Effect of Low Latitude Western Boundary Gaps on the Reflection of Equatorial Motions

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ABSTRACT

The western tropical Pacific is thought to be an important zone for generating El Nino : reflections at the boundary make it a potential source region of equatorial Kelvin waves. Calculations of the effect of a gappy western boundary on the reflection process are carried out in the framework of the low frequency limit of the shallow water equations and are highly idealized. The method is also applied to a schematic version of the flow through the Indonesian seas from the western Pacific to the Indian oceans.

The results indicate some strong sensitivities to the location of the gap and to the structure of the incoming flows. In addition, the results can be quite different depending on whether the zonal extent of the gap is assumed to be infinite or finite. (More precisely, the latter means that the extent of the gap is short compared with the zonal wavelength of the relevant free waves at that frequency.)

Due to the complexity of the results for even such a simplified model, it will be very difficult to be confident of any modeling study of the Indonesian throughflow short of a highly resolved numerical calculation with a detailed representation of the geometry and bathymetry. Nonetheless, we offer tentative conclusions concerning the efficiency of the western Pacific boundary as a reflector. Our results suggest that the realistic boundary will not greatly alter expectations on a simple solid boundary if the reflections important for El Nino are primarily in motion represented by low order Rossby modes. This is also consistent with observational evidence which tends to support that there is no anomalous throughflow during El Nino events.

1. Introduction

Recent studies have recognized that ocean dynamics have an important role in development of the phenomena known as El Nino and the Southern Oscillation (ENSO). In theory as well as in observations, the influence of equatorial Kelvin waves has been established. For example, White et al (1985), studying the increase of heat content, claimed that extra-equatorial Rossby waves generated by anomalous wind curl, after impinging on western coasts, could generate an equatorial Kelvin wave which in turn would trigger an El Nino event. However Battisti (1988, 1989) tempers their conclusions and finds that off-equatorial waves (poleward of 6° from equator) Rossby waves could not be invoked as a triggering mechanism in his model, although he also points out the importance of western boundary reflections for event terminations. Zebiak (1989), in the context of a linear dynamical model, discusses the role of Rossby waves and western boundary reflections : some of the equatorial transport near the western boundary originates at higher latitude in the interior basin due to the structure of Rossby wave induced circulation, but the zonal convergence/divergence at the western boundary determines the nature of the Kelvin wave signal.

Wether of not extra-tropical influences are important depend on how well the western boundary can reflect disturbances into equatorial Kelvin waves. In oceanic models and coupled models as well, reflection of Rossby waves at the western boundary is the basic mechanism to overcome the feedback tending to raise sea level. In numerical models (with the exception of the NORDA model; Kindle et al, 1989), the western Pacific boundary is a nice smooth wall, but in reality, it consists of irregular island chains. Therefore, it could be a poor wave reflector, allowing Rossby waves to pass through its many passages.

Our motivation for the work was to know how efficient the Pacific western boundary is at generating equatorial Kelvin waves by reflection of mean flux incident at the boundary. How efficient is the "gappy" western boundary of the Pacific Ocean at generating equatorial Kelvin waves by reflections of incident mass flux ? Does the irregular, "porous", western boundary support the simple planetary wave reflections which occur in numerical models (with a full boundary) ?

We focus this investigation on the effects of leaky boundaries on basin wide adjustment processes rather than on details of boundary layers in the vicinity of coasts or gaps. Our approach uses extensively techniques and results of Cane and Sarachick (1977, 1981) and is an extension of Cane and du Penhoat (1982) and du Penhoat, Cane and Patton (1983). Calculations are carried out in the framework of the linear shallow water equations that we solve for the asymptotic flow which results when a source is switched on at $t=0$ and remains steady thereafter. The results of this the problem are easily generalized to all low frequency flows.

2. The basic problem

At $t=0$, a flow comprised of Rossby waves impinges on a western boundary at $x=0$. The flow has a $H(t)$ form, a step function in time. The coast is oriented north-South, thus presenting a gaps of width $d=a\text{-}b$. We are interested in low frequency motion for which frequency is small compared to the equatorial scaling frequency. We seek the asymptotic response as $t \to +\infty$. Cane and Sarachick (1976) have shown that for large t, the asymptotic motions are of three kinds (all variables have been scaled by the usual equatorial scaling) :

(i) Long eastward propagating Kelvin waves, with zonal velocity u and height h proportional to ψ_0, the zeroth order Hermite function :

$$(u,v,h) \sim \left[\psi_0(y),0,\psi_0(y)\right] \quad \text{with} \quad \psi_0(y) = \pi^{\frac{-1}{4}} e^{-\frac{y^2}{2}}$$

(ii) Long Rossby waves, propagating energy westward. Both the equatorial Kelvin wave and the long Rossby waves satisfy the meridional geostrophic relation.

