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Effects of Systematic Biases in the Stratosphere on the Tropospheric Climate

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Abstract

Numerical models are important tools for studying the tropospheric climate of the present, past or future. Traditionally the atmospheric general circulation models used for climate studies are not very detailed in the representation of the stratosphere since it is assumed that stratospheric feedbacks to the troposphere are negligible for the representation of the tropospheric climate, its mean state, variability or change. This work uses specific experiments to demonstrate that the model representation of the stratosphere has effects on the mean state of temperature and wind in the troposphere, even if the model representation of the troposphere remains unchanged. The analysis of the dynamical forcing terms in the stratosphere shows that differences in stratospheric wave dissipation resulting from differences in the horizontal diffusion schemes, required to account for the different vertical resolutions, affect the Brewer-Dobson circulation and the Hadley circulation. Consequently the radiative balance is modified. Hence coupled atmosphere ocean simulations, which differ in the stratospheric representation, result in different equilibrium states for the tropospheric climate.

1. Introduction

Climate science makes use of observations, theory, and modelling to understand better the functioning of the climate system on Earth in present and past conditions, and to explore possible future climates. Comprehensive climate models developed for this purpose integrate the knowledge on the processes that are relevant for the dynamics of the climate system. Naturally climate models include a number of simplifications at the level of the underlying equations, the range of represented processes, or in the numerical implementation. This study is concerned with the simplification in the representation of the stratosphere in comprehensive climate models, as typically used for climate change studies. Typical climate models, understood here as coupled atmosphere ocean circulation models, include atmospheric general circulation models resolving the atmosphere from the surface up to the middle stratosphere. Typically, the uppermost full pressure level has been placed between about 1 and 10 hPa (Randall et al., 2007, Table 8.1). Even if the upper boundary of the model, i.e. the pressure at the upper boundary of the uppermost layer is 0 hPa, the circulation above the uppermost full level cannot be resolved. If processes above the uppermost model level have significant effects on the circulation in the resolved model domain, then plausible corrections are necessary or the effects will result in systematic errors. A second problem arises from the effects of the artificial upper model boundary on the dynamics in the uppermost layers, which requires special attention in the numerical formulation of the model, if numerical artefacts are to be avoided.

In the construction of early atmospheric general circulation models, which were developed for tropospheric applications, the problem of where to set the upper "lid" of the model was solved based on the concept that the stratosphere was largely driven by the tropospheric dynamics, while stratosphere to troposphere influences seemed unimportant, so that modelling the troposphere would be possible in models having an upward truncation of the atmosphere somewhere in the stratosphere. Condition for such a solution was however to control adverse effects from a lid placed in the stratosphere on upward propagating planetary waves and their effects in the vertically resolved atmosphere. Practically, a number of AGCMs were developed with an uppermost level at or near 10 hPa and only a few stratospheric levels, of which typically the uppermost few layers are exposed to increased numerical diffusion of atmospheric waves. This increased numerical diffusion on the one hand side approximates unresolved processes that dissipate resolved waves propagating upward into this region of a GCM, and on the other hand side are imposed to avoid numerical instabilities and wave reflection.

Such AGCMs have been widely used for climate research with focus on the troposphere, coupling with the ocean and the production of simulations without surface climate drift being the major concerns. However, early works showed nevertheless that the kind of numerical treatment of waves near the upper lid of classical AGCMs can have significant impacts on the simulated dynamics of the troposphere (Boville, 1984). In recent years, research on the dynamics of the atmosphere has shown in recent years that there is indeed a two-way dynamical coupling between the troposphere and the stratosphere, diagnosed from observations, analysed data, and model simulations. This motivates to revisit the problem of where to place the upper lid of AGCMs for tropospheric studies. The question addressed here is how much the representation of the stratosphere influences the tropospheric mean climate, as modelled in coupled atmosphere ocean models. A further question (to be addressed in a later work) is if the relative success in modelling the tropospheric climate in models with an upper atmospheric lid in the middle stratosphere also holds for other climate states, for

example for future conditions, when the stratospheric circulation is expected to be different from the current one, for which the wave damping near the lid has been developed empirically.

This work analyzes the effect of different stratospheric model representations for the ECHAM5 AGCM by comparing two model versions, which are identical in their formulation of processes in the troposphere and in their spatial resolution of the atmosphere between the surface and 100 hPa, in the horizontal as well as in the vertical. The two models however differ in the vertical resolution of the atmosphere above the 100 hPa level, i.e. in the stratosphere and mesosphere, and in the formulation of dissipative processes that affect the simulation of upward propagating waves, and hence wave mean flow interaction close to the model top. These two atmospheric model versions are used for climate simulations, i.e. for coupled atmosphere ocean simulations to test the zero hypothesis that stratospheric model differences have no impact on the tropospheric climate, and in atmosphere only simulations, where the tropospheric climate is largely fixed by given prescribed lower boundary conditions. While the former set of climate simulations allows interactions between stratosphere, troposphere and the troposphere and the ocean, the latter set of simulations suppresses such feedbacks and hence more clearly shows the regions where the model differences in the stratosphere cause systematic differences in circulation and temperature, and where they might extend to the troposphere.

