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STRATOSPHERIC PROCESSES AND THEIR ROLE IN CLIMATE
A Project of the World Climate Research Programme

# The 26th Session of the Joint Steering Committee of the WCRP

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The 26th session of the Joint Steering Committee (JSC) was held at the Escuela Superior Politecnica del Litoral (ESPOL) in Guayaquil, Ecuador. A. O'Neill and N. McFarlane attended on behalf of SPARC. Although an important aspect of the JSC sessions involves reviewing progress toward achieving WCRP objectives, particularly within the core projects (CliC, CLIVAR, GEWEX, and SPARC), these sessions also provide an ideal opportunity to examine new approaches and ideas. An important part of this year's meeting involved consideration of the report of the Task Force that was set up at the 25th session to further development of the WCRP strategic framework entitled Coordinated Observation and Prediction of the Earth System (COPES). A Modelling Panel and a Working Group on Observation and Assimilation of the Climate System were also set up to support the COPES and facilitate coordination of modelling and observational activities across the WCRP.

A number of special topics were also placed on the agenda for discussion at the JSC session this year. The topic of Atmospheric Chemistry and Climate was among these. SPARC, in collaboration with IGAC, has developed a high level of expertise and experience in this area and consequently was asked to lead the discussion. To facilitate discussion, a short report entitled "Chemistry-Climate Interactions" was submitted by the SPARC Co-Chairs for consideration of the JSC prior to the session. In his presentation on this topic A. O'Neill emphasized the importance and breadth of atmospheric chemistry and climate interactions as a WCRP issue, noting among other things that (a) the link between air pollution and climate is a key issue for society, (b) chemical processes affecting atmospheric composition are, in general, coupled and nonlinear. He then posed the following questions for consideration by the JSC in regard to Atmospheric Chemistry and Climate (AC&C):

- **What are the concrete objectives: chemical weather and climate prediction and scientific underpinning within the seamless prediction of COPES?**
- **Besides SPARC, what will other WCRP Projects do?**
- **Should SPARC continue to take the lead? What about tropospheric chemical modelling?**
- **How is the link with IGAC working? IGBP view?**
- **Is progress fast enough?**
- **What should happen now?**

The JSC thanked A. O'Neill for his presentation on AC&C. In reply, it reaffirmed the importance of AC&C issues to WCRP's objectives and stressed the need for devel-

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oping a roadmap for chemistry-climate models, observations and process studies. For this purpose, the JSC proposed the establishment of a Joint WCRP-IGBP Task Force (TF) involving WCRP core projects and working groups and IGBP (IGAC), led by SPARC and IGAC as the core-organizers. IPCC and possibly IHDP should be kept informed. The immediate task for the TF is to organize a workshop; outcomes of this workshop should be the terms of reference and suggestions for the way forward.

In his presentation to the JSC on SPARC, A. O'Neill summarized developments within the last year, noting that the Third SPARC General Assembly was very successful and that there was a smooth transition of the SPARC Office from Paris to Toronto. The main part of the presentation was devoted to highlighting a number of scientific and technical issues of concern to SPARC and putting before the JSC a number of questions pertaining to them. The first two of these questions relate to the SPARC theme of chemistry-climate interactions and to the earlier presentation and discussion on atmospheric chemistry and climate. They are also motivated in part by the current uncertainty in predictions of the Antarctic ozone minimum by chemistry-climate models (e.g. as reported in the paper by Austin *et al.*, 2003):

- **What is SPARC's role in the chemistry-climate initiative?**
- **Should SPARC's Chemistry-Climate Modelling Validation project lead to another AMIP-like experiment, and how should it be facilitated?**

The JSC affirmed that SPARC should play a leading role in the Joint WCRP-IGBP Task Force. It encouraged SPARC to open discussions with PCMDI to determine if it could help in facilitating an AMIP like experiment for GCMs with well resolved stratospheres and relatively comprehensive treatments of chemistry.

Validation of models places heavy reliance on global observational data sets and/or analyses. Changes in observing systems may give rise to spurious apparent changes in physical variables such as temperature, prompting the following question:

- **What needs to be done at a "high level" to ensure that the temperature record derived from satellites is cross-calibrated between satellites?**

The JSC noted that the importance of this calibration issue has been recognised by GCOS and the former WCRP satellite working group, leading to the "reprocessing project" now proposed under WOAP.

Analysis and modelling of patterns of climate change and variability are cross-cutting issues with the WCRP and are being examined within several of the core projects. These are also key issues within the COPES framework, and therefore prompt a question on how to foster desirable collaborative activities:

- **How should SPARC work with CLIVAR on "modes of atmospheric variability" and how they will change in a changing climate (and thereby respond to a COPES priority)?**

The JSC strongly encouraged joint SPARC and CLIVAR activities on "modes of atmospheric variability" and their change in a changing climate and was pleased that this topic is included in joint CLIVAR/SPARC session at the upcoming AMS meeting in June. The ideas exchanged there could be used to plan a joint SPARC/CLIVAR Workshop on Stratosphere-Troposphere coupling and modes of variability for early 2006, the scope also to be guided by the discussions at the SPARC SSG meeting.

A number of papers and posters were presented at the Third SPARC General Assembly dealing with key processes in the tropopause transition layer (TTL) using combinations of observational and modelling approaches. Use of cloud resolving models (CRM) is proving to be an innovative and potentially powerful approach to studying key processes within the TTL. Extensive use and analysis of the performance of CRMs for tropospheric applications has been carried out in the context of the GEWEX Cloud System Study and this suggests another avenue for fruitful collaborations within the WCRP:

- **Should SPARC partner with GEWEX on a new initiative to exploit cloud resolving models to understand processes in the tropopause transition layer?**

The JSC encouraged partnering of SPARC and GEWEX to develop a new initiative to exploit cloud resolving models to understand processes in the tropopause transition layer. The Pan-GCSS workshop in Athens provides an opportunity to begin development of this initiative.

An outcome of the SPARC Aerosol Assessment Project (ASAP) has been the development of a number of valuable data sets. The potential importance (and uncertainty) of the radiative effects of aerosols is widely known. However, the sensitivity of radiative transfer codes to the treatment of aerosols remains a potentially important source of uncertainty suggesting the question:

- **Should SPARC partner with GEWEX to initiate an intercomparison of how aerosols are treated in radiative transfer codes (with IPCC in mind)?**

The JSC also encouraged partnering of SPARC and GEWEX on issues pertaining to the radiative effects of aerosols. A merger of aerosol data sets is needed and a comparison of the treatment of aerosols in radiative transfer codes should be carried out.

Evaluating the influences of solar variability on climate is important for understanding climate change. A goal of the joint SPARC-CAWSES activity on this topic is to elucidate the effects of solar variability on atmospheric composition, for example on ozone. However the effect of solar variability on the radiation budget of the whole atmosphere and the surface and its impact on the oceans and cryosphere is of broader interest and has been regarded as an important factor in modelling and evaluation of climate change. The effects of changes in atmospheric composition are not addressed in most GCMs and this may be an important shortcoming. The question for the JSC prompted by this issue is:

- **What should SPARC do and how should we work with WGCM (with IPCC in mind)?**

In regard to solar forcing WGCM has focused on the radiative effects of long term variations in solar forcing at the top of the atmosphere (TOA) in the absence of directly attributable changes in atmospheric composition.

The JSC recommended that SPARC should work with WGCM in updating TOA solar forcing data while continuing to pursue its current interest and activities in regard to the effects of solar forcing variations on atmospheric composition.

Data Assimilation issues are important for SPARC for several reasons. There is now a wealth of stratospheric chemical data available from satellites. Stratospheric analyses are now being produced by major operational weather prediction centres and these have been used in off line chemical transport models. Some of the presentations given in the SPARC General Assembly highlighted encouraging developments as well as challenges associated with issues of bias, noise, and inaccuracies inherent in transport algorithms. The questions posed to the JSC in this regard were the following:

- **Should the SPARC Working Group on Data Assimilation (WGDA) combine with activity in the Working Group on Numerical Experimentation (WGNE)?**
- **What can the SPARC WGDA do for the WCRP Observations and Assimilation Panel (WOAP)?**

The issue of the cooperation between the SPARC WGDA, WGNE, and WOAP was left for later consideration. A variety of further actions need to be considered. Both of the SPARC presentations were very well received by the JSC. The consensus was that they were very timely in highlighting key issues for the WCRP as a whole and posing clear questions for discussion.

