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Detection of the Pinatubo volcanic heating signal in the lower stratosphere based on nudging assimilation and analysis increments.

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Detection of the Pinatubo volcanic heating signal in the lower stratosphere based on nudging assimilation and analysis increments[1]

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ABSTRACT

We investigate the possibility of using meteorological observations to estimate the anomalous heating in the lower stratosphere resulting from volcanic eruptions. Two methods are used: a simple estimation based on analysis increments and an estimation via assimilation of ERA data into a climate model using the nudging assimilation technique. It is indicated that the most accurate estimations are obtained from the analysis increment method. However, more experiments need to be carried out with the nudging technique before firm conclusions can be drawn. A discussion of anomalous atmospheric forcing based on tendency evaluation concludes that observations with low bias - or at least free of temporal inhomogeneities - are required to avoid misinterpretation of the results.

1. Introduction

Data assimilation of observed atmospheric dynamical variables, e.g., pressure, temperature and wind fields, into general circulation models provides information enabling estimation of temporal variations in external forcing of the atmosphere. By external forcing we here think of anomalous radiative heating or cooling of the atmosphere due to processes not occurring in the undisturbed climate. Conceptually we can think of data assimilation as a way to ensure that the assimilating atmospheric model is kept very close to observations over a long period of time. Ideally, at a given time within this period we should then have

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

where ψ is a prognostic variable (temperature, wind, pressure, humidity) at a given location in the atmosphere and subscript O denotes the instantaneous observed temporal tendency for a given observed 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. Using "perfect" (error free) observations and assimilation techniques the term R represents errors in the formulation of the assimilating atmospheric model which are due to problems in the parameterisation of unresolved physical (diabatic) processes and the use of inadequate numerical techniques to solve the basic adiabatic equations. Because R reflects instantaneous forcing errors in the model relative to the observations there is – in principle – no time for compensating response errors to develop during the assimilation period.

We can – assuming "perfect" observations and assimilation – divide R into three terms:

$$R = \overline{R} + R'_i + R'_e \qquad (2)$$

where $\overline{R}$ is the systematic (i.e., long term average) initial tendency errors of the assimilating model, R'_i is

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the temporal anomaly of initial tendency errors due to inability of the model to simulate processes *internal* to the atmosphere, i.e., processes which supposedly are build into the assimilating model, and R'_e is the temporal variations in the *external* forcing of the atmosphere, not build into the assimilating model. If, e.g., the atmosphere has been influenced by a volcanic external forcing over a year or two, and if this forcing is not explicitly coded into the assimilating model, the volcanic forcing anomaly will contribute to R'_e for temperature, i.e., in the first equation of thermodynamics.

It is important to note that the systematic initial tendency error ($\overline{R}$) is quite different from the systematic errors of the assimilating model. The former holds the true estimate of the model error, while the latter is the model error evolving over time in response to the tendency error. In this way we should consider $\overline{R}$ as the mean generic error. Due to energy dispersion, adjustments and feedbacks in the climate system the systematic errors of the assimilating model may be located in geographical regions quite remote from those where the systematic initial tendency errors occur. Therefore $\overline{R}$ constitutes a much better guidance than the systematic errors for improving atmospheric models. Examples and discussion of the usage of $\overline{R}$ for model improvement can be found in Klinker and Sardeshmukh (1992), D'Andrea and Vautard (2000), Kaas and Guldberg (2001) (GB) and Kaas et al. (2000).

Here we are mainly interested in the temporal anomalies $R'_i + R'_e$ of R in the case of temperature, i.e., $R'_i + R'_e$ represents the anomalous heating of the atmosphere. In particular, we aim at estimating R'_e, because it may assist in a detection of unknown climate forcing and/or in improving our quantification of known variations in climate forcing. In practise, however, there are a number of problems, which need consideration:

1. Uncertainties in the basic observations.

The available observations of the atmosphere are far from complete. A main problem in connection with detection of climate change and change in climate forcing is observational inhomogeneity (see discussion in KG). This problem is particularly severe for traditional satellite based infrared soundings of atmospheric temperature: large inter-satellite calibrations are needed, and often the time overlap between missions is insufficient to ensure a proper calibration. Furthermore, there are several creping inhomogenities related to instrumental drifts or orbital decay of the satellites. Obviously, sudden as well as creping inhomogeneities will influence estimations of external forcing when we use the technique outlined above.

2. Large temporal and spatial variations in the observational coverage.

If the data coverage in a certain region is sparse, there will be no local data to tell the story of an external heating anomaly. Therefore the effect of the forcing will not be detected correctly. In particular, when the anomalous forcing is detected by remote observations, there has been time for adjustments between wind and mass fields of the atmosphere. This means that we risk a misinterpretation of a heating anomaly as a forcing (drag) anomaly in the wind and visa versa.

3: Changing data-assimilation system (optimum interpolation, three – or four dimensional variational assimilation etc.)

Modern variational data assimilation is being improved and developed intensively at the numerical weather centres, with the European Centre for Medium-Range Weather Forecasts (ECWMF) being a lead actor. It is important to note, however, that any changes in the data-assimilation procedure or in the data-handling will lead to artificial jumps in *R*. Therefore to be useful for our purpose we must consider data being assimilated with a stationary assimilation system. Intentionally, the so-called re-analyses constitute the outcome of such systems. It is therefore needed to use e.g. the re-analysis products (ERA) from ECMWF or, alternatively, to develop a new system for assimilation of atmospheric observations and to perform a multi-year assimilation with a frozen version of this system.

4. Separation of R'_e and R'_i.

For estimation and detection of external forcing of climate we are not interested in the R'_i component of *R*. To filter this component we need to average over a fairly long time period. However, if this period becomes too long, we are also out-filtering R'_e, which we want to distil. For the ERA15 (Gibson et al., 1997) we have found (not shown here) that an averaging over a month is sufficient to remove most of the internal “noise” in the troposphere, however, still with some dependence of R'_i on sea surface temperature anomalies left over. At higher altitudes, the ERA15 constitutes a special problem since the assimilating model has a coarse resolution in the stratosphere not permitting adequate simulation of phenomena as the Quasi-Biennial Oscillation (QBO) and sudden stratospheric polar warmings. Therefore these phenomena give large contributions to R'_i. In Andersen et al. (2000) a method is used to remove the dependence of R'_i on the QBO, and a similar method may be applied in the case of sudden warmings. In estimations of *R* based on the new ERA40 these problems are anticipated much less severe because the assimilating model has a much higher stratospheric resolution.

