GIOTTO: A coupled atmosphere–ocean general-circulation model: The tropical Pacific

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SUMMARY

A new coupled general-circulation model (GIOTTO) has been developed. The individual components are composed of the atmosphere model, ECHAM-4, and the ocean model, MOM (Modular Ocean Model)-1.2. The model domain is global, and no flux correction is applied. The coupling is active between 60°N and 60°S. Poleward of 60° the atmosphere is forced by the climatological sea surface temperature (SST), and the ocean is relaxed towards the climatological SST and sea surface salinity. Further, the sea-ice coverage is prescribed. The coupling interval is set to two hours to resolve the diurnal cycle.

In this paper we describe the design of the model, and discuss results of a coupled 20-year integration. The representation of the mean state is realistic, although there is an overall cold SST bias of about one degree centigrade in the tropics, and a tendency to simulate a double Inter Tropical Convergence Zone. The annual cycle, as simulated in the equatorial Pacific, is too weak in the east Pacific and too strong in the warm-pool region. The phase, however, is well captured.

The SST variability in the equatorial Pacific is underestimated by about 30%, and the anomalies are too confined to the equator. The main features of El Niño–Southern Oscillation (ENSO) dynamics, like propagation of heat-content anomalies, reflection of equatorial Kelvin and Rossby waves, and westerly wind bursts, however, are correctly represented by the model. A variability analysis based on empirical orthogonal functions indicates that the ENSO mechanisms are simulated correctly. The model also appears to be well balanced with a remarkably low SST drift (0.5 degC decade⁻¹), and a realistic equatorial thermal structure. We are, therefore, confident that the model can be used for experimental seasonal predictions.

KEYWORDS: Numerical modelling  Seasonal forecasting

1. INTRODUCTION

The extremely strong El Niño event of 1997/98 has raised not only scientific interest, but also large public excitement over the assessment of the impact of El Niño on the climate and weather in other regions of the world, and over the possibility of predicting it. It is known that El Niño is only one aspect of a coupled ocean–atmosphere oscillation. Bjerknes (1969) was the first to link the oceanic El Niño phenomenon to the atmospheric Southern Oscillation (ENSO). Nowadays it is widely accepted that air–sea interactions and positive ocean–atmosphere feedbacks play a major role in ENSO dynamics (e.g. Wyrkki 1988; Philander 1990).

It is, then, strongly evident that coupled ocean–atmosphere models are needed in order to simulate and predict the ENSO cycle and related tele-connections. A large variety of statistical (e.g. Hasselmann and Barnett 1981), hybrid (e.g. Latif and

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Flügel 1991; Barnett et al. 1993), and physical models (e.g. Ji et al. 1994; Delecleuse et al. 1997; Frey et al. 1997; Latif et al. 1997; Schneider et al. 1997; Stockdale et al. 1998) have been developed in the past for this purpose.

In principle, a coupled general-circulation model (GCM) is the most faithful approximation to the governing laws of the system, and should, therefore, produce the best simulation or prediction of the system. However, practical difficulties prevent the coupled GCM approach from achieving its full potential. Small errors and inaccuracies in the parametrizations of sub-grid-scale processes or the numerical representation of the equations may be amplified due to positive air–sea feedbacks. Thus, the mean state of the coupled model may be rather different from the observed mean climate. This phenomenon is generally known as climate drift. Flux correction schemes (e.g. Cubasch et al. 1992; Stockdale et al. 1994) or anomaly coupling (Ji et al. 1994; Kirtman et al. 1997) have been applied in order to eliminate this drift. Such corrections, however, are sacrificing the models’ consistency by introducing ad hoc empirical corrections that cannot be justified physically. Neelin and Dijkstra (1995) showed that flux correction may add spurious feedback mechanisms which change the dynamical system. The nature of ENSO variability in a coupled model may be quite different with and without flux correction (Delecleuse et al. 1997). Therefore, most coupled GCMs used for ENSO studies avoid any sort of flux corrections. Anomaly coupling is an alternative method of avoiding model climate drift. However, interactions between the ENSO cycle and the mean state or the annual cycle may be suppressed.

Although a number of non-flux-corrected models exist worldwide (Mechoso et al. 1995) this is still an area under development. Most of the non-flux-corrected coupled models share similar deficiencies and problems (Mechoso et al. 1995). These difficulties include a mean sea surface temperature (SST) structure which is too symmetrical around the equator, a misrepresentation of the correct annual cycle of SST in the tropical Pacific and a substantial underestimation of ENSO-like interannual variability. To identify the reasons for these common error features it is important to have a variety of different coupled model configurations. Crucial aspects to look at are the parametrization of vertical mixing in the ocean, and the cloud and radiation schemes used in the atmosphere model (Mechoso et al. 1995; Frey et al. 1997).

In this paper we present GIOTTO, a new coupled ocean–atmosphere GCM without any flux corrections. The model consists of two components: ECHAM-4 (atmosphere), and MOM (Modular Ocean Model)-1.2 (ocean). The ECHAM-4 atmospheric model was developed at the Max Planck Institut in Hamburg (Roeckner 1995; Roeckner and Arpe 1995) and has proved to be one of the most accurate atmospheric models for climate studies. It uses a sophisticated parametrization of clouds and radiation and is optimized for climate studies, rather than short-term numerical weather predictions. For our coupled model it is used at T30 resolution with 19 vertical levels. Other coupled GCMs that use this atmospheric model are ECHO-2 (Frey et al. 1997) and the model described by Roeckner et al. (1996).