The following sections are organized as follows. Section 2 describes the models and the experiments. Section 3 compares the simulations of the coupled atmosphere ocean simulations using a low top and high top atmospheric GCM. Section 4 compares a pair of simulations using prescribed sea surface temperature and sea ice instead of a coupled ocean model to confirm the stratospheric origin of differences identified in section 3. Section 5 summaries and concludes.

2. Description of models and experiments

a) Vertical grids of the low top and high top atmospheric GCMs

The atmospheric GCM used for this study is the ECHAM5 model, which is used in two basic configurations: As tropospheric model with a top full level at 10 hPa (Roeckner et al., 2006), and as middle atmosphere model with a top full level at 0.01 hPa (Manzini et al., 2006), referred to as low top and high top model, respectively. Both models have a top half level at 0 hPa, hence include the full atmospheric mass, and both models are used here at horizontal resolution T63. Note that half levels are defined at the interface of atmospheric layers; while full levels are the mean pressure of the respective atmospheric layers.

The ECHAM5 GCM makes use of a vertical (Lorentz staggered) hybrid sigma pressure grid, which is terrain-following near the surface and purely pressure based above the lower stratosphere. The grid is defined by A and B coefficients representing the pressure and sigma portions of the pressure at half-levels $p_{h,k}=A_k +B_k*p_s$ at the interfaces between layers. Table 1 shows the coefficients defining the grids used here for the low top and high top atmospheres, respectively. The low top model resolves the atmosphere in 31 layers, of which 26 full levels are placed between the surface and approximately 100 hPa and 5 further full levels extend upward to 10 hPa. This 31 layer grid is identical to that used for the ERA-15 reanalysis, and has been used at MPI-M as standard vertical grid for tropospheric ECHAM5 simulations (Roeckner et al., 2006) and coupled ECHAM5/MPIOM climate simulations at horizontal

resolution T63 (Jungclaus et al., 2006). The high top, middle atmosphere ECHAM5 (MAECHAM5) model is used here with a 47 layer grid that has been derived from the 31 layer grid with two objectives: (1) to be identical with the 31 layer grid between the surface and 100 hPa, so that the vertical resolution of this part of the atmosphere is identical in both grids, and (2) to extend upward to a top full level pressure of 0.01 hPa, as done in earlier versions of the MAECHAM5 models based on different lower troposphere vertical grids (Manzini et al., 2006, Giorgetta et al., 2006). Hence, the 47 layer grid has the same 26 layers between the surface and approximately 100 hPa as the 31 layer grid. Higher up the vertical grid of the high top model has 21 more layers, of which 11 are between approximately 100 hPa and 10 hPa, and 10 layers between 10 hPa and 0.01 hPa. Both vertical grids have a top half level of 0 Pa, so that the full atmospheric mass is represented.

b) Resolved and unresolved wave mean flow interaction

While the low and high top models share the description of tropospheric processes, they differ in the description of dynamical processes impacting the simulation of the wave mean flow interaction in the stratosphere and mesosphere. Planetary waves are excited in both models and propagate upward into the middle atmosphere when dynamic conditions are fulfilled. In nature such waves propagate through the middle and upper atmosphere until they are destroyed through dissipative processes and hence act on the mean flow. Furthermore, gravity waves are central for the dynamics of the middle atmosphere, but their effects on the resolved flow (mean flow and resolved waves) of atmospheric GCMs generally needs to be parameterized. These dynamical processes are simulated differently in the low top and high top models.

In the low top model, planetary waves generally reach the upper lid before being dissipated. Hence their dissipation is numerically enforced by increasing horizontal diffusion of vorticity and divergence. Specifically the horizontal diffusion is computed as $dx/dt = -(-1)^q \cdot K_x \cdot 2^q \cdot x$, where $2q=8$ except in the uppermost 5 levels, where the order of hyper-diffusion is reduced to $2q=(6,4,2,2,2)$ at (90, 70, 50, 30, 10 hPa) so that the diffusion is less scale specific, i.e. more effective on larger scale waves near the upper lid. The coefficients and the orders of the diffusion operator are chosen empirically to avoid model instabilities and vertical wave reflection. The low top model includes a parameterization of the drag exerted by unresolved breaking gravity waves related to subgrid scale orographic features (Lott and Miller, 1999).