The equatorial Kelvin wave has meridional velocity $v = 0$ and long Rossby waves have $v \approx 0$. For these two kinds of asymptotic motions, the large t response is steady and independant of x.

(iii) Short Rossby waves (included the mixed Rossby-gravity wave). Sum of such modes is an ever thinning boundary layer trapped to $x=0$ (see Cane and Sarachick, 1979), thus forming a western boundary current.

In this work, we exploit the fact that at low frequency there is a streamfunction for the sum of eastward propagating short Rossby waves and that the only part of the reflection that propagates away from the boundary is the equatorial Kelvin wave.

Let $(\hat{u},\hat{h})$ be the low frequency incoming Rossby waves. The flow statisfies the geostrophic relation and is orthogonal to the Kelvin wave.

Two conditions have to be satisfied at $x=0$, the longitude of the coast :

1) $u=0$ at the coast, i.e., for $y>a$ and $y<b$

2) u and h are continuous across the gap, i.e, at $b<y<a$

When the incoming wave hits the coast, part is transmitted into the gap and part is reflected into an equatorial Kelvin wave of amplitude T^k.

The continuity condition at the gap longitude and the geostrophic balance east and west of the gap imply that there is no short Rossby waves in the gap. It also implies that discontinuities in h are thus possible at $y=a$ and $y=b$. Since west of the boundary u and h are in geostrophic balance, we must allow for an infinite zonal velocity at $y=a$ $(y=b)$ to balance the jump in h, i.e.

$$u=A\,\delta(y-a)+B\,\delta(y-b) \text{ at } y=a,b$$

Integration of the continuity equation leads to :

$$A=\int_a^{+\infty}\hat{u}\,dy+T^k\int_a^{+\infty}\psi_0(y)\,dy \text{ and } B=\int_{-\infty}^{b}\hat{u}\,dy+T^k\int_{-\infty}^{b}\psi_0(y)\,dy$$

and the boundary currents at the tips of the gap are simply fed the boundary currents along the North-South coast. The amplitude of the reflected Kelvin wave is then obtained by using the completness and orthogonality properties of the eigenfunctions of the shallow water equations (cf. Cane and Sarachick, 1977, 1981) :

$$T^k=-\frac{\int_a^b(\hat{u}(y)+\hat{h}(y))\psi_0(y)dy+\psi_0(a)\int_a^{+\infty}\hat{u}(y)dy+\psi_0(b)\int_{-\infty}^{b}\hat{u}(y)dy}{2\int_b^a\psi_0^2(y)dy+\psi_0(a)\int_a^{+\infty}\psi_0(y)dy+\psi_0(b)\int_{-\infty}^{b}\psi_0(y)dy}$$

Therefore, T^k, A and B determine the complete solution.

3. Results

All results are presented in nondimensional units. Distances are scaled by the standard equatorial radius of deformation (see values in Table I).

Results differ if the zonal extent of the gap is infinite (as supposed in calculations of section 1) or finite. More precisely, the latter means that the zonal extent of the gap is short compared with the zonal wavelength of the relevant free waves at that frequency. (It can also be treated equivalently as infinitesimally thin width; cf. Cane and du Penhoat, 1982). At low frequencies, it corresponds to the case of islands while the former case is relevant to a semi-enclosed sea.

On Fig. 1 are plotted the amplitude of the reflected Kelvin wave for different incoming Rossby waves. The results are sensitive to the location and width of the gap and to the meridional structure of the incoming flow.

For example, for the $n=1$ incoming Rossby wave, the reflected Kelvin wave is minimum if the gap is centered at the equator. However, the amplitude of the reflected Kelvin wave is not negligible. (On Fig. 1, the amplitude T^k has been normalized by the value of the Kelvin wave amplitude if the boundary were a full boundary). As the gap width decreases, wherever the gap is centered, the amplitude of the reflected Kelvin wave is over 0.9 : all the incoming mass flux is reflected back into the Kelvin wave and is equivalent to the full boundary case.

Plots for $n=2$ and $n=3$ incoming Rossby waves (Figs. 1b and 1c) show more structure, since as n increases, waves are less equatorially trapped and have a more complex oscillating structure. For $n=2$ incoming Rossby waves, which have an antisymmetric meridional structure, there is a weak reflected Kelvin wave as far as the gap is not centered on the equator. (For a full boundary, there is no reflected Kelvin wave; Cane and Sarachick, 1977; the amplitude of the reflected Kelvin wave is not normalized on Fig. 1b).