We note, that an estimate of *R* is different from the notation of radiative forcing of climate used by the IPCC (2001). The latter represents a radiative imbalance at the tropopause level (or alternative an imbalance in the total energy flux at the surface) and is measured in W/m^2. Due to the small heat capacity of the atmosphere, as compared to that of the oceans, it is not possible to

Data set	Source	Type	Period	levels
D1	ERA15	Monthly anomaly of 6 hour analysis increment of temperature	Jul. 1991 - Jun. 1993	30 hPa 50 hPa 70 hPa
D2	ARPEGE CLIMAT (forced around Eq.)	Monthly average of 24 hour analysis increments of temperature and wind	30 days (perp. Jan)	10-100 hPa
D3	ARPEGE CLIMAT (forced around 25N)	Monthly average of 24 hour analysis increments of temperature and wind	30 days (perp. Jan)	10-100 hPa
D4	ERA15	Monthly anomaly of 24 hour analysis increments of temperature	Jan. 1979 - Dec. 1993	30 hPa 100 hPa 850 hPa
D5	ERA15	Monthly anomaly of temperature	Jan. 1979 - Dec. 1993	30 hPa 100 hPa 850 hPa

Table 1. The five data sets (D1-D5) used for investigation of analysis increments. Note that D5 is the mean temperature and not the increments. The anomaly in D1 is with respect to the period Jan. 1985 - Dec. 1990.

estimate radiative forcing from pure meteorological data assimilation. This is because a given anomalous forcing of climate will be seen mainly as an energy input to the oceans, and since such energy input is not being monitored with sufficient accuracy (and assimilated) we cannot infer the forcing. We note, however, that the method proposed here may be applied to estimate external forcing of climate if it based on a data assimilation system for the *coupled* atmosphere ocean system. Because of the close coupling between ocean temperatures and the temperature in the troposphere, the methodology we suggest is of primary use in the stratosphere, which is less strongly related to the lower boundary condition of the atmosphere.

The present report deals with estimation of R'_e using two different techniques. In particular we are interested in identifying the lower stratospheric heating following the Pinatubo volcanic eruption in June 1991. The first method is to use the so-called analysis increments. These are differences between a meteorological analysis (i.e. an assimilated meteorological state), and a forecast valid at the same time and initialised from analysed data a few hours earlier (typically 3 or 6 hours). The analysis increment constitutes an approximation to R because it describes the initial drift of the model away from the observed (analysed) data. The use of analysis increments to estimate external forcing of the atmosphere was first suggested by Alpert et al. (1998) in a study aiming at quantifying the effect of tropospheric dust over the eastern tropical Atlantic. Andersen et al. (2001) (AKA) performed a pilot study demonstrating how the methodology can be used to estimate the stratospheric heating due to volcanic aerosol load.

In our second method we use a simple four-dimensional assimilation, the so-called nudging technique, of analysed data into a climate model. The nudging is a relaxation towards observed data (analyses in our case), which will force the model to stay quite close to the observations. Therefore, the relaxation term is an approximation to R. The methodology and its potential problems are described in KG.

The report is organised as follows. Section 2 describes the data used to estimate the temperature tendency residuals. Section 3 describes the residuals we obtain when using the analysis increment method while section 4 deals with residuals estimated via the nudging technique. In section 5 we describe a few idealised experiments designed to estimate the accuracy of the two methods. Finally, section 6 discusses a number of issues related to the methodology and concludes the report.

2. Data and data handling

As basic data we use the re-analyses from the first European re-analysis project ERA15, covering the period 1979-1993 (Gibson et al., 1997). For the analysis increments we have used the ERA15 data listed in table 1. To enable comparison with the nudging assimilation into ECHAM4 (see below), we consider as our first data set the anomalies in 6-hour temperature increments covering the period July 1991 to June 1993. Our second and third increment data sets are based on idealised experiments (to be described in section 5.1) with the ARPEGE/IFS climate model (Déqué et al., 1994) run at T42 resolution and with the same 31 vertical levels as used in the ERA15 assimilating model. In section 6 we also discuss a fourth data set: the anomalies in the 24-hour temperature analysis increments from the full ERA15 period. This longer increment data set is also listed in table 1 as well as the corresponding full temperature anomalies being used for inter-comparison in section 6.

For the nudging experiments we have assimilated the 6-hourly ERA15 data from the period July 1991 to June 1993 into the ECHAM4.6 model at T42 horizontal resolution. ECHAM4.6 is an intermediate model version between ECHAM4 (Roeckner et al. 1996) and the new model, ECHAM5. The assimilation was done with data interpolated from the ERA15 spatial representation

Experiment number	τ_T	τ_V	τ_D	τ_{lnPs}	τ_{ST}	τ_{SM}	**Correction of the diurnal cycle**
E1 (ECHAM4.6)	24 h	6 h	48 h	24 h	∞	∞	No
E2 (ARPEGE CLIMAT)	24 h	6 h	48 h	∞	48 h	48 h	Yes
E3 (ARPEGE CLIMAT)	12 h	12 h	12 h	∞	48 h	48 h	Yes

Table 2. Column 1 lists the experiment number and the model used, 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 Kaas and Guldberg, 2001 for more detailed explanaiton) from the assimilation.

(320x160 Gaussian grid points and 31 hybrid sigma-pressure vertical levels) to the representation used in ECHAM4.6 (i.e., 128x64 Gaussian grid points and 19 hybrid sigma-pressure levels). It is noted, that the ECHAM4.6 levels within the lower stratosphere are identical to those in ERA15. To be assimilated it was further needed to perform a temporal interpolation from the 6 hourly data to the individual time steps used in ECHAM4.6. The relaxation coefficients (see KG for details) used for the nudging are listed under experiment E1 in table 2. Our investigation of residual forcings based on this re-re-assimilation of ERA15 is limited to pressure levels at 30, 50 and 70 hPa [2].

Table 2 also lists a second and third experiment E2 and E3. These are idealised experiments to be described further in the section 5.2.

3. Residuals based on analysis increments

AKA used differences between analyses and 24-hour forecasts to estimate the volcanic heating at 30 hPa. These one-day forecast residuals (data set no. D4 in Table 1) are equivalent to 24-hour temperature increments. The main results from AKA are repeated in Fig. 1 in terms of Hovmöller diagrams of the zonal mean analysis increments from 50S to 50N. The top panel shows the deviation from the average annual cycle (1979-1993) of the increment. There is an overall tendency for a progressively stronger cooling of the lower stratosphere interrupted by the two heating events associated with the El Chichon (March 1982) and Pinatubo (June 1991) volcanic eruptions. However, the plot leaves the impression of a quite blurred picture. This is because it includes all the observational inaccuracies/noise and because of inability of the ERA15 assimilating model to simulate certain stratospheric phenomena as discussed in section 2. In the tropics it is mainly the QBO, which causes problems. AKA used a singular value decomposition (SVD) between stratospheric winds and the temperature increments to empirically isolate this effect of the QBO. The result of this is shown in the middle panel of Fig. 1. Note, that a time series covering several cycles of the QBO is needed to use this procedure. The lower panel finally shows the analysis increments but with the QBO effect removed. It can be seen that the direct 24-hour increment method leads to peak volcanic heating anomalies around 0.3 K/day for El Chichon and 0.25 K/day for Pinatubo. It is concluded in AKA, that this estimate is in fair agreement or somewhat to the low side and less persistent, as compared to estimates based on radiative transfer models (see e.g. Kinne et al., 1992; Stenchikov et al., 1998).