### Reference:

Austin, J. *et al.* Uncertainties and assessments of chemistry-climate models of the stratosphere. *Atmos. Chem. Phys.*, 3(1), 1-27, 2003.

# SPARC and the International Polar Year (IPY) 2007–2008

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## What is IPY?

Nearly 150 years ago, 13 nations joined forces for the first internationally co-ordinated programme of scientific exploration in the polar regions during the International Polar Year 1882–1883. Beyond the advances in science and geographical exploration, a principal legacy of the first IPY was that it set a precedent for international science co-operation. In 1904 the first permanent Antarctic station was established by the government of Argentina at Orcadas/South Orkneys to carry out research (including weather monitoring) and other activities in the region. This base still continues its scientific activities.



Following the successful IPY, the International Meteorological Organization promoted the second IPY in 1932–1933 to investigate the global implications of the newly discovered jet stream. Some 40 nations participated in the second IPY, which heralded advances in meteorology, atmospheric sciences, geomagnetism, and the "mapping" of ionospheric phenomena. In the years following World War II, many nations increased the number of permanent and semi-permanent stations and bases in both polar regions, thus providing many of the long term weather datasets that are crucial to our understanding of polar weather and climate processes and their relationship with the rest of the world.

Scientists again decided that an international science year was warranted to utilize the new technologies of the era. This time, the scope of the effort was global, and 67 nations participated in the International Geophysical Year (IGY) in 1957–1958. The IGY led to an increased level of research in many disciplines, and the scientific, institutional, and political legacies of the IGY endured for decades, in many cases to the present day. IGY helped consolidate the engagement of many countries in sustained Antarctic research.

Today, nations around the world are planning for a new International Polar Year in 2007–2008. This IPY will be far more than an anniversary celebration of the IGY or previous IPYs; it will be a watershed event and will use today's powerful research tools to better understand the polar regions. Automatic observatories, satellite-based remote sensing, autonomous vehicles, the Internet, and genomics are just a few of the innovative approaches for studying previously inaccessible realms. IPY 2007–2008 will be fundamentally broader than the IGY and past IPYs because it will explicitly incorporate multidisciplinary and interdisciplinary studies, including biological, ecological, and social science elements. Such a programme will not only add to our scientific understanding, but will also assemble a world community of participants with shared ownership in the results.

IPY 2007–2008 will provide a framework and impetus to undertake projects that normally could not be achieved by any single nation. It will allow us to think beyond traditional borders — whether national borders or disciplinary constraints — toward a new level of integrated, cooperative science. A coordinated international approach maximizes both impact and cost effectiveness, and the international collaborations begun today will build relationships and understanding that will bring long-term benefits. Within this context, IPY 2007–2008 will seek to galvanize new and innovative observations and research while at the same time building on and enhancing existing relevant initiatives, many of which have been subject to dwindling budgets and cancellations in recent years despite their relevance for long term monitoring. In addition, there is clearly an opportunity to organize an exciting range of educational and outreach activities designed to excite and engage the public, with a presence in classrooms around the world and in the media in varied and innovative formats.

## IPY 2007–2008 Organization

IPY 2007–2008 is organized around six themes. Its broad goals are:

- To determine the present environmental status of the polar regions by quantifying their spatial and temporal variability;
- To quantify, and understand, past and present environmental and human change in the polar regions in order to improve predictions;
- To advance our understanding of polar–global interactions by studying teleconnections on all scales;
- To investigate the unknowns at the frontiers of science in the polar regions;
- To use the unique vantage point of the polar regions to develop and enhance observatories studying the Earth's inner core, the Earth's magnetic field, geospace, the Sun and beyond; and
- To investigate the cultural, historical, and social processes that shape the resilience and sustainability of circumpolar human societies, and to identify their unique contributions to global cultural diversity and citizenship.

The IPY International Programme Office is hosted by the British Antarctic Survey in Cambridge, UK. The office provides planning, coordination, and guidance to 25 international organizations that support IPY, as well as 28 national IPY committees. In January 2005 the office received over 900 "Expressions of Intent" from the international research community, including one EoI from SPARC. SPARC-IPY was recognized by the IPY Joint Committee as having the "potential to make a major contribution to the IPY," and is expected to "clearly contribute to significant international collaboration." SPARC was invited to submit a full proposal, due in September 2005. While the IPY office will coordinate the research efforts, it does not fund the research. Funding is left to national and international funding agencies.

Although IPY 2007–2008 is oriented toward the polar surface environment, it also emphasizes connections to other regions as well as the solid Earth below and the atmosphere above.

## Connection Between Polar Climate and the Stratosphere

There is a strong dynamical connection between the circulation of the high-latitude stratosphere, and surface weather and climate. In particular, stratospheric wind anomalies tend to progress downward to the lowermost stratosphere (near 10 km), and then induce changes to the Arctic Oscillation (AO) pattern, which is similar to the NAO (North Atlantic Oscillation). Our understanding of the mechanisms is advancing, but it is still incomplete. Over the Arctic, the phase of the AO affects surface winds, temperature, sea-ice motion, and ice extent. There appears to be a memory in summer sea-ice of the previous winter's AO, and part of the observed thinning of sea-ice can be attributed to long-term changes in the AO.

In the Southern Hemisphere, observations and models show that the springtime stratospheric ozone hole has not only been linked to cooling of the lower stratosphere and strengthening of the circumpolar winds within the stratospheric polar vortex, but that these changes have induced surface circulation and temperature changes over Antarctica — changes lasting well into the summer. During spring and summer, the lower stratosphere at southern mid-latitudes has in turn undergone changes linked to the changes over Antarctica. Although chlorofluorocarbons (CFCs, now banned by international agreement) are largely responsible for current ozone depletion, increasing concentrations of greenhouse gases like carbon dioxide and methane may delay future ozone recovery.

As greenhouse gases increase, the circulation of the stratosphere will likely be affected. But we are not able to predict whether high-latitude stratospheric winds will become stronger or weaker a few decades from now. Such a trend — positive or negative — is expected to affect the AO at the Earth's surface. On climate-change timescales, stratospheric effects are potentially large, and understanding how the stratosphere will change, as well as how the stratosphere and troposphere are coupled, will contribute to reducing that uncertainty.

Trends and variability in the AO — including changes in the stratospheric circulation — would affect the lifetimes of natural greenhouse gases such as methane and $N_2O$, as well as of anthropogenic greenhouse gases such as CFCs or their replacements, since the removal of all of these gases requires, at least in part, transport through the stratosphere. Changes in lifetimes of greenhouse gases could themselves result in a forcing of the stratospheric circulation. It is uncertain to what degree ozone would be affected, since ozone acts not only as an ultraviolet filter but in a multifaceted manner as a greenhouse gas. Global, and in particular, polar ozone concentrations might respond sensitively to the circulation changes as well as to changes in ozone destroying trace gases resulting from the degradation of CFCs, $N_2O$ and methane.

Trends and variability in the AO are further reflected in ecosystem changes, and feedback from ecosystem changes could be manifested in the stratosphere as well as the troposphere. Here one research topic would be to study whether changes in permafrost and wetlands are consistent with AO signals, investigating the effects of temperature and precipitation changes on methane fluxes and then estimating how these fluxes will change with time based on AO trends, including the possibility of enhanced production of water vapour in the stratosphere due to methane oxidation.

This question of changes in stratospheric water vapour by any mechanism is extremely important because it has been posited that increased stratospheric water vapour due to either large methane releases or decreased latitudinal temperature gradients may have led to an increase in the frequency and thickness of polar stratospheric clouds and drastically changed the radiative balance in the Arctic during the Eocene (55–40 million years ago), with large feedbacks on high-latitude temperatures. Since both of these mechanisms for increasing stratospheric water vapour may be present in the climate of the near-future, it is essential to try to determine the magnitude of the changes that may take place.

Weather in southern-most countries like Australia, New Zealand, Chile, and Argentina is strongly influenced by Antarctic atmospheric processes. Even South Africa and Brazil occasionally suffer cold spells of Antarctic origin in winter. The future evolution of climate in those countries and over the southern oceans is thus strongly linked to the changes that will take place in the Antarctic troposphere and stratosphere. Some of the richest fisheries and marine ecosystems in the world could suffer major impacts from changes in Antarctic climate, affecting the current state of biogeochemical cycles, marine food chains and fishing activities.

