Northern hemisphere winter stratospheric variability in The Met. Office Unified Model

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(Received 3 December 1999; revised 27 March 2000)

SUMMARY

This paper uses the ensemble approach to study the simulated northern winter stratospheric variability in a tropospheric--stratosphere version of The Met. Office Unified Model (UM). The runs are for the December--March periods from 1979/80 to 1997/98; the ensemble for each winter has nine members. We use observed sea surface temperatures (SSTs), a fixed ozone climatology and fixed greenhouse gases. This paper concerns itself with the following questions: (1) Does the UM reproduce observed temperature trends with only SST variations? and (2) Can any significant trend be attributed to SST variations? The results of this paper suggest that the answer to question (1) is that the UM reproduces the observed temperature trend with only SST variations. In particular, northern winter in the simulated 1990s (1989/90 to 1997/98) is colder than northern winter in the simulated 1980s (1979/80 to 1988/89) at both 10 hPa and 50 hPa, with the model cooling trend (0.5--1.5 K between the decades for 30°N--70°N) comparable to that from observations (0.5--1 K decade$^{-1}$ for 30°N--70°N). The temperature differences between the simulated 1990s and 1980s generally are not significant at 10 hPa, but are generally significant at 50 hPa at the 95% significance level. The results in this paper provide a first step toward answering question (2) by suggesting that SST variations alone may explain the cooling trend at 50 hPa. Because the model results show that there is no significant cooling trend at 10 hPa outside a natural variability, one cannot rule this out as an explanation. A complete answer to question (2) requires the inclusion of all likely influences (in particular ozone and water vapour trends) and is beyond the scope of this paper.

KEYWORDS: Ensemble approach Major warmings Northern winter Stratospheric variability Temperature trends

1. INTRODUCTION

The ensemble approach (Palmer and Anderson 1994) where multiple realizations from a climate model are used to capture the range of variability in the atmosphere and other components of the Earth climate system, has become standard in studies of climate change (see, e.g., Tett et al. 1999), in seasonal forecasting (see, e.g., Stockdale et al. 1998), and in mid-range operational forecasting (see, e.g., Buizza et al. 1998). A key feature of the ensemble approach is that the squared error of the ensemble forecast is smaller than the mean squared error of all the individual forecasts (see, e.g., Brankovic et al. 1990).

By contrast, the ensemble approach has been very little used in the study of stratospheric variability. The traditional approach to studying stratospheric variability has tended to involve a single multi-annual realization from a climate model using one of the following: (a) observed sea-surface temperatures (SSTs), (b) climatological SSTs, or (c) perpetual SST conditions. Examples of this approach include Boville (1995), Hamilton (1995), Manzini and Bengtsson (1996), Langematz and Pawson (1997), and Yoden et al. (1999). Although a single multi-annual realization can provide, e.g., a good estimate of the interannual variability of the stratosphere, it cannot provide an estimate of the range of variability associated with a particular period. This means that important questions concerning the attribution of observed stratospheric changes to long-term 'natural variability' or to variability caused by anthropogenic effects are more difficult to answer with the traditional single realization approach than with the ensemble approach.

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As the ensemble approach provides a measure of the range of variability in climate simulations, it is particularly appropriate to address the questions which concern this paper:

(1) Does The Met. Office Unified Model (UM) reproduce observed temperature trends with only SST variations?
(2) Can any significant trend be attributed to SST variations?

A complete answer to question (2) requires the inclusion of all likely influences (in particular ozone and water vapour trends) and is beyond the scope of this paper. However, the results in this paper provide a first step toward answering question (2).

Reports of unusually cold northern winters in the lower stratosphere during the mid-1990s (Pawson and Naujokat 1999) and the absence of major stratospheric warmings throughout most of the 1990s (see, e.g., Manney et al. 1999), as well as other recent studies (Zurek et al. 1996, Coy et al. 1997, Pawson and Naujokat 1997 and Shindell et al. 1998) have raised the question of whether there is a clear trend in the persistence of the polar vortex into northern spring. Although this would be consistent with a stratospheric cooling caused by increases in well-mixed radiatively active gases or decreases in stratospheric ozone (e.g., Fels et al. 1980, Miller et al. 1992, McCormack and Hood 1994, Ramaswamy et al. 1996), the possibility of long-term natural variability cannot be ruled out. A cooling trend, as reported for the annual, hemispheric mean by Pawson et al. (1998) is likely to result in lower wintertime temperatures which, in turn, could lead to more polar stratospheric cloud formation and enhanced ozone destruction. This possibility has raised concerns about the effects of ozone change on (a) ecosystems via increases in ultraviolet radiation, and (b) the tropospheric energy balance via radiative feedbacks (Ramaswamy et al. 1996; Forster and Shine 1997).

This paper addresses the above two questions by applying the ensemble approach to study the northern winter stratospheric variability with a troposphere–stratosphere version of The Met. Office UM (Cullen 1993). The UM version used in this paper has 58 levels in the vertical, with the highest model full-level at 0.1 hPa (with the highest model half-level at 0.08 hPa) and good vertical resolution throughout the stratosphere (about 1.3 km). The tropospheric configuration of the 58-level (LS8) model is similar to that of the 30-level version of the UM (Pamment, personal communication). The horizontal resolution of the model is 2.5° latitude by 3.75° longitude. The model is run with a 15-minute physics time-step and two dynamics time-steps for every physics time-step (this is implemented to avoid instabilities developing from strong-wind cases). Similarly, a divergence criterion is applied at the highest full-model level. If a specified divergence limit is exceeded, the dynamics time-step is halved. This can be applied to any of the two dynamics time-steps in each physics time-step, but no further sub-division is allowed. As a result the model can have from two to four dynamics time-steps for every physics time-step.

The dynamical formulation of the model is based on the hydrostatic primitive equations. An efficient split-explicit scheme is used to solve the equations, which is designed to conserve mass, mass-weighted potential temperature, moisture and angular momentum (see Swinbank et al. 1998 for more details). As a consequence of the formulation of the dynamics, the model variables are carried on a staggered (Arakawa B) grid.

The representation of physical processes is provided by the so-called HADAM3 (HADley Centre Atmospheric Model version 3) physics (Pope et al. 2000). The model uses an orographic gravity-wave drag scheme (Shutts 1990; Milton and Wilson 1996) up to 20 hPa, where it is replaced by Rayleigh friction. The ozone field is fixed and is
given by the ozone climatology of Wang et al. (1995), which was used for the Second Atmospheric Modelling Intercomparison Project (AMIP II). The greenhouse gases are fixed.