As discussed in KG in the case of nudging, there is a risk of partly misinterpreting a heating error, i.e., a tendency residual in the first equation of thermodynamics as a residual in the wind forcing, i.e., an error in the wind drag or acceleration, and visa versa. For estimates based on nudging this problem becomes progressively more severe when a weaker relaxation is used. In the case of analysis increments, we will obviously have the same problem: due to dynamical adjustments, analysis increments based on relatively long lead times, as in AKA, will result in underestimation of the volcanic heating and in misinterpretation in terms of artificial wind forcing residuals. This problem is discussed further in section 5.1.

To enable comparison with the estimates based on nudging (see below), Fig. 2 shows anomalies of 6-hour analysis increments for the period July 1991 to June 1993 at three different pressure levels: 30, 70 and 100 hPa (data set no. D1 in Table 1). There are two contributors to the differences between the upper panel in Fig. 1 and the right panel in Fig. 2: 1) the use of 6-hour increments versus 24-hour increments, and 2) anomalies with respect to the mean annual cycle in the period January 1985 to December 1990 versus the period January 1979 to December 1993. We anticipate that the

[2] Our original intention within the ACE project as to re-assimilate all 15 years of ERA data into the ECHAM4 model using the nudging technique, i.e., a Newtonian relaxation towards ERA, and to estimate both the El Chichon and Pinatubo volcanic forcings. However, due to computation and storage limitations this turned out not to be possible within the financial frames of ACE. Instead we planned to analyse the volcanic signal from a previous re-assimilation of ERA into the ARPEGE/IFS climate model (Kaas et al., 2000). However, because of a very weak stratospheric nudging with relaxation (or e-folding) times going gradually to 0 at the model top in this assimilation, it was found that the identified forcing was unrealistically small (less than .05 K/day). Therefore, as a final solution, it was decided to consider only the Pinatubo case, and to perform a new set of ERA15 re-assimilations (experiment E1 in Table 2) into the ECHAM4.6 model including the period January 1985 to June 1993.

second contribution is most important, since observations in the lower stratosphere were available only once or twice per day for assimilation in ERA15, meaning that also the 6-hour increments will only reflect one or two daily observations, which makes them quite similar to 24-hour increments. Note that no attempts were made to remove the effect of the QBO in Fig 2. Fig. 2 clearly shows an estimated volcanic heating which is considerably stronger at 50 than at 70 and 30 hPa. This is a somewhat surprising result, since estimates based on radiative transfer models (Stenchikov et al., 1998) suggest a maximum heating closest to 30 hPa.

4. Residuals based on nudging

As for analysis increments based on relatively long forecasts (24 hours), and as discussed in KG, a nudging with relatively weak relaxation will underestimate the actual anomalous forcing of the atmosphere, but on the other hand a stronger relaxation will introduce other problems mainly related to the representation of the diurnal cycle in the interpolated analysis and to associated problems with dynamical adjustments. The relaxation used here (experiment E1 in table 1) is relatively weak. The resulting anomalous heating residual in the period July 1991 to June 1993 is plotted as Hovmöller diagrams in Fig. 3. The anomalies are calculated relative to the mean annual cycle of the nudging term in the period January 1985 to December 1990. No attempts were made to correct for the effect of the QBO as in Andersen et al. (2001).

As expected there are many similarities between the timing and the sign of estimates in Fig. 2 and Fig. 3. It is also seen, that the nudging, in agreement with the increment method, suggests a stronger heating at 50 hPa than at 30 hPa, However, the anomalous heating obtained with the nudging is generally much weaker than the increment anomalies suggest. We ascribe this to the weak nudging coefficients and to the coarser resolution (T42 versus T106 in the ERA15) used in E1. We will discuss this issue further in the section 5.

5. Estimation of accuracy

5.1 Analysis increments

To estimate if the increment method provides a reasonable estimate of the true heating we have performed idealised experiments resulting in data sets D2 and D3 listed in table 1. Both experiments are based on a perpetual January simulation with the ARPEGE/IFS climate model in T42 resolution. In this basic simulation we ran the model in its standard configuration, but in the first equation of thermodynamics a constant zonally symmetric heating in the lower stratosphere was added to the model's own heating resulting from the parameterised processes. The basic simulation is intended to mimic nature in a situation where an anomalous forcing, i.e., a volcanic forcing, influences the atmosphere. The simulation covered 60 days, and full model (so-called re-start) data were stored each 24 hours. Here we only consider the last 30 days to allow for model spin-up. Fig. 4a shows the artificial additional heating in the basic simulation. This heating mimics the heating rates estimated with radiative transfer models (see Stenchikov et al., 1998) in the case of typical volcanic eruptions near the Equator.

Based on the daily output from the basic simulation a total of 30 24-hour "forecasts" were made with exactly the same model, but in its standard configuration, i.e., without the additional forcing in the first equation of thermodynamics. These simulations mimic forecasts with an assimilating model not explicitly including a parameterisation of volcanic aerosols, as in the case of the ERA15 assimilating model. The ensemble average of the differences between the verifying basic simulation – the "truth" – and each 24 hour "forecast" constitute a measure of the accuracy we can expect to achieve when using 24 hour increments to estimate volcanic heating of the lower stratosphere. Fig. 4b shows this ensemble average and Fig. 4c the difference between the two. It is easily seen that the 24-hour increment method tends to weaken the heating signal by more than a third. Furthermore, and as expected, the signal is smeared out and covers higher latitudes than the true heating. This is related to the problem of adjustments between wind and mass fields, mentioned in section 3. Accordingly we should be able to see an artificial 24-increment in the wind field, although no wind forcing was added in the basic simulation. As an example of this the zonal average of the zonal wind residual is displayed in Fig. 4d. Although this forcing is quite weak with a maximum around 0.15 m/s/day it demonstrates the basic problem of potential misinterpretations.

The results plotted in Fig. 5 are equivalent to those in Fig. 4, except that the heating anomaly in the basic simulation was located at 25N. The location of a heating off the Equator does not improve or deteriorate the ability of the increment method to detect the volcanic heating, and also in this case we detect a spurious wind forcing.

We conclude that analysis increments constitute a fair method to detect varying anomalous heating of the lower stratosphere, but that the heating rates are underestimated by more than 30%.