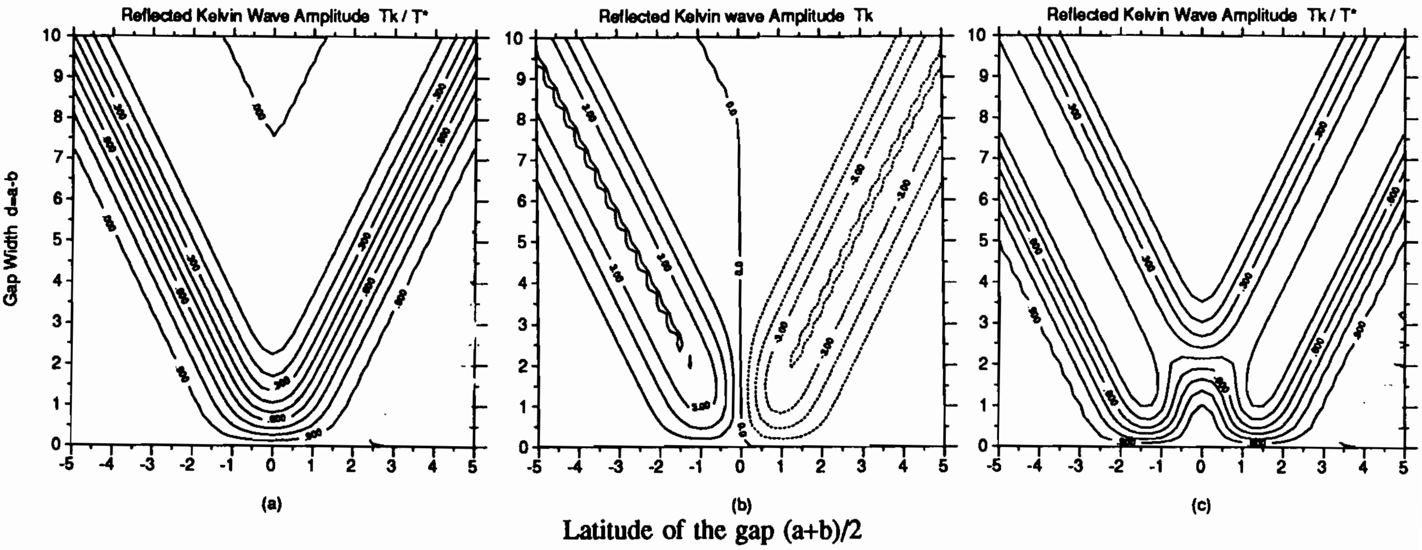

FIG. 1. Reflected Kelvin wave amplitude for a gap of infinite zonal length a)For $n=1$ incoming Rossby wave. Amplitude has been normalized by the reflected Kelvin wave amplitude for a full boundary. b) For $n=2$ Rossby wave, c) For $n=3$ incoming Rossby wave, normalized by the reflected Kelvin wave amplitude for a full boundary.

Results for a gap of finite zonal length are presented in Fig. 2. The methods used in section 1 apply. The solution differs as, in the downstream region west of the gap, the boundary condition $u=0$ at $y>a$ and $y<b$, together with the geostrophic relation leads to $h=D_a$ at $y>a$ and $h=D_b$ at $y<b$. D_a and D_b are constant. In this case, some mass flux must escape to set up a constant sea level on the western side of the gap. D_a and D_b are determined by projecting the Kelvin wave onto the solution west of the gap :

$$D_a=\hat{h}(a)-aA+T^k\psi_0(a) \quad \text{and} \quad D_b=\hat{h}(b)+bB+T^k\psi_0(b)$$

A and B have the same expression as in the former case. The amplitude of the reflected Kelvin wave is then :

$$T^k=-\frac{\int_a^b(\hat{u}+\hat{h})\psi_0 dy+\psi_0(a)\int_a^{+\infty}\hat{u}dy+\psi_0(b)\int_{-\infty}^b\hat{u}dy+\hat{h}(a)\int_a^{+\infty}\psi_0 dy+\hat{h}(b)\int_{-\infty}^b\psi_0 dy-a\int_a^{+\infty}\hat{u}dy\int_a^{+\infty}\psi_0 dy+b\int_{-\infty}^b\hat{u}dy\int_{-\infty}^b\psi_0 dy}{2\int_b^a\psi_0^2(y)dy+2\psi_0(a)\int_a^{+\infty}\psi_0(y)dy+2\psi_0(b)\int_{-\infty}^b\psi_0(y)dy+b\left(\int_{-\infty}^b\psi_0(y)dy\right)^2-a\left(\int_a^{+\infty}\psi_0(y)dy\right)^2}$$

Figs. 2 show the amplitude of the Kelvin wave for $n=1,2,3$ incoming Rossby waves. Though in some respects they present the same kind of pattern as in the former case, striking differences exist. First, this case is less efficient in reflecting waves. For example, for a gap centered at the equator and for $n=1$ Rossby wave, the amplitude of the reflected Kelvin wave is still minimum compared to an off-equatorial gap, but the amplitude of the reflected Kelvin wave is much smaller than in the infinite case : with $d=1$ and a gap centered at $y=1$, $T^k=0.3$ for the finite case and $T^k=0.7$ for the infinite case. With $d=1$ and a gap centered at the equator, $T^k=-0.1$ for the finite case and $T^k=0.45$ for the infinite case.