For each northern winter in the period 1979/80 to 1997/98, four month integrations were made with nine-member ensembles. An alternative approach would involve running a 19-year integration with a nine-member ensemble; however, the cost of such an experiment for L58 would be prohibitive (current computing constraints would preclude the use of an ensemble larger than three or four members). We thus sacrifice the long numerical integration approach to allow the use of a relatively large ensemble size.

The integrations were initialized on the nine consecutive days of the period 22–30 November with the 12 UTC analyses from ERA extended to cover the period beyond 1993. The initial data was extended in the vertical from 10 hPa to 0.1 hPa by ascribing to the temperature, and the zonal and meridional wind fields between 10 hPa and 0.1 hPa the values at 10 hPa. As radiative time-scales are relatively fast in the upper atmosphere, the circulation relaxes to realistic values over periods of time ranging from a few days to about a week. The observed SSTs and ice cover are based upon the Hadley Centre Global sea-ice and Surface Temperature (GISST) analyses (Rayner et al. 1996) for the winters 1979/80 and 1980/81 and upon the Reynolds Optimum Interpolation (OI) analyses (Reynolds and Smith 1994) for the winters 1981/82 to 1997/98.

The L58 version of the UM was used in the AMIP II experiment. Comparison between the L58 January climatology and the European Centre for Medium-range Weather Forecasts (ECMWF) Re-analysis (ERA; Gibson et al. 1997) extended to cover the period beyond 1993, indicates that L58 has a cold bias of about 5 K in the northern winter mid stratosphere. By comparison against the UKMO stratospheric analyses (Swinbank and O’Neill 1994) L58 typically has a stratospheric cold bias of 5–10 K.

A 49-level (L49) version of the UM, which has a stratosphere similar to that of L58, has been described in Swinbank et al. (1998) and in Pawson et al. (2000). These papers show that L49 provides a reasonable representation of the northern winter stratosphere by comparison against several climatologies derived from observations.

The L49 model tends to have a cold bias in the lower-middle stratosphere (100 hPa–30 hPa) by comparison against the Tiros Operational Vertical Sounder (TOVS) climatology, but shows a realistic annual cycle at 100 hPa by comparison against the ERA climatology. Note that the stratospheric bias in L58 is an improvement on the bias in the L49 version which incorporates HADAM2b physics (Swinbank et al. 1998). This improvement can be attributed to the use of the Edwards–Slingo (Edwards and Slingo 1996) radiation scheme in HADAM3. The L49 model also represents well the structure of the zonal winds for January by comparison against a climatology derived from the former National Meteorological Center (NMC) by Randel (1992). However, both L49 and L58 are unable to simulate the tropical quasi-biennial oscillation (QBO) and, although the model simulations show a semi-annual oscillation (SAO) near the stratopause, the westerly phase is very weak or absent.

Hansen et al. (1997) have studied the role of climate forcings and unforced variability via ensembles of climate simulations in which forcings are added one by one. One of their results is that observed SSTs alone provide a global-mean annual-mean cooling of about 2 K in the lower and mid stratosphere for the period 1979–95. Although there are similarities between the experiment described in this paper and run ‘Aa’ in Hansen et al. (which just considers SST forcings), there are two important differences: (a) the model used in this paper encompasses the whole stratosphere whereas that of Hansen et al., which has a model lid at 10 hPa, does not, and (b) the model used in this paper has higher vertical resolution in both the stratosphere and troposphere than that of
Hansen et al., which has nine levels in the vertical. There is a further difference between the experiment described in this paper and run 'Aa' of Hansen et al.; viz., Hansen et al. performs a five-member ensemble of 18-year runs (1979–96), whereas the experiment in this paper involves nine-member ensembles of four-month runs for each of the 19 northern winters in the period 1979/80 to 1997/98.

The paper is structured as follows. In section 2 we evaluate (a) the model variability for individual northern winters, and (b) the model's climatology and variability. Comparison is made against The Met. Office analyses and/or ERA. In section 3 we use the ensemble approach to provide a measure of the variability for individual northern winters and how it depends on the SST forcing. We then discuss the major warmings simulated by the model and evaluate the realism of the simulations. This is done by (a) studying a particular northern winter (chosen because it exhibits the typical features associated with major warmings), and (b) providing a summary of the simulated major warmings, including their frequency and distribution. In both cases comparison is made with observed major warmings. The results from sections 2 and 3 establish the model's variability and realism. In section 4 we compare the simulated northern winters in the 1980s and the 1990s and provide answers (with caveats) to the questions posed above. Finally, in section 5 we provide conclusions and suggestions for further work.

2. Evaluation of model variability

In this section we firstly compare results from the ensemble simulation with the equivalent data from The Met. Office analyses and, secondly, we evaluate the model's climatology and variability against The Met. Office analyses and/or ERA. The comparison between the model realizations and The Met. Office analyses is done for several northern winters (December–March) in the period 1991/92 to 1997/98. The comparison between the model and ERA is done for the January monthly mean climatology. The evaluation of the model climatology is done by averaging over all 171 model realizations of northern winter (19 winters from 1979/80 to 1997/98 times nine members in each ensemble). The evaluation of the model variability is done by calculating the standard deviation of all model realizations about the model climatology. The ensemble approach allows comparison of the model unforced variability against observations (section 2(a)), and provides a large number of realizations to calculate the model’s climatology and variability (section 2(b)).

(a) Comparison between model and The Met. Office stratospheric analysis

Figure 1 shows the 10 hPa temperature time-series at 90°N (a proxy for stratospheric variability) for the model and The Met. Office analyses. The left-hand side panels show the nine members of the model ensemble, plus the average of the ensemble, for 1995/96 (Fig. 1(a)) and 1997/98 (Fig. 1(c)). The right-hand panels show The Met. Office analyses for 1995/96 (Fig. 1(b)) and 1997/98 (Fig. 1(d)).

The Met. Office analyses for 1995/96 are generally encompassed by the model realizations, and the general behaviour of the ensemble mean (depicted by the thick line) follows the low frequency behaviour of the analyses; a slight decrease during the period December–January followed by a gradual increase over the period February–March. This general agreement between the model realizations and The Met. Office analyses is also seen for the zonal-mean zonal wind at 61.25°N and 10 hPa (not shown).

