Danish Climate Centre

DMI, Ministry of Transport

Using the nudging technique to estimate climate model forcing residuals.

A contribution to the ACE Scientific Support Study

Eigil Kaas and Annette Guldberg

Report 01-6

Using the nudging technique to estimate climate model forcing residuals. A contribution to the ACE Scientific Support Study

This report constitutes a contribution to the Scientific Support Study (Work Package 4100a) of the "Atmospheric Climate Experiment" (ACE). ACE is an Explorer Opportunity Mission in the ESA Living Planet Programme and its purpose is to monitor variations and changes in the climate of the Earth based on the GNSS-LEO radio-occultation (RO) technique (see Kursinski et al. 1997). At the time of publication of the present report, ACE has the status of a so called "hot stand by" mission.

Danish Climate Centre, Report 01-6

Eigil Kaas and Annette Guldberg

ISSN: 1399-1957 (online version)

ISBN: 87-7478-445-5

Danish Meteorological Institute
Lyngbyvej 100
DK-2100 Copenhagen Ø
Denmark

Phone: +45 3915 7500
Fax: +45 3927 1080
www.dmi.dk

Using the nudging technique to estimate climate model forcing residuals. A contribution to the ACE Scientific Support Study[1]

EIGIL KAAS, ANNETTE GULDBERG

Danish Climate Centre, Danish Meteorological Institute, Copenhagen, Denmark

ABSTRACT

This report includes an overview of the usage of the nudging technique for detection of forcing residuals in the atmospheric component of climate models. Such residuals can be used for two different purposes in climate research: improvement of climate models and quantification of changing external forcing of climate. A number of examples illustrating potential problems with forcing residuals determined this way are given. Finally a short discussion on the needed accuracy of the underlying analyses of the meteorological parameters is provided. It is concluded that homogeneity and small bias are needed key features for detection of variations in the external forcing of climate. The GNSS-LEO radio-occultation technique will provide such data. Furthermore, the high level of homogeneity of these data makes them ideal for classical climate monitoring on decadal or longer time scales including detection of trends.

1. Introduction

Variations and trends in the climate can be due to

1. internal chaotic processes in the climate system,
2. varying natural external forcing of climate (primarily from solar variations and variations in the volcanic activity)
3. varying anthropogenic forcing of climate (primarily changing the concentrations of the well mixed greenhouse gases and the concentration of tropospheric and stratospheric ozone as well as the radiative direct and indirect effects of tropospheric aerosols).

According to climate model simulations (see e.g., IPCC 2001) of the present day climate the internal climate processes (1.) are generally dominating the variations/trends from one year to the next. But when averaging over periods longer than 10-20 years or so, and in particular when considering global means, trends related to internal processes are considered to be smaller than the typical forced trends over similar periods. It should be noted, however, that past climates, e.g. during the last ice age, experienced much larger and longer lasting internal variations than today.

In the recent years with very strong focus on the anthropogenic impact on climate it is obviously of key importance to assess if the observed trends are due to human activities, or if they can be considered natural, i.e. due to chaotic processes and/or natural external forcing. Unfortunately, this distinction or separation is difficult to obtain from observations alone, because there is a large uncertainty in our knowledge of the magnitude of the three contributions listed above. Ideally, if we had an indefinitely long and accurate observational record and some fair estimate of the simultaneous variations in external forcing in the past, it would be possible to estimate the level and temporal behaviour of natural variability, and thereby to deduct if the recent variations and trends were "unnatural". In practice, however, we do not have such data available: a direct detection of anthropogenic climate change (e.g. an expected temperature trend of 0.1-0.2 K/decade) requires accurate and homogenous observations for decades in order to filter out the noise related to internal climate variability.

[1] This report constitutes a contribution to the Scientific Support Study (Work Package 4100a) of the “Atmospheric Climate Experiment” (ACE). ACE is an Explorer Opportunity Mission in the ESA Living Planet Programme and its purpose is to monitor variations and changes in the climate of the Earth based on the GNSS-LEO radio-occultation (RO) technique (see Kursinski et al. 1997). At the time of publication of the present report, ACE has the status of a so called “hot stand by” mission.

The lifetime of individual satellites is limited to a few years for known reasons. This implies a major and general problem when using satellite data for climate monitoring and in particular for attribution of trends to anthropogenic or natural forcing[2]: namely that data compiled from several consecutive missions cannot be considered homogenous[3]. To homogenize satellite data by calibration of overlapping consecutive mission periods is not a trivial task. This is basically because the vertical soundings (e.g., via the HIRS instrument on NOAA's TIROS satellites) are not self-calibrating and need a complicated calibration from one generation of satellites to the next. This cannot always be achieved in a proper way because of missing overlapping mission periods and because there are temporal drifts in the sensitivity of the instruments and in the orbits. In fact, only the Microwave Sounding Units (MSU instruments) have been calibrated with reasonable accuracy (Christy et al., 2000), although there have been many problems and considerable criticism of this work (Hurrell and Trenberth, 1998). A main problem with the MSU data is that there are considerable overlaps in the vertical weighting functions for the individual spectral channels. This strongly limits the effective number of degrees of freedom in the vertical. Obviously there is a need for improved satellite based climate monitoring with better vertical resolution, and which is free of measurement bias or at least with an *a priori* reproducible bias from one mission to the next enabling direct inter-comparison and fewer inter-calibration problems.

However, in climate research observed data may be used in ways additional to the classical estimation of means and trends: for minimization of forcing residuals in climate models and for detection of varying external *forcing* of climate. These applications will enhance our understanding of climate variability, assist in the attribution of climate change to human activity, and reduce the uncertainty in climate model simulations of potential future anthropogenic climate change. In both cases, it is needed to assimilate the observed data into a climate model and thereby to estimate the forcing residual of the model. One of the most obvious ways of assimilation is the simple nudging or Newtonian Relaxation towards numerical weather analyses – the so-called re-analyses.

In section 2 we discuss the usage of atmospheric data for estimation of forcing residuals in the atmospheric component of climate models and how this can be used as a guideline for improvement of the models. Section 3 presents results from a number of numerical experiments illustrating potential problems with forcing residuals obtained with the nudging technique. Section 4 includes a discussion of the potential use of forcing residuals in atmospheric models for estimation of varying external forcing of climate. In Section 5 we summarise the findings.

[2] This reservation is less important for the stratosphere which is responding quickly to changes in radiative processes partly because it is influenced less than the troposphere by the coupling to the world oceans.

[3] In climate research the notation homogenous is used to describe varying observed data reflecting real changes in nature as opposed to changes due to changing instrumental accuracies/drifts, local environmental conditions near the instrument, and analysis and retrieval routines (see e.g., Karl et al., 1993).