## SPARC's Contribution to IPY

Because IPY will occur over a short time period, SPARC will focus on details of the polar stratosphere in a programme called "The structure and evolution of the stratospheric polar vortices during IPY and its links to the troposphere."

The Antarctic ozone hole is one of the most recognized environmental issues of the $20^{th}$ century. Ozone changes in the Arctic, though lesser in magnitude, are equally important. Recent research has shown that the evolution of stratospheric ozone is tightly coupled to a wide range of processes acting within and outside the winter polar vortices. Much of this understanding has been achieved within the SPARC programme. The IPY programme offers a unique opportunity for SPARC to assemble a range of scientific expertise to study the Antarctic and Arctic Polar Vortices, the loci of key processes associated with ozone depletion and its eventual recovery, as well as contribute towards a better understanding of the coupling mechanisms between the troposphere and the stratosphere.

SPARC-IPY will co-ordinate the activities of the international SPARC community in relation to IPY. This co-ordination will be directed toward both satellite and ground-based experimental campaigns, as well as specific initiatives promoted by SPARC to increase understanding of the polar atmosphere. The services of the SPARC Data Center will be made available to facilitate acquisition and archiving of key data that will be used for projects or generated by them during the IPY period.

In addition to coordinating and facilitating IPY projects within the SPARC community, SPARC-IPY will promote specific initiatives directed toward the understanding of major features and processes in the polar middle atmosphere during the IPY period. These initiatives will include a range of research activities involving modelling, observations, and analysis, and will include workshops and meetings as needed or desirable to facilitate research and dissemination of results. These efforts will be carried out in context of the SPARC Project core thematic programmes of Stratospheric Chemistry and Climate, Stratosphere-Troposphere Coupling, and Detection, Attribution, and Prediction of Stratospheric Changes.

The dynamics, transport and chemistry of the polar vortices, as well as of properties relevant to microphysical processes, such as the formation of polar stratospheric clouds, will be documented as completely as possible. To achieve this detailed picture, SPARC-IPY will bring together available research and operational satellite data, as well as ground-based and aircraft data. SPARC-IPY will promote co-ordinated field campaigns to enhance the database, and encourage work on data assimilation and intercomparison of assimilated datasets to yield a unique synthesis of data on the polar vortices. Weather services carrying out routine radiosonde and ozonesonde measurements would be encouraged to increase the frequency of the observations and to store the data with full resolution.

Observations of a range of variables within the stratospheric polar vortex will be used, together with data assimilation, models and other analysis techniques to create a coherent and comprehensive picture of the current state of the stratosphere in the Arctic and Antarctic, and to elucidate further the interaction of the polar stratosphere with the underlying troposphere.

The project will involve the multi-national SPARC community and its affiliates, e.g. other WCRP research projects. SPARC-IPY will co-ordinate relevant field campaigns supported by national and international research programmes, and will seek, where possible, to promote additional campaigns where there are data gaps that need to be filled. As a core project within the WCRP, the SPARC Project relies on co-ordinated activities by agencies and groups within the affiliated community for the resources needed to carry out observational campaigns and research within its thematic programmes. SPARC-IPY will function within this general SPARC framework of co-ordination and collaboration with related and linked national and institutional IPY projects.

# Report on GRIPS March 14-17, Toronto, Canada

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## Introduction

The tenth and final workshop for the GCM-Reality Intercomparison Project for SPARC (GRIPS) was held in Toronto. For about a decade, GRIPS has been the modelling focus for SPARC, and has had the role of evaluating and comparing different dynamical models of the stratosphere from the international community. A number of tasks have been defined, ranging from basic evaluation and validation of GCMs, through studies that aimed to understand the realism and limitations of processes in the models, to longer model simulations designed to examine the consistency of how models respond to changes in forcing. Following the tradition of past GRIPS workshops, the programme included elements related to the formal tasks, alongside other presentations, ranging from developments in individual models to broader scientific questions of relevance to the community.



In order to meet the evolving demands on SPARC, the modelling strategy has been revised, to the extent that the Chemistry-Climate Model Validation (CCMVal) project will become the main focus. CCMVal has a broader mandate than GRIPS, with the intention of addressing issues of relevance to the topics of all three of SPARC's major themes (Stratospheric Chemistry and Climate, Stratosphere-Troposphere Coupling, and Detection and Attribution of Stratospheric Change). Many of the issues are directly related to some of the GRIPS tasks in explicit relationships (*e.g.*, model climate and the factors that influence it), and indirect relationships (*e.g.*, impact of parameterizations on climate and climate change). However, in order to answer these questions, SPARC needs to have a much broader scope for model evaluation and validation with formal tasks on details of chemistry and radiation. CCMVal will encompass detailed analyses of the chemistry (*e.g.*, photolysis) and transport, as well as the dynamics and radiation. To this end, CCMVal will likely involve detailed study of the radiative, chemical and dynamical aspects of current models, not necessarily restricting attention to GCMs.

Because of the relevance of many GRIPS activities to CCMVal, a main aspect of the 2005 Workshop was on how any unfinished tasks may be carried forward. It should be stressed that, even though no formal GRIPS tasks involving chemistry-climate models have been defined, much discussion at this and previous workshops has been devoted to such models. As these models reach a level of maturity that justifies cross-comparison, it is natural that SPARC's modelling activities should look for suitable tests of these models, just as GRIPS has explored the performance of the dynamical-radiative components of the full CCMs.

WCRP has led numerous modelling efforts with the Working Group on Coupled Models (WGCM) and the Working Group on Numerical Experimentation (WGNE). Projects such as AMIP, CMIP among others have fulfilled the needs of the community, with substantial investment in infrastructure. WCRP's objective is to unify these efforts into a common framework under the Coordinated Observation and Prediction of the Earth System (COPES), with two panels that will report directly to the Joint Scientific Committee (JSC): the WCRP Modelling Panel (WMP) and the WCRP Observations and Assimilation Panel (WOAP). The COPES WMP will consist of representatives from all WCRP programmes (S. Pawson will represent SPARC) and from all WCRP working groups.

## Model Developments

Several presentations discussed recent developments in climate models. Discussion of these model developments enabled speakers and audience to share common experiences with the various models.

**J. Scinocca** from CCCma and **B. Morel** from LMDz gave presentations on the ongoing development and application of global coupled climate models aimed at understanding climate change and variability. The focus at CCCma is primarily on improved representation of physical processes, but a major undertaking over the past year has been the set-up and execution of a large number of scenario runs for use in the upcoming IPCC Fourth Assessment Report. The CCCma global model is used operationally to produce seasonal forecasts, and because the model computes 'weather' at 20-minute time steps, one can use the model to say something about extreme events and their changing probabilities. The LMDz-Reprobus CCM has been evaluated with a 10-year climatology and compares well with observations. Sensitivity studies for the orographic gravity wave forcing have been performed, and imply that the introduction of the stratosphere has led to an increase in the surface AO persistence and predictability.

**A. Bushell** presented the efforts in extending the Met Office HadGAM1 'New Dynamics' Climate Model. The model has changed from non-hydrostatic to quasi-hydrostatic, and to a hybrid sigma-pressure grid, and includes new dynamics (*e.g.* mass flux convection, statistical cloud scheme, prognostic ice microphysics, non-local boundary layer), extra middle atmosphere physics such as methane oxidation (photolysis of water vapour at higher levels), spectral gravity wave parameterization (new dissipation and launch spectra, and a transparent upper boundary, hydrostatic non-rotating dispersion relation), and new added input data. The group is also looking to incorporate satellite data assimilation, increasing predictability at seasonal timescales and improving process representation, such as the QBO.

**R. Stolarski** and **M. Gupta** reported on results of the Goddard stratospheric chemical transport model (CTM) and on updating the NASA Goddard Coupled Chemistry-Climate model (GEOS-GCM). The credible performance of the CTM driven by GEOS-GCM meteorology in simulating the age of air, the life-cycle of the polar vortex and the observed ozone trends is the basis for development of a coupled model. The CTM transports source molecules, (*e.g.* CFCs, Halons, methyl bromide, methane, nitrous oxide) with specified surface mixing ratios, and chemical families (*e.g.* NOx, ClOx).