The high top model has a vertical domain that is deep enough to resolve refraction and dissipation of planetarywaves in the stratosphere. Hence horizontal diffusion is applied equally at all levels, i.e. $2q=8$ is used at all levels. The high top model parameterizes the orographic gravity wave drag as in the low top model, and in addition it includes a parameterization for the propagation and dissipation of unresolved gravity waves from atmospheric sources (Hines, 1997). These atmospheric gravity waves typically break and force the resolved flow in the upper stratosphere and mesosphere. This process is not represented in the low top model because the vertical grid does not resolve these layers.

c) Coupled atmosphere ocean models

For the 2nd part of this study, the atmospheric GCM ECHAM5 is coupled to the ocean GCM MPIOM for both vertical grids. The coupling is computed at daily intervals and technically implemented through the OASIS3 coupler. The coupling of the low top ECHAM5 model to

the MPIOM ocean model at 1.5° resolution is described in Jungclaus et al.(2006). This model has been used extensively for climate change studies, as those discussed in the IPCC AR4 report of Working Group 1. The coupled model including the high top atmosphere is identical to the previous one, except for the vertical grid above 110 hPa. Note that both coupled models use the same parameter setting for the tropospheric processes, so that the same climate is to be expected, if the stratospheric differences between the low and high top atmospheric models were non-relevant for the tropospheric circulation. If however a significant difference occurs between simulations using the low top and high top coupled models, then it is explained by the differences in stratospheric dynamics caused by the different vertical resolutions and the differently simulated wave mean flow interactions.

3. Coupled atmosphere ocean simulations

Coupled atmosphere ocean simulations were performed for constant pre-industrial boundary conditions. A 100 year portion of an existing preindustrial control simulations made for CMIP3/IPCC-AR4 is used here as a reference for the climate of the low top climate model. This 100 year data set is referred to as CM31. A new simulation was performed using the high top climate model. This simulation, CM47, was started from an ocean state of the low top CM31 climate simulation and a newly initialized atmosphere. Due to the differences in the atmospheric initial state it is expected that the CM47 simulation differs from CM31 in details, but not in a statistically significant manner if the zero hypothesis holds, that the model differences in the stratosphere have no effect on the tropospheric climate.

The CM47 simulation developed quickly a climate drift that lasted about 50 years and lead to a global warming of about 0.5 K compared to the CM31 climate. The CM47 simulation was then extended to 160 years so that the last 100 years, which do not expose any systematic drift, can be compared to the 100 years of data of the CM31 simulation. A first result consists in the statistically significant change of the global mean near surface temperature by 0.5°C, thus rebutting the zero hypothesis stated above. Consequently differences in the stratospheric climate, caused here by different model representations of the stratosphere, lead to changes in the troposphere and ocean.

The following comparison is based on the climatological averages of the 100 years of data of the CM31 and CM47 simulations. Annual and zonal mean temperature and zonal wind are displayed in Figure 1 and Figure 2, respectively, for both simulations and for their difference on the common vertical domain reaching from the surface to 10 hPa.

Figure 1 shows the climatological annual and zonal mean temperature of CM31 (upper left and right) and CM47 (lower left) and their difference (lower right). A strong polar warming (6 K at the North Pole and 12 K at the South Pole) in the stratosphere and cooling equatorward of about 50° latitude is apparent. This warming in CM47 vs. CM31 corrects the known strong stratospheric high latitude winter time bias in temperature of the low top ECHAM5 model used in the CM31 simulation (not shown). In the tropical latitudes the stratospheric temperature is decreased in CM47 compared to CM31, up to typically 2 K at 20 hPa and higher. This pattern of opposite temperature differences in high latitudes and low latitudes results from a change in the mean meridional circulation, as it can be diagnosed in the transformed Eulerian mean streamfunction (not shown). This requires systematic differences in the wave mean flow interaction, as it will be diagnosed more in detail in section 4.

The tropospheric annual and zonal mean temperatures are approximately 0.5 K warmer in CM47 than in CM31. Note the maxima in the upper tropical troposphere and in the Arctic surface temperature, where the difference reaches 1 K. This signature resembles the stronger climate warming of both regions compared to the global mean near surface warming as found in transient climate projections. The Arctic response is accompanied by a decrease in sea ice area and volume (not shown), thus amplifying the temperature response by a positive albedo climate feedback.

Figure 2 shows the climatological annual and zonal mean zonal wind of CM31 (upper left and right) and CM47 (lower left) and their difference (lower right). Largest differences occur in the stratosphere, where the westerlies are generally weaker in CM47, consistently with the mean temperature changes. The equatorial latitudes show a strong positive difference at 10 and 20 hPa, which results from the annual mean differences of the SAOs in CM47 and CM31. Important are the statistically significant positive and negative differences in zonal mean zonal wind that extend throughout the troposphere to the surface, indicating a small equatorward shift in the subtropical jets. These adjustments are part of the response of the tropospheric climate in the coupled atmosphere ocean system to the changed stratospheric dynamics.

The coupled atmosphere ocean simulations, using models with differences restricted to the stratosphere above the 100 hPa level, thus show a statistically significant response in wind and temperature of the troposphere down to the surface, and implied oceanic changes. Based on the design of the experiments, only the different stratospheric representations of the low top and high top model can explain these differences.