When the gap width gets infinitesimally small, we might expect the same answer as previously, namely all the mass is returned in the reflected Kelvin wave. In fact for the $n=1$ Rossby wave, the amplitude gets surprisingly small : all the mass flux is in the boundary currents and is used to set up the sea level on the other side of the gap. This effect depends on the latitudinal structure of the incoming flow as shown in Figs. 2.

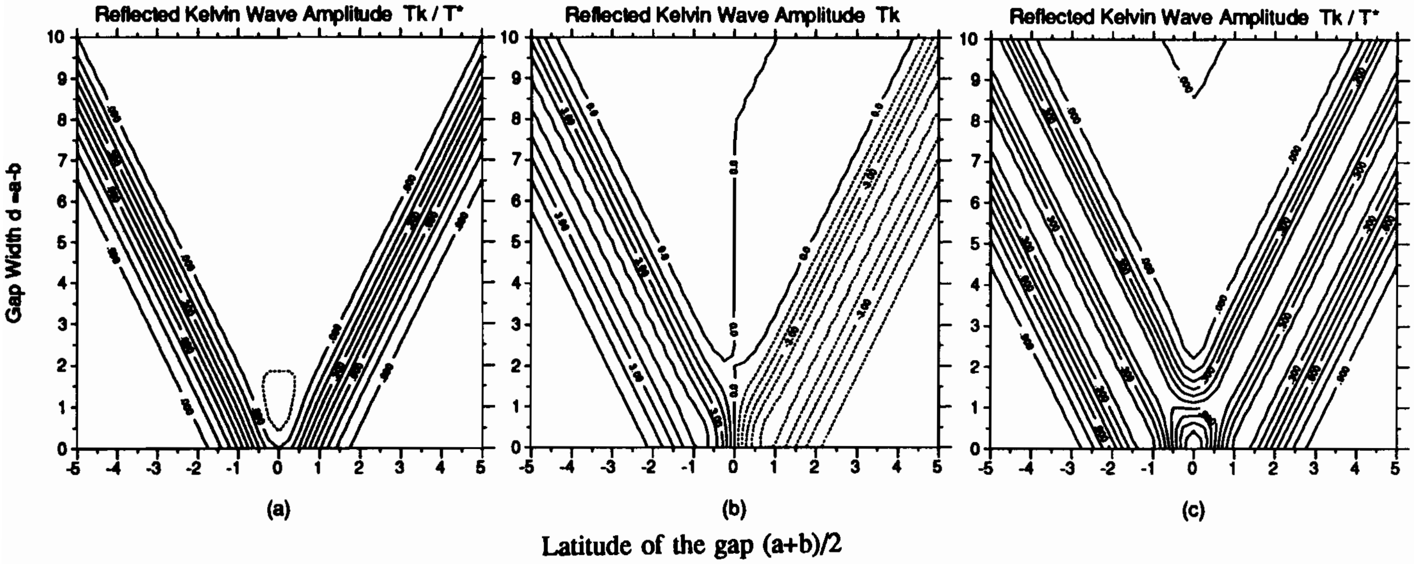

FIG. 2. As in Fig. 1 but for a gap of finite zonal length a) For *n=1* incoming Rossby wave, b) For *n=2* incoming Rossby wave, c) For *n=3* incoming Rossby wave.

4. Application to the western Pacific boundary

The western Pacific boundary consists of an irregular chain of islands. We now apply the method developed in previous sections to know how good it is as a reflector. We know from mass conservation that the mass flux through the Indonesian seas is just the difference between the incident flow (Rossby waves) and the reflected Kelvin wave flow. The partition between transmitted and reflected mass flux is determined by the way the pressure gradient adjusts to the geometry, i.e. to the pattern of the height field h.

We first must simplify the very complex geometry of this region. The assumptions made in our theory are certainly questionnable, but once they are made, certain topological invariances are obtained. Then the flow through the complex region is the same as that through an equivalent simple geometry.

Cane and Gent (1984) presented a theory for calculating the effect of a solid sloping western boundary on low frequency wave reflections : as the frequency becomes sufficiently small, the effect of the sloping coast disappears. Throughout our calculations, we have made the assumption that the frequency is low enough so that phase differences with longitudes may be ignored. Thus the sloping coast may be replaced by a zig-zagging coastline in the manner so familiar in numerical models, and finally by a straight coast. Only the gaps matter.

