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CLIMATE SIMULATIONS WITH THE GLOBAL COUPLED ATMOSPHERE-OCEAN MODEL ECHAM2/OPYC PART I: PRESENT-DAY CLIMATE AND ENSO EVENTS

by

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Climate Simulations with the Global Coupled Atmosphere-Ocean Model ECHAM2/OPYC

Part I: Present-day climate and ENSO events

by

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Abstract

In this paper the global coupled atmosphere-ocean general circulation model ECHAM2/OPYC and its performance in simulating the present-day climate is presented. The model consists of the T21-spectral atmosphere general circulation model ECAHM2 and the ocean general circulation model OPYC with a resolution corresponding to a T42 Gaussian grid, with increasing resolution towards the equator. The sea-ice is represented by a dynamic thermodynamic sea-ice model with rheology. Both models are coupled using the flux correction technique.

With the coupled model ECHAM2/OPYC a 210 year integration under present-day greenhouse gas conditions has been performed. The coupled model simulates a realistic mean climate state, which is close to the observations.

The model generates several ENSO events without external forcing. Using traditional and advanced (POP-technique) methods these ENSO events have been analysed. The results are consistent with the "Delayed Action Oscillator" theory. The model simulates both, a tropical and an extra-tropical response to ENSO, which are in good agreement with observations.

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1. Introduction

Coupled atmosphere-ocean models are designed to simulate the present-day climate and to estimate its future trends, or to analyse climate states of the past. A complete model of the climate system is not only based on the laws of fluid mechanics and thermodynamics but also requires the inclusion of chemical and biological processes. However, present models cover only dynamics and thermodynamics of the two subsystems, atmosphere and ocean. Even without this restriction, the development of models simulating the physical atmosphere-ocean system is a demanding task. Not only the models of the individual subsystems, atmosphere and ocean, have to be improved, also the current coupling techniques by means of an anomaly flux coupling or flux correction can only be accepted as an intermediate solution. The ultimate goal should be a replacement by a technique which couples the subsystems directly.

Due to improvements in the parameterization schemes for the unresolved physical processes and due to enhanced computer power within the recent years, general circulation models of the atmosphere and the ocean are presently able to produce a quite realistic simulation of the observed atmospheric and oceanic climate states, respectively (e.g. Schlesinger, 1988; Boer et al., 1992). Coupled atmosphere-ocean models have already been applied to series of problems. Since the pioneering work on coupling models (e.g. Manabe and Bryan, 1969; Bryan et al., 1975; Manabe et al., 1975; Manabe et al., 1979; Washington et al., 1980) global coupled models have been used to estimate the anthropogenic greenhouse warming (e.g. Stouffer et al., 1989; Manabe et al., 1990; Manabe et al., 1992; Cubasch et al., 1992; Murphy et al., 1992). In order to investigate the El Niño/Southern Oscillation (ENSO) phenomenon and the potential to forecast ENSO events, coupled atmosphere-ocean models were applied where either (at least) one subsystem (in most of the cases the ocean) was restricted to a part of the globe or where an interactive coupling of the two subsystems was only allowed in the tropics (see McCreary and Anderson (1991) or Neelin et al. (1992) for overviews).

In order to provide a realistic simulation of the observed climate, a coupled atmosphere-ocean general circulation model (GCM) should not only be able to reproduce the mean climate state. Also the natural variability of the coupled climate system has to be simulated realistically. The real climate system exhibits variability on any time-scale. Not regarding phenomena with very short characteristic times, which are on the micro- and meso-scale, the large scale variability includes short-time weather phenomena as well as long-term fluctuations such as the changes between ice- and warm-ages. The interannual variability with time-scales up to a few decades is of particular importance for the present coupled GCMs, which allow the integration up to a few hundred years of simulated time. The most prominent phenomenon on that time-scale is ENSO. Also, the variability of the monsoon, the changes in the sea-ice coverage, the variability of the North-Atlantic

sea surface temperature and the North-Atlantic Oscillation are of interest. However, the present global coupled models have not been able to reproduce the natural variability of the climate system satisfactorily.

In this paper the global coupled general circulation model ECHAM2/OPYC and its performance in simulating the present-day climate is presented. After a brief description of the atmosphere model ECHAM2 (Section 2) and the ocean model OPYC (Section 3) the coupling technique is discussed (Section 4). In section 5 the simulated mean climate state of a 210 year integration under present-day conditions is presented and compared with observations. The tropical interannual variability of the ECHAM2/OPYC model is analysed by means of traditional and advanced (POP-technique) methods (section 6). A concluding discussion of the results closes the paper (section 7).

2. The Atmosphere Model ECHAM2

Based on the numerical weather prediction model of ECMWF¹, the spectral atmosphere general circulation model ECHAM² has been developed jointly by the Meteorologisches Institut der Universität Hamburg and the Max-Planck-Institut für Meteorologie, Hamburg, Germany. Prognostic variables are vorticity, divergence, temperature, (logarithm of) surface pressure, humidity and cloud water (ice and water phase). The model contains parameterizations of radiation, cloud formation and precipitation, convection, and vertical and horizontal diffusion. Land surface processes are described by a five-layer heat conductivity soil model and by a hydrological model to determine evaporation and local runoff. Vegetation is prescribed.

The model is currently used in two different horizontal resolutions, T21 and T42. This corresponds to a horizontal homogeneous resolution of approximately 8.6° and 4.3°, respectively, for dynamical processes. The corresponding Gaussian grid for calculating the nonlinear and the diabatic terms have a resolution of approximately 5.6° and 2.8°, respectively. The model uses 19 vertical layers in a hybrid σ -p-coordinate system with a top height at 10 hPa. The annual and daily cycles of the solar radiation are included. For runs in the uncoupled-mode, the annual cycles of the sea surface temperature and the sea-ice distribution are prescribed as lower boundary condition. Currently three versions of the model exist (ECHAM1, ECHAM2, ECHAM3), representing successively more advanced versions. A comprehensive description can be found in Roeckner et al. (1992), which also contains a basic climatology of the model.

The performance of the model, especially its ability to reproduce the observed natural vari-

¹ECMWF=European Center for Medium Range Weather Forecast

²ECHAM=ECMWF model, Hamburg version

ability, has been analysed in a number of papers (e.g. Claussen and Esch, 1993; Lohman et al., 1993; König et al., 1993; Ponater et al., 1994). Uncoupled ECHAM simulations have been performed in order to investigate internal atmospheric processes or the sensitivity of the model to external forcing (e.g. Graf et al., 1993; Graham et al., 1994). Finally, ECHAM (version ECHAM1) served as atmospheric part in coupled atmosphere-ocean simulations (e.g. Bakan et al., 1991; Oberhuber et al., 1992; Latif et al., 1992; Cubasch et al., 1992).

