.2 dynamic meter are shaded.

As demonstrated using the intermediate geostrophic equation with a wind forcing (Yamagata et al., 1990), winds must be strong enough to excite the nonlinear IG eddies. Actual northers are very gusty like cold surges but the monthly mean winds we have adopted smooth out the strong gustiness effect. The present model results suggest that those sharp wind fluctuations with a period less than a month cannot be neglected in modeling the formation of nonlinear synoptic eddies.

SUMMARY

In order to obtain a coherent picture about how things evolve in the northeastern tropical Pacific, we have developed a regional ocean circulation model with fine horizontal resolution of 0.25° × 0.25° for a seasonal cycle and a Indo-Pacific basin model with less fine horizontal resolution of 0.5° × 0.5° for interannual changes. The seasonal model was forced by the Hellerman-Rosenstein winds and the interannual model was forced by the ECMWF winds. Both models are initialized by use of the Levitus climatology of temperature field. Several significant results enough to organize the past literature and to stimulate further studies are as follows.

The model Costa Rica Dome of which center is located near 10 °N, 90 °W and grows rapidly in late spring off the Gulf of Papagayo and matures in summer and early fall in accord with the northward migration of the ITCZ which strengthens the NECC. In winter strong northers converging the southernmost ITCZ from three passes in Central America excite three noticeable warm anticyclonic eddies confined in the upper layer. The anticyclones are identified as the IG eddies as suggested by Matsuura and Yamagata (1982). The Costa Rica Dome is eroded by those propagating warm eddies but, at the same time, a new embryo of the dome begins to evolve from the propagating cold eddies excited by the same northers. Without the embryo, the Costa Rica Dome cannot be generated by the winds associated with the northward migration of the ITCZ in summer. The Costa Rica Dome may be classified into a planetary nonlinear mode, as discussed by Yamagata et al.(1990), rather than a linear response to the local wind stress curl.

The interannual changes in the formation of the Costa Rica Dome from 1981 to 1988 are analyzed using a less fine version of our general circulation model forced by the ECMWF winds. The model demonstrates that the Costa Rica Dome cannot be generated well in some years like 1981, 1983 and 1988 because of lack of the elements of formation due to a larger scale climate variabilities, such as ENSO. The model also shows an ability to capture several interesting synoptic events. For example, the U.S. Navy's Geosat altimetry data shows that an unusually strong anticyclone was generated in the Gulf of Tehuantepec in October, 1987 and then moved westward keeping its identity. This synoptic event is clearly resolved in the model but the strength of the simulated eddy is less than observed one. This is partly due to adoption of monthly mean winds which smooth out strong gustiness of the northers.

REFERENCES

BROENKOW, W (1965) The distribution of nutrients in the Costa Rica Dome in the eastern tropical Pacific Ocean. Limnol Oceanogr, 10, 40-52.

CANE, M A (1979) The response of an equatorial ocean to simple windstress patterns. J Mar Res, 37, 253-299.

CLARKE, A J (1988) Inertial wind path and sea surface temperature patterns near the Gulf of Tehuantepec and Gulf of Papagayo. J Geophys Res, 93, 15491-15501.

CROMWELL, T (1958) Thermocline topography, horizontal currents and "ridging" in the eastern tropical Pacific. Inter-Am Trop Tuna Comm Bull, 3, 135-164.

GOLDENBERG, S O and O'BRIEN, J J (1981) Time and space variability of tropical Pacific wind stress. Mon Wea Rev, 109, 1190-1207.

NANSEN, D V and MAUL, G A (1991) Anticyclonic current rings in the eastern tropical Pacific Ocean. J Geophys Res, 96, 6965-6979.

HELLERMAN, S and ROSENSTEIN, M (1983) Normal monthly windstress over the world ocean with error estimates. J Phys Oceanogr, 13, 1093-1104.

HOFMANN, E E, BUSALACCHI, A J and O'BRIEN, J J (1981) Wind generation of the Costa Rica Dome. Science, 214, 552-554.

HURD, W E (1929) Northers of the Gulf of Tehuantepec. Mon Wea Rev, 57, 192-194.

LEGECKIS, R (1988) Upwelling off the Gulfs of Panama and Papagayo in the tropical Pacific during march 1985. J Geophys Res, 93, 15485-15489.

LEVITUS, S (1982) *Climatological Atlas of the World Ocean.* NOAA Prof. Paper 13, 173 pp., 17 microfiche, U. S. Govt. Printing Office, Washington, D. C.

MATSUURA, T and YAMAGATA, T (1982) On the evolution of nonlinear planetary eddies larger than the radius of deformation. J Phys Oceanogr, 12, 440-456.

MCCREARY, J P, LEE, H S and ENFIELD, D B (1988) The response of the coastal ocean to strong offshore winds: with application to circulations in the Gulfs of Tehuantepec and Papagayo. J Mar Res, 47, 81-109.

PACANOWSKI, R and PHILANDER, S G H (1981) Parameterization of vertical mixing in numerical models of tropical oceans. J Phys Oceanogr, 11, 1443-1451.

PHILANDER, S G H, HURLIN, W J and SEIGEL, A D (1987) Simulation of the seasonal cycle of the tropical Pacific Ocean. J Phys Oceanogr, 17, 1986-2002.

REED, R K (1983) Heat fluxes over the eastern tropical Pacific and aspects of the 1972 El Nino. J Geophys Res, 88, 9627-9638.

REED, R K (1985) An estimate of the climatological heat fluxes over the tropical Pacific Ocean. J Climate Appl Meteor, 24, 833-840.

RODEN, G I (1961) On the wind driven circulation in the Gulf of Tehuantepec and its effect upon surface temperatures. Geofis Int, 1, 55-72.

SHIBATA, A and KITAMURA, Y (1990) Geosat sea level variability in the tropical Pacific in the period from November 1986 to February 1989, obtained by collinear method.Oceanogr Mag, 40, 1-25.

STUMPF, H G (1975) Satellite detection of upwelling in the Gulf of Tehuantepec, Mexico. J Phys Oceanogr, 5, 383-388.

STUMPF, H G and LEGECKIS, R V (1977) Satellite observations of mesoscale eddy dynamics in the eastern tropical Pacific Ocean. J Phys Oceanogr, 7, 648-658.

THOMAS, W H (1979) Anomalous nutrient-chlorophyll interrelationships in the offshore eastern tropical Pacific Ocean. J Mar Res, 37, 327-335.

UMATANI, S and YAMAGATA, T (1991) Response of the eastern tropical Pacific to meridional migration of the ITCZ: The generation of the Costa Rica Dome. J Phys Oceanogr, 21, 346-363.

WILLIAMS, G P and YAMAGATA,T (1985) Geostrophic regimes, intermediate solitary vortices and Jovian eddies. J Atmos Sci, 41,453-478.

WYRTKI, K (1964) Upwelling in the Costa Rica Dome. Fish Bull, 63, 355-372.

YAMAGATA, T (1982) On nonlinear planetary waves: a class of solutions missed by the quasi-geostrophic approximation. J Oceanogr Soc Japan, 38, 236-244.

YAMAGATA, T, SAKAMOTO, K and ARAI, M (1990) Locally-induced nonlinear modes and multiple equilibria in planetary fluids. PAGEOPH, 133, 733-748.