Similarly, while the latitude of the gap is important, the longitude is not : the offset between islands has no effect for long low frequency waves and as far as we are concerned with the interior flow. We finally end up with a simplified geometry shown in Fig. 3, roughly deduced from the 1000m isobath. The coast north of a is representing the Mindanao coastline and b stands for the New Guinea, Halmahera and Celebes islands, assuming that there is no significant flow through Molucca Strait. The connection to the Indian ocean is through the Strait of Makassar. We also assume that there is no flow through the Arafura Strait. The latitude of the Indonesian islands is determined by d (see table I). An alternative representation is given by Clarke, 1989. Using results of previous sections, we are able to compute the different pieces which make up the solution but we need one more condition. This is obtained from the relation for h, which constrains the possible pressure gradient to drive the flow through. We also assume that there is no Kelvin wave incoming from the Indian ocean, i.e. west of g. Then we get for T_6 , the amplitude of the reflected Kelvin wave in the Pacific ocean :

$$T_6 = -\frac{\hat{h}(b) + b\int_{-\infty}^{b} \hat{u}(y)dy + \left(\int_{-\infty}^{+\infty} \hat{u}(y)dy\right)\left(\psi_0(g) - g\int_{g}^{+\infty} \psi_0(y)dy\right)\left(\int_{-\infty}^{+\infty} \psi_0(y)dy\right)^{-1}}{\psi_0(g) - g\int_{g}^{+\infty} \psi_0 dy + \psi_0(b) + b\int_{-\infty}^{b} \psi_0(y)dy}$$

This relation simply indicates that the resulting amplitude of the reflected Kelvin wave in the Pacific Ocean part depends only on the latitude of the Indian and Pacific Ocean entrances, i.e. g and b. This is consistent with conservation of mass and with our simplifying rules.

For our schematization of the western Pacific boundary (Fig. 3), we choose $b=3°N$ (representing Halmahera and Morotai islands) and $g=8.5°S$ standing for the latitude of the Lombok and Timor straits. Table I gives the computed amplitude of the reflected Kelvin wave in the Pacific Ocean for the first three baroclinic modes and for different incoming Rossby waves.

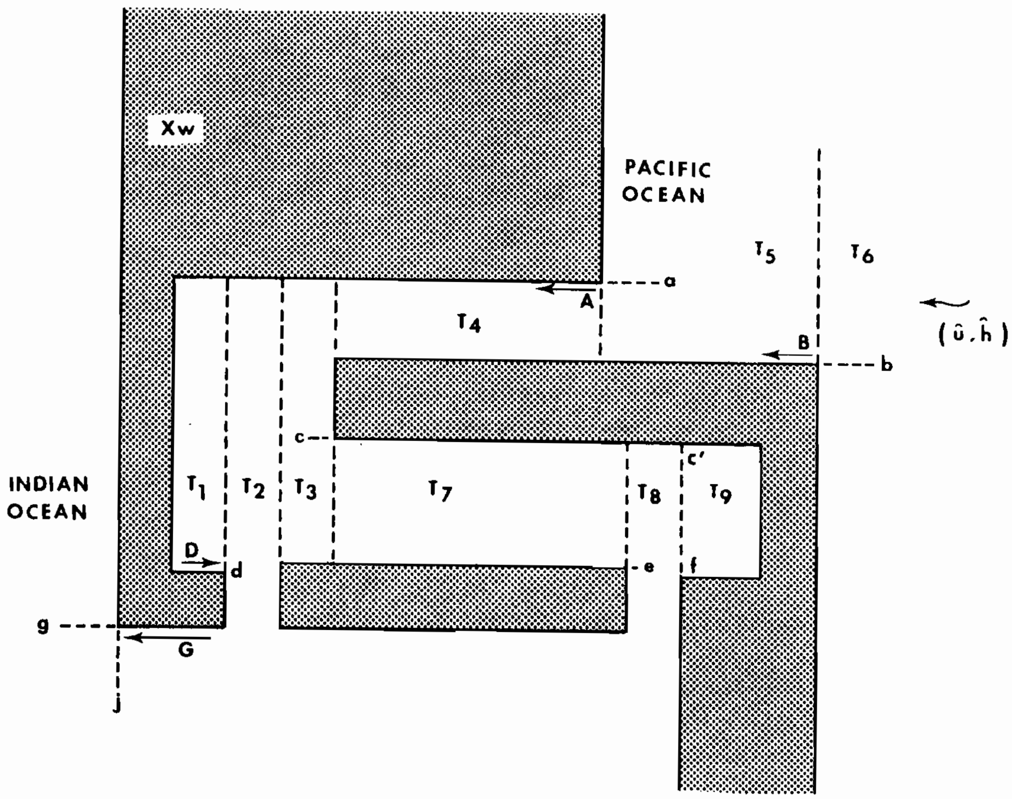

FIG. 3. Schematic geometry of the Western Pacific Boundary and of the Indonesian seas.