The MOM-1.2 ocean model is a further development of the original Geophysical Fluid Dynamics Laboratory (GFDL) code as configured by Rosati and Miyakoda (1988). For the vertical mixing a level-2.5 turbulence closure scheme is used (Mellor and Yamada 1982). The zonal resolution is 1°, the meridional resolution decreases from 1/3° at the equator to about 2° at the northern and southern boundaries. The domain is almost global (75°S–65°N) and the coupling is active between 60°S and 60°N. MOM is, worldwide, one of the most successful models for ocean studies. Examples of coupled GCMs that include the MOM model are the Center for Ocean–Land–Atmosphere Studies model (Schneider et al. 1997), the GFDL coupled model (Rosati et al. 1997),
the National Centers for Environmental Prediction coupled model (Ji et al. 1994), and the Bureau of Meteorology Research Center coupled model (Power et al. 1998).

This study is the first time that these two circulation models have been coupled to each other. In this configuration our coupled model, GIOTTO-1.5, is one of the very few models that combine an atmospheric model, optimized for climate studies, with an ocean model that uses a highly complex vertical mixing scheme.

The results presented here are from a 20-year coupled integration of GIOTTO-1.5. The mean state, the annual cycle, and the interannual variability in the tropics as derived from this run are discussed. During the model development it was found that the vertical resolution of the ocean model has a large impact on the representation of the ENSO dynamics, and thus on the level of interannual variability.

The paper is organized as follows. In section 2 the coupled model is described. In sections 3 and 4 we describe the mean state and the annual cycle, respectively. ENSO dynamics and interannual variability as represented by the model are presented in section 5. The conclusions are stated in section 6.

2. Coupled model configuration, GIOTTO-1.5

(a) Atmosphere model, ECHAM-4

The ECHAM-4 atmosphere model was developed at the Max Planck Institut für Meteorologie in Hamburg. The dynamics are based on the European Centre for Medium-Range Weather Forecasts (ECMWF) operational weather-forecast model, which is a global low-order spectral model. Physical parametrizations, however, were developed in Hamburg and are optimized for climate studies. Special emphasis was put on the radiative scheme and on the representation of clouds. Specifically, a better representation of the low-level stratus clouds and modifications to the atmospheric convection scheme clearly improved the model performance relative to the older version, ECHAM-3.6. The tendency to simulate a double Inter Tropical Convergence Zone (ITCZ), and an underestimation of the cloud coverage in the equatorial west Pacific were clearly reduced. These improvements were also maintained in coupled integrations in which the model was coupled to the Hamburg ocean model, HOPE-2 (Fischer et al. 1997; Frey et al. 1997). A description of the ECHAM-4 atmosphere model, and the modifications relative to older versions are given in a technical report (DKRZ 1995) and by Roeckner (1995) and Roeckner and Arpe (1995), the performance under prescribed SST conditions was investigated by Moron et al. (1998).

For our coupled model we use a version with a triangular truncation at wave-number 30 (T30). Nonlinear terms and physical parametrizations are calculated on a $96 \times 48$ Gaussian grid, which corresponds approximately to a horizontal resolution of 3.75°. There are 19 vertical levels which are defined on $\sigma$ surfaces in the lower troposphere and on $p$ surfaces on higher levels.

(b) Ocean model, MOM 1.2

The configuration of the ocean model is basically the same as described by Rosati and Miyakoda (1988). Different vertical mixing schemes can be chosen in this model. Rosati and Miyakoda showed a considerable improvement in the simulated mixed-layer depth using the Mellor–Yamada 2.5-level turbulence closure scheme which led us to the decision to use this vertical mixing scheme in the coupled configuration. A few modifications, however, were found to be necessary in order to achieve realistic vertical profiles of temperature, density, and velocities. These are a small modification of the stability function in the Mellor–Yamada turbulence closure scheme (Galperin et al. 1988),
a different distribution of the vertical levels, and a uniform time step for tracers, velocities, and stream function. These, seemingly small, modifications had a large impact on the ENSO dynamics in the model, and thus on the interannual variability, which is described later in this paper.

The model equations are solved on a nearly global grid (75°S–65°N). The zonal resolution is 1°; meridionally the resolution decreases from 1/3° at the equator to about 2° at the northern and southern boundaries. Since the model does not contain a sea-ice model, SST and sea surface salinity are relaxed towards observed climatology poleward of 60° with a relaxation time of fifteen days. The sea-ice coverage is prescribed by climatology.

In the original Rosati–Miyakoda version 15 unequally distributed vertical levels were used. In an uncoupled control integration, and also in a coupled run with this ocean model, an unrealistic split of the thermocline at the equator was found (not presented here). The propagation of equatorial Kelvin waves from the west to the east Pacific was basically not present, and a strong semi-annual cycle in SST was found in the equatorial east Pacific. ENSO-like variability was reduced to an unrealistic low level and a pronounced climate drift in terms of NINO-3 (150°W–90°W, 5°N–5°S) SST averages was encountered. These errors were traced back to numerical instabilities in the vertical mixing scheme due to the choice of the vertical levels.