2. Climate model improvement

Climate models are - at least intentionally - developed from basic (micro-scale) physical laws and principles. However, in practice the application on large scales of the micro-scale physical laws is a complicated matter requiring assumptions and integral constraints. This means that, although physically based, all climate models include a number of closure variables, which - again intentionally - are only weakly dependent on the actual climate (assuming no extreme excursions from the present climate). The closure variables generally have to be determined empirically. For some processes, e.g., boundary layer turbulence, local observational campaigns can deliver an appropriate tuning data set for estimating the closure constants. However, a determination at macro-scale/global is required for most processes such as formation and dissipation of clouds, their radiative properties, the water vapour processes, which control them, and the lower tropospheric radiative balance. Traditionally the macro-scale tuning has been achieved by comparing observed mean climate variables with corresponding variables produced in long simulations with the model to be tuned. In the IPCC's review on the validation of climate models (IPCC, 2001), it is clear that most climate model validation activities focus on this simple and naive matching of long term mean atmospheric states between model and data – normally supplemented by comparisons of the simulated and observed climatic variability. Therefore, improved model performance has been largely accomplished through the tuning of loosely-constrained "climate parameters" (the tuning parameters). As a result it is difficult to say whether we have any better understanding of the feedbacks internal to the atmospheric system largely related to clouds and radiation after having tuned the model - or if we simply obtained this apparent improvement through the introduction of other, partly compensating, errors. Without a better understanding of these feedbacks and any others internal to the climate system, we are left with models not fully suited for predicting future climate change.

There are, however, alternative approaches to improving models, which should reduce the risk of error compensation. Using assimilation of "observed" data (with errors) into a climate model it is, in principle, possible to distil the *forcing* in the prognostic differential equations (e.g. the first equation of thermodynamics) not already build in to the assimilating model. This can be done since there is typically a tiny but systematic offset (R) between the initial temporal development in atmospheric forecasts and in the analyses (observations):

$$\left.\frac{\partial \psi}{\partial t}\right|_O = \left.\frac{\partial \psi}{\partial t}\right|_M + R \qquad (1)$$

where ψ is a prognostic variable (temperature, wind components, humidity) at a given location in the atmosphere and subscript O denotes the instantaneous temporal tendency in the analyses (observations) for a given total atmospheric state vector. Similarly, M indicates the temporal tendency simulated by the assimilating model given the same observed atmospheric state vector. This means that R is a tendency residual (i.e., a residual forcing) not simulated by the model as it should have been, i.e. a proxy for the model deficiency. Improvement of atmospheric models by minimisation of R has been used by e.g., Klinker and Sardeshmukh (1992), Kaas et al., (1999) and D'Andrea and Vautard (1999). Due to different technical obstacles, e.g. related to misinterpretation of mainly moisture spin up in the assimilating model and to some extent to gravity wave noise, there are certain problems in calculating R at a given time for the three dimensional atmosphere. The European research project POTENTIALS (Kaas et al., 2000) investigated different possibilities to solve these problems. The results from POTENTIALS as well as other studies suggest that a four-dimensional data assimilation technique is needed, i.e. both time and space are considered simultaneously. The simplest technique is a nudging or relaxation technique which allows assimilation of data at non-synoptic hours and which was used for many years at the U.K. Met Office for data assimilation in numerical weather prediction (NWP). A variational data assimilation may, however, be relevant since it provides a very elegant and accurate way to estimate R. See e.g. D'Andrea and Vautard (1999) for a description of this method in the framework of a simple 3-level quasi-geostrophic model.

In addition to being useful for minimization of the forcing residual (R) in the dynamical equations of atmospheric models, a long term assimilation of observed data is useful for tuning and development of chemical schemes to be coupled into general circulation models of the atmosphere. This is because the actual model state can be kept very close to the varying observed state during the assimilation, i.e., one retains the integrity of the full model, but the environment is forced to be approximately as observed.

It is generally noted that systematic errors in climate models are considered a major source of uncertainty in model-based estimates of anthropogenic climate change. Therefore the improvement of atmospheric models via data assimilation techniques is an important climate research technique.

3. The nudging technique and its limitations.

As mentioned above the nudging technique constitutes a simple way to assimilate data into an atmospheric model. This technique was used intensively in the POTENTIALS project. The final report of that project (Kaas et al., 2000) includes a number of applications and a theoretical discussion on the usage of the technique. He we briefly describe the algorithm and then, in a number of model experiments carried out during the ACE Scientific Support Study demonstrate the strength and weakness of the methodology.

The nudging algorithm

In the present applications we always use the nudging technique to re-assimilate an existing set of gridded data, e.g. the re-analyses ERA15 from the European Centre for Medium Range Weather Forecasts (ECMWF) (Gibson et al. (1997) covering the period 1979-1993, into an atmospheric model, which may be the ECMWF model itself (the IFS model), the atmospheric component of a climate model, or any other atmospheric model.

In practise a first step always needed is to perform a space interpolation of all the model prognostic variables from the existing analyses to the spatial representation of the actual model. For a given prognostic variable ψ it is furthermore needed to temporally interpolate between the typically 6 hourly existing analyses to the individual time steps of the actual model. Obviously this pre-processing constitutes a huge amount of data storage and handling. It is important that the spatial interpolation is done in a way as consistent as possible to avoid a severe destruction of the approximate balances between wind and mass fields. Also for the surface parameters like soil wetness and temperature it is important to use a proper correction for any difference in surface topography between the model producing the existing analyses and the actual model.

Using t and Δt to denote the time and the size of the time stepping, there is an explicit

$$\psi(t+\Delta t) = \tilde{\psi}(t+\Delta t) + \Delta t \frac{\psi^{ERA}(t+\Delta t) - \tilde{\psi}(t+\Delta t)}{\tau} \qquad (2)$$

and an implicit

$$\psi(t+\Delta t) = \tilde{\psi}(t+\Delta t) + \Delta t \frac{\psi^{ERA}(t+\Delta t) - \psi(t+\Delta t)}{\tau}$$
$$\Leftrightarrow \psi(t+\Delta t) = \frac{\tau\tilde{\psi}(t+\Delta t) + \Delta t \psi^{ERA}(t+\Delta t)}{\tau + \Delta t} \qquad (3)$$

way to formulate the nudging assimilation of the prognostic variable ψ. In (2) and (3) $\tilde{\psi}(t+\Delta t)$ is a provisional one step forecast valid at time $t+\Delta t$ carried out with our re-assimilating model. The term $\psi^{ERA}(t+\Delta t)$ denotes the properly interpolated data set to be assimilated into our model. The relaxation time τ defines the strength of the nudging. A small value of τ keeps the re-assimilating model variable very close to the data is assimilates. Although the notation in (2) and (3) indicates that ERA re-analysis data are assimilated we could in principle assimilate any data set available in the same spatial and temporal representation as our model.

The equations (2) and (3) are formulated for a model employing a two time level scheme, i.e., typically a modern semi-Lagrangian, semi-implicit dynamical core as in the ECMWF model. For three time level models as most climate models including our applications Δt in front of the relaxation term has to be replaced by $2\Delta t$.

We have performed a number of sensitivity tests comparing the explicit (2) and the implicit nudging (3) but have found no noticeable differences. Therefore, because the explicit formulation is simpler we have used this formulation. Furthermore, it allows for a so called “hard nudging”, i.e., a full insertion of the interpolated analyses, by simply setting $\tau=\Delta t$ (or $=2\Delta t$ in case of a three time level time stepping scheme) which is needed in some applications (e.g. Kaas et al., 1999).