The model spread (a proxy for unforced variability) tends to be small during the first half of December, a period in which the model simulation is influenced by the initial conditions. Studies of the autocorrelation of the anomalies for each ensemble member
Figure 1. North pole temperatures (K) at 10 hPa for (a) the nine members of the model ensemble for northern winter 1995/96 (top left-hand panel); (b) The Met. Office analyses for northern winter 1995/96 (top right-hand panel); (c) the nine members of the model ensemble for northern winter 1997/98 (bottom left-hand panel); (d) The Met. Office analyses for northern winter 1997/98 (bottom right-hand panel). The means of the ensemble are depicted by the thick solid line.
with respect to the ensemble mean suggest that the influence of the initial conditions diminishes after 10–15 days in the troposphere and the stratosphere. This is supported by other results reported in this paper (see later). However, there could be an influence from the initial conditions on the range of flow regimes the model adopts on longer time-scales.

In contrast to 1995/96, The Met. Office analyses for 1997/98 are not generally encompassed by the model realizations, and the general behaviour of the ensemble mean (depicted by the thick line) does not follow that of The Met. Office analyses. This general lack of agreement between the model realizations and The Met. Office analyses is also seen for the zonal-mean zonal wind at 61.25°N and 10 hPa (not shown). In 1997/98 the model ensemble members have very little spread, i.e. variability, compared to 1995/95 and this is consistent with the differences in model behaviour between 1995/96 and 1997/98.

The only difference between the model set-ups for 1995/96 and 1997/98 are the SSTs used and the initial conditions. In particular, 1997/98 experienced very strong El Niño-Southern Oscillation (ENSO) conditions (Slingo 1998). Focusing on the individual ensemble members for 1997/98, it can be seen that they follow closely The Met. Office analyses until about mid-December (when the influence of the initial conditions diminishes). After this time, although there is one ensemble member which appears to follow The Met. Office analyses for some days, the ensemble members have little variability over January and February and do not reproduce the temperature variations shown by The Met. Office analyses. By March the ensemble members show increased variability as the final warming takes place; The Met. Office analyses show relatively high temperature values but not a clear increase in the temperature trend (as occurs during 1995/96—see Fig. 1(b)).

The rapid temperature changes seen over December in 1997/98 suggest rapid changes in the stratospheric flow regime. As discussed by Palmer and Anderson (1994) and Lahoz (1999), rapid changes in the flow regime tend to be associated with low model predictive skill. Conversely, steady flow regimes such as those suggested by the temperature variability seen during most of 1995/96 tend to be associated with high model predictive skill. This suggests that the conditions for December 1997 should be difficult to simulate by a single model realization and that they could represent a case unlikely to be captured even by a large number of realizations. In this case, the little variability exhibited by the ensemble members suggests that the putative rapid changes in the stratospheric flow regime strongly constrain the subsequent evolution of all ensemble members.

The behaviour of the ensemble members for 1997/98 suggests that while the initial conditions constrain the variability the members follow the changes in the flow regime, but when this influence diminishes, the members do not follow these changes. Although the model has deficiencies (e.g. it has a cold bias), the evidence suggests that the shortcomings in the simulation of 1997/98 may be due to the nature of the flow regime rather than to any inherent model deficiencies.

This contention is further strengthened by model simulations for winters where there are The Met. Office analyses available for comparison (1991/92 to 1997/98). For these winters, the ensemble spread tends to capture the observed behaviour of the 10 hPa 90°N temperature time-series throughout a substantial part of the December–March period (with the noted exception of 1997/98), although the model typically exhibits the aforementioned cold bias. The model ensemble members also exhibit a realistic behaviour of major warmings (see section 3(b)), and a realistic distribution and frequency of major warmings within each northern winter (see section 3(c)). Also, the
Figure 2. Time series (time v latitude) of the temperature (K) model climatology at 10 hPa. The contour interval is 2 K. (b) Time series (time v latitude) of the zonal wind (m s\(^{-1}\)) model climatology at 10 hPa. The contour interval is 2 m s\(^{-1}\).
Figure 3. Time series (time v latitude) of the temperature (K) model variability at 10 hPa. The contour interval is 1 K.

monthly averaged temperature and zonal wind fields for the model ensemble members represents the stratospheric circulation in a realistic manner (not shown).

(b) Model climatology and variability

Figures 2(a) and (b) show that the climatology of the model at 10 hPa over latitudes 30°N–90°N is realistic, with the lowest northern winter zonal-mean temperatures occurring at high latitudes and during January, and the highest zonal-mean zonal wind values occurring between 60°N and 70°N and during late January and early February (see the observational datasets described in Swinbank et al. 1998 and Pawson et al. 2000). For levels between 10 hPa and 1 hPa the location of the maximum zonal-mean zonal wind values is tilted slightly equatorward. This indicates that the climatological polar night jet (PNJ) of the model is tilted slightly equatorward. Note that observations (e.g. Pawson et al. 2000) indicate that the PNJ tilts equatorward with height, and that most models incorporating Rayleigh friction (as does the L58 version used in this paper) have a PNJ which does not tilt equatorward.

At latitudes around 30°N, where there is relatively low model variability (see Fig. 3), the model has a cold bias of 5–10 K with respect to The Met. Office analyses (compare with Fig. 10 of Swinbank et al. 1998). This model bias is consistent with that illustrated in Fig. 1. Figure 3 shows that the model variability at 10 hPa during northern winter and latitudes 30°N–90°N is realistic, with the largest values (up to about 15 K) occurring during March, which is the period of the final warming. The variability for zonal wind (not shown) is also realistic and shows a maximum between 60°N and 70°N, the location of the PNJ at 10 hPa. Comparison with the variability of the 10 hPa temperature field in The Met. Office analyses for the period 1991/91 to 1997/98 (not shown) indicates that the model and the analyses share common elements such as increased variability toward
Figure 4. Monthly mean (latitude v pressure) of temperature (K) for a January multi-annual mean for (a) the model runs in this experiment (January 1980–January 1998) and (b) the extended ERA climatology (January 1979–January 1998). The contour interval is 5 K. (c) is the difference between the model and the ERA multi-annual means for January. The contour interval is 1 K. The shading indicates that the model is colder than ERA.
high latitudes and relatively high variability in March. On the other hand, the analyses generally show greater short-term variability, particularly during mid winter (note that the analyses variability is calculated from a much smaller number of realizations than for the model).