We created a new set-up of MOM with 18 vertical levels. The uppermost layer thickness is 10 metres which then increases exponentially with a constant growth rate of 20% down to a depth of about 315 metres. Below 315 metres the growth rate is set to 50%, down to a maximum depth of 3000 metres. With this distribution of vertical levels the numerical representation of vertical mixing (i.e. second derivatives) is more accurate than in the original set-up (e.g. Ferzinger and Peric 1997). Furthermore, the numerical error is equally distributed over all layers. The 18-level version of the ocean model proved to be a good compromise between computational costs and a realistic representation of the ocean dynamics. This model represents the equatorial dynamics substantially better than the 15-level version. A further increase to 21 vertical levels, however, did not show further improvement. Tests with significantly higher vertical resolution were not performed due to excessive computational costs.

A second, but minor, difference is the modification of the so-called stability function in the Mellor–Yamada turbulence closure scheme as proposed by Galperin et al. (1988). With these modifications the model gives a much better representation of the vertical temperature distribution along the equator (Fig. 5). In the following we shall refer to this model as GIOTTO-1.5.

(c) Coupling interface, CONDUCTOR 1.0

The coupling interface, CONDUCTOR 1.0, was developed at Consiglio Nazionale delle Ricerche, Istituto per lo studio delle Metodologie Geofisiche Ambientali and uses Parallel Virtual Machine routines to exchange information between the individual components. The two models are coupled without applying any flux correction. Further, no fine tuning was applied in order to improve the coupled simulations. Model development was carried out using only uncoupled integrations. Over the ocean the lower-boundary conditions for the atmosphere model are formed by the SSTs as computed by the ocean model. Over land the surface temperatures are computed by a soil model which is part of the ECHAM-4 model. The ocean model is driven by wind stress, net heat flux (short-wave radiation excluded), net short-wave radiation, and net fresh-water flux. The fluxes are taken as computed by ECHAM-4. Short-wave radiation is allowed
to penetrate into the water. Following Paulson and Simpson (1977) the downward irradiance is parametrized as a linear combination of two exponential functions with e-folding scales of 0.35 metres and 23.0 metres, respectively (Rosati and Miyakoda 1988). Since MOM is a rigid-lid model, the fresh-water fluxes must be transformed into a boundary condition for salinity, this is done in the following way:

\[ \rho_0 c_p K_H \cdot \frac{\partial S}{\partial z} = (E - P) \cdot S^* . \]  

(1)

\( K_H \) is a vertical mixing coefficient as computed by the turbulence closure scheme. \( c_p \) and \( \rho_0 \) are the heat capacity and the mean density of water, respectively. \( S \) denotes the salinity and \( S^* \) its climatological value. \( (E - P) \) is the net fresh-water flux as computed by the atmosphere model, ECHAM-4. Poleward of 60° the surface temperatures and sea surface salinity are relaxed towards climatology. This limitation is necessary since the MOM ocean model does not contain a sea-ice model. The atmosphere model is integrated with a time step of 30 minutes, the ocean model uses a time step of 40 minutes. Although it is possible in MOM to use different time steps for the equations for velocities, tracers, and the stream function, a uniform time step for all equations was found to be crucial in obtaining realistic velocities.

Since the horizontal resolution of the two models is different, flux and SST fields must be interpolated from one grid onto another one. Interpolating from a coarser grid onto a finer one (atmosphere to ocean) is a rather straightforward process. A simple bi-linear interpolation scheme is used. The opposite way, however, (ocean to atmosphere) is more difficult. In many coupled models an averaging algorithm is used. This may lead to strong smoothing of the SST, as seen by the atmosphere model, which is not desirable, especially in the region of the equatorial cold tongue. The meridional temperature gradient directly north and south of the equator would be strongly smoothed. To avoid that, we also use, for the interpolation from the finer ocean grid to the coarser atmosphere grid, a bi-linear interpolation scheme. For each atmospheric grid point over the sea the four nearest ocean grid points are taken, such that the atmosphere point lies in a box formed by the four ocean points. Then the SST is bi-linearly interpolated to the atmosphere point.

Since the land–sea masks of the ocean and atmosphere model do not match, all land points must be filled with reasonable estimates of sea surface values before interpolating. For that purpose a Laplace equation is solved over the entire horizontal model grid, prescribing values on all ocean points as boundary conditions. This method ensures a smooth extrapolation of variables from ocean points to land points.

In principle, a bi-linear interpolation scheme is conservative. In our case, however, conservation of interpolated quantities is not exactly satisfied for two reasons. The first one is the mismatch between the land–sea masks of the two models. The second is the fact that during the interpolation from the finer ocean grid to the coarser atmosphere grid not all ocean points are used. This second point, however, averages–out with time. From the fact that there is very little drift in our coupled model we infer that these effects are small.