4. Atmospheric simulations with prescribed SST and sea ice conditions

The previous section demonstrated that differences in the model representation of the stratosphere result in a statistically significant change in circulation and temperature throughout the whole atmosphere and adjustments in the ocean (not shown). From the differences it is however not immediately clear to which degree their structure results from the initial differences in stratospheric dynamics caused by the model differences or from a troposphere stratosphere interaction. Therefore a second pair of simulations, AM31 and AM47, was performed using the low and high top atmosphere models, respectively, with prescribed sea surface temperature and sea ice data of 1979 to 1996 (AMIP2 boundary conditions). The use of prescribed boundary conditions instead of a dynamic coupled ocean model effectively prevents near surface climate change over the oceans and through advective effects also over land masses. Stratospheric effects on the mean state of the troposphere will generally be limited to the upper troposphere, and feedbacks from the troposphere back to the stratosphere are suppressed, so that the immediate effects of the model differences concerning stratospheric dynamics can be identified more directly.

Figure 3 shows the annual and zonal mean temperature differences between the high and low top models, for the coupled simulations CM47 and CM31 (left = Fig.1 lower right) and for the atmosphere only simulations AM47 and AM31 (right). Qualitatively, the differences are highly similar in the whole stratosphere, and quantitatively they are nearly identical between 10 and 70 hPa. Stratospheric quantitative differences exist in the extratropics, where the temperature differences are stronger between the atmosphere only simulations AM47 and AM31 than between the coupled simulations CM47 and CM31. Apparently the tropospheric

differences differ. As expected the atmosphere only simulations AM47 and AM31 are statistically equivalent close to the surface, where the same sea surface temperature and sea ice boundary conditions are prescribed. Note however the statistically significant signal in the high northern latitudes in the middle troposphere. This might results from the dynamic coupling of the wintertime high latitude stratosphere and troposphere by annular modes, and their modification by the imposed differences in stratospheric dynamics between the high and low top models. A general tropospheric warming is found only between CM47 and CM31, where the ocean is coupled interactively to the atmosphere. This coupling is necessary to investigate downward effects of systematic changes in the stratosphere on the troposphere. Here the systematic differences were introduced by changes of the modelled stratospheric dynamics, as discussed in section 2, and the tropospheric climate differences between the coupled high and low top models stabilized within 60 years of integration. The systematic differences in the stratospheric dynamics leading to this climate response were diagnosed in the atmosphere only simulations AM47 and AM31.

Figure 4 shows differences between AM47 and AM31 of annual and zonal mean tendencies in zonal wind, as diagnosed from the divergence of the EP flux (computed 6-hourly), the orographic gravity wave drag, the non-orographic gravity wave drag, and the horizontal divergence. Largest differences are found in the diagnosed EP-flux divergence, which is a direct consequence of the different damping of the resolved wave spectrum in the uppermost levels, due to the different horizontal diffusion operators, cf. section 2. Especially at 50 hPa the low top model damps waves significantly more than the high top model, resulting in stronger negative divergence of the EP flux and in a stronger negative drag. Hence the difference AM47-AM31 is strongly positive. At 30 and 10 hPa, in the upward shadow of the strong positive differences at 50 hPa, the differences are of opposite sign. Thus the negative drag by the divergence of the diagnosed EP-flux is stronger in the high top model than in the low top model, where resolved waves have already been dissipated at 50 hPa. This relatively strong wave mean flow interaction of the high top model contributes to the reduced zonal mean zonal wind found in Fig. 2. The different implementation of the horizontal diffusion scheme hence results in significant differences in wave dissipation and hence wave mean flow interaction in the stratosphere. Beyond this, the implementation of the horizontal diffusion scheme differs also in its effect on the zonal mean, which is damped towards 0 wind in the low top model, while this effect does not occur in the high top model (Fig. 4, lower left). This dynamical forcing difference mainly occurs in the Antarctic polar jet, where zonal winds are strongest, and may induce a spurious tropospheric stratospheric connection.

The orographic gravity wave drag, damping the strong westerlies at 10 and 20 hPa, is generally stronger in the low top model (except near 60°S, where no land exists). This difference occurs simply because the westerly winds at these levels and latitudes are stronger in the low top model. The non-orographic gravity wave drag exists only in the high top model and exerts a small negative drag at 50 hPa above the subtropical jets. Figure 4 shows that the different implementation of the horizontal diffusion scheme is a major cause of the differences in the stratospheric dynamics.