We have used the version ECHAM2 in T21 horizontal resolution for our coupled model, but we imposed a slight modification to the lower boundary condition in the presence of sea-ice (see section 4 for details). ECHAM2 differs from ECHAM1, which has been applied in the Hamburg CO₂ scenario simulations (Cubasch et al., 1992), in the choice of the orography (ECHAM2 uses a mean orography instead of an envelope orography) and in omitting the gravity wave drag parameterization. At T21 resolution, this choice of orographic forcing ensures a more realistic simulation of the atmospheric circulation, especially in the North-Atlantic sector (Sausen et al., 1993). In comparison to ECHAM3, the climate as simulated with ECHAM2/T21 is only slightly different, and it would be hard to decide which version simulates the observations better. The T21 horizontal resolution was selected as the only resolution, at which currently integrations are possible that are long enough to investigate interannual and interdecadal variability. The version selected for our study requires a time step of 40 minutes.

3. The Ocean Model OPYC

Like ECHAM, the ocean general circulation model OPYC³ (Oberhuber, 1993a; 1993b) is based on the primitive equations with the hydrostatic approximation. OPYC consists of three components: the deep ocean, the surface boundary layer and the sea-ice.

In the vertical, the deep ocean uses 8 layers in an isopycnal coordinate system, i.e. 8 layers of constant potential density. In the absence of vertical mixing due to turbulence, no flow goes across the layer boundaries. The layer thicknesses are prognostic variables. The choice of this coordinate system allows the maintenance of strong vertical gradients with a relatively small number of layers, and thus a realistic simulation of the thermocline. Temperature and salinity are treated as active tracers. Density is determined according to a realistic equation of state. Diapycnal mixing and convective transport is parameterized.

A surface boundary layer model is located on top of the deep ocean model. The surface boundary layer contains the same physical processes as the deep ocean model. However, the surface boundary layer is not a material layer like the layers in the deep ocean. The

³OPYC=Ocean model in isopycnic coordinates

thickness of the surface layer is determined from wind stirring, surface buoyancy fluxes due to heat and fresh water, penetrating solar radiation and shear stress in the uppermost ocean. The top of the model is formulated as a free surface, i.e. the total height of the ocean is given by the sum of the prognostically computed layer thicknesses.

The sea-ice model predicts the ice thickness, the ice concentration and the ice drift. A viscous-plastic rheology ensures a reasonable response of the ice momentum and the ice distribution to the forcing by the ice-ocean and ice-atmosphere stress. Freezing and melting is calculated from the heat flux into leads and due to heat conductivity. In order to allow the overflow across the North Pole, an open pole boundary condition is employed.

Horizontally OPYC uses an Arakawa B-grid. A horizontal resolution corresponding to a T42 Gaussian grid (approx. $2.8^\circ \times 2.8^\circ$) was chosen for the simulations presented in this paper. In the tropics the meridional resolution gradually increases towards 1° at the equator in order to resolve equatorial Kelvin waves. This results in a total of 128×72 longitude $\times$ latitude gridpoints.

A realistic land-sea mask and a realistic bottom topography are used. The model is forced by the wind stress, the total heat (sensible and latent heat plus short and longwave radiation) and fresh water fluxes and by the flux of turbulent kinetic energy. A semi-implicit time integration scheme allows steps of 1 day. The transport of potential temperature and salinity is computed with a simple semi-Lagrangian method while the momentum transport is carried out using a scheme that conserves potential vorticity and energy.

The ocean model OPYC has already been applied in a variety of studies (e.g. Miller et al., 1993; Holland et al., 1993; Miller et al., 1994).

4. The Technique of Coupling

As illustrated in Figure 1, the atmosphere and the ocean models are coupled by the air-sea fluxes of momentum (wind stress), total heat and mass (fresh water). The fresh water flux is given as precipitation minus evaporation plus the river runoff along the coastal boundaries. The latter is calculated according to a river transport scheme using the local runoff as input (Sausen et al., 1994). In addition to the fluxes listed above, the flux of turbulent kinetic energy is transferred into the ocean. This flux and a penetrative solar radiation ensure a realistic simulation of the oceanic mixed layer thickness, in particular during summer.

τ	wind stress
Q	total heatflux
TKE	turbulent kinetic energie input into the ocean
P	precipitation
E	evaporation
R	surface runoff
h_{ice}	sea-ice thickness
c_{ice}	sea-ice coverage
T_O	sea surface temperature
T_{ice}	sea-ice surface temperature
T_{eff}	effektive surface temperature

Figure 1: Coupling scheme for the global coupled general circulation model ECHAM2/OPYC.

The fluxes are calculated in ECHAM2, using as surface boundary conditions the sea surface temperature, the sea-ice thickness and the sea-ice concentration (the fraction of the grid-box covered by sea-ice) as provided by OPYC. However, ECHAM2 does not take into account a fractional sea-ice coverage. ECHAM2 rather assumes that a grid-box is either totally ice-covered (sea-ice concentration of 1) or totally free of ice (concentration of 0). Therefore, the sea-ice thickness h_{ice} , the sea-ice concentration c_{ice} (fractional sea-ice cover) and the sea surface temperature T_O , as simulated by the ocean model, have to be translated into an effective surface temperature T_{eff} for the atmosphere according to the following procedure:

1. A sea-ice surface temperature T_{ice} is determined for those gridpoints with $c_{ice} > 0$. We define $T_{ice} = T_O$ for $h_{ice} < 10$ cm. In the case of $h_{ice} \geq 10$ cm we assume that T_{ice} represents the temperature of the top 10 cm of the ice and that T_{ice} is given by the heat balance

$$c_p \frac{dT_{ice}}{dt} = Q_A - Q_O, \quad (1)$$

where c_p is the heat capacity of the top 10 cm of the ice, t is the time. Q_A is the total heat flux from the atmosphere into the ice (sum of solar and longwave radiation plus latent and sensible heat flux), and Q_O is the heat flux through the ice (coming from the ocean)⁴. The flux Q_O is parameterized as

$$Q_O = \frac{\beta}{h_{ice}} (T_{ice} - T_O) \quad (2)$$

with heat conductivity coefficient $\beta = 2$ W/(mK). Q_A in equation (1) depends on T_{ice} . In order to solve (1), a semi-implicit scheme is applied. Q_A is linearised with respect to T_{ice} at the reference point $T_{eff}^{(old)}$:

$$Q_A(T_{ice}^{(new)}) = Q_A(T_{eff}^{(old)}) + [T_{ice}^{(new)} - T_{eff}^{(old)}] \frac{\partial Q}{\partial T_{ice}}(T_{eff}^{(old)}) \quad (3)$$