Fluxes and SSTs are exchanged every two hours between the models. We think that such a relatively short coupling interval is important in order to properly resolve the diurnal cycle. It is obvious that the short-wave radiation exhibits a strong diurnal cycle with a maximum of about 1000 W m\(^{-2}\) at noon (local time), and a minimum of zero W m\(^{-2}\) at night. This induces increased vertical mixing in the upper ocean, since the hydrostatic stability is reduced at night. With a coupling interval of one day it is impossible to resolve this nonlinear effect. Also, the other fluxes depend on
the local time; in particular zonal wind stress and long-wave radiation vary strongly on time-scales of a few hours. The net heat flux (without short-wave radiation) averaged over the NINO-3 region varies by about 30 W m$^{-2}$ during one day, and the zonal wind stress may change its strength by a factor of two during a day. Both may strongly influence the vertical mixing in the upper ocean. The initiation of convection (in the atmosphere) is another process that is probably affected by the diurnal cycle. The simulated SST in the equatorial Pacific is about 0.2 degC higher in the afternoon (local time) than in the early morning. This is a small range but it might be important, since convection may suddenly start, if the SST exceeds a certain threshold.

3. MEAN STATE

A 20-year coupled integration of the G l o t o - 1.5 model was carried out, and results of that run (referred to as cop005) are presented. To create the atmosphere initial state the atmosphere model was integrated for one month with climatological January SSTs as lower-boundary conditions. The ocean initial state was obtained from a two-year spin-up, starting from Levitus (Levitus 1982) climatological temperatures and salinities. The other ocean prognostic variables were set to zero. This run was forced with ECMWF re-analysis fields. The ocean spin-up was started in January 1979, thus the ocean initial conditions of the coupled run correspond to 1 February 1981. A two-year spin-up is certainly at the lower limit of what is acceptable for the oceanic part. The adjustment time in the tropics, however, is of the order of a few months (Delecleuse et al. 1997). For control purposes we continued the forced ocean spin-up run for another 16 years and found that at least the ENSO related circulation was in balance after the two-year spin-up. For higher latitudes this is certainly not the case. Major sub-surface adjustments during the first five years of the coupled run were found in the central Pacific at about 25$^\circ$N and 20$^\circ$S. In the Indian ocean almost no initial drift was found. In the Atlantic ocean the major adjustment was found in the western part at about 10$^\circ$N.

The mean SST derived from the coupled 20-year run is displayed in Fig. 1 (upper panel). Below, the Levitus (Levitus 1982, 1994) mean SST is given as a reference, and in the lower panel the differences are displayed. Overall, the coupled-model mean state shows a cold bias in the tropics and mid latitudes. In the upwelling regions off the American west coast, and the Gulf of Guinea, and in the eastern part of the equatorial cold tongue, however, the coupled model is too warm. The cold tongue penetrates a bit too far into the west Pacific, and the SST patterns are too symmetrical around the equator. In particular, south of the equator, eastward of the date line, the SSTs are too zonally orientated. This is a common feature of many coupled GCMs which leads to the tendency to simulate a double ITCZ. We attribute this error to small SST errors which are then amplified by a coupled feedback mechanism. In reality a strong north–south asymmetry in terms of SST exists in the tropical Pacific (e.g. Philander 1990). North of the equator the surface water is between 1 degC (at about 160$^\circ$W) and 4 degC warmer than south of the equator. This strong temperature gradient causes more convection north of the equator, and, thus, northward cross-equatorial winds (Fig. 2). Northward winds cause an eastward water-mass transport north of the equator, and a westward transport south of the equator. Hence, warm water flows from west to east north of the equator, and cold water from east to west south of the equator. This mechanism maintains the observed north–south temperature gradient. If the north–south SST gradient is, for some reason, underestimated in the coupled integration this error can be amplified due to air–sea interactions. A smaller temperature gradient causes weaker northward wind
components and thus decreases the temperature gradient even more. In Fig. 3 we can see that the northward cross-equatorial wind component is much too weak in the coupled model. A possible reason for that error might be the simulated cloud coverage. The mean total-cloud coverage off the Peruvian coast is about 40% in our coupled model which is an underestimation by a factor of about 1.5 relative to observations (not presented here). This leads to an increased heat flux into the ocean due to more short-wave radiation, and thus decreases the north–south temperature gradient. Another reason could be the
vertical mixing in the ocean model. Reduced vertical mixing causes a small SST error which is then amplified by the mechanism described above.

Coupled ocean–atmosphere models tend to underestimate the upwelling of cold water at the South American coast, which also leads to a weaker-than-observed north–south temperature gradient. This is caused by weaker-than-observed off-shore winds, an error that is mainly a resolution problem in the atmosphere model, and is common to many other coupled GCMs.

The tropical Pacific SST is simulated much more realistically in GIOTTO-1.5 compared with an older version of our coupled model (GIOTTO-1.2). This indicates that the vertical mixing plays a crucial role, since the only differences between the two models are different vertical resolutions and a slightly modified vertical mixing scheme in the ocean.

The west–east SST difference between Indonesia and the Peruvian coast is about 2.0 degC smaller than observed. The reason for this is not completely clear. The erroneous cloud coverage certainly plays a role in the eastern part of the Pacific.