5.1 Nudging

To estimate if the nudging assimilation provides a reasonable estimate of the true heating we have performed the idealised experiments E2 and E3 listed in table 1. These experiments were performed with the ARPEGE/IFS climate model at T21 horizontal resolution and 31 vertical levels. Both experiments consist of assimilation of a simulation equivalent to the basic simulation used to investigate analysis increments in the previous section. This basic simulation is 30 days long, and run in perpetual January mode (at T21). A constant zonally symmetric heating in the lower stratosphere was added to the model's own heating. Fig. 6a shows the heating used. The heating is almost identical to that in Fig. 4a, but is displayed on model levels in Fig. 6a. The data from the basic simulation was stored each 6 hours.

To enable assimilation via nudging this model output was interpolated in time to the individual time steps of the model using a cubic spline interpolation. This was done to mimic the procedure used when ERA15 data are assimilated. We used the two sets of relaxation coefficients E2 and E3 listed in Table 2 to nudge a version of the model not including the additional heating in Fig. 6a towards the output from the basic simulation. Fig. 6b shows the 30 day average of the relaxation term (see AG for details). This average should be equal to the forcing in Fig. 6a, if the nudging method was perfect. But it is seen that the relaxation parameters E2 lead to a 50% underestimation of the heating. However, if we use the stronger relaxation E3 we obtain a better reproduction with only about 30% underestimation.

It is interesting to compare the findings in sections 5.1 and 5.2 with the results in Fig. 2 and 3. The assimilation into ECHAM4.6 at T42 horizontal resolution results in much weaker heating residuals than obtained with the increment method: for the Pinatubo peak the nudging method only gives about 0.1 K/day as compared to about 0.2 K/day for the increment method. In the idealised experiments with ARPEGE/IFS we obtain 0.15 K/day with the nudging technique in the experiment E2 (with relaxation identical to that in E1 in the stratosphere) while the increment method gives 0.2 K/day. The reasons for this are unclear and calls for further experimentation. We guess it could be related to the use of 24-hour increments in data sets D2 and D3 and to the difference in horizontal resolution: the difference between T42 (D2 and D3) and T21 (E2) is smaller than jump from T42 (E1) to T106 (D1).

6 DISCUSSION

The use of data assimilation for detection of changes in the atmosphere constitute a simple filter presumably removing most of the internal noise associated with chaotic atmospheric dynamics. When we consider "raw" observations, the large level of internal climate variability implies that a long period of observations is needed to obtain e.g. statistically significant trends. Based on the notation in section 1 the assimilation makes it possible to detect the change in atmospheric heating rates R'_e directly by removing the internal noise R'_i. However, a full removal of R'_i requires the use of a perfect atmospheric model in the data assimilation. We can demonstrate these issues further by comparing the temperature anomalies (Fig. 7) in the ERA15 data with the temperature increment anomalies (Fig. 8) at three different levels 30, 100 and 850 hPa, i.e., one level in the lower stratosphere, one near the tropopause and one in the lower troposphere. At 30 hPa the QBO temperature signal is clearly visible in Fig. 7, but unfortunately it can also be seen in Fig. 8. This is because the ERA15 assimilating model is non-"perfect" and has a coarse stratospheric resolution. At 100 hPa the assimilating model behaves better because it has more levels. Here the volcanic heating (Fig. 8) is weaker than at 30 hPa, but because of the much smaller R'_i the volcanic signal is seen more clearly. At 100 hPa it is also interesting that the response (Fig. 7) to the volcanic forcing at 100 hPa has a very different structure from that in Fig. 8 because of dispersive processes in the atmosphere. In the lower troposphere (lower panels) there are some similarities between heating and response with an apparent tropical cooling taking place around 1985/86. The large tropical anomalies in Fig. 7 are mostly related to the ENSO phenomenon. If the ERA15 model is non-perfect in the sense that it has a too weak response to the sea surface temperature anomalies, it will be seen in Fig. 8 (lower panel) as heating anomalies occurring largely at the same time as the ENSO events. There are some weak signs of this in the figure.

Both the analysis increment method and the nudging technique will lead to results that are influenced by the four general problems listed in the in section 1. Regarding the first item, one must generally be aware, that a considerable part of the anomalies in e.g. Fig. 7 and 8 could be artificial and due to inhomogenous (mainly satellite) data being assimilated in the ERA15 project. There is little doubt that data from the ACE mission would help reducing inhomogeneity problems because the observations will have small bias and furthermore, due to the technique used, the bias will be the same from one mission to the next. Therefore ACE data must be used in NWP data assimilation systems and in particular it must enter future re-analyses (item 3) to ensure their homogeneity.

Regarding item 2, we note that also here ACE is expected to improve the situation as it provides data with high vertical resolution and covering all regions of the globe.

It is expected that the new ERA40 re-analyses from the ECMWF will reduce the problems related to separation of the terms R'_e and R'_i (item 4). This is because the assimilating model is more advanced and has a higher resolution, particularly in the vertical, than the ERA15 model. A better model will generally reduce R'_i, as mentioned above.

We have found that both the nudging method and the increment method can be used to detect and quantify varying external heating in the atmosphere. Both methods underestimate the heating, apparently with the increment method being the best of the two. More experiments with other relaxation coefficients are needed, however, to verify this first impression.

We note that to some extend both methods depend on the actual model used. Since it is strong dynamical adjustments and not so much the speed of radiative adjustment which leads to underestimation we consider dependence on model physical parameterisation a minor issue. However, the spatial resolution of the model used is of key importance. To test this importance we plan to repeat our idealised experiments in section 5 but with the same model resolution as in ERA15, i.e. with the horizontal resolution enhanced to T106. Furthermore, these experiments should include tests of the difference between 6 and 24-hour increments, noting again as in section 3, that for real applications based on few observations per day for a given location there should be relatively little difference between 6 and 24-hour increments.