Here $T_{ice}^{(new)}$ denotes the sea-ice surface temperature at the new time $t + \Delta t$ and $T_{eff}^{(old)}$ is the effective surface temperature one time step before at time t . After inserting (3), the discretized form of (1) is

⁴We use the convention, that downward fluxes are positive und upward fluxes are negative.

$$c_p \frac{T_{ice}^{(new)} - T_{ice}^{(old)}}{\Delta t} = Q_A(T_{ice}^{(old)}) + \frac{\beta}{h_{ice}} [T_O - T_{ice}^{(new)}] + [T_{ice}^{(new)} - T_{eff}^{(old)}] \frac{\partial Q_A}{\partial T_{ice}}(T_{eff}^{(old)}) \quad (4)$$

where Δt is the length of the time step. $T_{ice}^{(new)}$ follows immediately from eq. (4).

2. The new effective surface temperature is obtained by blending the ice surface temperature and the sea surface temperature according to

$$T_{eff}^{(new)} = c_{ice} T_{ice}^{(new)} + (1 - c_{ice}) T_O \quad (5)$$

To avoid a climate drift of the coupled system a flux correction is applied (Sausen et al., 1988). This is equivalent to coupling the atmosphere and the ocean by the anomalies of the fluxes computed relatively to the equilibrium states of the uncoupled subsystems (Sausen and Cubasch, 1989). Flux correction terms are applied to the effective surface temperature of the ocean and the sea-ice, which is transferred into the atmosphere, and to the fluxes of heat, momentum, fresh water and turbulent kinetic energy, from the atmosphere into the ocean. Some special treatment is applied to the flux correction for the positive defined turbulent kinetic energy input into the ocean (Lunkeit, 1993) and the solar radiation (see Section 5).

Both models are integrated synchronously, but with their individual time steps. Thus the fluxes computed at each 40 minutes time step by the atmosphere model are averaged over one full time step of the ocean (1 day) and then transferred to the ocean model, whereas the oceanic input into the atmosphere is fixed for 36 atmospheric time steps (= 1 model day).

Despite of the flux correction, previous integrations with the model ECHAM1/OPYC (Sausen, 1991) or the integration with the coupled model ECHAM1/LSG (Sausen et al., 1989) exhibited a residual climate drift, which very gradually started. This secondary drift could be caused by an inadequate spin-up of the uncoupled subsystems, resulting in inadequate flux correction terms. The spin-up of the uncoupled subsystems did not account for the variability of the forcing terms, which is present in the coupled run. This variability could lead to different mean climatological states of the models.

In principle, it is not possible to account for the variability of the coupled run in the uncoupled spin-up integrations. As a first guess to include appropriate variability in the uncoupled runs, we carried out a four-step preparation procedure before starting a coupled run itself (Figure 2):

Figure 2: Flow chart of the preparation procedure.

1. A 10-years reference run was performed with the uncoupled atmosphere model ECHAM2. The climatological annual cycles (i.e. 10 identical years of observed monthly mean data) of the sea surface temperature and sea-ice coverage were used as lower boundary conditions. An idealised sea-ice thickness of 2 m was assumed. Grid-boxes were either totally covered by ice or free of ice. Based on the output of this run we computed 10 annual cycles of daily anomalies of the air-sea fluxes relative to the simulated climatological (10 years) monthly means.
2. A 200-years spin-up run was performed with the uncoupled model OPYC using a climatological annual cycle of the observed forcing fields (ECMWF and COADS data).
3. The oceanic spin-up integration was continued for another 50 years, but now superimposing the daily anomalies of the ECHAM2 air-sea fluxes from step one as additional forcing terms. The mean forcing (same as in step two) assured that no long-term drift occurred, as this forcing terms contain negative feedbacks via relaxation terms in the formulae for the air-sea fluxes of heat and fresh water. However, on shorter time scales from days to months (and even up to years at high latitudes) the ocean model was able to develop considerable variability. Due to non-linearities in OPYC such as in the equation for the mixed layer depth or in the sea-ice dynamics, the simulated mean climatological state of the ocean is quite different from the state that is achieved if the ocean is driven by a monthly mean forcing only. This uncoupled OPYC run serves as a reference run for the later performed coupled run.
4. A second (10 years) integration was performed with ECHAM2 using the identical forcing as in step one. Using the same algorithm as in the coupled model (description above), the effective surface temperature was diagnostically computed from the actual atmospheric state and input data from OPYC (sea surface temperature, sea-ice thickness and sea-ice concentration). However, the OPYC data were not included in the forcing of this ECHAM2 run. The air-sea fluxes diagnosed in this run served as one part in the flux correction formulae for the correction seen by the ocean model.

The difference between the prescribed surface temperature of the ECHAM2 reference run (used as forcing in step 4) and the effective surface temperature diagnosed in step 4 yields the flux correction for the (effective) surface temperature. Zonal averages of the prescribed surface temperature, the diagnosed effective surface temperature and the resulting flux correction are shown in Figure 3 for the Northern Hemispheric winter (December to February) and Northern Hemispheric summer (June to August), respectively. The diagnosed effective surface temperature is close to the observations. Larger differences occur mostly in the winter hemispheres and at high latitudes (sea-ice covered regions). In general, the

effective surface temperature diagnosed from the OPYC data is warmer than the observed temperatures used as lower boundary condition in the uncoupled ECHAM2 reference run.

Figure 3: Zonally averaged effective surface temperature prescribed in the uncoupled ECHAM2 reference run (solid curve), diagnosed from the uncoupled OPYC reference run (dotted curve) and the resulting flux correction (dashed curve). Upper panel: December to February, lower panel: June to August.