Additionally, GEOS-GCM simulations using different scenarios of SST and CTM produced ozone distributions that have shown reasonable behaviour. Chemistry and radiation in GEOS-CCM are presently coupled only through stratospheric and mesospheric ozone. Future plans include invoking the radiative coupling of chemically modified water vapour between 380K surface and top of the model domain, extensive evaluation of the model with UARS and AURA observations, introducing a combined tropospheric-stratospheric-mesospheric photochemical mechanism, and coupling with an ocean-ice model.

## Stratospheric Forecasting

**G. Roff** presented the study from WGNE on stratospheric prediction. Better skill in the stratosphere is expected since the dynamics is dominated by a quasi-stationary polar vortex, unlike the troposphere, which is influenced by transient and synoptic scale waves. Therefore, it is to better test our skill in predicting "stratospheric weather" when the polar vortex is undergoing strong changes over a short period of time, such as sudden warmings. The results show that stratospheric forecasting in the Northern Hemisphere (NH) and Southern Hemisphere (SH) shows similar characteristics. The stratospheric forecasting performance at six days is comparable to three days in the troposphere, but there exists large variability in the forecast skill at six days, often depending on how active the planetary waves are and thus how quickly the vortex distorts. Skill is increased by increasing stratospheric vertical resolution and by raising the lid.

**T. Hirooka** showed results from a study examining the predictability of Stratospheric Sudden Warmings (SSWs) in the NH inferred from ensemble forecast data. Two case studies (Mukougawa and Hirooka (2004, Mon. Wea. Rev.), Mukougawa *et al.* (2005, GRL, submitted)) found predictability times of two weeks to one month for SSW events in December of 1998 and 2001. It was found that different ensemble members showed a high sensitivity of the prediction skill to the initial conditions. This case study looks at SSWs in 2002/03 and 2003/04. It was found that the predictability of SSWs in these cases was approximately two to three weeks, and that the predictability is dependent on occurrences of SSW events.

While stratospheric forecasting has been peripheral to the main aims of GRIPS, several presentations have addressed it at various workshops. This issue has been addressed by the WGNE group, but with the reorganization of SPARC's modelling activities and the attempts by the WCRP to unify their modelling work through COPES, the topic remains relevant to SPARC. Discussions of stratospheric forecasting were thus based on this premise. It was noted that various groups are now using the same models for analysis/forecasting as for climate studies. This means that study of model performance when constrained by atmospheric data will likely emerge as an important diagnostic for the climate models. Such activities will likely be coordinated through SPARC's data assimilation group, in conjunction with CCMVal and the WCRP-COPES panels.

## Polar Vortices, Warmings and Annular Modes

**G. Roff** presented the results from the comparison of polar vortices between models and analyses (GRIPS task 1i). The NMC dataset indicates that the typical characteristics for the SH polar vortex are; a rapid and deep onset throughout the depth of the atmosphere, a high correlation between its size and the maximum wind speed, it is generally polar centred and symmetric, and during its demise there is gradual decay from aloft. All the models tend to capture the main features of the vortex, but the vertical extent is greatly affected by model characteristics (*e.g.* those models with a sponge layer are forced to close the polar vortex near the top levels, and those with no gravity wave parameterization tend to have very strong, deep vortices). Elliptical diagnostics show that models that do not extend high enough have polar vortices that are artificially curtailed aloft.

**L. Polvani** presented a new climatology of SSWs. From observations we can determine their frequency, amplitude, type and distribution, and from models we can learn about their dynamical behaviour. Using the WMO definition (easterly winds at 60N and 10hPa), SSWs can be classified into two main groups according to their evolution: vortex displacement and vortex splitting. In the NCAR/NCEP and ERA-40 analyses, it was found that about three SSW events occurred every four years, and that vortex displacements accounted for two thirds of these events. These numbers should serve as benchmarks for models. Other results are that the largest variability is in January-March, splitting events are concentrated in January and February, and there is little evidence of trends in the number of warmings. He also noted that the Baldwin and Dunkerton picture of downward propagation of the AO signal might be misleading in that the weak vortex events are actually SSWs, and the strong vortex events are really non-events.

Aspects of the circulation that have received much attention are the annular modes in the two hemispheres. In a study using the MRI Chemical GCM, **Y. Kuroda** examined the impacts of solar variability on the annular mode in the SH and compared the model response to observations. Regarding the Northern Annular Mode (NAM), an important question remains: what drives the daily variability of the NAM? There is no 'NAM tendency' equation. **A. Haklander** presented an analysis of mid-latitude stratospheric wind variations, using the zonal-mean momentum equation, to investigate changes in normalized cross-covariances of eddy forcing terms with the zonal mean wind tendency. Using the Transformed Eulerian Mean formulation, the resulting daily absolute angular momentum changes yield both a change in relative and planetary angular momentum, each of similar magnitude. There is a downward propagation of low-frequency variations of the wind tendency not visible in the resolved eddy forcing terms, suggesting that gravity wave drag may play a role in the downward propagation of variations in the upper stratosphere and lower mesosphere.

Kodera (2002, 2003) found that the signal associated with the NAO in winter extends to hemispheric scale and into the upper stratosphere in High Solar (HS) years (K-phenomenon). Ogi *et al.* (2003) also found that the signal associated with the winter

NAO tends to persist until next summer in HS years (O-phenomenon). **K. Kodera** presented a study using ERA-40 data, to determine whether the Southern Annular Mode (SAM) exhibits similar behaviour. A correlation analysis was performed on the October-November mean SAM index separately on HS and LS years. Both phenomena were found in SAM, and SAM is very persistent in HS years. Observations, and model runs with and without chemistry suggest that ozone plays a key role for persistence of SAM signal. The high extension of the SAM signal toward the upper stratosphere in late winter (K-phenomenon) is important factor for the driving of ozone to polar-lower stratosphere for the persistence of the AM (O-phenomenon).

**J. Perlwitz** presented a study on the impact of stratospheric climate change on the troposphere by stratosphere-troposphere dynamical coupling. The key processes in stratosphere-troposphere dynamical coupling are the upward propagation of planetary Rossby waves from the troposphere to the stratosphere, and the absorption of wave activity in the stratosphere which changes the mean flow, which in turn changes the region of strongest interaction of the waves due to these changes in the basic state. The zonal mean flow perturbation progresses downward and poleward. Wave activity may also be reflected back down into the troposphere such that the structure of tropospheric waves is modified, but there is little effect on the zonal mean.

To study the downward propagation of the Northern Annular Mode (NAM), the leading EOFs were calculated of the daily zonal mean height, and the time-lagged correlation coefficients for DJF relative to 10 hPa. It was found that the downward progression of the NAM was captured in all models studied, except in the GISS model, but that the relationship between the stratospheric and near-surface NAM fields is stronger in models than in reanalysis. Models also show a stronger persistence of NAM in the stratosphere.

Discussion on stratospheric variability and its coupling to the troposphere was lively. While early GRIPS tasks studied warmings and interannual variability of the coupled troposphere-stratosphere system, these model evaluations were limited by the lack of long runs. Since it is now becoming possible to run multi-decadal simulations (even with interactive chemistry), it is timely to revisit these questions. Reviving the early GRIPS Tasks, in the context of examing 20-50 year simulations with present models, is thus possible. The presentations by G. Roff and J. Perlwitz, as well as many ideas raised in discussion, pointed to the continued value of such evaluations, for both fundamental understanding of GCMs (or CCMs) and for insight into how changes in the circulation into the 21st century may be impacted by the baseline circulation statistics of the models. Such work is encouraged in the GRIPS-CCMVal transition period.

## Gravity Wave Drag

A realistic simulation of the climate of the middle atmosphere requires the transfer of angular momentum by unresolved gravity waves (GW). Without a gravity wave drag (GWD) parameterization scheme most GCMs will not produce many of the basic features of the middle atmosphere (e.g., cold summer mesopause, zonal wind reversal of the mesosphere, QBO). A number of parameterizations are currently used in GCMs, with the fundamental difference between them being how the waves “break” and deposit their momentum to the background flow. However, due to the lack of observational evidence, the parameters of a GWD parameterization are typically “tuned” to obtain a reasonable mean climate.