For the first Rossby mode, the amplitude of the reflected Kelvin wave is 85% of the amplitude if the boundary were a full boundary. It is over 90% for higher baroclinic modes which are more equatorially trapped. For antisymmetric (n even) Rossby modes, the presence of a gap allows a reflected Kelvin wave to exist whereas a full boundary would not. Therefore as far as this theory is concerned, the western Pacific boundary acts as a substantial reflector.

Table I : Results for a schematic Pacific western boundary as in Fig.3
Latitude : b = 3 N g = 8.5 S
$t \rightarrow \infty$

Baroclinic mode #	l = 1			l = 2			l = 3		
Radius of deformation	356.9 km			278.8 km			222.6 km		
	Rossby #			Rossby #			Rossby #		
	1	2	3	1	2	3	1	2	3
T^k	0.15	-0.016	0.026	0.16	-0.011	0.027	0.168	-0.007	0.0287
$\frac{T^k}{T^\infty}$	0.85	-	0.74	0.91	-	0.76	0.95	-	0.79

5. Conclusion

The analytical techniques used in this note are an extension of those used by Cane and du Penhoat (1982) to study the effect of low latitude islands. The results show that the amplitude of the reflected equatorial Kelvin wave is sensitive to the location of the gap and the structure of the incoming flow. In addition, the results can be quite different depending on whether the zonal extent of the gap is assumed to be infinite or finite. (More precisely, the latter means that the extent of the gap is short compared with the zonal wavelength of the relevant free waves). A coastline with a finite zonal length gap is a less efficient wave reflector than an infinite zonal length gap. In particular, the difference is non-negligible close enough to the equator for $n=1$ Rossby wave.

In carrying out a calculation as idealized as the one here, one hopes the results will be robust enough so that their possible application to the real world is obvious. We are not concerned with the details of the boundary flows, but did hope to deduce some general rules about the strength of the interior flow, i.e., the amplitude of the reflected Kelvin wave. The complexities of even such a simplified model convinces us that it will be very difficult to be confident of any modeling study of the Indonesian flowthrough short of a highly resolved numerical calculation with a detailed representation of the geometry and bathymetry. Nonetheless, we offer a few tentative conclusions concerning the efficacy of the western equatorial Pacific "boundary" as a reflector. As the greatest sensitivity in our results concerns the region very near the equator and since the western Pacific is effectively blocked within one or two radius of deformation off the equator, low n Rossby modes will be largely reflected. Thus, if as proposed by Zebiak (1989) and Battisti (1989) among others, the reflections important for El Nino are primarily in motions well represented by the low n Rossby modes, then this result suggests that the realistic boundary will not greatly alter the expectations based on a simple solid boundary.

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TABLE OF CONTENTS

ABSTRACT i

RESUME iii

ACKNOWLEDGMENTS vi

INTRODUCTION

1. Motivation 1
2. Structure 2

LIST OF PARTICIPANTS 5

AGENDA 7

WORKSHOP REPORT

1. Introduction 19
2. Working group discussions, recommendations, and plans 20
   a. Air-Sea Fluxes and Boundary Layer Processes 20
   b. Regional Scale Atmospheric Circulation and Waves 24
   c. Regional Scale Oceanic Circulation and Waves 30
3. Related programs 35
   a. NASA Ocean Processes and Satellite Missions 35
   b. Tropical Rainfall Measuring Mission 37
   c. Typhoon Motion Program 39
   d. World Ocean Circulation Experiment 39
4. Presentations on related technology 40
5. National reports 40
6. Meeting of the International Ad Hoc Committee on TOGA COARE 40

APPENDIX: WORKSHOP RELATED PAPERS

Robert A. Weller and David S. Hosom: Improved Meteorological Measurements from Buoys and Ships for the World Ocean Circulation Experiment 45

Peter H. Hildebrand: Flux Measurement using Aircraft and Radars 57

Walter F. Dabberdt, Hale Cole, K. Gage, W. Ecklund and W.L. Smith: Determination of Boundary-Layer Fluxes with an Integrated Sounding System 81