5. Conclusions

Past climate simulations generally neglected the stratospheric dynamics, usually because of limited computational resources. The stratosphere was considered as a boundary region, where typically horizontal diffusion schemes were modified near the upper lid to avoid numerical problems and to simulate a possibly acceptable mean stratospheric circulation, thus

creating a reasonable upper boundary for the troposphere. The ECHAM5 model in its traditional low top setup (uppermost full level at 10 hPa) is a typical model of this kind. The increased understanding of the dynamic coupling of the troposphere and stratosphere by annular modes of the high latitudes makes this traditional approach however questionable, especially in the context of a changing climate, with different dynamical signals in the troposphere and stratosphere. Therefore this study compares two coupled atmosphere ocean simulations, which differ only in the vertical resolution of the atmosphere above 100 hPa and in the representation of resolved and unresolved wave mean flow interaction in this part of the atmosphere. The active coupling of atmosphere and ocean is an important choice, because only this way the dynamical feedback from the stratosphere can be expected to cause changes in tropospheric climate, in response to changes in stratospheric dynamics, if such effects exist. The mean tropospheric climate of the coupled high top model indeed deviated statistically significantly from the mean climate of the coupled low top model, thus showing that dynamic differences in the stratosphere, caused here by model differences restricted to the stratosphere, cannot be neglected.

The comparison of the corresponding atmosphere only simulations shows highly similar differences in the stratospheric mean circulation and temperature pattern, but no tropospheric climate change. This shows that dynamic feedbacks from the stratosphere to the troposphere can only unfold fully if the lower troposphere is not locked to a climate defined by prescribed lower boundary conditions.

The analysis of the dynamical forcing terms finally shows that the different implementation of the horizontal diffusion scheme in the low and high top model is a major cause of the diagnosed changes in the stratospheric circulation and temperature. These differences in the horizontal diffusion schemes are a direct consequence of the different vertical grids, where the horizontal diffusion scheme of the low top model is modified to control the resolved waves near the upper model lid. This enforced wave damping is however non-physical. The high top model resolves the stratosphere and partly the mesosphere, hence it resolves the large scale dynamical variability of the stratosphere, such as sudden stratospheric warmings, virtually lacking in the low top model (Cagnazzo and Manzini 2009). Additionally, the high top model includes a physically based parameterization of non-orographic gravity wave propagation and dissipation, so that wave mean flow interaction and its downward effects are not affected by spurious influences of non-momentum conserving parameterization of mechanical dissipation (Shaw and Shepherd, 2007. A non-physical modification of the horizontal diffusion scheme can thus be avoided.

Based on this study it is recommendable to use atmospheric models that resolve the dynamical stratosphere reasonably well and represent wave mean flow interaction in a physically sound way also for climate studies. Not only will the stratospheric circulation be better represented, but also the dynamical interaction of the troposphere and stratosphere. This study focused on the global effects of extratopical dynamical processes, because none of the models simulates the quasi-biennial oscillation of the tropics. Further feedbacks of the stratosphere to the troposphere may therefore be expected once a climate model includes an atmospheric model that simulates the QBO as an internal mode of variability. While the ECHAM5 model can simulate the QBO at sufficiently high vertical resolution, this has not been used for this study on purpose. The tropospheric portion of the high resolution vertical grid differs from the tropospheric portion of the low top model, and hence would not have allowed the clean separation of the tropospheric and stratospheric model representations.

References

Cagnazzo, C. and E. Manzini, Impact of the Stratosphere on the Winter Tropospheric Teleconnections between ENSO and the North Atlantic and European Region, J. Climate, 22, 1223-1238, 2009.

Giorgetta, M. A., E. Manzini, E. Roeckner, M. Esch, and L. Bengtsson, Climatology and forcing of the quasi-biennial oscillation in the MAECHAM5 model, J. Climate, 19, 3882-3901, 2006.

Hines, C. O., Doppler-spread parameterization of gravity-wave momentum deposition in the middle atmosphere. Part 1: Basic formulation, Journal of Atmospheric and Solar-Terrestrial Physics, 59, 371-386, 1997.

Lott, F. and Miller, M. J., A new-subgrid-scale orographic drag parameterization: Its formulation and testing. Quart. J. Roy. Meteor. Soc., 123, 101-127, 1997.

Manzini, E., M.A. Giorgetta, M. Esch, L. Kornblueh, and E. Roeckner, The influence of sea surface temperatures on the Northern winter stratosphere: Ensemble simulations with the MAECHAM5 model, J. Climate, 19, 3863-3881, 2006.

Randall, D.A., R.A. Wood, S. Bony, R. Colman, T. Fichefet, J. Fyfe, V. Kattsov, A. Pitman, J. Shukla, J. Srinivasan, R.J. Stouffer, A. Sumi and K.E. Taylor, Climate models and their evaluation. In: Climate change 2007: The physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M.Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, 2007.

Roeckner, E., R. Brokopf, M. Esch, M. Giorgetta, S. Hagemann, L. Kornblueh, E. Manzini, U. Schlese, and U. Schulzweida, Sensitivity of simulated climate to horizontal and vertical resolution in the ECHAM5 atmosphere model, J. Climate, 19, 3771-3791, 2006.

