3.2 LARGE SCALE PRESSURE COUPLING BETWEEN THE ATMOSPHERIC PLANETARY BOUNDARY LAYER AND THE SOUTHERN OCEAN

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1. INTRODUCTION

The instability processes which act on the general circulations of the atmosphere and ocean separately are well known. Baroclinic instability in the atmosphere gives rise to growing Rossby waves which evolve into our weather systems, whereas in the ocean, a similar process generates mesoscale ocean eddies on a time scale of months. The joint instability of the coupled ocean-atmosphere in which the potential energy of the mean flow is converted into potential energy of growing waves has received scant attention. In this paper, we present an instability analysis for the large density contrast system of air and water.

2. THE BAROCLINIC INSTABILITY OF THE TWO AND A HALF LAYER COUPLED OCEAN-ATMOSPHERE

Consider the two and a half layer coupled ocean-atmosphere shown in Fig.1, in which the oceanic structure is the well-known one and a half layer ocean model, and suppose that there exist constant zonal velocities (U_i) in the three layers, and that infinitesimal perturbations of meridional geostrophic velocity (v_i) occur in the atmospheric and oceanic upper layers (the two active layers) together with infinitesimal displacements (η_i) of the layer surfaces. Then, following the analysis for small density contrast systems (Pedlosky, 1964), extended to large density contrast systems (Bye, 1992), we derive the potential vorticity equations for the atmospheric layer and the oceanic upper layer, from which the necessary conditions for instability are obtained.

______________________________ η_1
ATMOSPHERIC PLANETARY BOUNDARY LAYER (U_1, v_1, ρ_1, h_1)
________ Sea surface ____________ η_2
OCEANIC UPPER LAYER (U_2, v_2, ρ_2, h_2)
______ Main thermocline ___________ η_3
OCEANIC LOWER LAYER (U_3, ρ_3)

Figure 1 The two and a half layer model

On constructing the characteristic equation for the wavespeed, and locating the wavenumber and growth rate of the most unstable wave (Bye, 2002), we obtain the following results. The most unstable wave is an atmospheric divergent Rossby wave, steered by the zonal wind (U_1) and propagating at the deep ocean velocity (U_3), with the phase speed,

$$C_r = U_1 - \beta/(f_0^2/gh_1 + (2\pi/\lambda)^2) = U_3 \qquad (1)$$

in which $f_0 = 2\Omega \sin\phi$ is the Coriolis parameter, where Ω is the angular velocity of rotation of the Earth and ϕ is latitude, and $\beta = 2\Omega \cos\phi/a$, a being the radius of the Earth, and h_1 is the depth of the atmospheric layer and g is the acceleration of gravity. The wavelength and period of the wave are respectively,

$$\lambda = 2\pi (gh_1)^{1/2} / f_0 \qquad (2)$$

and,

$$T = 2\pi (gh_1)^{1/2} / (f_0 U_3) \qquad (3)$$

from which we obtain,

$$U_1 - U_3 = \beta gh_1 / f_0^2 \qquad (4)$$

The growth time constant of the most unstable wave is,

$$\tau = 2 / ((U_3 - U_2)\beta\rho_1\Delta\rho/\rho^2)^{1/2} \qquad (5)$$

where ρ_1 is the density of air, ρ is a standard ocean density, and $\Delta\rho = \rho_3 - \rho_2$ is the density contrast in the one and a half layer ocean. The interpretation of eqn.(5) is that for $U_3 - U_2 > 0$, potential energy may be extracted from the ocean. The greater the positive oceanic shear, the more rapid is the wave growth.

3. ENERGY FLOW

On evaluating the kinetic energy (KE) and potential energy (PE) of the perturbation in each fluid, we find that most of the energy lies in the atmospheric perturbation, and that the KE and PE/unit width are approximately of equal magnitude, i.e. the potential

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energy released from the ocean through the relaxation of the thermocline crosses the sea surface and enters the atmosphere. The upward rate of energy transfer across the sea surface,

$$W = p_a \, \partial \eta_2 / \partial x \, (U_1 - U_3) \qquad (6)$$

where p_a is the atmospheric pressure, and $\partial \eta_2 / \partial x$ is the sea surface slope, and for a growing perturbation, $W>0$.