The basic philosophy behind our usage of the nudging to estimate the forcing residual R in (1) is the following: when the relaxation time τ is short, the model is kept very close to the data it assimilates throughout the period of assimilation. Thus the last term in (2) is the term additional to the provisional forecast, which is needed to keep the model close to the assimilated data. Thereby

$$R(t+\Delta t) \approx \frac{\psi^{ERA}(t+\Delta t) - \tilde{\psi}(t+\Delta t)}{\tau} \qquad (4)$$

when observed data (e.g., re-analyses) are assimilated.

Simple theoretical considerations (see Kaas et al., 2000) show as expected that this approximation is reasonable provided the time constant used in the relaxation is chosen sufficiently short compared to the typical atmospheric error growth. For most practical applications (see the following subsection) it is, however, questionable to choose such a short τ value. Therefore as discussed in Kaas et al. (2000) one has to consider temporal averages over a period from a week to a month of the relaxation term. Such temporal averages do, at least in the simple framework of a single prognostic equation, approximate the tendency residual R provided the τ value is not chosen too long, e.g., comparable to the radiative adjustment e-folding times (days).

The choice of τ

In practice, i.e., in a model with several coupled prognostic equations and using “real” data, the estimation of tendency errors by assimilation is a much more difficult task than e.g. in simple energy balance models with only one equation. This is particularly evident in unbalanced models based on e.g. the primitive equations as essentially all atmospheric components of climate models. In such models geostrophic adjustment processes (accomplished via the gravity waves) play a very important role in establishing the delicate balances between the wind and mass (pressure and temperature) fields of the atmosphere. Such imbalances are constantly created in the (model) atmosphere via the dynamical (e.g. potential to kinetic energy conversion) and physical (e.g. release of latent heat) processes. In the atmosphere as well as in models of it, this means that e.g. a cooling at a given location in the atmosphere is accompanied by adiabatic heating via the setup of indirect circulations. Therefore, when estimating tendency errors, it is a risk, at least partly, to misinterpret a heating error, i.e., a tendency residual in the first equation of thermodynamics, as an error in the wind forcing (e.g. a gravity wave drag problem or a subgrid scale vertical transport of momentum in Cumulus towers), and vice versa. Also the problem of moisture spin up constitutes a major problem and a source of error when estimating tendency errors.

Due to the problems and potential sources of error we have to perform a number of idealised experiments and sensitivity tests to define useful values of τ. We have performed three types of idealised experiments, including:

- Twin experiments, i.e., re-assimilation of synthetic data produced by the model under investigation. In this case the residual forcing obtained obviously should be as close as possible to 0.
- Identification via re-assimilation of specified anomalous forcing anomalies in the model under investigation.
- Long simulations using the identified forcing as an anomalous constant "flux correction"

In these experiments we have used the ARPEGE/IFS spectral climate model (Déqué et al., 1994) in a version with 31 hybrid pressure-sigma vertical levels and with a coarse horizontal resolution of T21. The dynamic prognostic variables are vorticity (V), divergence (D), temperature (T) and log surface pressure ($\ln Ps$). In our nudging procedure it is the spectral expansion coefficients of these variables, which are assimilated, and the experiments we perform aim at identifying a useful set of τ values. To avoid potential moisture spin-up problems we do not assimilate humidity as suggested by Jeuken et al. (1996). In this way the model physics are quite free, but they are forced to operate in the dynamical environment of the data, which are assimilated. To avoid a drift of the model land surface, we perform a weak assimilation of the upper most soil layer temperature and soil wetness. Obviously, since the soil model operates in grid points only, this part of the assimilation is performed in grid point space.

We have limited our tests to situations where the same τ value is used for all wave numbers for a given prognostic variable. All assimilations have been performed with synthetic data produced in one month long simulations with the same model (ARPEGE/IFS, T21/L31). In these simulations all data were stored each 6 hours to mimic observed (re-analysed) data.

In the following we describe the experiments and the results obtained.

<u>Twin experiments</u>

The purpose of these simulations is to reveal fundamental problems with the nudging technique and the coding of the assimilation. In particular, we are interested in artificial forcing residuals and how these depend on the relaxation time.

In all twin experiments we assimilate the model data from a "control" simulation. This simulation is performed with the same model we want to use for assimilation. Thus we do not need to perform a spatial interpolation before we assimilate. Since the data from

Relaxation number	τ_T	τ_V	τ_D	τ_{lnPs}	τ_{ST}	τ_{SM}	Correction of diurnal cycle
A1	24 h	6 h	48 h	24 h	48 h	48 h	Yes
A2	24 h	6 h	48 h	24 h	48 h	48 h	No
A5	6 h	12 h	12 h	∞	48 h	48 h	Yes
A6	6 h	12 h	12 h	∞	48 h	48 h	No
A8	3 h	3 h	12 h	3 h	48 h	48 h	Yes
A11	12 h	12 h	12 h	∞	48 h	48 h	Yes

Table 1. Column 1 lists the numbering of the relaxations, column 2-7 the relaxation times for atmospheric temperature τ_T, atmospheric vorticity τ_V, atmospheric divergence τ_D, the log of atmospheric surface pressure τ_{lnPs}, the temperature in the upper most soil layer τ_{ST}, and wetness of the upper most soil layer τ_{SM}. The last column indicates whether or not we have removed the average diurnal cycle (see text) from the assimilation.

the control simulation are stored only each 6 hours we do, however, need to perform the same temporal interpolation as if we assimilated observed data. In practise this is achieved using a cubic spline in time, i.e., involving 4 neighbouring instants, covering a 24 hour period.

We have tested 12 different combinations of τ values. Here we mainly report the results from four of these tests demonstrating the basic findings. These choices of relaxations are numbered A2, A5, A8 and A11. The τ values for each combination and each prognostic variable are listed in Table 1. Table 1 also include 2 other relaxation choices, A1 and A6, illustrating other issues to be discussed in the following. Relaxations A2 should be considered a kind of baseline, as it has already been used in a full assimilation of the ERA15 (Gibson et al., 1997) data into ARPEGE in the winter months (Kaas et al., 2000).

The twin test of the A2 experiment revealed a problem related to the diurnal cycle. The problem is seen most clearly when using a somewhat stronger relaxation than in A2. Therefore we only present the result for the assimilation experiment A6: Fig. 1a is a Hovmoeller diagram of the difference between the assimilated surface pressure in the tropics and the surface pressure in the control experiment. For the purpose of plotting the Hovmoeller diagram both control and assimilated data were stored each time step (i.e. 30 minutes). It is easily seen that this difference is dominated by a 12 hourly wave travelling around the globe. Another way to illustrate the problem is to calculate the spatial distribution of the long-term difference and of the root mean square difference between the 6 hourly (00, 06, 12, 18UTC) assimilated and control data. This is shown in figure 2b upper and lower panel respectively. A number of considerations have led to the conclusion that the problem arises because an internal atmospheric 12 hourly mode is triggered during the assimilation due to problems in resolving the diurnal cycle in the control data when only 6 hourly data are used. Since we have to use such data in practical applications we needed to invent a cure. There are at least two possible cures. One (not reported here) is to use time-dependent relaxation values (so-