Shaw, T. A., and T. G. Shepherd, Angular momentum conservation and gravity wave drag parameterization: Implicaservation and gravity wave drag parameterization: Implications for climate models, J. Atmos. Sci., 64, 190-203, 2007.

Table 1. Pressure and sigma coefficients of the 31 layer grid and the 47 layer grid of the low and high top ECHAM5 GCM used in this study. k is the half level index starting at the top of the model (k=0). A is the pressure value in [Pa], and B is the sigma coefficient in [Pa/Pa]. A and B values in italics are identical in both grids.

| 31 layer grid of low top model			47 layer grid of high top model		
k	A [Pa]	B [Pa/Pa]	k	A [Pa]	B [Pa/Pa]
			0	*0*	*0*
			1	1.989185	0
			2	6.572090	0
			3	15.67390	0
			4	30.62428	0
			5	54.54572	0
			6	92.55883	0
			7	150.5047	0
			8	235.3274	0
			9	356.1003	0
			10	523.9195	0
			11	751.0429	0
			12	1051.137	0
			13	1438.988	0
			14	1930.177	0
			15	2540.697	0
0	*0*	*0*	16	3286.553	0
1	2000.00	0	17	4199.574	0
2	4000.00	0	18	5303.957	0
3	6000.00	0	19	6624.704	0
4	8000.00	0	20	8187.185	0
5	*9976.137*	*0.0004*	21	*9976.137*	*0.0004*
6	*11820.54*	*0.0029*	22	*11820.54*	*0.0029*
7	*13431.39*	*0.0092*	23	*13431.39*	*0.0092*
8	*14736.36*	*0.0203*	24	*14736.36*	*0.0203*
9	*15689.21*	*0.0370*	25	*15689.21*	*0.0370*
10	*16266.61*	*0.0595*	26	*16266.61*	*0.0595*
11	*16465.00*	*0.0879*	27	*16465.00*	*0.0879*
12	*16297.62*	*0.1220*	28	*16297.62*	*0.1220*