This can also be seen from the geographical distribution of the annually averaged flux correction (Figure 4). Almost everywhere the correction is negative. However, large differences only appear in a few regions. The large values at the east coasts of the continents of the Northern Hemisphere refer to the strong temperature gradients in these regions. Small errors in the representation of the oceanic currents (Gulf-Stream and Kuroshio current) due to the resolution of the models lead to large differences. The differences at the sea-ice margin are related to the inconsistency of the sea-ice coverage in the uncoupled oceanic and atmospheric reference runs. In the North Atlantic and the North Pacific, OPYC simulates a sea-ice margin, which lies slightly southwards from the sea-ice margin used in the ECHAM2 reference run. In the Southern Hemisphere the sea-ice concentration simulated by OPYC is too small in order to get the same cold effective surface temperatures as in the observations.

Figure 4: Geographical distribution of the annually averaged flux correction for the effective surface temperature. Isolines for $\pm 15.$, $\pm 1.$, $\pm 5.$, $\pm 4.$, $\pm 3.$, $\pm 2.$, $\pm 1.$, ± 0.5 °C, negative contours are dashed.

These inconsistencies at the sea-ice margin can also be identified in the correction for the total heat flux (Figure 5). In regions, where OPYC simulates less sea-ice than observed, the flux correction shows large negative values. Here the heat flux out of the ocean used as a forcing term in the OPYC reference run is greater than the corresponding heat flux simulated by ECHAM2. East of Greenland the differences have a maximum. Differences of the land-sea-mask between OPYC and ECHAM2, a bias in the sea-ice simulation

by OPYC or the representation of sea-ice in ECHAM2, could be responsible for these inconsistencies. As in ECHAM2 a grid box is either totally covered with sea-ice or free of ice, the effect of leads on the heat flux cannot be taken into account.

Figure 5: Geographical distribution of the annually averaged flux correction for the total surface heat flux. Isolines for $\pm 500.$, $\pm 250.$, $\pm 100.$, $\pm 25.$ W/m^2 , negative contours are dashed.

The most striking differences occurred in the fresh water flux in the Northern Hemisphere, especially in the Arctic Basin (Figure 6). Partly, these differences been identified as an artifact of the ocean model formulation caused by a too strong salinity diffusion near the pole: In the Arctic Ocean strong horizontal salinity gradients are observed (e.g. Levitus 1982). The Newtonian formulation of the fresh water flux in the ocean model⁵ produces also strong gradients in the uppermost ocean layer, and the model diffusion causes a too strong horizontal transport of salt. This had to be compensated by surface fluxes, which are diagnosed as relaxation terms and which therefore strongly differ from the corresponding observed fresh water flux.

The wind stress as simulated in the ECHAM2 reference run shows quite the same pattern as the observations, but the amplitudes are smaller. This had to be compensated by the additive flux correction.

⁵A relaxation of the salinity of the uppermost ocean layer towards the observations is used as boundary condition in the uncoupled OPYC run.

After the preparation procedure described above a coupled simulation was started using the final states of atmosphere and ocean models as achieved at the end of steps 4 and 3, respectively, as initial conditions. With the coupled model a 210-years control integration for present-day conditions (present-day greenhouse gas concentration and present-day solar radiation) has been performed. Atmospheric data were archived every 12 simulated hours, oceanic data every 30 simulated days (a model month), except for mixed layer and sea-ice data, where the achieve interval is every 24 hours. The integration required about 2500 CPU hours of a single processor of a Cray YMP. The model output amounts after densely packing about 100 Giga Byte for the whole integration.

Figure 6: Geographical distribution of the annually averaged flux correction for the fresh water flux. Isolines for $\pm 50.$, $\pm 25.$, $\pm 10.$, $\pm 5.$, $\pm 1.$ mm/day, negative contours are dashed.

5. The Mean Climate

Due to the careful preparation procedure and due to the flux correction the coupled run exhibits no initial shock. Figure 7 shows the time evolution of the global annual mean near surface (2m) temperature difference⁶ between the coupled integration and the

⁶The near surface temperature was selected to show the behavior of the coupled model for two reasons: First, it is a variable of common interest and usually studied in global warming simulations. Second, as the variable is located at the interface of the two components forming the coupled model, it is a rather

uncoupled ECHAM2 reference run. The time series shown in Figure 7 starts immediately after the coupling, no spin-up period was skipped. As the near surface (2 m) temperature does not depart too far from the surface temperature, especially over the ocean and as the sea surface temperature in the reference run is prescribed according to observations, the global mean of the near surface temperature of the reference run is close to observations.

The climate drift of the coupled system is rather small. In the first 100 years the global mean temperature decreases exponentially with a time scale of about 25 years by about 0.5 K. In the 11th decade a spontaneous cooling occurs where the global annual mean near surface temperature drops about 0.4 K within a few years. After this “cold event” the system slowly recovers. This cooling is connected with an oceanic phenomenon. An outbreak of fresh and cold water out of the Arctic Basin cools the entire North Atlantic. During this “cold event” the deep convection in the North Atlantic was switched off and never recovered to the initial strength. A second, but weaker fresh water outbreak appeared about 50 years later. The physical mechanisms leading to the long-term variability in the North Atlantic have not yet been analysed in detail. A sudden reduction of the Arctic ice cover occurred shortly before the fresh water outbreaks. (For more details on this phenomena see Lunkeit et al., 1994b). At the end of the integration (210 years) the global mean of the near surface temperature is significantly higher than its minimum and only 0.4 K below its initial value.

Figure 7: Time evolution of the globally averaged annual mean near surface (2 m) temperature difference between the coupled simulation and the climatological mean of the uncoupled ECHAM2 reference run.

good indicator for a climate drift.

Figure 8 presents the time evolution of the zonal mean near surface temperature in a Hovmoeller diagram, which presents a five-year running mean of the temperature difference between the coupled integration and the uncoupled atmospheric reference run. The temperature of the coupled run remains rather close to its initial value, especially in the tropics where the derivation is less than 0.5 K. In the Northern Hemisphere the cooling connected with the fresh water outbreak in the 11th decade is the dominant signal. The temperature drop amounts up to 2 K. A low frequency oscillation can be observed in the Southern Hemispheric extra-tropics. The temperature decreases about 2 K within the first 100 years, but returns to the initial state after about 200 years.

Figure 8: Hovmoeller diagram of the zonally averaged annual mean near surface temperature difference (5-year running mean) between the coupled simulation and the climatological mean of the uncoupled ECHAM2 reference run. The contour interval is 0.5 K. The zero line is omitted.