Figure 4 shows that the model January monthly mean temperature climatology has a similar pattern to the ERA climatology, and that the model has a cold bias of 5–10 K in the northern winter lower and mid stratosphere (for latitudes poleward of 45°N). Overall, the results from section 2 (see Figs. 1–4) suggest that the model is a useful tool for studying northern winter stratospheric variability.

Given the large number (i.e. greater than 30) of model realizations used to calculate the model variability, we regard this variability as being representative of the model’s natural variability for northern winter. Thus, we regard the multiple realizations as independent and coming from an approximately normal distribution (describing the simulated northern winter) with mean \( \mu_o \) and standard deviation \( \sigma_o \), where \( \mu_o \) is calculated model climatology and \( \sigma_o \) is the calculated natural variability. The behaviour of the ensemble members in each northern winter supports this view (e.g. the proportion of the ensemble members lying within 1 standard deviation from the mean on 1 February at 10 hPa and 90°N is close to the value expected from a normal distribution). We can then compare the climatologies for different sub-periods within the 1979/80 to 1997/98 period and assess if the difference between them is significant by comparing it against the variability of each sub-period. If the difference is larger than the variability considered, the climatologies for the sub-periods would be regarded as being significantly different. An example would be testing the null hypothesis that, given that the means and standard deviations of each sub-period are, respectively, \( \mu_1 \) and \( \mu_2 \), and \( \sigma_1 \) and \( \sigma_2 \), the means of each sub-period are the same, i.e. \( \mu_1 = \mu_2 \). We will make use of this idea when comparing the simulated northern winter in the 1980s and the 1990s in section 4.

3. Model Intraseasonal Variability

In this section we study the simulated stratospheric variability for individual northern winters and its dependence on the SST forcing. We then discuss the major warmings simulated by the model and evaluate the realism of the simulations. These results provide information on model variability and realism which contributes toward providing an answer to the questions posed in this paper.

In section 3(a) we study the behaviour of the simulated 10 hPa temperature time-series at 90°N for 1982/83 (a northern winter which experienced ENSO conditions) and 1984/85 (a northern winter which experienced La Niña conditions). In the period December 1979–March 1998 five northern winters (1982/83, 1986/87, 1991/92, 1992/93 and 1997/98) experienced ENSO conditions and one northern winter experienced La Niña conditions. All simulated northern winters which experienced ENSO conditions (except 1997/98) have at least one ensemble member deviating significantly from the others during mid winter, and we choose 1982/83 as representative of this behaviour.

To evaluate the realism of the model simulated major warmings we look at a case study for northern winter 1988/89 in section 3(b). This northern winter is chosen because it exhibits the typical features associated with major warmings. We follow the World Meteorological Organization (WMO) and define a major warming as occurring when a reversal in the zonal-mean zonal wind occurs down to 10 hPa and poleward of 60°N. For this definition we consider the time-series at 10 hPa, 5 hPa and 1 hPa, and latitudes 61.25°N to 88.75°N inclusive. Note that other definitions of a major warming can be used. In this case study we focus on time series of zonal wind and temperature and
Figure 5. North pole temperatures (K) at 10 hPa for the nine members of the model ensemble for northern winter 1982/83. The mean of the ensemble is depicted by the thick solid line.

synoptic maps of simulated geopotential height. To evaluate further the model's realism we provide in section 3(c) a summary of the simulated major warmings which occurred during December–March in the period 1979/80 to 1997/98. In both the case study and the summary we compare the simulations against observations.

(a) Dependence of variability on SSTs

Figure 5 shows the nine members of the model ensemble plus the ensemble mean (depicted by the thick line) for the 10 hPa temperature time-series at 90°N for 1982/83. The most striking feature of the time-series is the large spread, i.e. variability, in the ensemble members (note that this is in marked contrast with 1997/98, which also experienced ENSO conditions; see Fig. 1).

The features of these time-series worth noting are as follows. Firstly, the low spread exhibited over the initial two weeks of December, a period during which the impact of the initial conditions is significant (see section 2(a)). Secondly, the high spread exhibited throughout the rest of the winter, and not just the during the final warming where such high spread is to be expected. Thirdly, the large sudden warmings experienced by individual ensemble members, and which, in general, are not replicated by the other members. Fourthly, the large day-to-day variability exhibited by all members. This large day-to-day variability confirms the difficulty of making predictions in the winter stratosphere at time-scales beyond about ten days (see, e.g. Lahoz 1999).

The results for 1982/83 illustrate the advantages of the ensemble approach over the deterministic (i.e. one member) approach at the medium- and long-term time-scales. In particular, the average of the ensemble has a smoother variation than the individual ensemble members (and is thus more likely to be closer to the model climatology), and as seen for 1995/96, the spread of the ensemble can capture most of the observed
variability when, in general, the individual members cannot. Note that although the ensemble mean provides the best forecast, the mean accuracy of the ensemble mean forecast is only that represented by the correlation between the ensemble mean and the individual climate realizations. This is because the real world only runs through the 'experiment' once.

Figure 6 shows the nine members of the model ensemble plus the ensemble mean (depicted by the thick line) for the 10 hPa temperature time-series at 90°N for 1984/85. The most striking feature of the time-series is the low spread, i.e. variability, in the ensemble members, especially in the period between mid December and the beginning of March (after the influence of the initial conditions diminishes and before the final warming). During this period there is only one ensemble member which has significant variability.

The results for 1982/83 and 1984/85 suggest that the northern winter stratospheric variability may be influenced by the SSTs (as the only difference between the winters is the SSTs and the initial conditions). Results for other northern winters (see Fig. 1) support the view that SSTs may influence the northern winter stratospheric variability.

The model results for the northern winters which experienced ENSO and La Niña conditions suggest that the response of the stratosphere to the phase of ENSO is not universal. In particular, 1982/83 exhibits a relatively high degree of variability among the ensemble members, whereas 1984/85 and 1997/98 exhibit a relatively low degree of variability. The northern winters of 1986/87, 1991/92 and 1992/93 exhibit a relatively high degree of variability (not shown).