a **b**

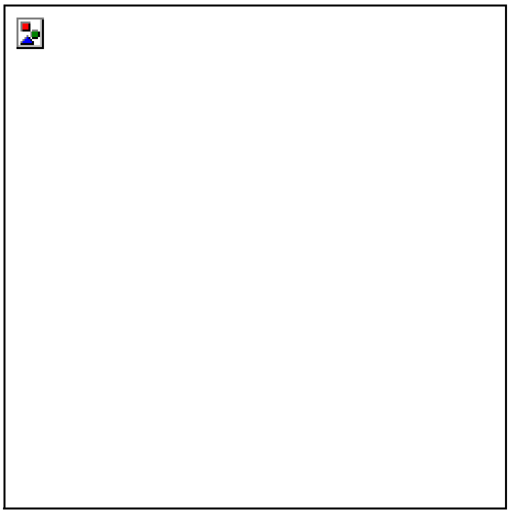


Fig. 1. Hovmoeller diagrams of differences in zonal mean surface pressure in the latitude band ±30°. The differences are between the result of assimilations and the original data ("the truth"). The left panel is for experiment A6 while the right is for A5. Horizontal axis: longitude in degrees. Vertical axis: time step number (30 minutes each). The units are hPa.

called nudging with window). The other is to assimilate only anomalies from the long-term average diurnal cycle. This can be achieved by re-formulating equation (2):

$$\begin{aligned}\psi &= \tilde{\psi} \\ &+ 2\Delta t \frac{\psi^{ERA} - \tilde{\psi}}{\tau} \\ &- 2\Delta t \frac{\psi^{DIERA} - \psi^{DCTR}}{\tau}\end{aligned} \qquad (5)$$

where we for simplicity have omitted the time referencing. In (5) superscript DIERA denotes the climatology of the diurnal cycle of the temporally interpolated data being assimilated, and DCTR denotes the raw (non-interpolated) climatology of the diurnal cycle of the interpolating model. Both DIERA and DCTR depend on the annual cycle and have to calculated before the assimilation. DCTR has to be taken from a long control simulation. Note that we have formulated (5) for a three time level scheme since we use an Eulerian version of ARPEGE/IFS employing the standard "leap frog" time stepping scheme. Figs. 1b and 2a show that our cure does a nice job. However, we have no guarantee that this is also the case when assimilating real data. However, it should be noted that this not a very severe problem as compared to other problems described below.

Fig. 3 shows the long-term (i.e., 30 day) averages of the zonal mean temperature tendency residuals for the twin experiments corresponding to relaxations A1, A2, A5, A8 and A11 (see table 1). These tendency should ideally be zero. If is first noted that there are some differences between A1 and A2 illustrating that our removal of the diurnal cycle from the assimilation produces some difference in the zonal mean tendency residual. But this effect is relatively small. The relaxations A5 and in particular A8 produces very large artificial temperature residuals, indicating that strong relaxations should be used with care because they trigger adjustments between the wind and mass fields. Although we only assimilate anomalies from the average diurnal cycle in the A5 and A8 relaxations we consider it most likely, that the artificial tendencies are due to the temporal interpolations between each “observational” hour. This is because the tendencies we obtain (not shown) when assimilating the raw data (i.e., data from the control run stored each time step) are small, also for A5 and A8. The problems are probably not related to a moisture spin up, as the precipitation (not shown) is very similar for all experiments. The adjustments and the implications in terms of artificial tendencies can also be seen in Figs. 4 and 5 for the zonal mean of the zonal and meriodional wind. Generally the artificial tendencies occur in the same regions for the 3 variables, and simple considerations regarding adiabatic heating rates related to the zonal and meridional flows support the point of view that twin experiments with strong nudging induces secondary circulations which needs to be counteracted by artificial heating rates.

It is seen from table 1 that several of the experiments have been performed with no assimilation of the surface pressure. This is because one may argue that surface pressure is adjusted to the vertically integrated divergence, which is as already being assimilated, and that we would over specify the data, risking introduction of artificial noise. We have performed a number of experiments (not shown) to investigate the impact of assimilating the surface pressure or not, but none of these have shown any noticeable impact. Fig. 2a demonstrate that assimilation of surface pressure is unnecessary since the root mean square deviations between surface pressure in the original data and in the assimilated data are very small, even when we don’t assimilate this variable as in A5.

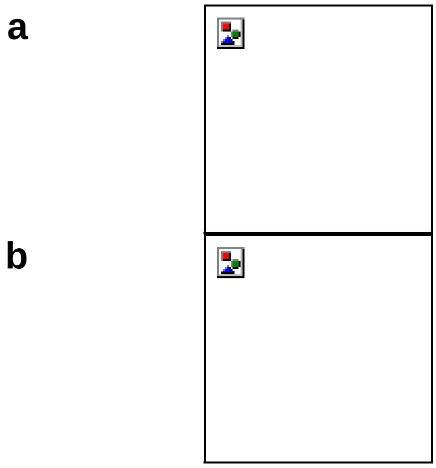


Fig. 2. The two upper plots in panels “a” and “b” show the average of the long-term (30 days) difference between the air surface pressure in the control simulation ("the truth") and in the result of assimilations. The two lower plots shows the corresponding root mean square difference. Panel “a” corresponds to experiment A5 and panel “b” to A6. The plots are based on 4 times daily data at 00, 06, 12 and 18 UTC.