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FIGURES

Figure 1

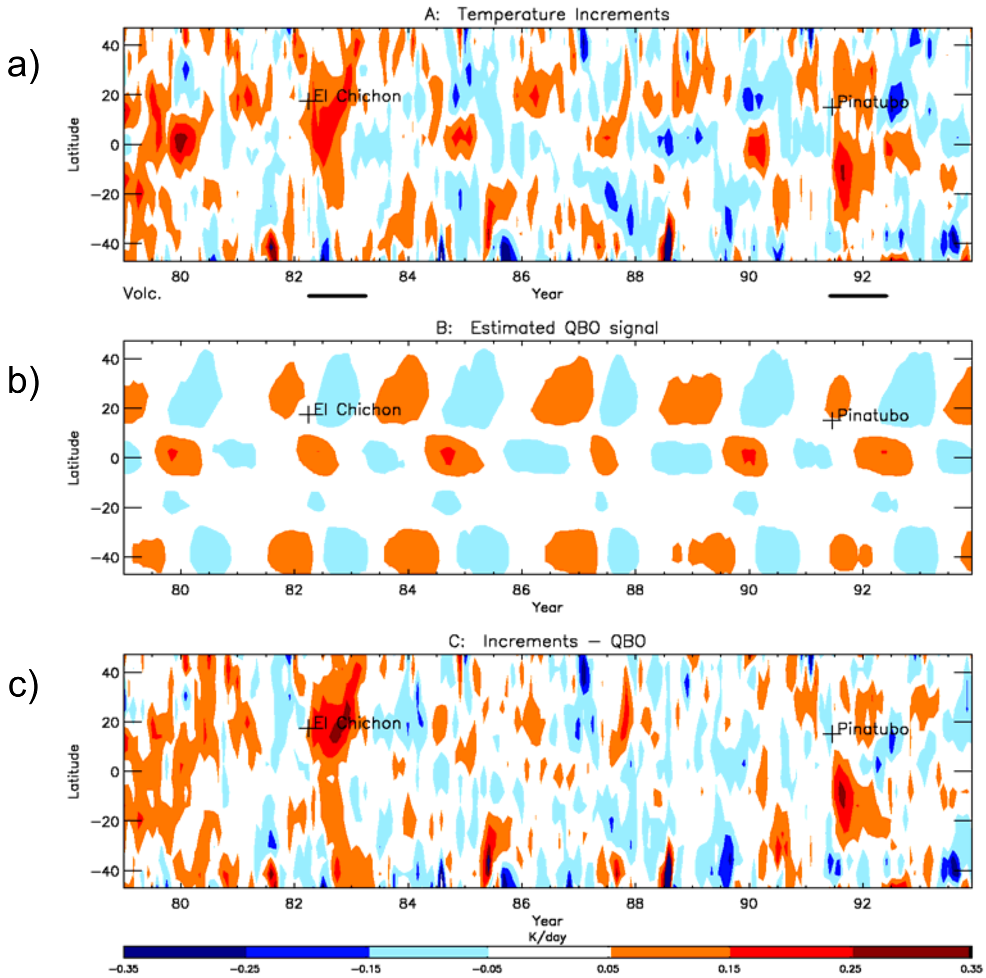

Figure 1. a) Hovmöller diagrams of 24 hour zonally averaged temperature analysis increments at 30 hPa. The line beneath the diagram indicates the periods that were excluded during calculation of SVD (see Andersen et al., 2001 for details). b) The estimated QBO signal of the increments calculated by using the first two patterns for the increments and the coefficients for the zonal winds. c) The increments with the estimated QBO signal removed. The color scale (in K/day) applies to all three Hovmöller diagrams. The horizontal axis displays time and the vertical axis the latitude from 50S to 50N.

Figure 2

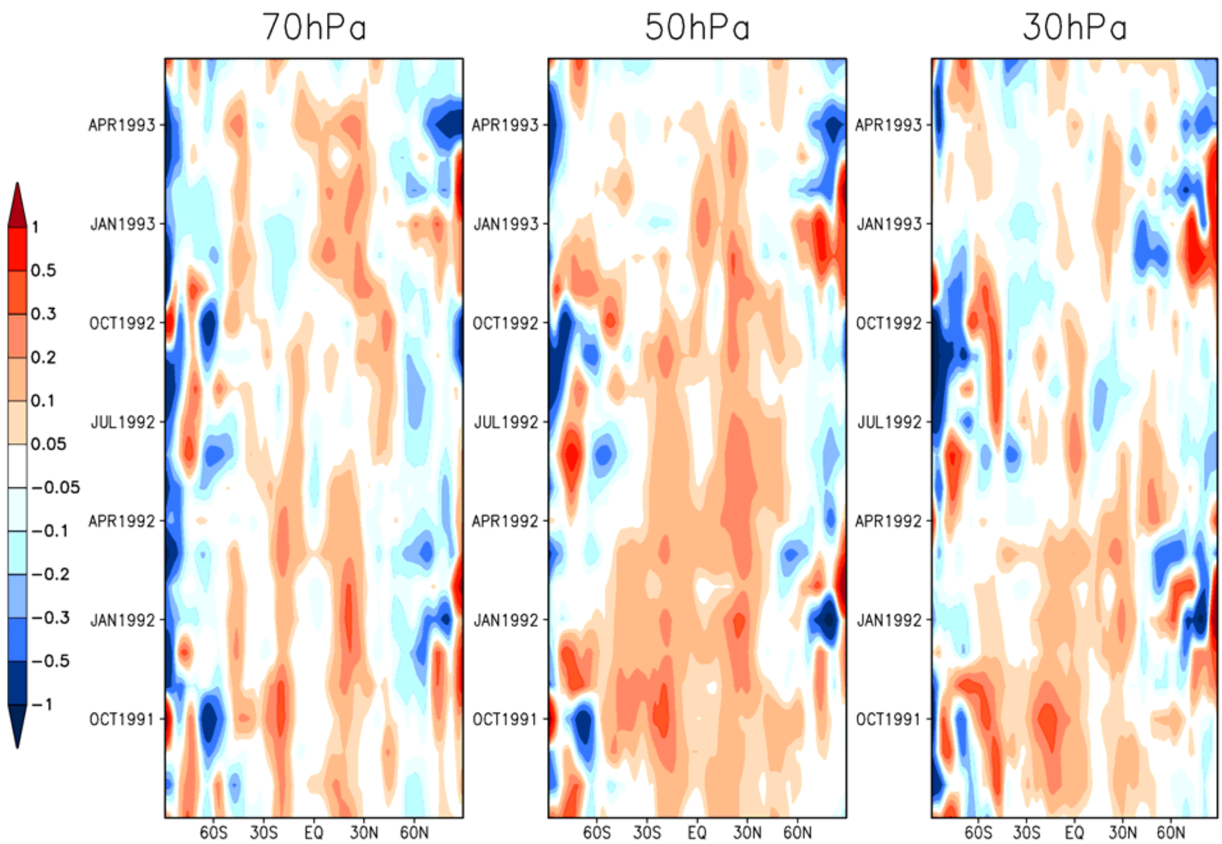

Figure 2. Hovmöller diagrams of anomalies of zonally averaged 6-hour temperature analysis increments at 70, 50 and 30 hPa in the period July 1991 to June 1993. The horizontal axes shows the latitude and the vertical the time. Units: K/day.