While gravity wave properties at source level are the greatest uncertainty, typically they are specified so that a reasonable middle atmosphere results from numerical simulations. However, this feature does not allow for changes of source properties in different climate change scenarios. Using the WACCM, **F. Sassi** tested three GW generation schemes: the base case (with a zonally uniform GW source that operates continuously with an intermittency factor), the Charron and Manzini (2002) scheme for GW production due to frontal development, and the Beres (2004) scheme for production of GW by convection. The results show that model consistent GW schemes are in general preferable to *ad hoc* solutions. Frontogenesis and convection introduce realistic seasonal and spatial variability, however, these schemes introduce a new level of tuning that can be arbitrary and model dependent, and needs further exploration.

**C. McLandress** addressed the question of how the differences in the parameterization of wave breaking actually affect climate simulations. Gravity wave dissipation can occur in two ways: nonlinear dissipation, and critical level (CL) dissipation. It is the nonlinear dissipation that is different in each of the GWD parameterizations. Using the standard settings from Hines, Warner-McInytre (WM) and Alexander-Dunkerton (AD) will produce large differences in the GCM responses. However, by modifying the saturation threshold for WM and AD, it is possible for all three parametrizations to deposit their momentum at the same height. The wind response is nearly identical. This suggests that it is height at which the GWs are dissipated, and not the details of the nonlinear dissipation mechanisms, that is the crucial factor in determining the GCM response.

**T. Shaw** addressed the question of spurious downward influence of the mesosphere on the stratosphere due to using GWD parametrizations that are not constrained by momentum conservation, either in principle or in the way in which they are implemented. Momentum conservation implies that GWD induced downwelling (and heating) at a given altitude depends only on the gravity wave momentum flux through that altitude (Haynes *et al.*, 1991), such that GWD feedbacks from changes in the zonal wind above a given level are restricted to the region above. Therefore, zonal wind changes above a level cannot affect the circulation below it via GWD feedbacks. To respect the momentum constraint and avoid spurious downward influence, any nonzero parameterized momentum flux at the model lid must be deposited in the model domain. Dynamical feedbacks from parameterizations could be falsely interpreted as stratospheric and mesospheric effects on climate.

**S. Pawson** showed runs from the GEOS-4 GCM that were capable of producing a QBO. The AD GWD parametrization was used and the control run produced a reasonable SAO and a low-frequency oscillation in the tropical stratosphere, with some resemblance to the QBO with a period of about 20 months. GWD by a “convective spectrum” in the AD scheme does the work, however, the spectral parameters need to be quite different from those recommended by Alexander and Rosenlof

(AR). It was found that in order to produce the QBO shorter wavelengths (100 km vs. 4000 km) were used than in the standard AR scheme, a much narrower spectral width was needed, and a stronger momentum flux at launch level was needed. It was also found that the QBO period increased with slightly weaker GW forcing and there was a better downward extent of the QBO, especially in the easterly anomalies. Increasing the vertical resolution also led to tighter shear zones and perhaps better downward propagation into the lowest part of the stratosphere.

## Issues in Chemistry-Climate Modelling

**R. Stolarski** gave a presentation on the need for, and the problems with producing, high-quality, long-term datasets. Long-term datasets, such as the total ozone data from TOMS and SBUV, enable us to test how processes act together to give decadal-scale responses. The ultimate prediction test is our hindcast capability so that decadal scale hindcasts compared to long-term data sets provide additional evaluations of models beyond process-oriented evaluation.

A long-term dataset is comprised of data from many satellites, and each instrument has calibration and drift issues. It is necessary to estimate trend uncertainty due to instrument drift and to remove known systematic errors from long-term datasets. However, residual errors could remain that lead to drift uncertainty. In addition, the introduction of a new instrument into a dataset introduces a new uncertainty, so it is important to have enough overlap to properly calibrate instruments and improve uncertainties. Ground stations are useful in that they may be calibrated as needed, but statistical uncertainty dominates the trend uncertainty, and there are still drift uncertainties in individual ground stations. Using a combination of both satellite and ground-based measurements is better.

Extratropical ozone builds up in winter/spring due to transport from the tropics, and decreases in late-spring/summer due to photochemistry (more so in NH than in SH). It tends to be that years that are high/low in spring are high/low in summer. **T. Shepherd** showed that the correlation coefficient (of detrended data) between different months from 35° to 60° shows a remarkable seasonal persistence of ozone anomalies until fall. In both hemispheres, 35°-80° springtime ozone is a good predictor for *both* 35°-60° and 60°-80° summertime ozone. In fact it is a better predictor of 60°-80° summertime ozone than is 60°-80° springtime ozone. Once the vortex breaks down, the correlations become coherent throughout the extratropics (Fioletov and Shepherd, 2005). The SH midlatitude ozone variability seems to be slaved to NH variability. This memory could be in tropical zonal winds, i.e. the "flywheel" (Scott and Haynes, 1998).

**V. Fomichev** presented the result from an ongoing study to diagnose the impact of increasing greenhouse gas (GHG) concentrations in the Canadian Middle Atmosphere Model (CMAM). A series of multi-year runs has been performed with double $CO_2$, with and without interactive chemistry, and with sea surface temperatures (SSTs) prescribed for both $1xCO_2$ and $2xCO_2$ climate. In response to $CO_2$ doubling, the middle atmosphere cools ~10K with maximum impact near the stratopause. The ozone radiative feedback (through both solar and IR heating) reduces the $CO_2$ induced cooling by up to ~4.5K. The main impacts of SST changes on the middle atmosphere are a higher and warmer tropopause, cooling just above the tropopause (in both the tropics and extratropics), and a decrease in ozone near the tropopause, both possibly related to the "tropopause shift". However, the winter polar regions can be highly variable, so long runs are required to increase confidence in the results.

On behalf of A. Scaife and N. Butchart, **A. Bushell** presented the results from the Level 4 GRIPS task on how climate change affects the Brewer-Dobson circulation. The troposphere-stratosphere mass exchange determines the lifetimes of key trace gases, and transports ozone down into the troposphere. Therefore, changes in the mass circulation will affect both ozone and climate predictions. The Unified Model predicts that an increase in GHGs will cause a systematic increase in troposphere-stratosphere exchange (Butchart and Scaife, 2001), and this study is an attempt to assess the robustness of this predicted change in mass exchange and the role of planetary wave driving using 14 different model runs from 11 different groups.

| | | | |
|---|---|---|---|
| **Level 1 Tasks :** | | | |
| 1a | Documentation | S. Pawson | |
| 1b | Climatology | S. Pawson | |
| 1c | Troposphere-Stratosphere Connection | K. Kodera | |
| 1d | Sudden Warmings | W. Lahoz | complete |
| 1e | Travelling Waves/Tides | K. Hamilton | complete |
| 1f | Tropical Waves | T. Horinouchi | complete |
| 1g | Stratosphere-Troposphere Exchange | - | |
| 1h | Spatial Wavenumber | J. Koshyk | Complete |
| 1i | Polar Vortices | G. Roff | ending, report this meeting |
| 1j | Transport | - | |
| **Level 2 Tasks :** | | | |
| 2a | Radiation Scheme Comparison and Validation | P. Forster, V. Fomichev, U. Langematz | First stage almost complete |
| 2b | Inferred GWD | S. Pawson | complete |
| 2c | Impacts of Mesospheric Drag | S. Beagley, B. Boville | complete |
| 2d | GWD Evalution | C. McLandress | complete |
| **Level 3 Tasks :** | | | |
| 3a | PINMIP : Impact of Mt. Pinatubo aerosols | G. Stenchikov, A. Robock | ongoing |
| 3b | Response to solar forcing anomalies | K. Kodera, Matthes | complete |
| 3c | Response to ozone trends | U. Langematz | near completion |
| 3d | Response to CO2 change | - | |
| **Level 4: Tasks :** | | | |
| 4a | Changes in residual circulation due to climate change | A. Bushnell, A. Scaife, N. Butchart | report this meeting |

*Table 1: The GRIPS tasks, leaders and stages of completion.*

The results of the study are that all models have a Brewer Dobson circulation with upwelling in a "tropical pipe" extending to about ±30° (annual mean), and the pipe is displaced by 10°-20° toward the summer pole. However, at 70 hPa, the upwelling mass flux varies by up to a factor of 2 between models. The maximum upwelling is in DJF and is often 50% larger than in JJA. All models confirm a positive trend in the mass flux across the tropopause due to climate change. This occurs throughout the year and is consistent with an increase in planetary wave driving at upper levels. The increase is about 2% per decade. This feedback could amount to a ~20% decrease in the lifetime of $N_2O$ and CFCs, for example, by 2100 and is not included in standard climate models.