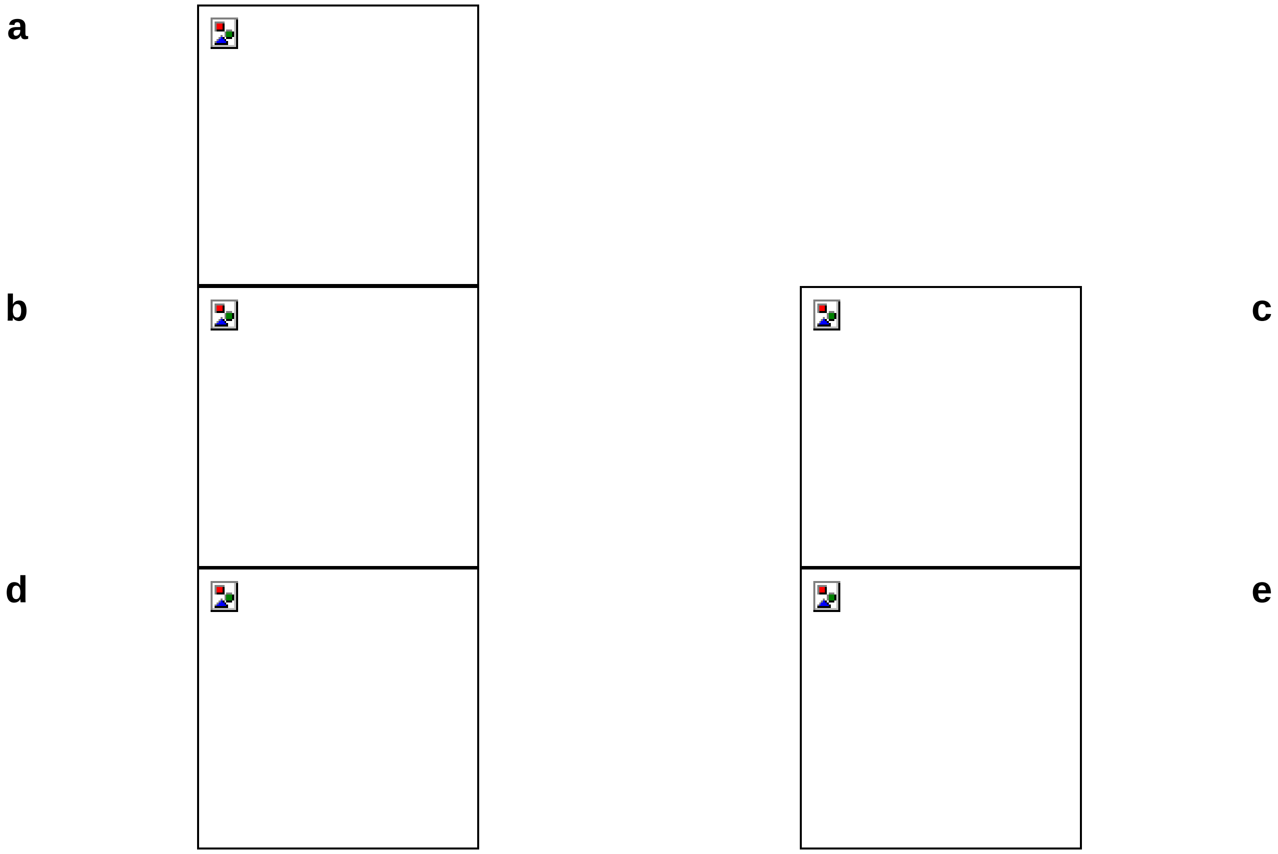


Fig. 3. Monthly average of the zonal mean tendency residual for temperature in the twin experiments, corresponding to the relaxation coefficients A1 (panel a), A2 (panel b), A5 (panel c), A8 (panel d) and A11 (panel e). The vertical axis shows the model levels with level 1 the top most level and level 31 the lower most level. Note the different contour levels in each panel. Units: K/day.

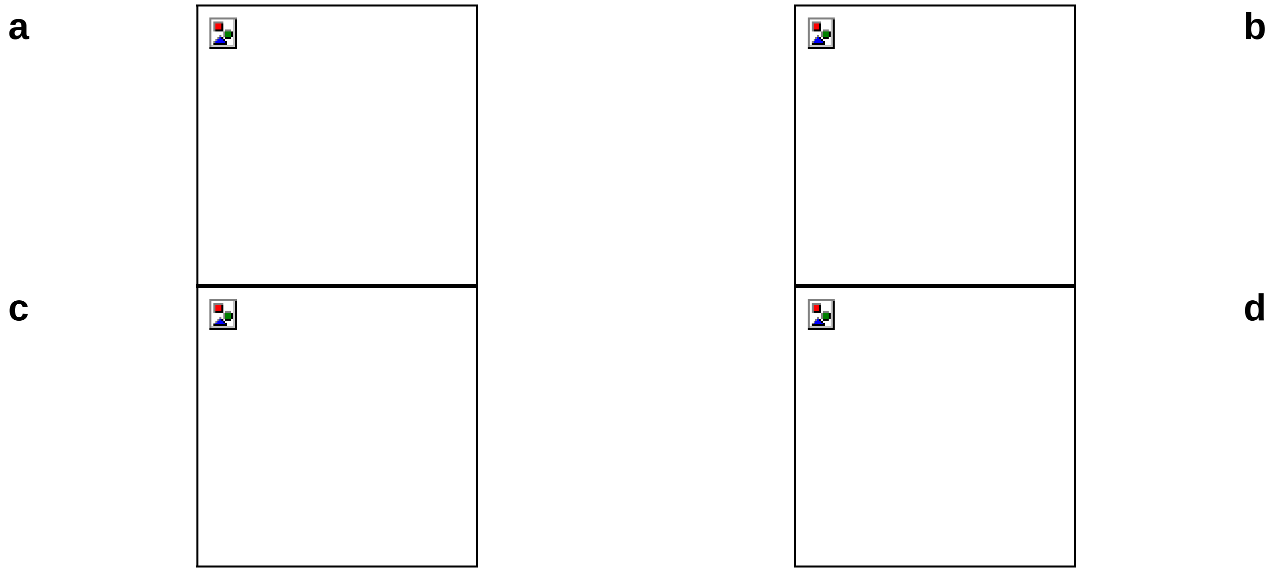


Fig. 4. As figure 3, but for the zonal mean zonal wind. Note the different contour levels in each panel. Units: m/s/day.

The choices of relaxation parameters in A2 (/A1) and in A11 seem to be preferable as they lead to relatively small artificial residual tendency estimates.

The overall conclusion from the twin experiments is that small relaxations result in small artificial tendencies.

a b

c d

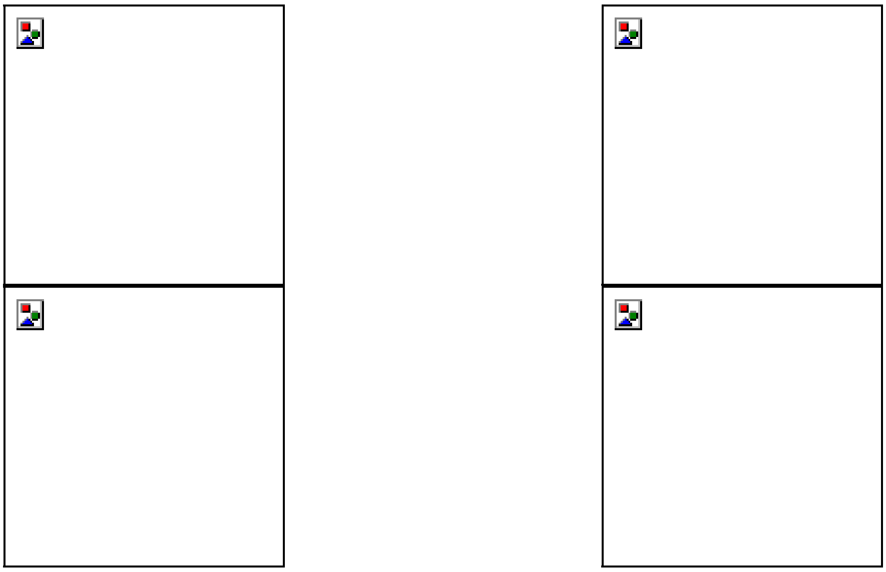


Fig. 5. As figure 3, but for the zonal mean of the meridional wind. Note the different contour levels in each panel. Units: m/s/day

Identification of specified anomalous forcing

The next test is to identify a known pre-specified forcing anomaly via the assimilation, i.e., what we ideally want to do using real data. The first step is to perform an integration with our model after the introduction of an artificial heating in the first equation of thermodynamics or a drag in the momentum model equations. Here, since it is most relevant for the ACE mission, we only discuss the heating case, although a number of corresponding experiments have been performed with the model gravity wave drag scheme turned off. Fig. 6 shows the constant heating we have introduced in a 30 day simulation with data stored each 6 hours as in the case of twin experiments. The heating we have used is an estimate of the volcanic forcing due the Pinatubo volcanic eruption (see e.g. Kirchner ….). In the second step we assimilate the resulting temporally interpolated 6 hourly data (mimicking observed data) using the nudging technique and the relaxation parameters A2, A5, A8 and A11.