13	*15791.60*	*0.1614*	29	*15791.60*	*0.1614*
14	*14985.27*	*0.2057*	30	*14985.27*	*0.2057*
15	*13925.52*	*0.2542*	31	*13925.52*	*0.2542*
16	*12665.29*	*0.3062*	32	*12665.29*	*0.3062*
17	*1126.123*	*0.3611*	33	*1126.123*	*0.3611*
18	*9771.406*	*0.4182*	34	*9771.406*	*0.4182*
19	*8253.211*	*0.4767*	35	*8253.211*	*0.4767*
20	*6761.340*	*0.5359*	36	*6761.340*	*0.5359*
21	*5345.914*	*0.5951*	37	*5345.914*	*0.5951*
22	*4050.718*	*0.6536*	38	*4050.718*	*0.6536*
23	*2911.569*	*0.7106*	39	*2911.569*	*0.7106*
24	*1954.805*	*0.7654*	40	*1954.805*	*0.7654*
25	*1195.890*	*0.8172*	41	*1195.890*	*0.8172*
26	*638.1489*	*0.8650*	42	*638.1489*	*0.8650*
27	*271.6265*	*0.9077*	43	*271.6265*	*0.9077*
28	*72.06360*	*0.9442*	44	*72.06360*	*0.9442*
29	*0*	*0.9730*	45	*0*	*0.9730*
30	*0*	*0.9923*	46	*0*	*0.9923*
31	*0*	*1.000*	47	*0*	*1.000*
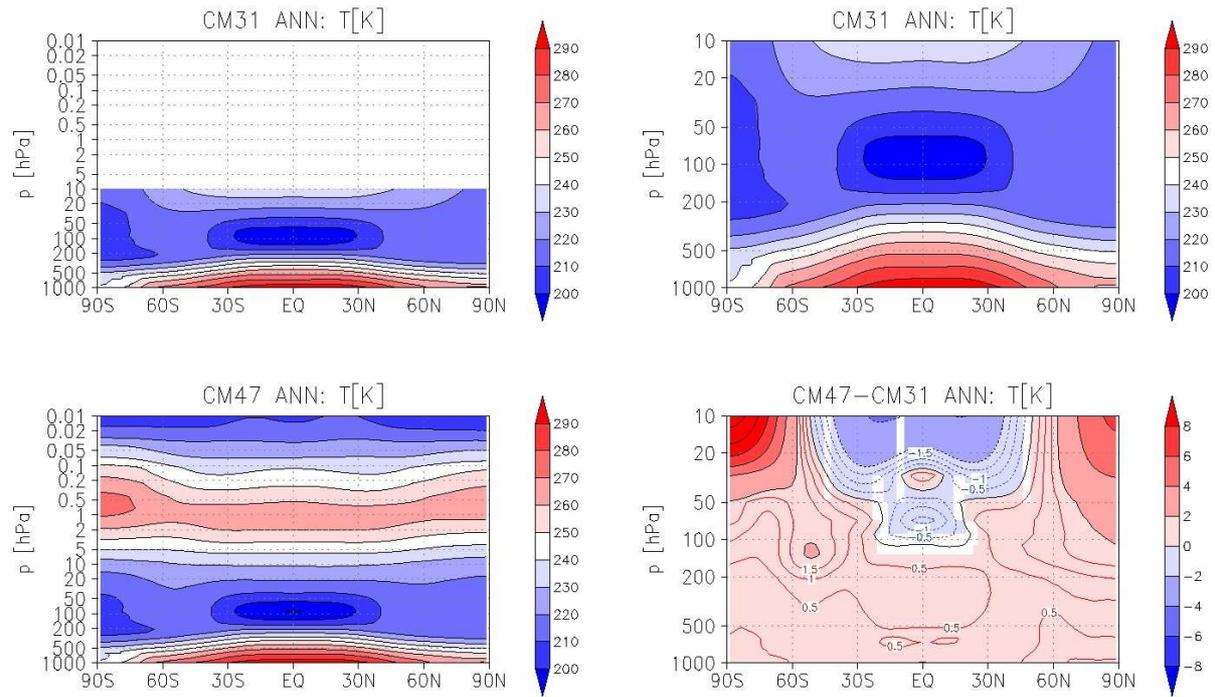
Figure 1. Annual and zonal mean temperature T in [K] in the coupled ECHAM5/MPIOM climate model (CM). Upper left: 31 layer atmosphere up to 10 hPa; Lower left: 47 layer atmosphere up to 0.01 hPa; Upper right: as upper left, but in the range of 1000 to 10 hPa; Lower right: difference between the 47 and 31 layer atmospheres in the range of 1000 to 10 hPa. Color shading indicated statistically significant differences following a two-sided Student t-test at a 95% confidence level.
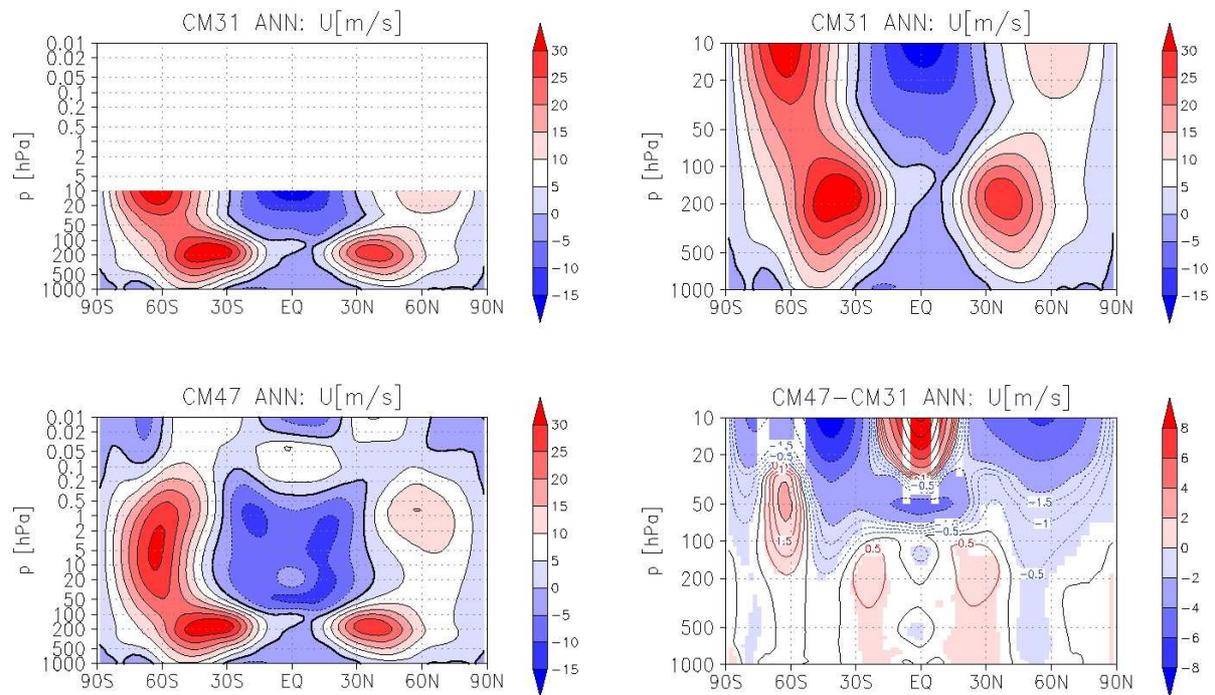