MEETING COLLECTED PAPERS

WATER MASSES, SEA SURFACE TOPOGRAPHY, AND CIRCULATION

Klaus Wyrtki: Some Thoughts about the West Pacific Warm Pool 99
Jean René Donguy, Gary Meyers, and Eric Lindstrom: Comparison of the Results of two West Pacific Oceanographic Expeditions FOC (1971) and WEPOCS (1985-86) 111
Dunxin Hu, and Maochang Cui: The Western Boundary Current in the Far Western Pacific Ocean 123
Peter Hacker, Eric Firing, Roger Lukas, Philipp L. Richardson, and Curtis A. Collins: Observations of the Low-latitude Western Boundary Circulation in the Pacific during WEPOCS III 135
Stephen P. Murray, John Kindle, Dharma Arief, and Harley Hurlburt: Comparison of Observations and Numerical Model Results in the Indonesian Throughflow Region 145
Christian Henin: Thermohaline Structure Variability along 165°E in the Western Tropical Pacific Ocean (January 1984 - January 1989) 155
David J. Webb, and Brian A. King: Preliminary Results from Charles Darwin Cruise 34A in the Western Equatorial Pacific 165
Warren B. White, Nicholas Graham, and Chang-Kou Tai: Reflection of Annual Rossby Waves at The Maritime Western Boundary of the Tropical Pacific 173
William S. Kessler: Observations of Long Rossby Waves in the Northern Tropical Pacific 185
Eric Firing, and Jiang Songnian: Variable Currents in the Western Pacific Measured During the US/PRC Bilateral Air-Sea Interaction Program and WEPOCS 205
John S. Godfrey, and A. Weaver: Why are there Such Strong Steric Height Gradients off Western Australia ? 215
John M. Toole, R.C. Millard, Z. Wang, and S. Pu: Observations of the Pacific North Equatorial Current Bifurcation at the Philippine Coast 223

EL NINO/SOUTHERN OSCILLATION 1986-87

Gary Meyers, Rick Bailey, Eric Lindstrom, and Helen Phillips: Air/Sea Interaction in the Western Tropical Pacific Ocean during 1982/83 and 1986/87 229
Laury Miller, and Robert Cheney: GEOSAT Observations of Sea Level in the Tropical Pacific and Indian Oceans during the 1986-87 El Nino Event 247
Thierry Delcroix, Gérard Eldin, and Joël Picaut: GEOSAT Sea Level Anomalies in the Western Equatorial Pacific during the 1986-87 El Nino, Elucidated as Equatorial Kelvin and Rossby Waves 259
Gérard Eldin, and Thierry Delcroix: Vertical Thermal Structure Variability along 165°E during the 1986-87 ENSO Event 269
Michael J. McPhaden: On the Relationship between Winds and Upper Ocean Temperature Variability in the Western Equatorial Pacific 283

John S. Godfrey, K. Ridgway, Gary Meyers, and Rick Bailey: Sea Level and Thermal Response to the 1986-87 ENSO Event in the Far Western Pacific 291

Joël Picaut, Bruno Camusat, Thierry Delcroix, Michael J. McPhaden, and Antonio J. Busalacchi: Surface Equatorial Flow Anomalies in the Pacific Ocean during the 1986-87 ENSO using GEOSAT Altimeter Data 301

THEORETICAL AND MODELING STUDIES OF ENSO AND RELATED PROCESSES

Julian P. McCreary, Jr.: An Overview of Coupled Ocean-Atmosphere Models of El Nino and the Southern Oscillation 313

Kensuke Takeuchi: On Warm Rossby Waves and their Relations to ENSO Events 329

Yves du Penhoat, and Mark A. Cane: Effect of Low Latitude Western Boundary Gaps on the Reflection of Equatorial Motions 335

Harley Hurlburt, John Kindle, E. Joseph Metzger, and Alan Wallcraft: Results from a Global Ocean Model in the Western Tropical Pacific 343

John C. Kindle, Harley E. Hurlburt, and E. Joseph Metzger: On the Seasonal and Interannual Variability of the Pacific to Indian Ocean Throughflow 355

Antonio J. Busalacchi, Michael J. McPhaden, Joël Picaut, and Scott Springer: Uncertainties in Tropical Pacific Ocean Simulations: The Seasonal and Interannual Sea Level Response to Three Analyses of the Surface Wind Field 367

Stephen E. Zebiak: Intraseasonal Variability - A Critical Component of ENSO ? 379

Akimasa Sumi: Behavior of Convective Activity over the "Jovian-type" Aqua-Planet Experiments 389

Ka-Ming Lau: Dynamics of Multi-Scale Interactions Relevant to ENSO 397

Pecheng C. Chu and Roland W. Garwood, Jr.: Hydrological Effects on the Air-Ocean Coupled System 407

Sam F. Iacobellis, and Richard C.J. Somerville: A one Dimensional Coupled Air-Sea Model for Diagnostic Studies during TOGA-COARE 419

Allan J. Clarke: On the Reflection and Transmission of Low Frequency Energy at the Irregular Western Pacific Ocean Boundary - a Preliminary Report 423

Roland W. Garwood, Jr., Pecheng C. Chu, Peter Muller, and Niklas Schneider: Equatorial Entrainment Zone : the Diurnal Cycle 435

Peter R. Gent: A New Ocean GCM for Tropical Ocean and ENSO Studies 445

Wasito Hadi, and Nuraini: The Steady State Response of Indonesian Sea to a Steady Wind Field 451

Pedro Ripa: Instability Conditions and Energetics in the Equatorial Pacific 457

Lewis M. Rothstein: Mixed Layer Modelling in the Western Equatorial Pacific Ocean 465

Neville R. Smith: An Oceanic Subsurface Thermal Analysis Scheme with Objective Quality Control 475