Fig. 7 shows the long term (i.e., 30 day) average temperature tendency residual. We can see that the heating is identified in all experiments, but also, that there are large differences between the assimilations. The assimilation employing a strong relaxation (A8) identifies – as expected – the forcing with highest accuracy. However, this is at the expense of an unacceptably large artificial forcing elsewhere, which by comparison with Fig. 3 can be seen to be almost identical to the artificial twin forcing. Therefore, in a practical application assimilating real data, we have to use a fairly weak nudging because we cannot estimate a quantity corresponding to the twin forcing to be subtracted.

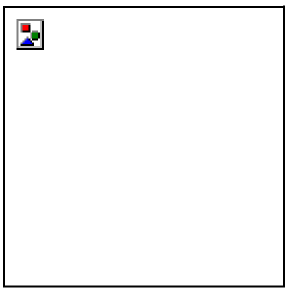


Fig. 6. Zonal average of an artifical heating introduced in the first equation of thermodynamics. A 30 day simulation is performed with this heating on. Units: K/day.

The tendency residual for the zonal mean of the zonal and meridional wind is shown in Figs. 8 and 9. In this case the relaxation coefficients A8 lead to the smallest artificial momentum forcings directly associated with the heating anomaly we want to identify. But again for this set of relaxation parameters the strong artificial forcing elsewhere is unacceptable. For the weaker nudging in A11 and in particular in A2 we see that the lack of full assimilation leads to rather large artificial momentum residuals. Between A11 and A2 there is little doubt that A11 is preferable since it captures a larger fraction (about two thirds versus only the half)of the forcing without introducing too much noise.

We are left with a situation where the short relaxation times are the only useful possibilities. This may lead to the question if nudging can be used at all to estimate heating residuals. It is, however, important to emphasize, that although temperature tendency residuals obtained with a weak nudging are too small, they provide a very good hint as to where and when a tendency residual occurs and with what sign. Heating is the important driving mechanism of the atmosphere and in models of it, and therefore information as that in Fig. 7 is highly valuable for improvement of climate models. In practice, one will need to progress in an iterative way by estimating the residuals for a basic model version. Then one uses the tendency residuals as a guideline suggesting which physical processes needs improvement in the model formulation. The assimilation with the improved model can be used to suggest if the improvement in the physical parameterisation actually cured the problem. As mentioned above in section 2 this way of a b c d

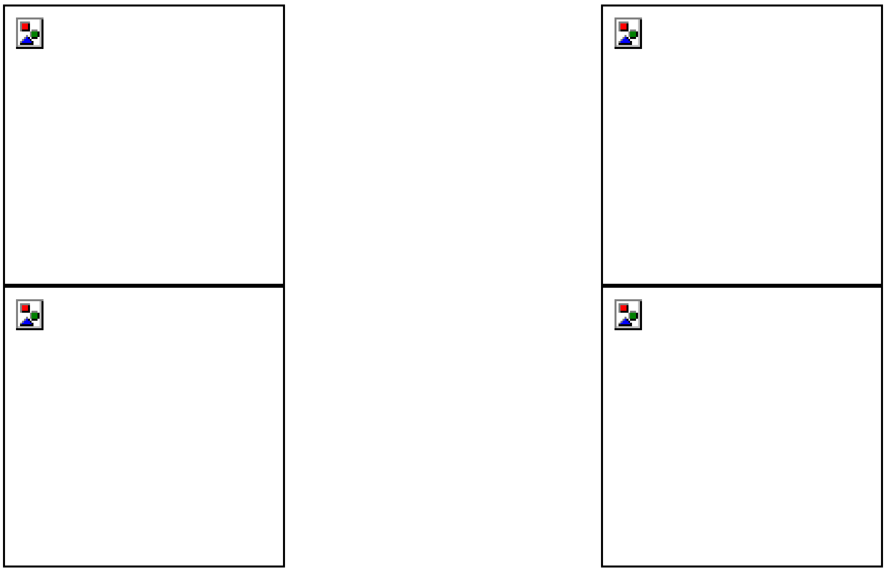


Fig. 7. Monthly average of the zonal mean temperature tendency residual in idealised forcing experiments, corresponding to the relaxation coefficients A1 (panel a), A2 (panel b), A5 (panel c), A8 (panel d) and A11 (panel e). The vertical axis shows the model levels with level 1 the top most level and level 31 the lower most level. Units: K/day.

a b c d

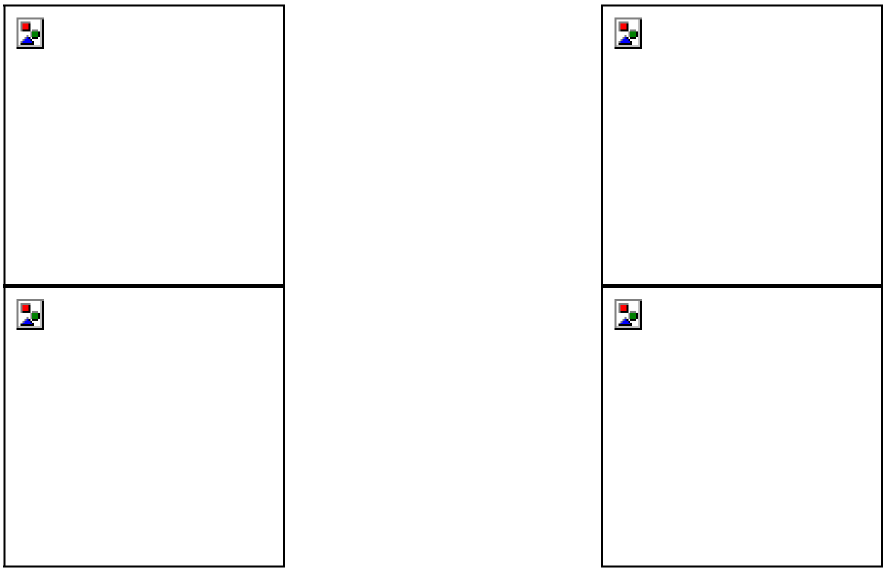


Fig. 8. As figure 7, but for the zonal mean zonal wind. Units: m/s/day.

improving/tuning models is highly preferable from the simple comparison of a long term simulation with a corresponding observed data set because is reduces the risk of introduction of compensating errors.

Fig. 10 shows an example from the POTENTIALS project where we have assimilated the ERA15 re-analyses into the same version of ARPEGE as used above, although at T42 horizontal resolution, and with the A2 set of relaxation parameters. The plot shows the estimated zonal mean tendency residual averaged over all Januaries in the period 1979-1993. From these residuals we obtain a strong suggestion that the convection and possibly its interaction with radiation need adjustments. Actually, a later and improved version of the model (see Kaas et al., 2000) showed a much reduced tendency error in the same regions. Long simulations with the improved version produced a climate much closer to the observed climatology.

Long "flux corrected" simulations.