Not only the zonal mean of the near surface temperature of the coupled model remains close to its initial value, also the geographical distribution of this variable is realistically simulated. Figure 9 shows the geographical distribution of the 10-year mean of the near surface temperature as simulated by the coupled model ECHAM2/OPYC during the last ten years of the integration (mean of the years 201-210) and the corresponding differences to the uncoupled atmospheric ECHAM2 reference run. The simulated climate is quite realistic both with regard to the absolute values and to the pattern compared with the uncoupled ECHAM2 reference simulation. In most of the areas of the globe the differences

relative to the uncoupled ECHAM2 climate are less than 1 K. Larger values are only observed around Antarctica, in the Arctic basin and in some small localized areas in the tropics and mid-latitudes.

Figure 9: Geographical distribution of the near surface temperature of the coupled integration. Upper panel: Annual mean of the last decade (years 201-210). Contour interval: 5 °C) Lower Panel: Difference to the climatological annual mean of the uncoupled ECHAM2 reference run. Isolines for $\pm 15.$, $\pm 1.$, $\pm 5.$, $\pm 4.$, $\pm 3.$, $\pm 2.$, $\pm 1.$, ± 0.5 °C), negative contours are dashed.

The Northern Hemisphere shows negative anomalies in general. These negative anomalies are a relict of the “cold event” in the 11th decade. The largest differences occur in the northern North-Atlantic east of Greenland. Here the cold anomalies of the near surface temperature are associated with a southward shift of the sea-ice margin by about 5° to 10° (1 to 2 ECHAM grid points). In the Southern Hemisphere positive as well as negative anomalies can be observed. These are also connected to anomalies of the sea-ice coverage. Finally, we would like to mention that the interdecadal variability in polar regions is rather large, especially in those regions exhibiting a larger difference in Figure 9 (lower panel).

After the discussion of the near surface temperature we will study three other atmospheric key-variables in order to demonstrate that the coupled ECHAM2/OPYC model is able to simulate a realistic present-day climate. We present an intercomparison of the last decade (years 201-210) of the coupled ECHAM2/OPYC integration, the uncoupled ECHAM2 reference run (ten year mean) and observations (ECMWF level IIb analysis of year 1979-1989 and Legates and Willmot (1990) precipitation data) by means of the zonal mean zonal wind (Figure 10), the Northern Hemisphere 500-hPa geopotential height field (Figure 11) and the precipitation (Figure 12), respectively. Here only the Northern Hemispheric winter (December, January, February) is considered. The corresponding fields for June, Juli, August lead to the same conclusions and are not discussed in detail.

In general, the differences between the coupled ECHAM2/OPYC simulation and the uncoupled ECHAM2 reference run are smaller than the differences between the model results and the observations. Like the uncoupled model ECHAM2, the coupled model ECHAM2/OPYC simulates a too weak subtropical jet in the Southern Hemisphere, which is located too far equatorwards. The Southern Hemispheric westerlies are in general underestimated in both models. In the Northern Hemisphere, ECHAM2/OPYC simulates a stronger subtropical jet than the uncoupled ECHAM2, the latter being closer to the observations.

The simulated 500-hPa geopotential height fields (Figure 11) in both models are pretty much the same. Compared with the observations, both models underestimate the North Atlantic trough. Also, the simulation of the precipitation is quite similar in both models (Figure 12). The simulated patterns are in a good agreement with the observations. However, due to the uncertainties of the observations themselves, especially over the oceans, the errors of the models are hard to evaluate.

Figure 10: Climatological December to February mean of the zonally averaged zonal wind for the last decade (years 201-210) of the coupled integration (upper panel), the uncoupled ECHAM2 reference run (middle pane) and ECMWF-observations (lower panel). Contour interval: 5 m/s, negative contours are dashed.

Figure 11: Climatological December to February mean of the northern hemisphere 500-hPa geopotential height field for the last decade (years 201-210) of the coupled integration (upper panel), the uncoupled ECHAM2 reference run (middle pane) and ECMWF-observations (lower panel). Contour interval: 8 gpm.

Figure 12: Climatological December to February mean precipitation for the last decade (years 201-210) of the coupled integration (upper panel), the uncoupled ECHAM2 reference run (middle pane) and observations (lower panel). Isolines for 0.5, 1., 2., 4., 8., 16. mm/day.

So far, only atmospheric variables have been discussed. Due to its longer characteristic times the oceanic component of our coupled system exhibits a different behavior in some aspects. Of course, the sea surface temperature shows the same features as the near surface temperature which has already been discussed. However, in the deep ocean a drift of the temperature can be observed. Figure 13 shows the horizontal mean of the ocean temperature anomaly as function of depth and time. The anomaly is calculated relative to the long term (210 years) mean of the data. During the 210 years the deep ocean warms by about 1 K and achieves a final value of approximately 3 °C. This quasi-linear trend was already visible in the uncoupled ocean reference run and originates from the spin-up of the deep ocean. Unfortunately, due to a lack of computing time it was not possible to extend the uncoupled ocean spin-up integration until an equilibrium state was achieved.

Figure 13: Hovmoeller diagram of the globally averaged annual mean ocean temperature anomalies (difference to the long term means) in the coupled simulation. Contour interval: 0.1 K, negative contours are dashed.

One of the most critical parameters in a coupled global general circulation model is the sea-ice. Most of the coupled models suffer from a strong drift of the sea-ice coverage during the coupled control integration (e.g. Meehl and Washington, 1990; Cubasch et al., 1992). Due to the high non-linearity of the ocean-ice-atmosphere system the success of the flux correction method in its original version is limited. The correction terms estimated from the uncoupled reference runs (where an oceanic grid point is either totally ice covered or free of ice) are only meaningful for this climate state (ice or no ice). If in

the coupled model integration the climate state changes (ice to no ice conditions or vice versa) the working-point of the coupled system changes and the flux correction becomes useless if the flux correction field of a certain flux exhibits the sea-ice margin by e.g. a strong gradient. Especially, the solar radiation into the ocean changes drastically if the ice coverage changes, and therefore the flux correction terms for the solar radiation are no more valid under changed sea-ice conditions.

To take this facts into account, we only correct the incoming solar radiation at the bottom of the atmosphere in the coupled model ECHAM2/OPYC, but not the reflected part. The incoming solar radiation does not depend directly on the sea-ice coverage. Thus, the flux correction calculated from the uncoupled reference integrations is valid for both the ice and the no ice conditions. Together with the preparation scheme described in Section 4, the sea-ice distribution in the coupled ECHAM2/OPYC integration remains realistic and close to the uncoupled reference state. Figure 14a shows the climatological sea-ice distributions in both hemispheres for summer and winter conditions estimated from the last 10 years of the integration (year 201 to 210). For comparison Figure 14b displays the corresponding distributions of the uncoupled OPYC reference run (mean of the last ten years before coupling).