Duane E. Stevens, Qi Hu, Graeme Stephens, and David Randall: The hydrological Cycle of the Intraseasonal Oscillation 485

Peter J. Webster, Hai-Ru Chang, and Chidong Zhang: Transmission Characteristics of the Dynamic Response to Episodic Forcing in the Warm Pool Regions of the Tropical Oceans 493

MOMENTUM, HEAT, AND MOISTURE FLUXES BETWEEN ATMOSPHERE AND OCEAN

W. Timothy Liu: An Overview of Bulk Parametrization and Remote Sensing of Latent Heat Flux in the Tropical Ocean 513
E. Frank Bradley, Peter A. Coppin, and **John S. Godfrey**: Measurements of Heat and Moisture Fluxes from the Western Tropical Pacific Ocean 523
Richard W. Reynolds, and **Ants Leetmaa**: Evaluation of NMC's Operational Surface Fluxes in the Tropical Pacific 535
Stanley P. Hayes, Michael J. McPhaden, John M. Wallace, and **Joël Picaut**: The Influence of Sea-Surface Temperature on Surface Wind in the Equatorial Pacific Ocean 543
T.D. Keenan, and **Richard E. Carbone**: A Preliminary Morphology of Precipitation Systems In Tropical Northern Australia 549
Phillip A. Arkin: Estimation of Large-Scale Oceanic Rainfall for TOGA 561
Catherine Gautier, and **Robert Frouin**: Surface Radiation Processes in the Tropical Pacific 571
Thierry Delcroix, and **Christian Henin**: Mechanisms of Subsurface Thermal Structure and Sea Surface Thermo-Haline Variabilities in the South Western Tropical Pacific during 1979-85 - A Preliminary Report 581
Greg. J. Holland, T.D. Keenan, and **M.J. Manton**: Observations from the Maritime Continent : Darwin, Australia 591
Roger Lukas: Observations of Air-Sea Interactions in the Western Pacific Warm Pool during WEPOCS 599
M. Nunez, and **K. Michael**: Satellite Derivation of Ocean-Atmosphere Heat Fluxes in a Tropical Environment 611

EMPIRICAL STUDIES OF ENSO AND SHORT-TERM CLIMATE VARIABILITY

Klaus M. Weickmann: Convection and Circulation Anomalies over the Oceanic Warm Pool during 1981-1982 623
Claire Perigaud: Instability Waves in the Tropical Pacific Observed with GEOSAT 637
Ryuichi Kawamura: Intraseasonal and Interannual Modes of Atmosphere-Ocean System Over the Tropical Western Pacific 649
David Gutzler, and **Tamara M. Wood**: Observed Structure of Convective Anomalies 659
Siri Jodha Khalsa: Remote Sensing of Atmospheric Thermodynamics in the Tropics 665
Bingrong Xu: Some Features of the Western Tropical Pacific: Surface Wind Field and its Influence on the Upper Ocean Thermal Structure 677
Bret A. Mullan: Influence of Southern Oscillation on New Zealand Weather 687
Kenneth S. Gage, Ben Basley, Warner Ecklund, D.A. Carter, and **John R. McAfee**: Wind Profiler Related Research in the Tropical Pacific 699
John Joseph Bates: Signature of a West Wind Convective Event in SSM/I Data 711
David S. Gutzler: Seasonal and Interannual Variability of the Madden-Julian Oscillation 723
Marie-Hélène Radenac: Fine Structure Variability in the Equatorial Western Pacific Ocean 735
George C. Reid, Kenneth S. Gage, and **John R. McAfee**: The Climatology of the Western Tropical Pacific: Analysis of the Radiosonde Data Base 741

Chung-Hsiung Sui, and Ka-Ming Lau: Multi-Scale Processes in the Equatorial Western Pacific 747
Stephen E. Zebiak: Diagnostic Studies of Pacific Surface Winds 757

MISCELLANEOUS

Rick J. Bailey, Helene E. Phillips, and Gary Meyers: Relevance to TOGA of Systematic XBT Errors 775
Jean Blanchot, Robert Le Borgne, Aubert Le Bouteiller, and Martine Rodier: ENSO Events and Consequences on Nutrient, Planktonic Biomass, and Production in the Western Tropical Pacific Ocean 785
Yves Dandonneau: Abnormal Bloom of Phytoplankton around 10°N in the Western Pacific during the 1982-83 ENSO 791
Cécile Dupouy: Sea Surface Chlorophyll Concentration in the South Western Tropical Pacific, as seen from NIMBUS Coastal Zone Color Scanner from 1979 to 1984 (New Caledonia and Vanuatu) 803
Michael Szabados, and Darren Wright: Field Evaluation of Real-Time XBT Systems 811
Pierre Rual: For a Better XBT Bathy-Message: Onboard Quality Control, plus a New Data Reduction Method 823