Figure 3

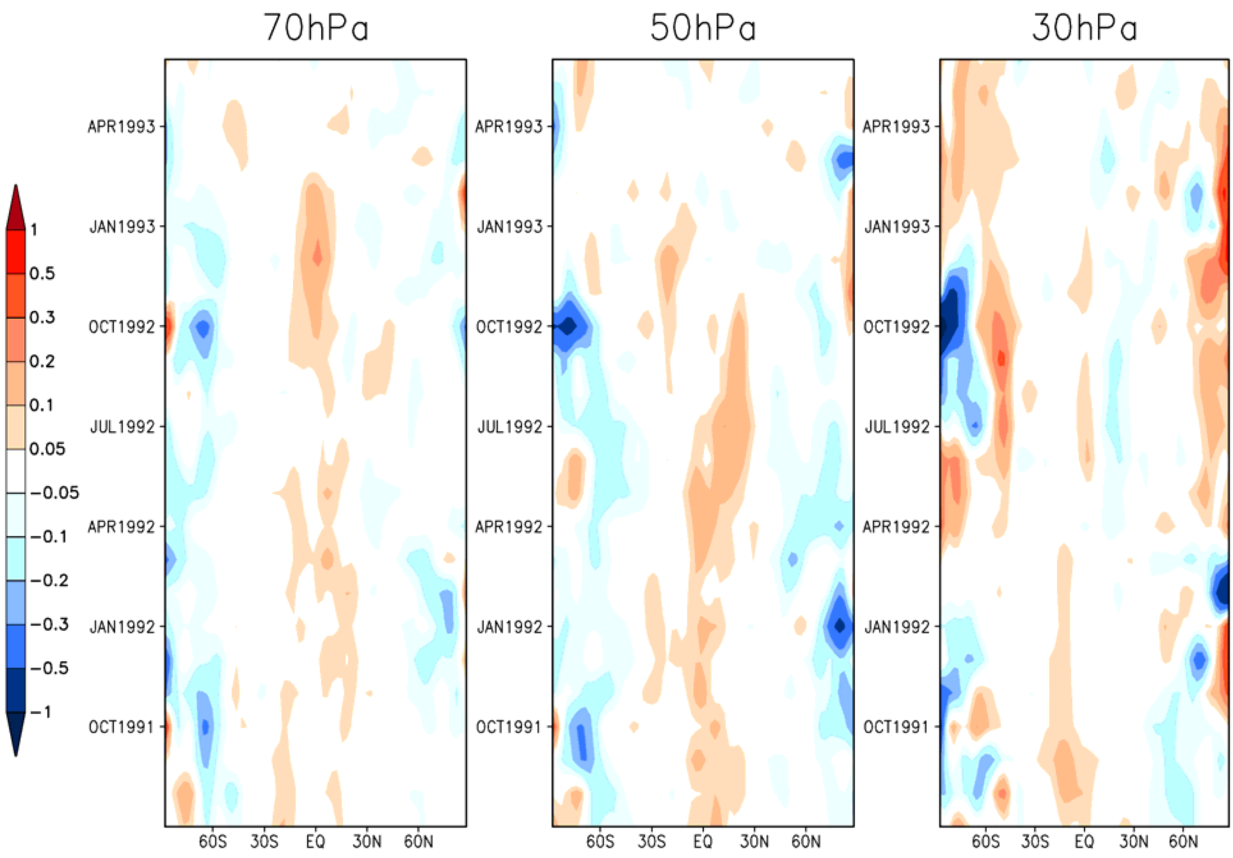

Figure 3. Hovmöller diagrams of zonally averaged tendency residual anomalies in the period July 1991 to June 1993 at 70, 50 and 30 hPa estimated via the nudging technique with relaxation coefficients as listed in Table 1 (E1). The horizontal axes shows the latitude and the vertical the time. Units: K/day

Figure 4

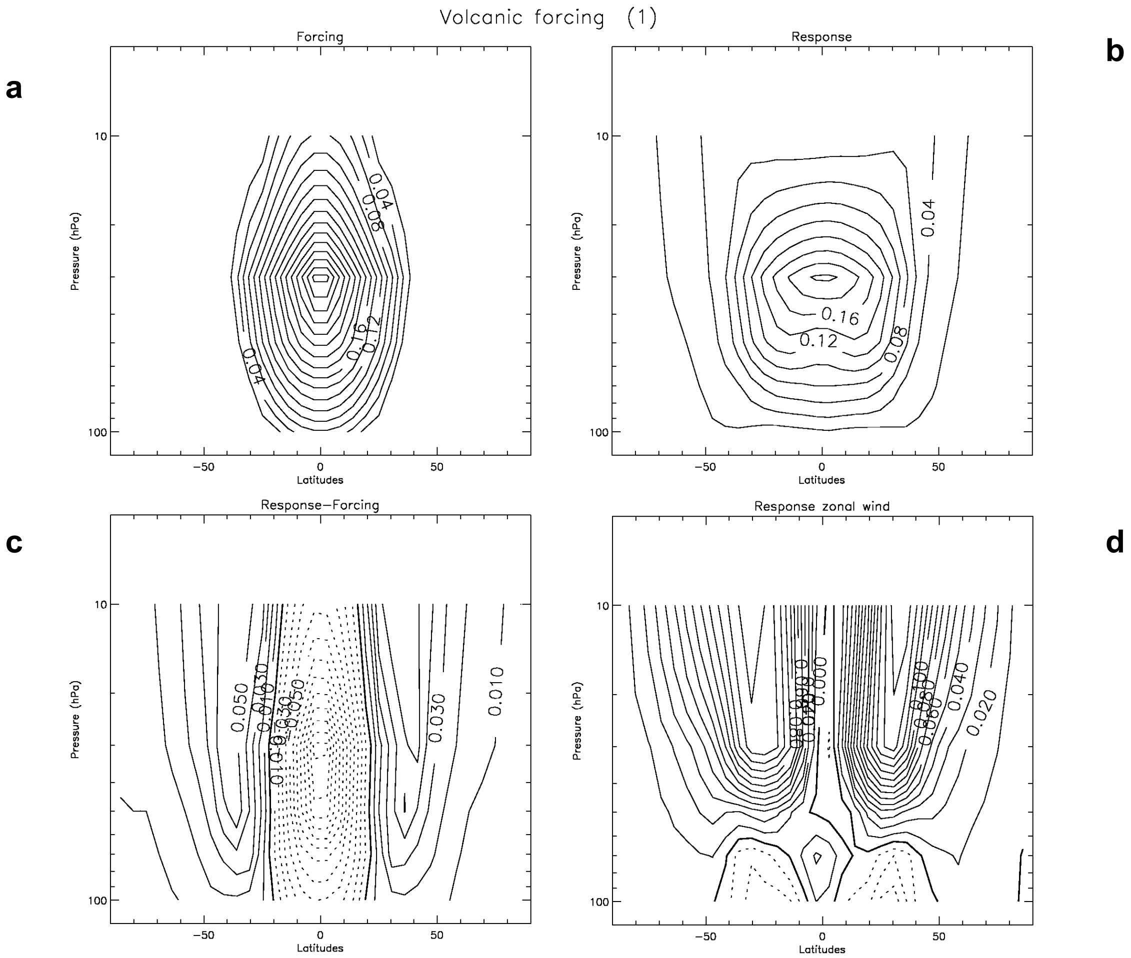

Figure 4. Zonally averaged results from idealised experiments testing the accuracy of the increment method (dataset D2 in table 1). Panel a shows the anomalous heating in the basic simulation (see text for details). Panel b shows the detected heating obtained as an average of 30 individual 24-hour temperature increments. Panel c is a difference plot between panel b and a. Panel d is similar to b, but for zonally averaged zonal wind increments.