In the west Pacific the trade winds along the equator are slightly too strong. Therefore, the equatorial upwelling is increased, and thus the equatorial cold tongue penetrates too far into the west. The precipitation patterns (Fig. 4) are consistent with the SST deficiencies in the equatorial Pacific. North of the equator, in the region of the
Figure 3. u-component (upper panel) and v-component (lower panel) of the mean 1000 mb wind fields as derived from the coupled model. Positive values are shaded and correspond to eastward and northward flow for u and v, respectively.

Figure 4. Mean total precipitation as computed from the simulation. The contour interval is 2 mm day$^{-1}$. Regions with a mean precipitation above 4 mm day$^{-1}$ are shaded.

ITCZ, the simulated rainfall is less than observed, in agreement with the cold SST bias in that region. The simulated South Pacific Convergence Zone extends too far to the east, and its orientation is more zonal than observed. The rainfall along the equator is reasonably well simulated. Although the equatorial cold tongue extends too far into the
Figure 5. Mean vertical temperature distribution of the upper 300 metres along the equator as derived from the simulation cop005 (see text) (upper panel), and the Levitus dataset (middle panel). The differences are plotted in the lower panel. Contour intervals are 1 degC.
west, the model does not simulate a dry region over the equator west of the date line. The minimum precipitation does not fall below 4 mm day$^{-1}$ in that region.

A sharp equatorial thermocline is crucial for successful ENSO simulations. In Fig. 5 we present a vertical temperature section from the sea surface down to a depth of 300 metres. In the first and second panels the mean temperature sections, as derived from the cop005 run and the Levitus dataset (Levitus 1982, 1994) respectively, are displayed. The last panel shows the difference between them. Overall GIOTTO-1.5 gives a realistic representation of the temperature structure at the equator in the Pacific. The slope of the thermocline is reasonably well simulated although it is a bit too weak. The thickness of the mixed layer is well captured. However, in the central equatorial Pacific vertical mixing seems to be underestimated in the model. A weak point is a too diffuse thermocline which is suggested by the positive values in the difference plot east of the date line. The mean equatorial current system of the Pacific is well represented (Fig. 6). At the surface, westward flow prevails driven by the trade winds. The mean equatorial undercurrent reaches a maximum speed of about 0.8 m s$^{-1}$ which is in reasonable agreement with observations, and the eastward extension of the core of the undercurrent is well captured. These features improved substantially in the GIOTTO-1.5 run relative to an older version of the model with a different vertical-level distribution in the ocean.

4. **SEASONAL CYCLE**

Although the external solar forcing has a six-month period, an annual cycle is observed in the eastern equatorial Pacific. Sea surface temperature, the strength of the equatorial undercurrent, and the trade winds exhibit a clear annual cycle in this region. Many physical processes including air–sea interactions are responsible for that, but it has not been fully understood yet (Mechoso et al. 1995). Thus, the seasonal cycle is a very challenging test case for a coupled ocean–atmosphere GCM.

As already pointed out in the previous section, vertical mixing in the ocean and the representation of clouds play an important role in the successful simulation of the mean state of the equatorial Pacific. This holds true even more for the seasonal cycle. In Fig. 7 a time–longitude plot of the seasonal cycle of the equatorial Pacific SST in terms of deviation from the respective annual mean is presented. The data are averaged over
$2^\circ$N–$2^\circ$S around the equator. The mean seasonal cycle as computed from the simulation and from the GISST (Global sea ice coverage and sea surface temperature data) dataset (Parker et al. 1995) is given in Fig. 7.

The model exhibits an annual cycle, although only in the extreme eastern equatorial Pacific. The phase of the annual cycle is correctly captured. The maximum temperature is obtained in April, the minimum is reached in late September. The amplitude, however, is substantially underestimated. We do not have a complete explanation for that. Recent results obtained with two versions of the coupled GCM ECHO-2 indicate that the type of systematic error of the mean state found in GIOTTO-1.5 also causes a weaker-than-observed annual cycle in this region (Latif 1998, personal communication). A reduced zonal SST gradient causes weaker-than-observed trade winds, which may indicate reduced variance. Thus, the variation of equatorial upwelling is reduced which implies, as a direct consequence, a weaker seasonal SST signal. In the central Pacific the model exhibits a semi-annual cycle which is in contrast to observations. The SST climatology in the western Pacific is, in reality, governed by a very weak annual cycle which also contains a semi-annual component. Maximum temperatures are observed in spring, the
minimum is reached in late (northern hemisphere) summer, with a secondary maximum and minimum in November and February, respectively. The phase of this cycle is basically captured by the model, although the simulated amplitude is about twice as large as observed, when averaged over the NINO-4 region (150°E–150°W, 5°N–5°S). This feature can be explained by the penetration of the cold tongue into the warm-pool region, which was already seen in the mean state. In the central Pacific the coupled model exhibits an annual cycle with a pronounced semi-annual component which is in strong contrast to observations.