Work by Van Loon and Labitzke (1987) and Labitzke and Van Loon (1989) suggests that the response of the stratosphere to the phase of ENSO is not universal. In particular, they find that 1963/64 and 1982/83 (ENSO years) do not fit the mean pattern of an anomalous weak polar vortex and an intense Aleutian high. In these papers they suggest
that the untypical behaviour is due to the injection into the stratosphere of large amounts of gases and aerosols from the volcanic eruptions of Agung (1963) and El Chichón (1982). As the L58 model used in this paper does not take into account volcanic eruptions, the non-universal nature of the simulated response of the stratosphere to the phase of ENSO cannot be ascribed to the impact of gases and/or aerosols of volcanic origin.

(b) Case study of major warmings: 1988–89

Figure 7 shows the nine members of the model ensemble for the 10 hPa zonal-mean zonal wind time-series at 61.25°N for 1988/89. There are two members which experience a major warming (members 3 and 6); member 3 is depicted by the thick solid line and member 6 by the thick dashed line. There is also one member which just fails to experience a reversal of the zonal-mean zonal wind at 10 hPa (member 4—not highlighted). The most striking feature of the time-series is the behaviour of the members experiencing major warmings: whereas prior to the warmings they behave similarly to the other members, subsequently they behave very differently. This behaviour is also seen at 5 hPa and at 1 hPa.

The behaviour of the 10 hPa 90°N temperature (Fig. 8) for member 3 shows a strong warming of about 70 K associated with the mid February major warming. For member 6 it shows a strong warming of about 40 K associated with the late February major warming. The warming experienced by member 3 occurs over a shorter period than that for member 6. Member 4 (not highlighted) shows a strong warming of about 60 K in early February and is almost a major warming.
Figure 8. North pole temperatures (K) at 10 hPa for the nine members of the ensemble for northern winter 1988/89. Ensemble member 3 is depicted by the thick solid line and ensemble member 6 is depicted by the thick dashed line. See text for further details.

Figure 9. Time series (time v latitude) of the zonal wind (m s\(^{-1}\)) at 10 hPa for northern winter 1988/89 for member 3 of the ensemble.
Prior to the major warmings, members 3 and 6 show almost simultaneous weaker warmings of about 10 K in early February followed by almost simultaneous coolings of about 5 K. Synoptic maps of geopotential height at 10 hPa for member 3 (see later) indicate that the cooling is associated with the development of the wavenumber-2 pattern in the major warming. Synoptic maps of geopotential height at 10 hPa for member 6 (not shown) indicate that the cooling is associated with a slight displacement of the polar vortex toward the pole prior to the main development of the major warming.

Figure 9 shows the time-latitude evolution of the zonal-mean zonal wind at 10 hPa for member 3. The PNJ is located between 60°N and 70°N (which is the model’s climatological location of the PNJ at 10 hPa—see Fig. 2(b)) prior to the wind reversal seen in mid February. For member 6 (not shown), the location of the PNJ is less well-defined and tends to be located very slightly equatorward of the model’s climatological PNJ. For members 3 and 6, just before the wind reversal the PNJ first weakens and then strengthens over a short period of time. This reflects the behaviour of the zonal wind and temperature seen in Figs. 7–8.

Synoptic maps of geopotential height at 10 hPa for member 3 indicate that 10–15 days prior to the peak of the major warming, the polar vortex was not unusually small (it extended to 30°N) but was distorted by the presence of a strong Aleutian high and a weak anticyclone extending from the sub-tropical Atlantic to the Arabian peninsula. The Aleutian high, and an Atlantic anticyclone which had formed in the meantime, then strengthen. At the peak of the warming both anticyclones move toward each other and split the polar vortex asymmetrically, with the Atlantic anticyclone dominating the Aleutian high (Fig. 10). This warming exhibits the characteristics of a ‘wavenumber-2’ type warming (see, e.g. Swinbank et al. 1998). After the split the cyclone over Europe strengthens, and that over the North Pacific decays over a period of 2–3 weeks. The Atlantic anticyclone moves away from the pole toward North America and the Aleutian high decays. By early March, the cyclone over Europe moves poleward and the Atlantic anticyclone moves to the climatological location of the Aleutian high. This warming shares many of the wavenumber-2 characteristics exhibited by the major warming observed in the northern winter of 1984/85 (Fairlie and O’Neill 1988).

Synoptic maps of geopotential height at 100 hPa for member 3 (not shown) indicate that 10–15 days prior to the peak of the major warming, the cyclonic circulation at 100 hPa was strong and centred near the pole, and extended to 30°N. As the warming developed, the cyclonic circulation at 100 hPa split into two cyclones of unequal strength, with the weaker one situated at the location of the climatological East-Asia low. The split at 100 hPa occurs a few days before the polar vortex at 10 hPa splits into two cyclonic circulations. Throughout the period of the ‘wavenumber-2’ type warming, the wavenumber-2 pattern at 10 hPa extends down to 100 hPa, with an eastward tilt with decreasing height. This behaviour is also seen during the major warming observed in the northern winter of 1984/85 (Fairlie and O’Neill 1988).

The synoptic maps of geopotential height at 10 hPa and 100 hPa indicate that the warming experienced by member 3 extends down to the upper troposphere and suggest that the ‘wavenumber-2’ type behaviour at 10 hPa is strongly influenced by the behaviour of the cyclonic circulations at 100 hPa.

The major warming simulated by member 6 (not shown) was mainly of a ‘wavenumber-1’ type (see, e.g. Swinbank et al. 1998) and toward the later stages, exhibited the characteristics of a ‘wavenumber-2’ type warming. The synoptic features at 10 hPa were to some extent reflected at 100 hPa, especially when the warming exhibited ‘wavenumber-2’ type characteristics. In particular, the polar vortex at 100 hPa (initially situated at the climatological location of the East-Asia low) split into two cyclones
Figure 10. Synoptic maps of member 3 of the ensemble for 10 hPa geopotential height (km) for (a) 4 February 1989, (b) 6 February 1989, (c) 8 February 1989, (d) 10 February 1989, (e) 12 February 1989, and (f) 14 February 1989. The contour interval is 100 m.
before the polar vortex at 10 hPa split into two cyclones. As the polar vortex at 10 hPa
split into further cyclones, the polar vortex at 100 hPa became more distorted, but the
synoptic structure at 100 hPa differed from that at 10 hPa. This warming shares some
features with that observed during December 1998 (Manney et al. 1999), in particular
the Aleutian high pushing a weakened and distorted polar vortex off the pole.