Figure 14a: Simulated climatological mean sea-ice coverage in the coupled simulation (years 201-210) for the northern (left column) and southern (right column) hemisphere for December to February (upper panels) and June to August (lower panels). Contour interval: 0.1.

Figure 14b: As Figure 14a but for the uncoupled OPYC reference run

6. El Niño/Southern Oscillation

El Niño/Southern Oscillation (ENSO) is the most prominent phenomenon with a strong atmosphere-ocean interaction on time scales of a few years. The close connection between the atmospheric phenomenon Southern Oscillation and the oceanic phenomenon El Niño was first described by Bjerknes (1969). The Southern Oscillation is characterized by the seesaw in surface pressure, which involves opposite changes in the eastern and western hemispheres (Berlage, 1957). These surface pressure anomalies represent changes of the Walker Circulation, the thermodynamically direct atmospheric circulation parallel to the equator. The Southern Oscillation can be described by the Southern Oscillation Index (SOI), which is defined as the pressure difference between Tahiti and Darwin/Australia.

El Niño denotes the occurrence of anomalous warm surface water in the tropical east and central Pacific (Rasmusson and Carpenter, 1982). The anomalies exist for about one year and have their maxima near Christmas time. The opposite phenomenon, the occurrence of anomalous cold surface water, is known as “La Niña”. El Niño and La Niña appear in irregular intervals with a time scale of about 2 to 10 years. The El Niño and the Southern Oscillation phenomena are not independent. The SOI is highly negatively correlated with the sea surface temperature in the equatorial East Pacific.

Due to its impact on the regional and global climate (Rasmusson and Carpenter, 1982; Cane, 1983; Glantz et al., 1991), the ENSO phenomenon has been intensively studied using observational data by means or models of various complexity (overviews are given by e.g. Philander, 1990; McCreary and Anderson, 1991; Neelin et al., 1992). Meanwhile, it is well established that ENSO can be described as an oscillation between a warm phase (El Niño) and a cold phase (La Niña). However, the mechanisms responsible for the maintaining of this cycle are presently not well understood.

The discussion of ENSO as simulated by the model ECHAM2/OPYC will concentrate on the years 21 to 100 of the integration. The first 20 years are omitted to account for the spin-up. Figure 15 shows a Hovmoeller diagram of the simulated sea surface temperature in the equatorial Pacific for the year 21 to 80. The temperatures are extracted directly from the ocean model, i.e. before the transformation to the coarser atmospheric grid and before adding the flux correction. The model simulates a realistic temperature gradient across the Pacific as well as a realistic annual cycle. Compared with observations (e.g. Philander, 1990; his Figure 1.12) the temperature in the western Pacific is slightly too warm (about 1-2 K), the minima of the annual cycle in the eastern Pacific are about 1 K too warm. The 28 °C isotherm is shifted about 10° to the east.

Figure 15: Hovmoeller diagram of the monthly mean sea surface temperature in the Pacific along the equator for the years 21 to 80 of the coupled simulation. Unit: $^{\circ}\text{C}$, Contour interval: 1°C .

A careful inspection of Figure 15 reveals a considerable interannual variability. This becomes more obvious, if the mean annual cycle is removed from the data (Figure 16). In order to exclude long-term variability the mean annual cycle is removed by subtracting the 20-year running mean. ENSO-like variability with cold and warm events in the Eastern

and central Pacific can be observed. The amplitudes of the anomalies reach up to 2 K. The typical time-scale of this variability is about 2 to 10 years. As in observations the simulated El Niño varies in amplitude as well as in the frequency of occurrence. Compared with the observations the simulated El Niño is somewhat too weak and the anomalies are somewhat too restricted to the eastern part of the Pacific.

Figure 16: Hovmoeller diagram of the monthly mean sea surface temperature anomalies in the Pacific Ocean along the equator for the years 21 to 80 of the coupled simulation. Isolines: ± 1.5 , $\pm 1.$, ± 0.3 °C, negative contours are dashed.

The pattern related to the modeled El Niño is shown in Figure 17a. Plotted is the loading of the first Empirical Orthogonal Function (EOF) of the sea surface temperature of the equatorial Pacific estimated from the years 21 to 100 of the integration. The first EOF explains 26% of the total variance. The loading of the second EOF is shown in Figure 17b (9% of the total variance). Figure 18 presents the corresponding principal components. Remarkable is the high coherence of the first and second EOF on time-scales of about 3 years with a phase lag of -90° (Figure 19). Together, both EOFs describe a westwards moving pattern. This westward propagation of the anomalies can also be seen in the Hovmoeller diagram (Figure 16).

Figure 17a: Loading of the first EOF of the sea surface temperature in the equatorial Pacific as simulated in the years 21 to 100 of the coupled integration.

Figure 17b: Loading of the second EOF of the sea surface temperature in the equatorial Pacific as simulated in the years 21 to 100 of the coupled integration.

Figure 18a: Principal Components of the first EOF of the sea surface temperature in the equatorial Pacific as simulated in the years 21 to 100 of the coupled integration.

Figure 18b: Principal Components of the second EOF of the sea surface temperature in the equatorial Pacific as simulated in the years 21 to 100 of the coupled integration.

Figure 20: Hovmoeller diagram of the simulated monthly mean anomalies of the zonal wind stress (left panel; Isolines for ± 6 , ± 3 , ± 1 mPa), the sea surface temperature (middle panel; Isolines for ± 1.5 , ± 1 , ± 0.3 K) and the sea level (right panel; Isolines for ± 2 , ± 1 , ± 0.3 cm) in the Pacific along the equator. Negative contours are dashed.

An elegant method to illustrate the horizontal pattern and the time-dependent behaviour is the Principal Oscillation Pattern (POP) analysis (Hasselmann, 1988; von Storch et al., 1988). POPs are the normal modes of a linear system where the system-matrix is estimated from the data. POPs are in general complex with the real part Re and the imaginary part Im . The complex amplitudes satisfy the standard damped harmonic oscillator equation with a period T and a damping time τ . The evolution of the system in the two-dimensional POP-space can be described by the following cyclic sequence.

Using the anomalies of the sea level, the sea surface temperature and the zonal wind stress at the equatorial Pacific from year 21 to year 100 of our integration, a simultaneous POP analysis had been carried out⁷. The leading POP derived from this data explains about 21% of the total filtered variance. It has a rotation period T of 35 months and a damping time τ (e-folding time) of 180 months. The imaginary part (Figure 21) and the real part (Figure 22) of this POP can be identified as the transition and the extreme phase of ENSO, respectively.