**J. Austin** presented a study on the age of air and the meridional circulation in a GFDL coupled chemistry-climate model. Two 21-year runs were used: one with 1960 forcings and one with 1980 forcings. In time-slice simulations water vapour increased by 4% per decade from 1960 to 2000, entirely consistent with $CH_4$ oxidation. The tropical upwelling increased at the equivalent rate of 1.7% per decade. The decrease is approximately balanced by a 2.7% per decade decrease in the age of air. Model results are in reasonable agreement with observations for tropical upwelling, but underpredict age of air by over 30%, implying that there is too much mixing. A summary of GFDL's strategy for chemistry-climate evolution was also outlined.



**J. de Grandpré** presented a 20 year run from the CMAM including heterogeneous chemistry, interactive ozone, and current chlorine loading. The findings indicate that CMAM has a realistic representation of the residual circulation, but that the homogeneity of temperature, and that of long-lived constituents is too strong in the NH winter period. Ozone flux from the stratosphere to the troposphere has an upper limit of 750 Tg/year, and methane and nitrogen loss in the stratosphere is 13.6 Tg/year and 16.1 Tg/year respectively.

Discussion of chemistry-climate simulations tackled several aspects. There is a need to balance between studies that address the mechanisms of interaction and the necessity of producing multi-decadal integrations that span the period of around 1950-2100, in order to meet requirements for the forthcoming ozone assessment. Key scientific questions for CCM evaluation are the detection, attribution and prediction of trends in atmospheric ozone in the interactive chemistry-climate environment, the future recovery of stratospheric ozone within the context of human-induced and natural variabilities, the effects of changing chemical composition on the coupled tropospheric-stratospheric dynamics and radiation budget and vice-versa, and the effects of natural perturbations (*e.g.* volcanic) on the interactive chemistry-climate environment. All of these are priorities in SPARC research.

Cross-evaluation of multiple CCMs has not progressed substantially since the previous ozone assessment, and this is a task that will be undertaken within CCMVal. Many aspects of the validation may draw from GRIPS studies, but chemical evaluation will require different diagnostics and a different set of expertise than has been available for GRIPS.

In many ways, interpretation of chemistry-climate prediction studies is not a fully mature field. The discussions showed that the various research groups have different levels of accountability to funding agencies and, on this basis, have different priorities for model experiments. While the baseline scientific questions are essentially well defined, neither the strategy for numerical experimentation nor the analysis methods are well established. This means that more basic research is required, which may contradict the perceived requirement that all groups run their simulations under identical, controlled conditions. While there was consensus that emission scenarios for anthropogenic "climate" and "chemistry" gases will be adopted from established scenarios (from 3-D climate models and 2-D chemistry models), there is less agreement on issues such as: which scenarios would be used; what boundary conditions should be used (*e.g.*, future sea-surface temperatures); whether or not to impose solar cycles in incoming irradiance. The treatment of aerosols remains quite primitive, with models run with low aerosol loading at the moment, although this is a very important climate issue. In the prediction context, it is necessary to understand how major eruptions at different times will affect climate response and the detection of ozone changes. The role of bromine gases was also the focus of interesting discussion, motivated by the cameo appearance of **R. Salawich** at the meeting. This discussion especially underscored the importance of scientific experimentation, in order to better understand the fundamental issues in chemistry-climate change.

The lack of agreement on these issues is somewhat contradictory to the need for SPARC to contribute to the next WMO/UNEP ozone assessment. A major point raised was the time needed for groups to complete useful climate runs, and the fact that IPCC scenarios were not set in time for groups to respond. While the timeline allowed for 2-D runs to be performed, it will not be adequate for most groups to complete full 3-D CCM runs. The need to lead the science in defining acceptable scenarios, rather than lag behind many groups running different scenarios is desirable. This implies the need to quickly agree on scenarios for the upcoming assessements. Even then, some groups are already well into their simulations and these groups have not coordinated their scenarios.

## Discussion on GRIPS Tasks and Other Issues

The GRIPS tasks are in various stages of completion; some tasks can be brought to closure, and others should carry forward into CCMVal. A summary of the tasks and their status is given in Table 1.

Plans to update the model evaluation (Tasks 1a,b) will essentially be superseded by CCMVal. Studies of troposphere-stratosphere connections are ongoing and will be part of the transition from GRIPS to CCMVal, and although task 1d (Sudden Warmings) was completed, it may be revived. Level-2 tasks, aimed at understanding the parameterizations used in models and their impacts on the simulated climate, are mostly complete. P. Forster has almost completed the comparison stage of the radiation codes (Task 2a) and the results were presented at the SPARC General Assembly 2004.

Level-3 tasks are aimed at understanding how the climate models respond to different physical forcing mechanisms. The PIN-MIP experiments (Task 3a) will likely evolve into CCMVal tasks, along with task 3d. The solar forcing task (3b) is complete and will be reported on in this meeting. Substantial progress has also been made with the "1980 – 2000" ozone project, and it is almost complete (U. Langematz could unfortunately not attend the workshop). GRIPS had one Level-4 task, an examination of the changes in residual circulation in a changing climate, which was discussed.

While GRIPS is now ending, many of the remaining questions in "meteorological" modelling of the middle atmosphere will carry forward into CCMVal, where they will be complemented by questions of relevance to transport, chemistry and radiation. CCMVal should draw from the experiences of GRIPS, but will face new challenges. One of these concerns data management. Because of the possibilities for longer model runs, with chemical as well as meteorological output, CCMVal will require more resources. The SPARC Data Centre has already made some plans to cope with the amount of data that will be necessary to handle for some of the planned projects. Another challenge for CCMVal will be community involvement, so that the tasks do not fall on certain individuals. CCMVal is taking an "open conference" approach to many of its activites, but there will remain a need for focused, workshop-style activities within SPARC. The key will be to effectively combine COPES and CCMVal activities and procure funding.

## Acknowledgements

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# Overview of planned coupled chemistry-climate simulations to support upcoming ozone and climate assessments

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## Introduction

SPARC has established a new validation activity, CCMVal, for coupled chemistry-climate models (CCMs). The activity is based on the framework developed at the SPARC workshop on process-oriented CCM validation held in Grainau, Germany in November 2003 (Eyring *et al.*, 2004, 2005) and draws upon the experiences within the SPARC GCM-Reality Intercomparison Project (GRIPS) (See the Report on GRIPS in this issue). As more climate models include chemical components, the time has arrived for formal comparisons of these coupled chemistry-climate models. Within SPARC, this new activity will be one of the supporting "pillars" of the integrated themes.

The goal of the new activity is to improve understanding of CCMs and their underlying GCMs (General Circulation Models) through process-oriented validation. One outcome of this effort is expected to be improvements in how well CCMs represent physical, chemical, and dynamical processes. In addition, this effort will focus on understanding the ability of CCMs to reproduce past trends and variability and providing predictions from ensembles of long model runs. Achieving these goals will involve comparing CCM constituent distributions with (robust) relationships between constituent variables as found in observations. This effort is both a model-model and model-data comparison exercise. At the Grainau workshop, a set of key diagnostics was defined for evaluating CCM performance with respect to radiation, dynamics, transport, and stratospheric chemistry and microphysics (**http://www.pa.op.dlr.de/CCMVal/**). This approach allows modellers to decide (based on their own priorities and resources) which diagnostics to examine in any particular area. The CCMVal activity will help coordinate and organize CCM model efforts around the world.