Figure 2. Annual and zonal mean zonal wind U in [m/s] in the coupled ECHAM5/MPIOM climate model (CM). Upper left: 31 layer atmosphere up to 10 hPa; Lower left: 47 layer atmosphere up to 0.01 hPa; Upper right: as upper left, but in the range of 1000 to 10 hPa; Lower right: difference between the 47 and 31 layer atmospheres in the range of 1000 to 10 hPa. Color shading indicated statistically significant differences following a two-sided Student t-test at a 95% confidence level.
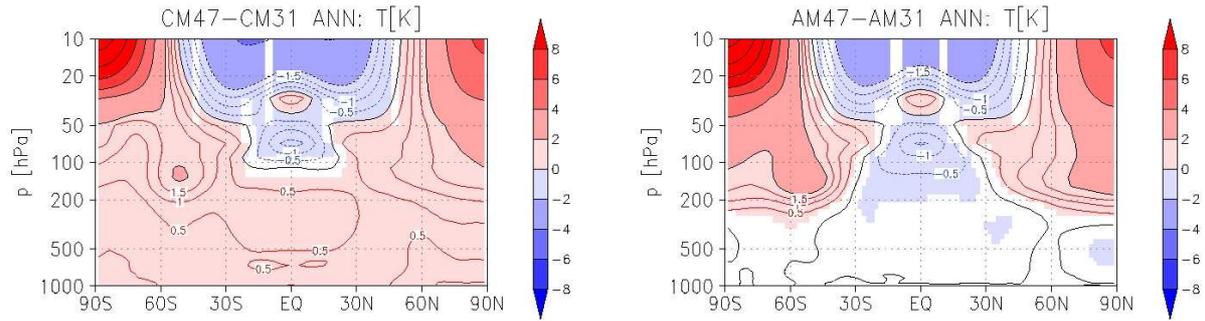
Figure 3. Annual and zonal mean temperature differences in [K]. Left: difference between the coupled ECHAM5/MPIOM simulations CM47 and CM31; Right: difference between the atmosphere only ECHAM5 simulations AM47 and AM31. Color shading indicated statistically significant differences following a two-sided Student t-test at a 95% confidence level.
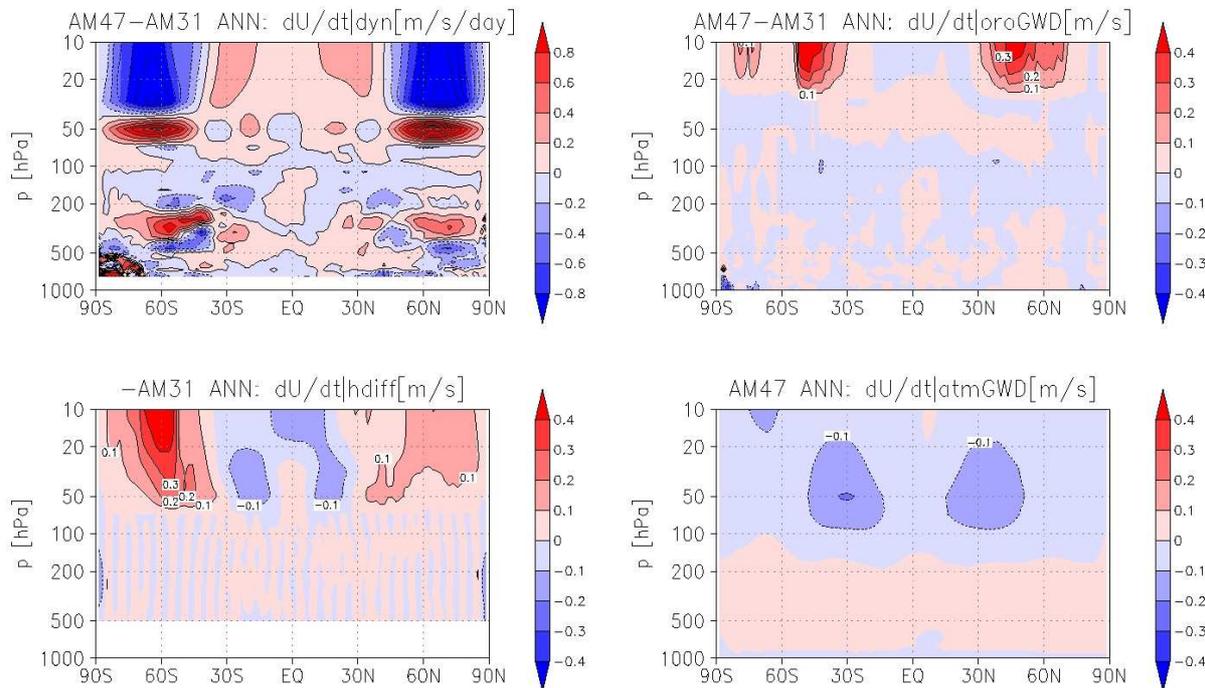

Figure 4. Differences of annual and zonal mean tendencies dU/dt between the 47 layer and 31 layer ECHAM5 simulations of resolved dynamics (=…) (upper left), horizontal diffusion (lower left), orographic gravity wave drag (upper right), and atmospheric gravity wave drag (lower right).