In the transition phase (positive imaginary part) before a warm (El Niño) event no signal in the zonal wind stress is present. The sea surface temperature shows a small positive anomaly in the eastern equatorial Pacific. This pattern is similar to the second EOF of the sea surface temperature (Figure 17b). A positive anomaly of the sea level (heat content) can be observed in the western part of the Pacific.

A quarter period (about 8 months) later the state has changed to a fully developed El Niño (positive real part). The El Niño phase is characterised by the well-known temperature anomaly in the eastern and central equatorial Pacific. The temperature anomaly is accompanied by a positive anomaly of the zonal wind stress located to the west of the temperature anomaly. The sea level (heat content) shows a negative anomaly in the western and a positive anomaly in the eastern part of the Pacific. The ENSO-cycle is completed by the negative imaginary part (transition phase) and the negative real part (La Niña).

⁷Before applying the POP analysis the data have been bandpass-filtered, maintaining periods from 1.5 to 15 years, and normalized with their individual area-averaged standard deviations.

Figure 21: Imaginary part (transition phase) of the dominant POP for the zonal wind stress (upper panel), the sea surface temperature (middle panel) and the sea level (lower panel).

Figure 22: Real part (extreme phase) of the dominant POP for the zonal wind stress (upper panel), the sea surface temperature (middle panel) and the sea level (lower panel).

This results are in a good agreement with results of similar POP-analysis carried out by Latif et al. (1992, 1993) using observations and output from a coupled general circulation model. In that model the atmospheric general circulation model ECHAM1 was coupled to an high resolution ocean circulation model, where the coupling was restricted to the tropical belt (30° North to 30° South).

Since the observed and simulated ENSO varies in structure as well as in the time-dependent behaviour, the result of the POP-analysis varies depending on the time-interval used. The pattern of the extreme phase and the sea level pattern of the transition phase is rather stable, while the rotation period, the damping time and the explained variance have large uncertainties.

Figure 23 shows Hovmoeller diagrams of the sea level anomalies in the Pacific at 10° North and 10° South. Again, bandpass-filtered data maintaining periods between 1.5 and 15 years of the years 21 to 50 were used. Following a warm (cold) event a positive (negative) sea level (heat content) anomaly penetrates from east to west across the Pacific. The anomaly needs about 2-3 years to cross the Pacific. In the western part the anomaly remains and builds a pool for a new El Niño/La Niña event. The crossing-time indicates Rossby-wave dynamics. In addition, an intensification of positive (negative) sea level anomalies in the western part of the Pacific can be seen during a negative (positive) event in the eastern part.

All results are consistent with the “Delayed Action Oscillator” mechanism, which was first proposed by Schopf and Suarez (1988). In this theory ENSO can be described as a self-sustaining oscillation between the warm phase El Niño and the cold phase La Niña. The warm temperature anomaly in the eastern and central part of the Pacific during the El Niño extreme phase causes positive anomalies of the zonal wind stress in the western and central Pacific. The slope of these anomalies forces upwelling off the equator causing negative anomalies in the upper ocean heat content. As a Rossby wave these anomalies propagate westward. The Rossby wave reflects at the western boundary into a downwelling Kelvin wave, which propagates eastwards along the equator. The El Niño event is cancelled and a La Niña event starts to develop. Since the thermocline in the western part of the Pacific is deep, the anomalies of the heat content do not result in anomalies of the sea surface temperature there. However, in the eastern part of the Pacific, where the thermocline is shallow, the heat content anomalies can affect the sea surface temperature. Once there is a temperature anomaly, air-sea interaction can cause a growing and a westward propagation of the temperature signal by advection.

However, due to the complexity of the general circulation model ECHAM2/OPYC processes cannot be clearly isolated from other processes occurring in the model. Especially, the interpretation in terms of Kelvin and Rossby waves is difficult. In the coupled atmosphere-ocean system, many wave modes are excited. Thus, the propagation speed

of the signal could not be expected to be in exact agreement with the theoretical phase speed of single oceanic Kelvin or Rossby wave modes. In our discussion of the mechanism, “Kelvin” and “Rossby” wave stands for phenomena which can be identified to some extent with these theoretical types of ocean dynamics (e.g. Oberhuber, 1993a).

Figure 23: Hovmoeller diagrams of the simulated sea level anomalies in the Pacific along 10° North (left panel) and 1° South for the years 21 to 50 of the coupled integration. Isolines for ± 2 , ± 1 , ± 0.3 cm, negative contours are dashed.

In the following the atmospheric component of ENSO, the Southern Oscillation, and the tropical and extra-tropical atmospheric response to ENSO events will be discussed. While doing this one has to remind that the dynamical resolution of the atmosphere model ECHAM2 in the tropics is only about 1/8th of the meridional resolution of OPYC. As the amplitude of equatorial Kelvin waves rapidly decreases with distance from the equator, equatorial anomalies as “seen” from the atmospheric model are somewhat weaker than the corresponding anomalies in the oceanic grid. Since the anomalies in the oceanic grid were already smaller than observed the expected response of the atmosphere model to ENSO will be weaker than observed. In addition, experiments with the uncoupled model ECHAM show that the atmospheric response to prescribed sea surface temperature forcing (or a perfect oceanic simulation of El Niño events) is weaker than the observed response (König and Kirk, 1991).

Figure 24 shows the Southern Oscillation Index (SOI) and the SST-anomalies averaged over the Niño-3 region (Figure 25) as simulated by the ECHAM/OPYC model in the years 21 to 100. Both time series are bandpass-filtered, maintaining periods from 1.5 to 15 years, and normalized with their long term standard deviation. Since data from single gridpoints of a spectral general circulation model like ECHAM2 seem not to be appropriate, the model SOI is defined as the difference between the area-mean of the surface pressure anomalies around Tahiti and around Darwin (Figure 25). Like for observations, a strong negative correlation (≈ -0.7) between the SOI and the Niño-3 temperature is also detected in the model. The anomalies amount up to three standard deviations. This is comparable to the observations and proofs that the tropical atmosphere-ocean interaction is realistically simulated by the model ECHAM2/OPYC.