Another way to verify that the residuals we have estimated via the nudging technique are reasonable is to re-inject them as an empirical forcing in the model prognostic equations. This requires a small modification of the model where we at each time step add a small tendency which is constant in time (although in practice varying with the annual cycle) but varying in space. If the identified forcing is reasonable the climatology of a long run with this "flux corrected" model should be close to the climatology of the observed data that were used to estimate the tendency residual. Fig. 11 shows an example of such empirical reduction of systematic model errors from the POTENTIALS project. We

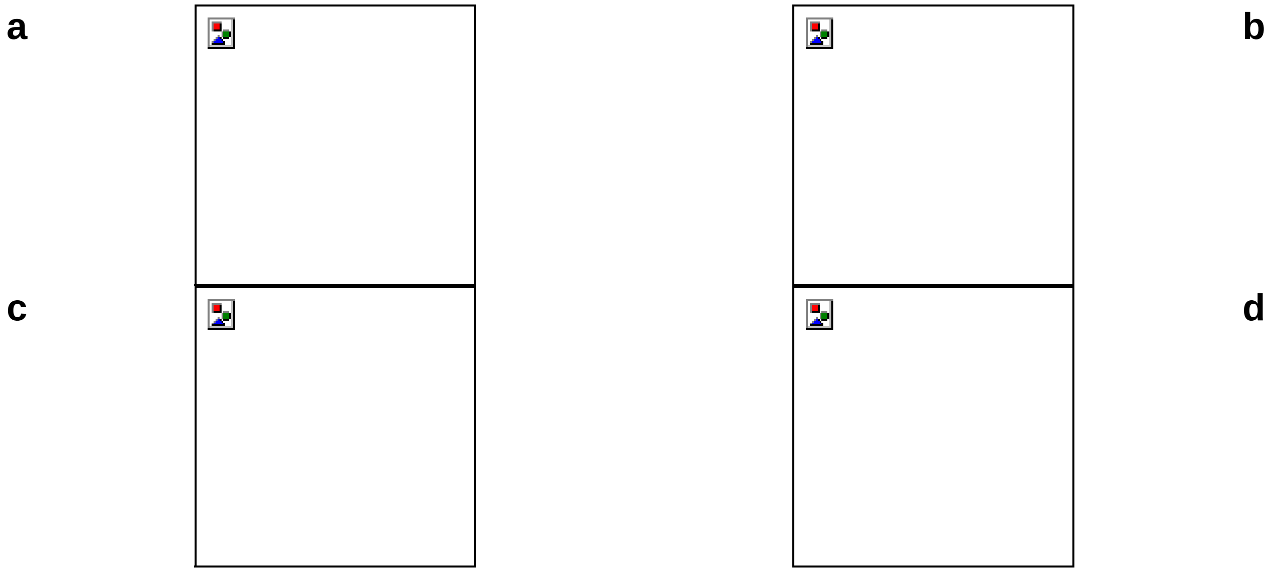


Fig. 9. As figure 7, but for the zonal mean of the meridional wind. Units: m/s/day.

performed 9 ensemble simulations for each of the winters 1979/80-1992/93 using the ARPEGE/IFS model in flux corrected mode. This means that e.g., the average empirical heating in Jan. was equal to that shown in Fig. 10, although an empirical forcing was introduced in all the prognostic equations being relaxed during the assimilation. Fig. 11 shows the systematic error in the zonal mean temperature in Jan., i.e., the difference between the climatology of ERA15 and the ensemble average of all 14 simulated winters. The left panel shows the error in the non-flux corrected simulation (i.e. the basic version of ARPEGE/IFS), while the right panel shows the systematic error in the empirically corrected version. It can be seen that the empirical forcing strongly reduces the model long term error. We thus conclude that the overall forcing residual obtained via the nudging technique provides a realistic estimate of the model error. This conclusion is supported by investigation of other variables. A striking example of strong improvement can be seen in Fig. 12 illustrating the reduction in systematic error of the 500 hPa height.

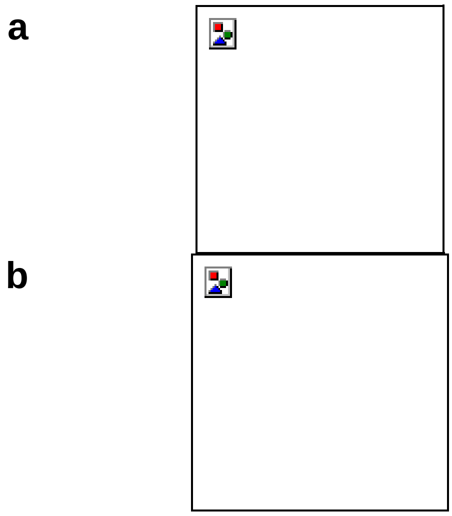


Fig. 10. Average residual tendency for temperature in January as obtained by assimilating the ERA15 data into ARPEGE/IFS (T42 Eulerian version). The set of relaxation coefficients used was A2. Panel a shows the zonal average while panel b shows the average distribution in model level 12, i.e., ~ 250 hPa. Units: K/day.

4. Estimation of varying external forcing of climate.

In addition to climate model improvement, forcing - or rather heating - residuals may also be used to estimate temporally varying external forcing of climate. This is because one can use a temporal anomaly of the heating residual (i.e., the model error in the first equation of thermodynamics) to estimate the magnitude of the anomalous heating by comparing with the residual in other periods. This approach is not to be confused with classical climate fingerprinting, the main goal of which is to detect and attribute changes in anticipated climate *response* to various external forcings. Instead of detecting changes in the basic meteorological quantities like temperature the method detects the more fundamental anomalous forcing of climate which can be due to e.g. varying solar activity, changed greenhouse effect, stratospheric ozone depletion or volcanic activity. The method may in this way be considered a filter which excludes the impact of the non-externally forced climate variations.

5. Discussion

A problem needing consideration is identification of the most appropriate type of data for detection of forcing residuals. For assimilation into climate models, we suggest that one should assimilate processed data such as the re-analysed data from the European Centre for Medium Range Weather Forecasts (ECMWF). Future re-analysed data sets including RO data should be sufficiently accurate and homogenous for estimation of climate model forcing residuals even in the a

b

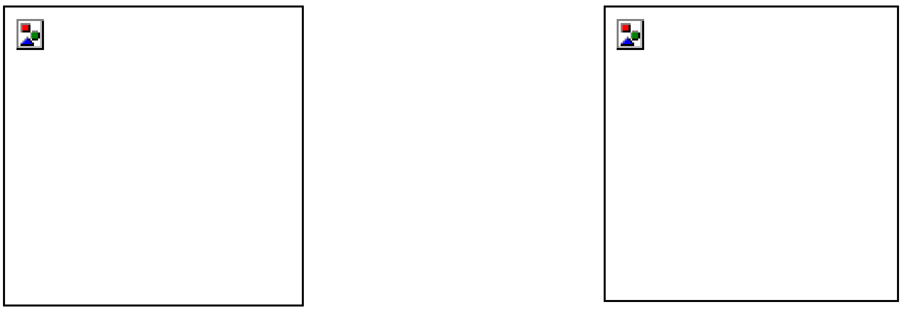
Fig. 11. Long term zonal mean systematic errors of temperature in control simulations (left) and in empirically forced simulations (right) as compared to ERA data. The figure is based on data from Dec, Jan and Feb in winters 1979/80, …, 1992/93. Contour interval: 1°C with negative contours dashed.

stratosphere, which is not possible with e.g. the existing ERA15 data due to lack of vertical resolution. It is anticipated too costly to assimilate e.g. RO data directly into climate models because to make any sense this would require simultaneous assimilation of all other types of observations, at least all types of wind data. The problems encountered so far in the ongoing ERA40 project demonstrates that building a homogenous and accurately assimilated data set is far from simple. It would be waste of time to re-invent the wheel by mimicking what has been done in the ERA project.