Figure 5

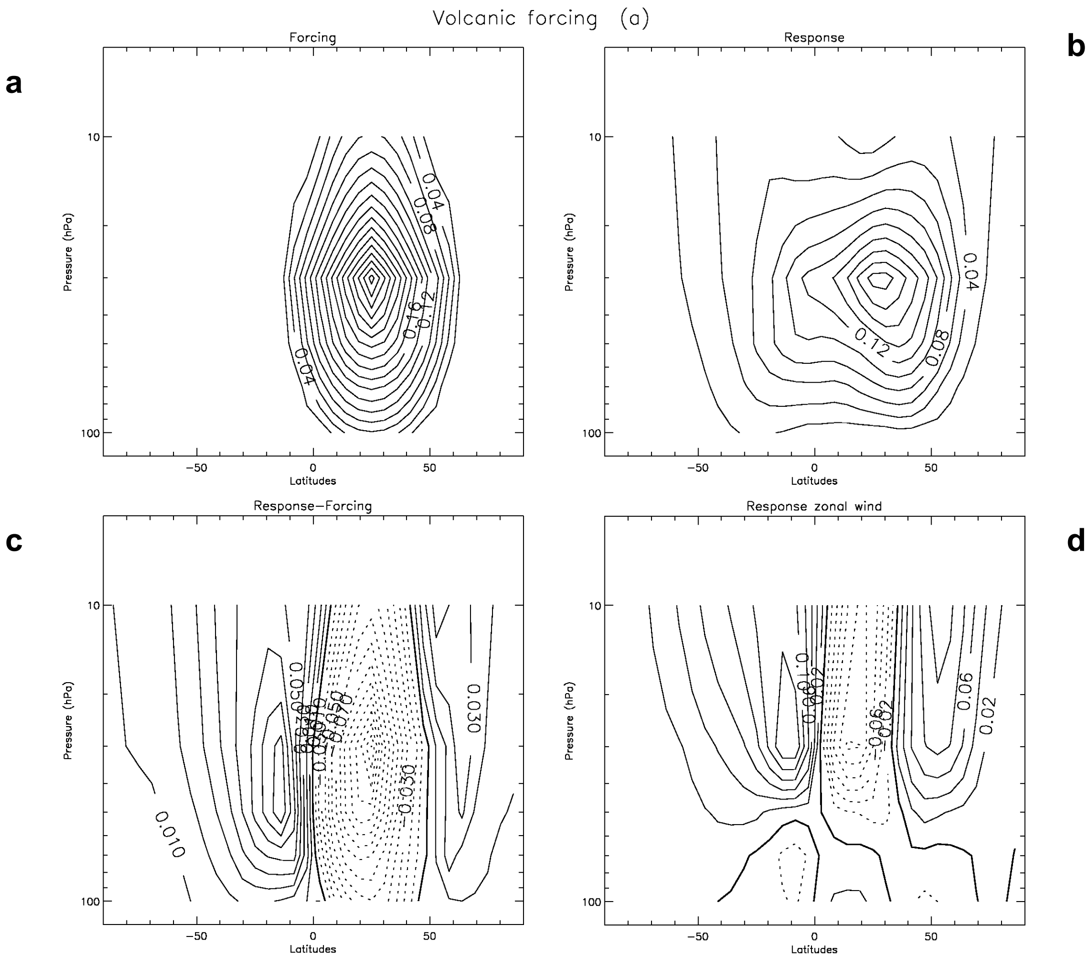

Figure 5. As Fig. 4, but with the test forcing located at 25 N (panel a).