The 1988/89 major warmings simulated by the model (especially that by member 6)
share some features with the major warming observed in 1988/89 (Erlebach et al.
1996). Like the observations, the major warming simulated by member 6 occurs in late
February and barely recovers during March, experiences a warming of about 40 K, and
exhibits characteristics of a ‘wavenumber-2’ type warming.

The above results indicate that the conventional criteria for ‘pre-conditioning’ and
subsequent development of a major warming (see, e.g. Labitzke 1981) were not met for
both members 3 and 6. In particular, the PNJ tended to be displaced equatorward rather
than poleward, and although the polar vortex was distorted 10–15 days prior to the peak
of the warmings, the polar vortex was not unusually small, especially for member 3 (see
Fig. 10). Furthermore, prior to the major warmings it is not obvious from the temperature
and zonal wind time-series (Figs. 7 and 8) that these members would experience a major
warming. In particular, member 3 has the strongest winds and coldest temperatures of
all nine members during late January. Results for at least one other northern winter (not
shown) tend to support the general conclusions concerning ‘pre-conditioning’ drawn for
1988/89. Note that a lack of evidence for the conventional criteria for ‘pre-conditioning’
has been reported for observed major warmings (Fairlie and O’Neill 1988; Manney et al.
1999).

(c) Summary of major warmings

Table 1 provides a summary of the major warmings simulated during December–
March in the period 1979/80 to 1997/98. We use the WMO definition of a major warm-
ing. Warming events which do not satisfy this criterion are classed as minor warmings
and excluded. We only consider events which take place between 15 December and 15
March to avoid periods associated with significant influence from the initial conditions
(early December) or the final warming (late March). We bin the warmings into four
periods: 15–31 December, January, February and 1–15 March. Events are binned by the
period in which the zonal-mean zonal wind at 61.25°N and 10 hPa becomes easterly.
We exclude events where, after wind reversal, the zonal-mean zonal wind at 61.25°N
and 10 hPa does not return to a westerly before 31 March. We also exclude events
where, after wind reversal, the zonal-mean zonal wind at 61.25°N and 10 hPa returns to
a westerly before 31 March, but does not do so at either 5 hPa or 1 hPa.

In compiling Table 1 we consider the nine ensemble members for each northern
winter in the period 1979/80 to 1997/98 (note that due to difficulties in the archiving
and subsequent retrieval of data, only eight members could be considered for 1982/83,
1983/84 and 1994/95). Besides the number of simulated major warmings, we indicate
the phase of ENSO and whether there were major warmings and/or an early final
warming in the observations.

Table 1 shows that the distribution of major warmings simulated by the model
is realistic. In particular, major warmings can occur throughout the winter, most of
the major warmings occur during mid and late winter (February and March), and the
polar vortex can recover after a warming simulated in mid winter—features which are
observed in The Met. Office analyses (Swinbank et al. 1998). Table 1 also shows that
the number of ensemble members which exhibit major warmings throughout the period
1979/80 to 1997/98 is small (about 15%). This suggests that the model is not excessively
active in simulating major warmings. Table 1 does not show a clear relation between the phase of ENSO and the frequency and distribution of major warmings (in agreement with the results of section 3(a)).

Table 1 shows that if the model simulates a major warming it is successful (i.e. a major warming or an early warming were observed) in eight out of 14 northern winters, and that if it does not simulate a major warming it is successful in four out of five northern winters. The model is always successful when two ensemble members exhibit a major warming in February or the first half of March. These results indicate that the model success in predicting major warmings is better than 50%, and that its performance is very good when (a) at least two ensemble members exhibit a major warming within a particular period, (b) no major warmings are simulated. The last two results are to be expected. If no ensemble members exhibit a major warming, conditions will not favour major warmings. If two ensemble members exhibit a major warming during a particular period, conditions will slightly favour major warmings.

Prior to 1991, major warmings typically occurred approximately once every two northern winters in the observations. A major warming in December 1998 (Manney et al. 1999) was the first since February 1991 (see, e.g. Pawson and Naujokat 1999); another major warming occurred in February 1999. Table 1 shows that during the 1980s (1979/80 to 1988/89) a major warming is simulated almost every northern winter, and that only in one northern winter did the model not simulate a major warming. The model is very successful in predicting the occurrence of major warmings during the 1980s, being correct in nine out of ten northern winters.

By contrast, during the 1990s (1989/90 to 1997/98; note that the simulations discussed in this paper do not include the 1998/99 winter), very few major warmings occurred in the observations. The model is less successful in predicting the occurrence of major warmings during the 1990s, being correct in four out of nine northern winters.

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Unsuccessful outcomes generally involve the model incorrectly predicting a major warming in a northern winter.

Twice as many major warmings are simulated in the 1980s compared to the 1990s. (Note, however, the clustering of simulated major warmings during the last three northern winters of the 1980s and the first three northern winters of the 1990s: about 50% of the simulated major warmings occur during this period.) This contrast between the 1980s and the 1990s remains unchanged if we relax the definition of a major warming and require that the wind reversal poleward of 60°N only take place at 10 hPa. This contrast suggests that changes in SSTs or the initial conditions or both may be affecting the frequency of simulated major warmings.

As the only difference between the simulated northern winters is the SSTs and the initial conditions, the difference in the frequency of simulated major warmings between the 1980s and the 1990s must be due to the SSTs, the initial conditions or both. Results reported in section 2(b) suggest that the influence of the initial conditions diminishes after 10–15 days in the troposphere and the stratosphere (note, however, that there could be an influence from the initial conditions on the range of flow regimes the model adopts on longer time-scales). This suggests that changes in SSTs could be sufficient to change the frequency of major warmings over the period 1979–98 and contribute toward a downward trend in stratospheric temperatures which, in turn, could lead to a trend toward a more persistent polar vortex in northern spring (see, e.g. Shindell et al. 1998).

There are several caveats concerning the results in Table 1. Firstly, as indicated in section 1, the model is unable to simulate a tropical QBO, and when initialized in a QBO westerly phase it loses the westerly winds at the equator over a period of 2–4 months. The generally realistic behaviour of the model, especially at northern winter extra-tropical latitudes, suggests that for the studies described in this paper this shortcoming is not serious. Secondly, the model typically has a cold bias of 5–10 K in the stratosphere (see Figs. 1, 4; see Swinbank et al. 1998 and Pawson et al. 2000). Although this bias could affect the frequency of major warmings (in particular, it could reduce their frequency), it would tend to affect each winter equally, so comparisons between individual northern winters or a collection of northern winters should be robust.