| Model | Horizontal resolution | No. Vertical Levels/ Upper Boundary | Group and location | References | Contacts |
|---|---|---|---|---|---|
| **AMTRAC** | 2°x 2.5° | 48 / 0.0017 hPa | GFDL, USA | Anderson *et al.* (2004); Austin (2002) | J. Austin |
| **CCSR/ NIES** | T21 | 30 / 0.06 hPa | NIES, Tokyo, Japan | Nagashima *et al.* (2002); Takigawa *et al.* (1999) | H. Akiyoshi, T. Nagashima, M. Takahashi |
| **CMAM** | T32 or T47 | 65 / 0.0006 hPa | MSC, University of Toronto and York University, Canada | Beagley *et al.* (1997); de Grandpré *et al.* (2000) | T.G. Shepherd |
| **E39/C** | T30 | 39 / 10 hPa | DLR Oberpfaffenhofen, Germany | Dameris *et al.* (2005) | M. Dameris, V. Eyring, V. Grewe, M. Ponater |
| **ECHAM5 / MESSy** | T42 | 39 / 0.01 hPa | MPI Mainz, MPI Hamburg, DLR Oberpfaffenhofen, Germany | Jöckel *et al.* (2004); Roeckner *et al.* (2003); Sander *et al.* (2004) | C. Brühl, M. Giorgetta, P. Jöckel, E. Manzini, B. Steil |
| **FUBCMAM** | T21 | 34 / 0.0068 hPa | FU Berlin, MPI Mainz, Germany | Langematz, et al. (2005) | U. Langematz |
| **GCCM** | T42 | 18 / 2.5 hPa | Univ. of Oslo, Norway; SUNY Albany, USA | Wong *et al.* (2004) | M. Gauss, I. Isaksen |
| **GEOS CCM** | 2° x 2.5° | 55 / 80km | NASA/GSFC, USA | In preparation | A. Douglass, P.A. Newman, S. Pawson, R. Stolarski |
| **GISS** | 4° x 5° | 23 / 0.002 hPa | NASA GISS, New York, USA | Schmidt *et al.* (2005a) | D. Rind, D. Shindell |
| **HAMMONIA** | T31 | 67 / 2.10-7 hPa | MPI Hamburg, Germany | Schmidt *et al.* (2005b) | G. Brasseur, M. Giorgetta, H. Schmidt, |
| **LMDREPRO** | 2.5° x 3.75° | 50 / 0.07 hPa | IPSL, France | In preparation | S. Bekki, D. Hauglustaine, L. Jourdain |
| **MAECHAM /CHEM** | T30 | 39 / 0.01 hPa | MPI Mainz, MPI Hamburg, Germany | Manzini *et al.* (2003); Steil *et al.* (2003) | C. Brühl, M. Giorgetta, E. Manzini, B. Steil |
| **MRI** | T42 | 68/0.01 hPa | MRI, Tsukuba, Japan | Shibata and Deushi (2005); Shibata *et al.* (2005) | K. Shibata |
| **SOCOL** | T30 | 39 / 0.01 hPa | PMOD/WRC and ETHZ, Switzerland | Egorova *et al.* (2004) | E. Rozanov |
| **ULAQ** | 10° x 20° | 26 / 0.04 hPa | University of L'Aquila, Italy | Pitari *et al.* (2002) | E. Manzini, G. Pitari |
| **UMETRAC** | 2.5° x 3.75° | 64 / 0.01 hPa | UK Met Office, UK NIWA Lauder (NZ) | Austin (2002); Austin and Butchart (2003) | G. Bodeker, N. Butchart, H. Struthers |
| **UM SLIMCAT** | 2.5° x 3.75° | 64 / 0.01 hPa | University of Leeds, UK | Tian and Chipperfield (2005) | M.P. Chipperfield, W. Tian |
| **UMUCAM** | 2.5° x 3.75° | 58 / 0.1 hPa | University of Cambridge, UK | Braesicke and Pyle (2003 and 2004) | P. Braesicke, J.A. Pyle |
| **WACCM3** | 2° x 2.5° | 66 / 140 km | NCAR, USA | Sassi *et al.* (2005) | B. Boville, R. Garcia, A. Gettelman, D. Kinnison, D. Marsh |

*Table 1: Main features of current coupled chemistry-climate models (CCMs). CCMs are listed alphabetically. The horizontal resolution is given in either degrees latitude x longitude (grid point models), or as T21, T30, etc., which are the resolution in spectral models corresponding to triangular truncation of the spectral domain with 21, 30, etc., wavenumbers, respectively. All CCMs have a comprehensive range of chemical reactions except the UMUCAM model, which has parameterized ozone chemistry. The coupling between chemistry and dynamics is represented in all models, but to different degrees. All models include orographic gravity wave drag schemes (O-GWD); most models additionally include non-orographic gravity wave drag schemes (NonO-GWD).*

In this way, the CCM community can provide the maximum amount of useful scientific information for WMO/UNEP and IPCC assessments.

As a first step, the CCM community has defined two reference simulations and a set of model forcings to support the upcoming WMO/UNEP Scientific Assessment of Ozone Depletion. The forcings are defined by natural and anthropogenic emissions based on existing scenarios, on atmospheric observations, and on the Kyoto and Montreal Protocols and Amendments. In the following sections, we describe current models and proposed model simulations, and discuss several special issues related to the use of CCMs.

## Participating models

During the last few years, a number of new CCMs have been developed, which significantly deepens the pool of available models. In comparison with the models used in support of the last WMO/UNEP ozone assessment (Austin *et al.*, 2003; WMO, 2003), current CCMs generally have improved representations of physical processes, and modelling groups have greater computational resources. Table 1 gives an overview of current coupled-chemistry climate models around the world.

## Multi-model simulations to support the upcoming WMO/UNEP assessment

CCMs represent both natural dynamical variability and the dynamical response to forcings such as sea surface temperatures (SSTs). As a result, a meaningful comparison of different CCM results requires a proper analysis of statistical significance and a careful representation of natural and anthropogenic forcings. To address these issues, a set of questions has been set up for the community to decide on possible model simulations and forcings. The draft was opened for discussion within the CCM community (see Table 1) and with several other experts.

The proposed scenarios were developed to address the following key questions outlined by the WMO/UNEP Steering Committee to be of significance to the upcoming assessment: (1) How well do we understand the observed changes in stratospheric ozone (polar and extra-polar) over the past few decades during which stratospheric climate and constituents (including halogens, nitrogen oxides, water, and methane) were changing? (2) What does our best understanding of the climate and halogens, as well as the changing stratospheric composition, portend for the future? (3) Given this understanding, what options do we have for influencing the future state of the stratospheric ozone layer?

In order to address questions (1) and (2), two reference simulations (REF) have been proposed.

### Reproduce the past: Reference simulation 1 (REF1), Core time period 1980 to 2004

**REF 1** is designed to reproduce the well-observed period of the last 25 years during which ozone depletion is well recorded, and allows a more detailed investigation of the role of natural variability and other atmospheric changes important for ozone balance and trends. This transient simulation includes all anthropogenic and natural forcings based on changes in trace gases, solar variability, volcanic eruptions, quasi-biennial oscillation (QBO), and sea surface temperatures (SSTs). SSTs in this run are based on observations. Depending on computer resources some model groups might be able to start earlier. We highly recommend reporting results for REF1 between 1960 and 2004 to examine model variability. Forcings for the simulation and a detailed description can be downloaded from the CCMVal website (http://www.pa.op.dlr.de/CCMVal/Forcings/CCMVal_Forcings.html). They are defined for the time period 1950 to 2004.

SSTs in REF1 are prescribed as monthly means following the global sea ice and sea surface temperature (HadISST1) data set provided by the UK Met Office Hadley Centre (Rayner *et al.*, 2003). This data set is based on blended satellite and *in situ* observations.

Both chemical and direct radiative effects of enhanced stratospheric aerosol abundance from large volcanic eruptions are considered in REF1. The three major volcanic eruptions (Agung, 1963; El Chichon, 1982; Pinatubo, 1991) are taken into account, i.e., additional heating rates and sulfate aerosol densities are prescribed on the basis of model estimates and measurements, respectively. A climatology of sulfate surface area density (SAD) based on monthly zonal means derived from various satellite data sets between 1979 and 1999 has been provided by **David Considine** (NASA Langley Research Center, USA). Details on how to represent the sulfate SAD before 1979 are described on the CCMVal web site.

The QBO is generally described by zonal wind profiles measured at the equator. While the QBO is an internal mode of atmospheric variability and not a “forcing” in the usual sense, at the present time most models do not exhibit a QBO. This leads to an underestimation of ozone variability, and compromises the comparison with observations. While some of the models internally generate a QBO, for the others it has been agreed to assimilate observed tropical winds. Assimilation of the zonal wind in the QBO domain can add the QBO to the system, thus providing, for example, its effects on transport and chemistry. Radiosonde data from Canton Island (1953-1967), Gan/Maledives (1967-1975) and Singapore (1976-2000) have been used to develop a time series of measured monthly mean winds at the equator (Naujokat, 1986; Labitzke *et al.*, 2002). This data set covers the lower stratosphere up to 10 hPa. Based on rocket wind measurements near 8 degree latitude, the QBO data set has been vertically extended to 3 hPa. The software package to assimilate the QBO by a linear relaxation method (also known as “nudging”) as well as the wind data sets have been provided by **Marco Giorgetta** (MPI Hamburg, Germany).