The GIOOTTO-1.2 results (not presented here) did show a clear semi-annual cycle in the east Pacific which is in strong contrast to observations. We attribute this mainly to the vertical mixing in the ocean. Due to the choice of the vertical layers in that model version, an almost homogeneously mixed layer is obtained at the top of the water column which is then followed by an unrealistically strongly stratified layer. The equatorial undercurrent is very weak in the eastern equatorial Pacific and Kelvin waves do not propagate so far in this version of the model. Thus, the upper mixed layer is basically decoupled from the water masses underneath, and the temperatures are dominated by solar radiation which, as a direct consequence, forces a semi-annual cycle. The new vertical resolution in the oceanic part of GIOOTTO-1.5, relative to GIOOTTO-1.2, improves the representation of the seasonal cycle substantially.

5. INTERANNUAL VARIABILITY

Our main goal in developing this coupled model was the simulation, investigation, and prediction of El Niño and La Niña events. Therefore, we are especially interested in the ENSO-like interannual variability of this model. Although 20 years might be at the lower limit necessary in order to evaluate the interannual variability we think that the main features can be assessed. A commonly used index for ENSO studies is the SST anomaly with respect to the annual cycle averaged over the so-called NINO-3 region. The time series of this index is presented in Fig. 8. The monthly mean SST anomalies, as computed from the 20-year coupled integration, are presented. The anomalies are given with respect to the mean annual cycle as computed from the coupled 20-year run. As a reference, the observed NINO-3 SST anomalies for the period 1975 to 1995 are also shown (thin line). The first remark is that there is not any obvious coupling shock at the beginning. This was also confirmed by a small set of preliminary hindcast experiments, carried out with this model (not presented here). Furthermore, it is hard to detect any drift in terms of NINO-3 SST averages. A linear regression fitted to the first fifteen years of the NINO-3 SST time series yields a mean cooling of about 2 degC in 100 years. This is a remarkably low value for a coupled model that does not apply flux corrections. Both the absence of a strong coupling shock and the small drift are important if the model is intended to be used for short-range climate predictions.

The 20 years of integration are certainly insufficient for a spectral analysis. However, from a first view it seems that the observed dominant ENSO period of about four years is reasonably well reproduced by the model. In the first ten years two El Niños and two La Niñas with anomalous NINO-3 SSTs beyond 1 degC are simulated. In year 16 another warm event occurs, and in year 18 a very strong La Niña event with anomalous NINO-3 SSTs of −3 degC is simulated. Relative to observed NINO-3 time series the model underestimates the interannual SST variability by about 30%.

An empirical orthogonal function (EOF) analysis of anomalous tropical Pacific SSTs gives more insight into the interannual SST variability (Fig. 9). Monthly mean SST anomalies between 120°E–80°W and 30°S–30°N were used. The data were taken
Figure 8. Monthly mean sea surface temperature (SST) anomalies, averaged over the NINO-3 index region (see text), as derived from the coupled model (cop005, see text). The anomalies are computed with respect to the mean annual cycle, obtained from the 20-year coupled integration. As a reference the observed anomalous SST is also presented for the period 1975–1995 (thin line).

Figure 9. First empirical orthogonal functions (EOFs) of tropical Pacific sea surface temperature (SST) anomalies obtained from the model (cop005, see text) (upper panel) and the GISST dataset (lower panel). The simulated EOF was computed from 20 years of integration, whereas the observed EOF was computed from 35 years of SST observations (1960–1994). The explained variances of the simulated and observed EOFs are 27% and 38%, respectively. The data were not filtered or detrended.
directly from the experiment without any filtering or subtraction of trends. As a reference we also display the corresponding EOF computed from the GISST dataset (Parker et al. 1995). Note that the GISST data is on a $1^\circ \times 1^\circ$ grid, 35 years of sea surface temperatures (1960 to 1994) were used (Fig. 9, lower panel). Both EOFs are normalized such that the variance of the corresponding time series is equal to one. Thus, the magnitude of the EOF patterns can be regarded as degree centigrade. The anomalies were computed relative to the mean annual cycle deduced from the coupled integration or the GISST data. The first EOF of the simulation explains 27% of the total variance, whereas the observed EOF explains 38%. Both EOFs clearly represent the ENSO related SST variability in the tropical Pacific. The maximum of the simulated SST variability is obtained in the eastern Pacific at the equator, which is in good agreement with observations. Weak points are that the strength of the interannual variability is underestimated by about 30%, and that the variability is too confined to the equator. Further, the characteristic off-equatorial SST anomalies of opposite sign are not as pronounced as in the observations, but they are noticeable. The corresponding time series of the principal components (not presented here) is very similar to the NINO-3 time series and nicely reflects the warm and cold events that were already seen in the NINO-3 SST time series (Fig. 8, lower panel). Overall, the interannual SST variability is reasonably well captured in the GIOTTO-1.5 model.