The results from section 3 indicate that (a) changes in SSTs or initial conditions or both may influence the model simulated stratosphere, (b) that the influence of the initial conditions diminishes after 10–15 days (although there could be an influence on the range of flow regimes the model adopts on longer timescales), (c) the model simulation of individual major warmings is realistic, (d) the model is reasonably successful in predicting the occurrence of major warmings, and (e) the frequency of simulated major warmings in the 1990s is much smaller than in 1980s (note, however, that there is an element of arbitrariness in how one partitions the years between 1979 and 1998). These results (together with those from section 2) establish the model’s variability and realism. We are now in a position to answer the questions posed in this paper by comparing the simulated northern winters in the 1980s and the 1990s.


In this section we answer the questions posed in this paper by using the ensemble approach to compare the simulated northern winters in the 1980s and the 1990s. We partition the northern winters into 1979/80 to 1988/89 (the 1980s) and 1989/90 to 1997/98 (the 1990s). We compare the mean and standard deviation of the ensemble members for the 1980s with their analogues for the 1990s. The mean and standard
deviation are computed in the same way that the model climatology and variability are computed (see section 2). We focus on temperature at latitudes between 30°N and 90°N.

In comparing the temperature for the northern winters in the 1980s and the 1990s, we test the null hypothesis that, given that the means and standard deviations of the 1980s and the 1990s are, respectively, $\mu_1$ and $\mu_2$, and $\sigma_1$ and $\sigma_2$, the means for the 1980s and the 1990s are the same, i.e. $\mu_1 = \mu_2$. The periods chosen have large (i.e. greater than 30) sample sizes of $N_1 = 89$ and $N_2 = 81$ (note that due to difficulties in the archiving and subsequent retrieval of data, only eight members could be used for 1982/83). We assume that the datasets are independent and approximately normally distributed. This null hypothesis addresses the question of whether the model simulates a downward trend in northern winter stratospheric temperatures over the period 1979/80 to 1997/98.

Figure 11(a) shows the difference plot between the temperature means for northern winter in the 1980s and the 1990s at 10 hPa. We shade the regions for which this difference is significant at the 95% confidence level. The four features to note in Fig. 11(a) are: (1) northern winter in the 1990s is cooler than in the 1980s (by 0.5–1.5 K for 30°N–70°N), (2) this difference is generally not significant, (3) the difference is significant for the first few days of the runs (this difference can only come from the different initial conditions, as the time-scale is too short for SSTs to have an influence in the mid stratosphere), and (4) after about one month the difference attains its minimum absolute value, suggesting that by this time the influence of the initial conditions has ceased to be significant (see also section 2(a)). At the 99% confidence level, the difference between the 1980s and in the 1990s at 10 hPa is generally not significant (not shown).

Figure 11(b) is the 50 hPa analogue of Fig. 11(a). The features to note in Fig. 11(b) are as for Fig. 11(a) with the exception that the difference between the 1990s and the 1980s (the former cooler by 0.5–1.5 K for 30°N–70°N) is significant at the 95% confidence level. At the 99% confidence level (not shown), the difference between the 1980s and 1990s at 50 hPa is generally significant but for shorter periods than for the 95% case.

The pattern of differences between northern winter in the 1980s and the 1990s at 70 hPa and 100 hPa, and their significance at the 95% confidence level (not shown), is very similar to that at 50 hPa (Fig. 11(b)). This suggests that the simulated cooling trend in the northern winter lower stratosphere is robust. The results from Figs. 11(a) and (b) suggest that the model simulates a cooling trend of 0.5–1.5 K between the decades at 10 hPa and 50 hPa for 30°N–70°N, and that this trend is significant at 50 hPa but not at 10 hPa.

Comparison of the periods 1979/80 to 1983/84 and 1993/94 to 1997/98 shows that the latter is cooler than the former (by 0.5–1.5 K for 30°N–70°N) at both 10 hPa and 50 hPa (not shown). These results are less robust than those for Figs. 11(a) and (b) (as the model realizations used to evaluate the trend with the five year periods are smaller in number) but still show a cooling trend. The main difference with Figs. 11(a) and (b) is that the 10 hPa trend using the five year periods is largely significant at the 95% confidence level. Together with Figs. 11(a) and (b), these results support the view that the cooling trend at 50 hPa is generally significant whereas that at 10 hPa is not.

Annual mean temperature trends for the period 1979–97 calculated from Stratospheric Sounding Unit/Mesospheric Sounding Unit (SSU/MSU) radiance data indicate a cooling trend of 0.5–1 K decade$^{-1}$ for 30°N–70°N (SORG 1999). In general, over the period between 1979 and 1994, both satellite and radiosonde data indicate a global and annual mean cooling of the lower stratosphere of 0.6 K decade$^{-1}$, with the cooling
Figure 11. (a) Latitude-time cross-section of the difference between the climatology for northern winter in the 1990s (1989/90 to 1997/98) and northern winter in the 1980s (1979/80 to 1988/89) at 10 hPa, temperature (K; contour interval 0.5 K). Negative values indicate that northern winter in the 1990s was cooler than in the 1980s. The shaded area indicates regions where this difference is significant at the 95% level. See text for further details.

(b) As Fig. 11(a) but at 50 hPa.
being 0.75 K decade⁻¹ for 30°N–60°N (SORG 1999). Comparison of the results from Figs. 11(a) and (b) with these observations suggest that the answer to question (1) is that the UM reproduces the observed temperature trends with only SST variations.

The only difference between the northern winters simulated by the model is the SSTs and the initial conditions. Results presented in this paper suggest that beyond 10–15 days into the runs the influence of the initial conditions diminishes (see section 2(a) and Figs. 11(a) and (b)). (Note that, however, there could be an influence from the initial conditions on the range of flow regimes the model adopts on longer time-scales.) Thus, the results in this paper provide a first step toward answering question (2) by suggesting that SST variations alone may explain the cooling trend at 50 hPa. One might ask whether the observed temperature trend at 10 hPa must be forced by ozone or greenhouse gas changes if it is not forced by SST changes. Because the model results show that there is no significant trend outside of natural variability, one cannot rule out natural variability as an explanation. A complete answer to question (2) requires the inclusion of all likely influences (in particular ozone and water vapour trends) to establish that the model is capable of reproducing the observed trends under realistic conditions, and is beyond the scope of this paper.