For estimation of model heating residuals, it is extremely important that the underlying observed data are homogenous and with very small bias. These are some of the key advantages of RO data. It is therefore concluded in KGa that RO will provide a positive contribution to climate modelling research, if they are assimilated in future re-assimilation projects.

For simple detection of observed climate trends based on "raw" (i.e. non assimilated) observations,

a



b

Fig. 12. Long term systematic errors of 500 hPa height in control simulations (left) and in empirically forced simulations (right) as compared to ERA data. The figure is based on data from Dec, Jan and Feb in winters 1979/80, …, 1992/93. Contour interval: 20 m with negative contours dashed.

directly inverted RO data will also be highly useful because of their small bias. This is a particular advantage in the future long term climate monitoring because bias correction and inter-mission calibration is a smaller problem, relatively, than for most other data types.

REFERENCES

D'Andrea, F. and R. Vautard, 2000a: Reducing systematic errors by empirically correcting model errors. *Tellus*, **52a**, 21-41.

Christy, J. R., R.W. Spencer and W. D. Braswell, 2000: MSU tropospheric temperatures: dataset construction and radiosonde comparisons. *J. Atmos. Ocean. Tech*., **17**, 1153-1170.

Déqué, M., C. Dreveton, A. Braun and D. Cariolle, 1994: The ARPEGE-IFS atmosphere model: a contribution to the French community climate modelling. *Clim. Dyn*., **10**, 249-266.

Gibson, J. K., P. Kållberg, S. Uppala, A. Hernandez, A. Nomura and E. Serrano, 1997: ECMWF Re-Analysis Project Report, 1. ERA description. European Centre for Medium-Range Weather Forecasts, Reading, UK, 66pp. (available on http://www.ecmwf.int/research/era/)

Hurrell, J. W. and K. E. Trenberth, 1998: Difficulties in obtaining reliable temperature trends: Reconciling the surface and satellite microwave sounding unit records. *J. Climate*, **11**, 945-967.

IPCC, 2001: Climate Change 2001: The Scientific Basis. Contribution of Working Group I to "The Third Assessment Report" of the Intergovernmental Panel on Climate Change (IPCC). Cambridge University Press, 2001.

Jeuken, A.B.M., P.C. Siegmund, L.C. Heijboer, J. Feichter, and L. Bengtsson, 1996. On the potential of assimilating meteorological analyses in a global climate model for the purpose of model validation. *J. Geophys. Res.*, **101**, 16939-16950.

Kaas, E., A. Guldberg, W. May and M. Déqué, 1999: Using tendency errors to tune the parameterisation of unresolved dynamical scale interactions in atmospheric general circulation models, *Tellus*, **51A**, 612-629.

Kaas, E., A. Guldberg, M. Déqué, A. Braun, J.-P. Piedelievre, J. F. Guérémy, B. Machenhauer, I. Kirchner, F. D’Andrea, R. Vautard, S. Corti and F. Molteni, 2000:. Final report of the POTENTIALS project (Project On Tendency Evaluations using New Techniques to Improve Atmospheric Long-term Simulations). Available from the EU-Commision (DGXII). 93 pp.

Karl, T.. R., R. G. Quayle, and P. Y. Groisman, 1993: Detecting climate variations and change: New challenges for observing and data management systems. *J. Climate*, **6**, 1481-1494.

Klinker, E., and P. D. Sardeshmukh, 1992: The diagnosis of mechanical dissipation in the atmosphere from large-scale balance requirements. *J. Atmos. Sci.*, **49**, 608-627.

The Danish Climate Centre

The Danish Climate Centre was established at the Danish Meteorological Institute in 1998. The main objective is to project climate into the 21 st century for studies of impacts of climate change on various sectors and ecosystems in Denmark, Greenland and the Faroes.

The Climate Centre activities include development of new and improved methods for satellite based climate monitoring, studies of climate processes (including sun-climate relations, greenhouse effect, the role of ozone, and air/sea/sea-ice interactions), development of global and regional climate models, seasonal prediction, and preparation of global and regional climate scenarios for impact studies.

The Danish Climate Centre is organised with a secretariat in the Research and Development Department, and it is co-ordinated by the Director of the Department. It has activities also in the Weather Service Department and the Observation Department, and it is supported by the Data Processing Department.

The Danish Climate Centre has established the Danish Climate Forum for researchers in climate and climate related issues and for others having an interest in the Danish Climate Centre activities. The Centre issues a bi-annual newsletter "KlimaNyt" (in Danish).

DMI has been doing climate monitoring and research since its foundation in 1872, and establishment of the Danish Climate Centre has strengthened both the climate research at DMI and the national and international research collaboration.

Previous reports from the Danish Climate Centre:

- Dansk Klimaforum 29.-30. april 1998. (Opening of Danish Climate Centre and abstracts and reports from Danish Climate Forum workshop). *Climate Centre Report 98-1* (in Danish).
- Danish Climate Day 1999. *Climate Centre Report 99-1*.
- Dansk Klimaforum 12. april 1999. Workshop: Klimatisk variabilitet i Nordatlanten på tids-skalaer fra årtier til århundreder. *Climate Centre Report 99-2* (in Danish).
- Luftfart og den globale atmosfære, Danmarks Meteorologiske Instituts oversættelse af IPPC's særrapport "Aviation and the Global Atmosphere, Summary for Policymakers". *Climate Centre Report 99-3* (in Danish).
- Forskning og Samarbejde 1998-1999. *Climate Centre Report 00-1* (in Danish).
- Drivhuseffekten og regionale klimaændringer. *Climate Centre Report 00-2* (in Danish).
- Emissionsscenarier, Danmarks Meteorologiske Instituts oversættelse af IPPC's særrapport "Emissions Scenarios, Summary for Policymakers". *Climate Centre Report 00-3* (in Danish).
- Metoder mødes: Geofysik og emner af samfundsmæssig interesse. Dansk Klimaforums Workshop 15.-16. maj 2000. *Climate Centre Report 00-4* (in Danish).

- A time-slice experiment with the ECHAM4 A-GCM at high resolution: The simulation of tropical storms for the present-day and of their change for the future climate. *Climate Centre Report 00-5*.
- The climate of the 21 st century: Transient simulations with a coupled atmosphere-ocean general circulation model. *Climate Centre Report 00-6*.
- Changes in the storm climate in the North Atlantic / European region as simulated by GCM time-slice experiments at high resolution. *Climate Centre Report 01-1*.