Beside the tropical phenomena, also extra-tropical response on ENSO can be found in the model. Figure 26 shows a correlation map between the annual mean sea surface temperature in the Niño-3 region and the 500-hPa geopotential height field for the Northern Hemisphere winter (December-February). Here the annual mean is defined as the mean from May to April. The correlations are calculated from bandpass-filtered (periods from 1.5 to 15 years are maintained) time series from year 21 to 100 of the integration. A relationship between ENSO and a Pacific North American (PNA)-like pattern in the northern hemisphere can be observed, which is in good agreement with observations (e.g. Horel and Wallace, 1981).

Figure 24: Simulated Niño-3 sea surface temperature index (solid curve) and Southern Oscillation index (dashed curve) for the years 21 to 100 of the coupled integration.

Figure 25: Areas used to estimate the Niño-3 sea surface temperature index and the Southern Oscillation index.

Figure 26: Correlation between the northern hemisphere winter (December to February) means of the 500-hPa geopotential height field and the annual mean (May to April) sea surface temperature averaged over the Niño-3 region.

We have restricted our analyses of model ENSO to the years 21-100 of the simulation. After year 100 of the integration the variability of the tropical Pacific decreases. The cold pool in the eastern part of the Pacific gets warmer while the warm pool in the western part cools. At the end of the integration (in the 21st decade) the state is similar to a permanent El Niño. So far, it could not be decided whether this drift is connected to the “cold event” in the 11th decade (e.g. by a weakening of the trade winds) or whether it is a result of a non-linear residual drift of the coupled model. Nevertheless, it is remarkable that a drastic change of the variability in the tropical Pacific occurs even if the mean climate changes only little. That highlights the importance of a correct representation of the mean climate in order to get a realistic interannual variability, at least in the ECHAM2/OPYC model.

7. Conclusions

The global coupled atmosphere-ocean general circulation model ECHAM2/OPYC was presented. The model consists of the atmosphere general circulation model ECHAM2 with T21 horizontal resolution and the ocean general circulation model OPYC with a resolution of approx. $2.8^\circ \times 2.8^\circ$, which increases towards the equator. The sea-ice is represented by a dynamic thermodynamic sea-ice model with rheology.

Both models were synchronously coupled by the air-sea fluxes of heat, mass and momentum. In order to avoid a climate drift the flux correction technique and a four-step preparation procedure were applied. In this procedure, the variability of the atmosphere model is used in the spin-up run of the ocean.

A 210-years present-day climate simulation was performed with the coupled model. Due to the flux correction and the preparation procedure the coupled model shows only a small climate drift. The simulated mean climate remains realistic and close to the climate of the uncoupled reference integrations and to the observations.

The model generates several ENSO-events without external forcing. The model ENSO events are rather realistically simulated, both in structure and in amplitude. The amplitudes of the simulated sea surface temperature anomalies amounts up to 2 K. The time-scale of the simulated ENSO is 2 to 10 years.

Traditional and advanced methods (POP-technique) have been used to analyse the simulated ENSO related interannual variability. The results are consistent with the generalized "Delayed Action Oscillator" theory. Kelvin and Rossby wave dynamics lead to a cycle with cold "La Niña" and warm "El Niño" events.

The model simulates the tropical as well as extratropical response to ENSO realistically. In contrary to the opinion of some other authors (e.g. Latif et al., 1993), the results demonstrate that a flux correction does not necessarily reduce the interannual variability in coupled models, especially those related to ENSO. In the coupled model ECHAM2/OPYC the realistic simulation of the annual cycle in the equatorial Pacific and a sufficient meridional resolution seem to be a necessary prerequisite for a successful reproduction of the observed variability.

However, flux correction could only be accepted as an intermediate solution in order to get realistic and reliable results in coupled simulations. One way to reduce the magnitude of the correction terms which have to be added could be the following preparation procedure:

1. An integration with the uncoupled atmosphere model is performed. In this integration, monthly mean observed sea surface temperatures and sea-ice distributions serve as lower boundary conditions. The fluxes of heat, momentum and mass into the ocean are extracted from this run.
2. A reference run with the uncoupled ocean model is performed. In this run the fluxes extracted in step 1 are used as forcing terms for the ocean. In addition, relaxation terms for the temperature and salinity are used in order to get a realistic climate close to the observations used in step 1 as boundary conditions for the atmosphere. The sea surface temperatures and the sea-ice distributions are extracted from this run.

3. A second integration (the reference run) with the uncoupled atmosphere model is performed. In this run the lower boundary conditions are the sea surface temperatures and sea-ice distributions simulated in the oceanic reference run (step 2).

If the ocean model is tuned carefully, the climatology of the sea surface temperatures and sea-ice variables simulated in the oceanic reference run (step 2) will be very close to the observations used in step 1. Therefore, the climate simulated by the atmosphere model using these terms as boundary conditions will not differ too much from the climate simulated in step 1. The difference between the fluxes of heat, momentum and mass simulated in step 3 and used in step 2 (including the relaxation terms) yields the flux correction. Since the sea surface temperature and the sea-ice distributions are the same in step 2 and step 3, no flux correction is needed for these terms.

In Part II of our paper, we present results from a transient greenhouse warming experiment, where the greenhouse gas concentration is increased according to the IPCC⁸ Scenario A. Changes of the mean climate state as well as changes of the interannual variability will be discussed. In Part III we will study the North Atlantic variability connect

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⁸IPCC=Intergovernmental Panel on Climate Change

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Figure 19: Coherence spectrum for the Principle Components of the first and second EOFs of the simulated sea surface temperature in the equatorial Pacific.

In order to demonstrate the mechanism related to the simulated El Niño events, Figure 20 shows three Hovmoeller diagrams for the years 21 to 50 of the integration: the anomalies of the zonal wind stress, the sea surface temperature anomalies and the anomalies of the sea surface elevation along the equatorial Pacific. The latter are a good measure of the heat content of the upper ocean. The sea level and the wind stress anomalies are bandpass-filtered maintaining periods from 1.5 to 15 years. The sea surface temperature anomalies are obtained as in Figure 16.

Again the non-periodic anomalies of the sea surface temperature can be seen. 1 to 2 years before a warm (cold) event occurs, a positive (negative) anomaly of the sea level (or the heat content) is located in the western part of the Pacific. In the ENSO year the heat content anomaly propagates within a few months across the Pacific and the temperature anomaly develops. The phase speed (about 2 m/s) of the propagation corresponds to the phase speed of an equatorial Kelvin wave. The temperature anomaly increases and propagates slowly into the central Pacific. There it decreases and vanishes after a few months. The positive (negative) sea surface temperature anomaly is accompanied by a positive (negative) anomaly of the zonal wind stress in the western and central Pacific.