Figure 6

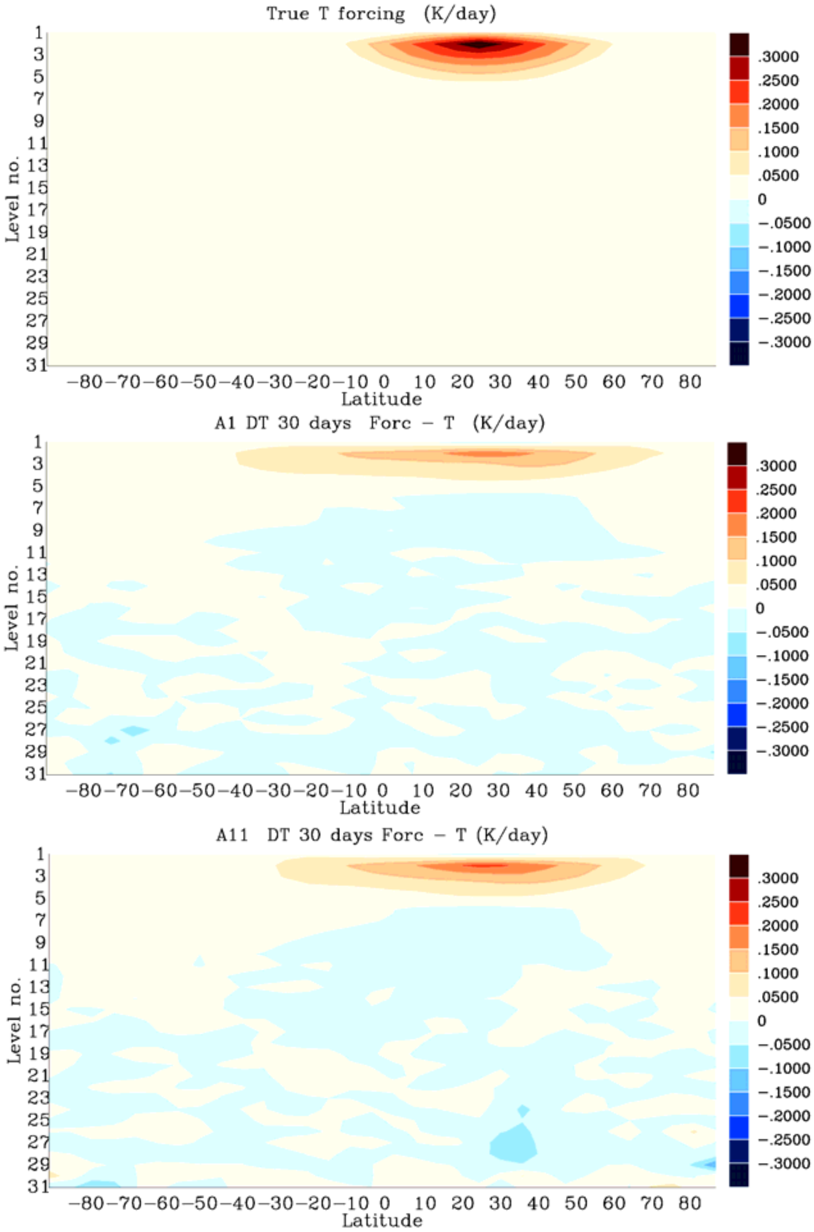

Figure 6. Zonal mean temperature forcings in idealised nudging experiments with the ARPEGE/IFS climate model run at T21 horizontal resolution. The upper panel shows a prescribed fixed zonal mean heating anomaly in a 30 day simulation with the model, and the middle panel shows the estimated heating residual when the synthetic data from the simulation are re-assimilated via the nudging coefficients in experiment E2 (see table 1). The lower panel is as the middle, but for nudging coefficients in experiment E3. The vertical axes show model level with level 1 at the model top. Units: K/day.

Figure 7
Figure 8

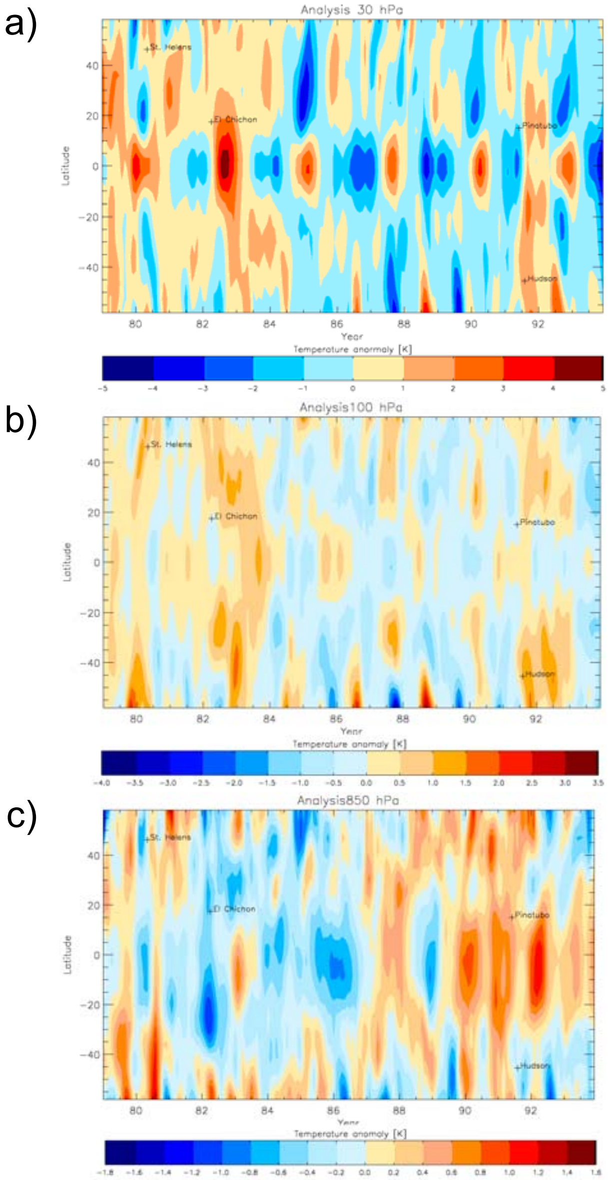

Figure 7. Hovmöller diagrams of zonally averaged temperature anomalies in the ERA15 data at levels 30 hPa in panel a), 100 hPa in panel b), and 850 hPa in panel c). The horizontal axis displays time and the vertical axis the latitude from 50S to 50N. The unit is temperature. Note the different colour scales in each panel.

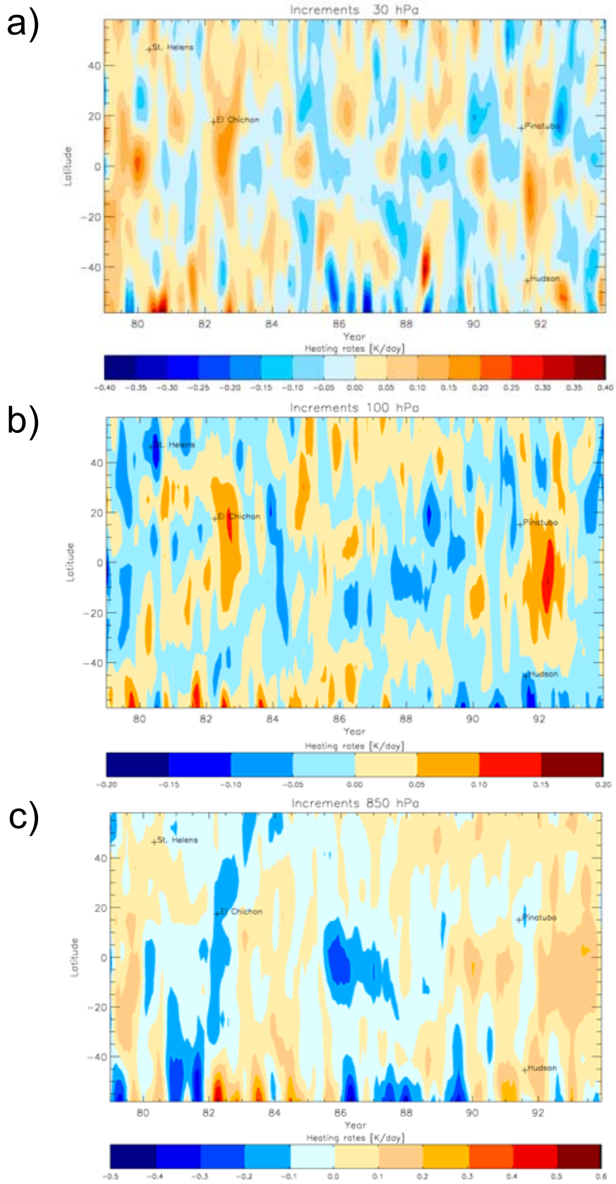

Figure 8. As Fig. 7, but for 24-hour analysis increments. Units: K/day. Note that panel a) equal to panel a) in Fig. 1, except for the different colour scale.

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).
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- 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*.
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- Synthesis of the STOWASUS-2100 project: Regional storm, wave and surge scenarios for the 2100 century. *Climate Centre Report 01-3*.
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