To explain the near-periodic nature of ENSO a conceptual model, the Delayed Action Oscillator was proposed by two groups at nearly the same time (Schopf and Suarez 1988; Suarez and Schopf 1988; Battisti 1989; Battisti and Hirst 1989). This model summarizes the fundamental zeroth-order physics involved in ENSO. For a comprehensive review of ENSO theory see the article of Neelin et al. (1997). Neelin et al. emphasize the importance of slowly propagating upper-ocean heat-content anomalies along the equator to interannual variability. Furthermore, they point out that the most realistic ENSO oscillation regime is in agreement with the proposed mechanism of Cane and Zebiak (1985) and Battisti (1988) in having a standing oscillation in SST with subsurface memory carrying the oscillation between phases. In Fig. 10 we present a time-longitude diagram of anomalous heat content and SST along the equator. The anomalies were computed from monthly means with respect to the mean annual cycle obtained from the 20-year experiment. As a measure for the upper-ocean heat-content we used the temperature averaged over the upper 315 metres of the ocean model. SSTs and upper-ocean heat content were averaged between 5°N and 5°S. The years labelled at the vertical axis only indicate that the run was initialized in February 1981. The propagation of heat-content anomalies from the west into the east Pacific can be seen. Large-scale heat-content anomalies cross the Pacific in about one year which corresponds well with observations (McPhaden 1993). Cold and warm anomalies show a distinct and almost constant propagation. In contrast to that the SST anomaly patterns are only oscillating in strength, their position is basically fixed. Both features agree with observations, and are consistent with the mechanism described in Neelin et al. (1997). They are crucial to the understanding and simulation of ENSO variability. In the older GIOTTO-1.2 run these characteristics were not present. We could not detect equatorial Kelvin waves crossing the whole Pacific, and heat-content anomalies were mainly oscillating, not propagating (not presented here). The interannual variability was reduced to a very low level, and the model did show a very severe drift of more than 1 degC (NINO-3 SST) within the first fifteen months.

ENSO may be regarded as a trajectory in a two-dimensional phase space which is spanned by two basis vectors that we call intermediate phase and extreme phase. In such a phase space the extreme phase can be identified as an El Niño event, whereas the
negative of the extreme phase corresponds to a La Niña event. The intermediate phase and its negative may be regarded as precursor patterns for an extreme warm or cold event, respectively (Latif et al. 1997). To identify such patterns in the coupled model integration we performed a combined EOF analysis of anomalous SST, upper-ocean heat content and wind stress. In this case the leading pair of EOF’s is interpreted as a single ENSO mode. This assumption is justified if the explained variance of the two EOF’s is similar, and if the two time series show significant cross-correlations with the maximum
Figure 11. Leading sea surface temperature (SST) and heat-content patterns of a combined empirical orthogonal function (EOF) analysis of SST anomalies, upper-ocean heat-content anomalies and anomalous wind stresses. The data were taken from the coupled 20-year run cop005 (see text).
at a non-zero time lag. The domain is restricted to the tropical Pacific. The first two years of the integration were not used in order to minimize the influence of initial drift. Since the upper-ocean heat content exhibits a non-negligible drift, we found it necessary to detrend the combined dataset pointwise. Along the equator the drift of heat content is very small and is comparable to the drift of SST. Off the equator, however, a drift of up to 0.35 degC per year in terms of average temperature over the upper 250 metres was found. This drift was present for the first five years of the coupled integration. The anomalies were computed with respect to the mean annual cycle, deduced from the coupled run. Each individual dataset was normalized by its total variance. Thus, the combined EOF analysis weights each data type equally. After the analysis, the combined patterns are separated and scaled back to normal units, so that EOF patterns of SST, heat content, and wind stress are obtained.

The patterns are presented in Fig. 11 and Fig. 12, the corresponding time series for the intermediate and the extreme phase are displayed in Fig. 13. Both EOFs together explain about 20% of the total variance. The two time series related to the patterns elucidate that the intermediate phase patterns were present between six and twelve months before an extreme event (represented by the first principal component). The lag correlation between the time series of the first and second EOF reaches a maximum of about 0.5 at a time lag of five to six months, which is too short relative to the 10 to 12 months obtained from observations. The stronger events (years 9, 16, and 18), however, seem to be characterized by a longer time lag, which is in better agreement with observations. The only exception is the El Niño event in year six which had
Figure 13. Time series of the combined empirical orthogonal function (EOF) analysis for the patterns of the intermediate and extreme phases.

almost no precursor. The near periodicity of ENSO is a well known fact, deduced from observations. Since we have integrated the model for only 20 years, it is very hard to obtain a reliable estimate of the dominant ENSO period in the model. For a spectral analysis the time is certainly too short, however, it might give an indication. A cross-spectrum analysis of the two EOF time series yields a peak at about 50 months, and also the autocorrelation of the two time series shows a secondary maximum at about 50 months. Both are indications that the ENSO period of the model is in the correct range, we could only prove it conclusively if we had longer time series.

The spatial patterns agree well overall with the observations (Latif et al. 1997; Fischer 1998), and are consistent with the thermocline depth patterns presented in Neelin et al. (1997). The patterns referred to as extreme phase represent nicely the typical El Niño anomaly patterns. We see strong SST and heat-content anomalies in the equatorial east Pacific, and a substantial weakening of the trade winds in the central Pacific. The intermediate SST pattern hardly shows any signature, which is in agreement with observations, and which also may be explained by the deep thermocline in the western equatorial Pacific. Thus, we may infer that the SST is mainly an oscillating pattern which changes its amplitude, but which does not propagate. The heat-content intermediate pattern, on the other side, is characterized by strong anomalies in the western and central equatorial Pacific which then propagate into the eastern equatorial Pacific. The strong signal at about 10°S in the central Pacific in the intermediate pattern does not agree with the corresponding patterns obtained from observations. It might be related to the strong off-equatorial wind-stress anomalies, seen in the intermediate wind-stress pattern. Besides that, the intermediate wind-stress pattern shows the characteristic wind precursor, the so-called westerly wind bursts in the equatorial west Pacific, north of Papua New Guinea.