The above results suggest that changes in SSTs could be sufficient to explain cooling trends in the lower stratosphere. They thus help answer the question of whether there is a clear trend in the persistence of the polar vortex into northern spring, and which has been raised by the studies of, e.g. Zurek et al. (1996), Pawson and Naujokat (1997), Coy et al. (1997) and Shindell et al. (1998).

Hansen et al. (1997), using the ensemble approach with a nine-level model with a lid at 10 hPa, obtain a cooling of about 2 K in the mid and lower stratosphere for SST forcings alone (this cooling is reflected in the change of the global-mean annual temperature for 1979–1995). When Hansen et al. include other forcings (stratospheric aerosols, greenhouse gases, ozone and solar irradiance) they get a cooling of about 5 K in the mid and lower stratosphere (this is the cooling reflected in the change of the global-mean annual temperature for 1979–95). Hansen et al. (1997) typically perform ensembles of multi-year runs, whereas the experiments in this paper are nine member ensembles of four month runs for each of the 19 northern winters in the period 1979/80 to 1997/98. The qualitative agreement between the results in this paper and those of Hansen et al. suggests that the former are robust.

Concerning the significant cooling trend in the northern winter lower stratosphere (50 hPa) simulated by the model, a question is whether SST changes would cause this cooling through changes in the circulation, through in situ changes in the water vapour distribution (see, e.g. Forster and Shine 1999), or by a combination of both processes. As no water vapour trends have been imposed in the model simulations, the above results suggest that the main contribution to the cooling would come from the changed circulation as a result of SST changes.

5. Conclusions and Further Work

In this paper we used the ensemble approach to study the simulated northern winter stratospheric variability in a troposphere-stratosphere version of the UM. The runs were for the December–March periods from 1979/80 to 1997/98 and the ensemble for each winter had nine members. We used observed SSTs, a fixed ozone climatology and fixed greenhouse gases. The UM was compared against the extended ERA dataset and The Met. Office stratospheric analyses. Although the model had a cold bias with respect
to both set of analyses, it tended to behave realistically (see sections 1, 2). Important elements of the simulated stratospheric variability such as major warmings were also generally realistic (see sections 3(b), 3(c)). This paper is concerned with the following questions:

(1) Does the UM reproduce observed temperature trends with only SST variations?
(2) Can any significant trend be attributed to SST variations?

The results of this paper suggest that the answer to question (1) is that the UM reproduces the observed temperature trend with only SST variations. In particular, northern winter in the simulated 1990s is colder than northern winter in the simulated 1980s at both 10 hPa and 50 hPa, with the model cooling trend (0.5–1.5 K between the decades for 30°N–70°N) comparable to that from observations (0.5–1 K decade⁻¹ for 30°N–70°N). The temperature differences between the simulated 1990s and 1980s generally are not significant at 10 hPa, but are generally significant at 50 hPa at the 95% significance level.

The results in this paper provide a first step toward answering question (2) by suggesting that SST variations alone may explain the cooling trend at 50 hPa. Because the model results show that there is no significant cooling trend at 10 hPa outside of natural variability, one cannot rule this out as an explanation. A complete answer to question (2) requires the inclusion of all likely influences (in particular ozone and water vapour trends) and is beyond the scope of this paper.

A number of assumptions are made in providing these answers. Firstly, that differences between individual northern winters are mainly due to differences in the SST forcing. Results presented in this paper (see sections 2(a) and 4) suggest that the influence of the initial conditions in the troposphere and stratosphere diminishes after 10–15 days and support this assumption. (Note that, however, there could be an influence from the initial conditions on the range of flow regimes the model adopts on longer time-scales.) Secondly, that the ensemble datasets for each period are independent and approximately normally distributed. The behaviour of the ensemble members (see, e.g. section 2(b)) supports this assumption.

There are two caveats concerning the answers to questions (1) and (2). Firstly, the arbitrary partition of the dataset between the 1980s and 1990s. This is done for convenience, but as shown in section 3(c), 50% of the simulated major warmings cluster during the last three northern winters of the 1980s and the first three northern winters of the 1990s. Alternative partitions of the 1980s and 1990s (see section 4) support the view that the cooling trend at 50 hPa is significant, whereas at 10 hPa it is not. Secondly, the experiments in this paper are nine-member ensembles of four month runs for the 19 northern winters in the period 1979/80 to 1997/98, as opposed to a multi-year ensemble integration. Note, however, that the qualitative agreement between these results and those of Hansen et al. (1997) (who typically used multi-year ensembles) suggests that the results described in this paper are robust.

Besides the above, there are other features worth noting: (1) The northern winter stratosphere is generally not very predictable on a day-to-day basis (see section 2(a)). This confirms what is already known. (2) SSTs appear to influence the northern winter stratosphere. For any individual northern winter there can be considerable variability between the ensemble members, but its degree can differ significantly between winters experiencing very different SST forcings (see section 3(a)). (3) The evidence to support the conventional criteria for ‘pre-conditioning’ and subsequent development of a major warming is not always met (see section 3(b)).
The results of this paper indicate the power of the ensemble approach and suggest it is the way forward to study stratospheric issues such as variability, simulation and attribution of past trends, and prediction of future trends. Although the application of this approach to the stratosphere is relatively novel, there is evidence that, with the increased realization of the nonlinear nature of the stratosphere and the increased availability of powerful computing facilities, it will become the standard in the near future.

This study is only the initial phase of detailed investigations of stratospheric variability which will involve the UM. Further work will investigate the role that ozone changes (as well as other forcings) may play in stratospheric temperature trends.

ACKNOWLEDGEMENTS

The runs described in this paper were carried out at the Hadley Centre, The Met. Office, UK (thanks to M. Harrison, R. Graham and K. Robertson). Thanks to the associate editor and two reviewers for their comments which helped to improve the paper. Thanks to the following people for helping to transfer and manipulate data: R. Brugge and J. Cole. Thanks to the following people for useful discussions: M. Ambaum, M. Blackburn, A. O’Neill, D. B. Stephenson and M. Baldwin. Thanks to A. Dethof for helping to improve some of the figures.

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