4. APPLICATION TO THE CLIMATE SYSTEM

The important question is whether this instability plays any role in the coupled ocean-atmosphere dynamics. Table 1 shows the predicted properties of the unstable wave for $h_1 = 1$ km (since the upward

$\lambda = 6200$ km,	$C_r = 0.05$ ms^{-1},	$T = 4$ years
$\tau = 42$ years,	$U_2 = -0.05$ ms^{-1},	$U_3 = 0.05$ ms^{-1}
$h_1 = 1$ km,	$U_1 = 10$ ms^{-1},	$\rho_1 = 1.2$ kgm^{-3}
$\rho_2 = 1028$ kgm^{-3},	$\Delta\rho = 1$ kgm^{-3},	$g = 9.8$ ms^{-2}
$f_0 = 10^{-4}$ s^{-1},	$\beta = 2\ 10^{-11}$ m^{-1}s^{-1}	

TABLE 1 Properties of the most unstable wave for typical controlling parameters

energy transfer is a part of the total frictional flux, which is trapped in the atmospheric planetary boundary layer), an ocean shear, U_3-$U_2 = 0.1$ ms^{-1} , and typical values of the other parameters. The growth time (τ) is an order of magnitude greater than the period (T) of the unstable wave, which indicates the possible existence of slowly growing fields, which inject structure into the climate spectra centered on a period of about 4 years.

Favourable oceanic shears, in which the upper layer flow is westward and the lower layer flow is eastward occur south of Australia (Rintoul and Bullister,1999), where subtropical water is transported to the west from the Pacific Ocean into the Indian Ocean, although the winds are westerlies and the deep ocean flow of the Antarctic Circumpolar Current is eastward, and the time and space scales of the most unstable wave appear to be similar to observations of the Antarctic Circumpolar Wave (White and Peterson, 1996).

The relative unimportance of the oceanic upper layer dynamics is consistent with the sea surface temperature being a slave to the atmospheric anomalies, even though the atmospheric anomalies are generated by the coupled instability, i.e. the slow eastward progression of the atmospheric Rossby wave would be expected to impress a positive temperature anomaly on the ocean in the poleward wind phase, and a negative temperature anomaly in the equatorward wind phase, giving rise to a poleward oceanic heat transport. For an atmospheric pressure wave amplitude of 1 hPa, the meridional particle excursions during the passage of the wave in the atmosphere and ocean are found to be, respectively, 40 000 km and 6 km. Hence, bearing in mind that the instability only occurs over a limited meridional range, the atmospheric wave would be expected to be of significant latitudinal extent, whereas the oceanic wave would only extend over a very small latitude band.

5. CONCLUSION

In summary, the instability of the coupled ocean-atmosphere outlined above is a generic process in the manner of the baroclinic instability of the atmosphere and ocean separately. It is characterized by an upward transfer of energy from the ocean to the atmosphere across the sea surface, brought about by the sea surface displacement wave, which over the Southern Ocean, may generate a free Rossby wave (eqn.(1)) in the atmosphere, propagating eastward at the speed of the deep ocean current

6. REFERENCES

Bye, J. A. T., 1992: Barocoupling of the atmosphere and ocean, *Tellus*, **44A**, 67-78

Bye, J. A. T., 2002: The pressure coupling between the ocean and the atmospheric planetary boundary layer: the unstable modes, *J. Atmos. Sci.* (submitted)

Pedlosky, J., 1964: The stability of currents in the atmosphere and the ocean: Part 1, *J. Atmos. Sci.,* **21**, 499-528

Rintoul, S. R. and J. L. Bullister, 1999: A late winter hydrographic section from Tasmania to Antarctica, *Deep-Sea Res.*, **46**, 1417-1454

White, W. B. and R. G. Peterson, 1996: An Antarctic circumpolar wave in surface pressure, wind, and temperature and sea-ice extent, *Nature,* **380**, 699-702