From the results presented in this section, we conclude that our coupled model, Giotto-1.5, captures the ENSO dynamics reasonably well, although there are weak points which must be addressed. The main features are all present and are consistent with the widely accepted ENSO theories.

6. Conclusions

We have developed a new coupled ocean–atmosphere GCM, Giotto-1.5. The individual components are formed by the atmosphere model, ECHAM-4, and the ocean
model, MOM-1.2. The model domain is global, and no flux correction is applied. The coupling is active between 60°N and 60°S, and the coupling interval is set to two hours to resolve the diurnal cycle.

A coupled 20-year experiment was carried out, and we discussed the mean state, the annual cycle and the interannual variability in the tropical Pacific. The climate drift, in terms of equatorial SSTs, is extremely small. A linear regression yields a drift of about 1 degC in 46 years.

The mean state is captured reasonably well. In most regions the errors relative to observations do not exceed 1 degC, and only in the equatorial upwelling regions off the Peruvian coast and in the Gulf of Guinea does it exceed 2 degC. Like many other coupled GCMs our model has the tendency to simulate a double ITCZ. We speculate that this is due to insufficient cloud coverage south of the equator, in the eastern Pacific. Positive feedbacks, due to air–sea interactions may cause a warming of the SST south of the equator, and thus lead to more symmetrical rainfall patterns around the equator. The vertical temperature distribution in the ocean in GIOTTO-1.5 is much more realistic than an older version of our model (GIOTTO-1.2). This improvement is due to increased vertical resolution and a modified distribution of the vertical ocean levels. It was found that this had a very large positive impact on the coupled-model performance. The mean state as simulated by the model gives a reasonable representation of the real state.

The seasonal cycle of the model faces some problems in the equatorial Pacific that must be addressed. In the east Pacific it underestimates the seasonal variations of SST by a factor of two, compared with observations, whereas they are overestimated by the same factor in the west Pacific. The problems in the east are probably related to the cloud coverage that was already mentioned above. The equatorial cold tongue penetrates too far into the west Pacific which causes the errors there. This is probably related to too strong equatorial winds, created by ECHAM-4. The annual cycle improved, like the mean state, substantially, after we modified the distribution of the vertical levels in the ocean.

The same holds for the interannual variability in the tropical Pacific. The main features of ENSO dynamics like propagation of heat-content anomalies, reflection of equatorial Kelvin and Rossby waves, and westerly wind bursts are present in the model. Therefore, we are optimistic that the model can be used for ENSO predictions. This is already indicated by some very preliminary forecast results. The number of experiments, however, is at this point much too small for a quantitative assessment of the predictive skill of the model. Weak points are the amplitude of SST anomalies which are underestimated by about 30%, and the interannual variability is too confined to the equator.

An important conclusion is that the vertical discretization of the ocean model must be treated with great care, since it can feed back with the mixing scheme and generate large errors at depth and at the surface. Though our results can be applied strictly only to this case, they raise an issue that also needs to be addressed for other mixing schemes. However, if appropriate care is taken to make the discretization and the mixing scheme consistent, then even with relatively low vertical resolution it is possible to obtain reasonable simulations.

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REFERENCES


Moron, V., Navarra, A., Ward, M. N. and Roeckner, E.

1998 Skill and reproducibility of seasonal rainfall patterns over tropical land areas in GCM simulations with prescribed SST. *Clim. Dyn.*, 14, 83–100


Neelin, J. D. and Dijkstra, H. A.


Paulson, E. A. and Simpson, J. J.


Philander, S. G. H.

1990 *El Niño, La Niña, and the Southern Oscillation*. Academic Press, San Diego, California, USA

Power, S. B., Tsetikitin, F., Coleman, R. A. and Sulaiman, A.


Roeckner, E.


Roeckner, E. and Arpe, K.


Roeckner, E., Oberhuber, J. M., Bacher, A., Christoph, M. and Kirchner, L.

1996 ENSO variability and atmospheric response in a global coupled atmosphere-ocean GCM. *Clim. Dyn.*, 12, 737–754

Rosati, A. and Miyakoda, K.


Rosati, A., Miyakoda, K. and Gudgel, R.

1997 The impact of ocean initial conditions on ENSO forecasting with a coupled model. *Mon. Weather Rev.*, 125, 754–772

Schneider, E. K., Zhu, Z., Giese, B. S., Huang, B., Kirtman, B. P., Shukla, J. and Carton, J. A.


Schopf, P. S. and Suarez, M. J.


Stockdale, T. N., Anderson, D. L. T., Alves, J. and Balmaseda, M.


Stockdale, T. N., Latif, M., Burgers, G. and Wolff, J. O.


Suarez, M. J. and Schopf, P. S.


Wyrtki, K.