The influence of the 11-year solar cycle on photolysis rates is parameterized according to the intensity of the 10.7 cm radiation of the sun (which is a proxy to the phase of the given solar cycle). The spectral distribution of changes in the observed extra-terrestrial flux is based on investigations presented by Lean *et al.* (1997) (see http://www.drao.nrc.ca/icarus/www/sol_home.shtml for details).

### Making predictions: Reference simulation 2 (REF2), Core time period 1980 to 2025

**REF 2** is an internally consistent simulation from the past into the future. The proposed transient simulation uses the IPCC SRES scenario A1B(medium) (IPCC, 2000). REF 2 only includes anthropogenic forcings; natural forcings such as solar variability are not considered, and the QBO is not externally forced (neither in the past, nor in the future). Sulfate surface area density is consistent with REF1 through 1999. Sulfate surface area densities beyond 1999 will be fixed at 1999 conditions (volcanically clean conditions). Changes in halogens will be prescribed following the Ab scenario (WMO, 2003; Table 4B-2). SSTs in this run are based on coupled atmosphere-ocean model-derived SSTs. Depending on computer resources some model groups might be able to run longer and/or start earlier. We recommend reporting results for REF2 until 2050. The forcings on the website are defined through 2100.

*Fully coupled atmosphere-ocean CCMs* that extend to the middle atmosphere and include coupled chemistry, will use their internally calculated SSTs. *CCMs driven by SSTs and sea ice distributions from the underlying IPCC coupled-ocean model simulation* could use the model consistent SSTs. One constraint is to make the SST dataset consistent with the SRES greenhouse gas (GHG) scenario A1B (medium). All other CCM groups will run with the same SSTs, provided by a single IPCC coupled-ocean model simulation. These simulations have good spatial resolution, so the data-sets should be suitable for all the CCMs participating in the WMO/UNEP assessment.

### Sensitivity simulations

Scenarios for sensitivity experiments to address question (3) will be defined later. Possible sensitivity experiments could be:

**SCN 1 (REF 1 with enhanced $Br_y$):** An additional simulation is being developed to represent the known lower stratospheric deficit in modelled inorganic bromine abundance. This simulation will be identical to REF 1, with the exception of including source gas abundances that will increase the stratospheric burden of $Br_y$. Details of this simulation will be made available shortly.

**SCN 2 (REF 2 with natural forcings):** A sensitivity simulation has been defined similar to REF1, with the inclusion of solar variability, volcanic activity, and the QBO in the past. Future forcings include a repeating solar cycle and QBO under volcanically clean aerosol conditions. SSTs are based on REF2. Greenhouse gases and halogens will be the same as in REF2.

A summary of the proposed CCMVal reference and sensitivity simulations is given in Table 2.

A web site containing descriptions of the model simulations, as well as relevant forcings (past and present), can be found at http://www.pa.op.dlr.de/CCMVal/Forcings /CCMVal_Forcings.html. The forcings for the specified simulations may be downloaded from this website.

## Discussion

In the effort to select the simulations in Table 2, several issues arose that required a special action or decision. A brief perspective on these issues is presented in this section.

### Greenhouse Gases and Halogens

**Historical and Future Trends:** It has been agreed that the simulations to represent the past (REF1) should not stop in the year 2000, but should be extended until 2004. Between the years 2000 and 2004 new measurements of trace gases from the SCIAMACHY, MIPAS, AURA, ACE and ODIN satellite instruments became available. In addition, new data from existing satellite instruments, such as TOMS, GOME, and HALOE, are also available for CCM intercomparisons. In response, **Stephen Montzka** (NOAA Climate Monitoring and Diagnostics Laboratory, USA) has offered to update the datasets of halogen and other greenhouse gas observations to 2004. Datasets of sea surface temperatures up to 2004 are available on the Hadley Centre web site (see http://www.hadobs.org/).

For the prediction simulation (REF 2) the community agreed to run with the GHG scenario SRES A1B(medium) with the halogen scenario Ab from WMO (2003). However, **Paul Fraser** (CSIRO, Australia) mentioned that the SRES reference scenario A1B is very unlikely to be realistic for $CH_4$ over the next 20 years. A1B requires $CH_4$ to increase from 1760 ppb in 2000 to 2026 ppb in 2020, *i.e.*, with a growth rate of 13-14 ppb per year. This growth rate has not been observed since the late 1970s. In contrast, the growth rate in the Southern and Northern Hemispheres for the past five years has been less than 1 ppb per year, with the current globally averaged concentration at 1750 ppb (2004).

In the proposed simulation REF2, the CCMs are driven by SSTs and sea ice distributions from coupled ocean-atmosphere model simulations using IPCC SRES GHG scenarios. To be consistent with the IPCC simulations, the GHG scenarios must be the same as in the coupled ocean-atmosphere model simulations. Therefore, it has been decided that $CH_4$ emissions during this phase of the assessment process will not be changed from the IPCC SRES GHG scenarios for REF2.

**Inorganic Bromine Deficit:** Model representations of inorganic bromine radicals in the lower stratosphere and comparisons with observations have recently been documented (see WMO, 2003, Chapters 1 and 2; Salawitch *et al.,* 2005 and references within). Results from these comparisons strongly suggest that models greatly underestimate the total inorganic bromine ($Br_y$) in this region (up to 6 pptv). Furthermore, it is clear that using time-dependent boundary conditions as prescribed in the Ab scenario (WMO 2003) will not correct the modelled $Br_y$ distribution. It is believed that this discrepancy occurs because very short-lived (VSL) bromine-containing source gases are not included in the models. Incorporating these species into CCMs will require understanding of the magnitude and geographic distribution of the sources of these VSL gases and their loss processes in the atmosphere. Based on input from **Ross Salawitch** (Jet Propulsion Laboratory, USA), **Martyn Chipperfield** (University of Leeds, UK), and **Stephen Montzka,** and considering the time constraints of including VSL species and related processes into CCMs, it was decided that no attempt should be made to address the $Br_y$ deficit in the reference simulations (REF1 and REF2). This means that the reference CCM experiments will necessarily underestimate stratospheric $Br_y$ by about 25%. This will impact their ability to reproduce, for example, polar ozone loss quantitatively, and predict the future ozone changes; this caveat needs to be remembered when analyzing the results. However, a sensitivity simulation is being developed to examine this issue (SCN1). Quantification of the effect of enhanced $Br_y$ on ozone trends for the WMO/UNEP 2006 ozone assessment will most likely be done with 2D and 3D chemical-transport models.

| Scenario | Period | Trace Gas | Halogens | SSTs | Background & Volcanic Aerosol | Solar Variability | QBO | Enhanced $BrO_Y$ |
|---|---|---|---|---|---|---|---|---|
| **REF1** | **1980-2004** If possible 1960 to 2004 | **OBS** GHG used for WMO/UNEP | **OBS** used for 2002 WMO/UNEP 2002 runs. Extended until 2004 | **OBS** HadISST1 runs. | **OBS** Surface Area Density data (SAD) provided by David Considine | **OBS** MAVER data set, observed flux | **OBS or internally generated** | |
| **REF2** | **1980-2025** If possible until 2050. | **OBS + A1B(medium)** | **OBS** + Ab scenario from WMO/UNEP 2002 | **Modelled SSTs** | **OBS / SAD** from 1999 | - | **Only internally generated** | - |
| **SCN1** | **1980-2004** | **OBS** | **OBS** used for WMO/UNEP 2002 runs | **OBS** | **OBS** | **OBS** | **OBS or internally generated** | **Included** Based on Salawitch *et al.* [2005] |
| **SCN2** | **1980-2025** | **OBS + A1B (medium)** | OBS + Ab 2002 scenario from WMO/UNEP 2002 | **Modelled SSTs** | **OBS /** SAD from 1999 | **OBS** repeating in future | **OBS / in future or internally generated** | - |

*Table 2: Summary of proposed CCMVal simulations.*

## QBO and Solar Variability

Solar activity, as well as the QBO, has a strong influence on ozone variability. Some CCMs with high horizontal and vertical resolutions are able to internally generate a QBO. However, the majority of CCMs do not generate a QBO. Consequently these models simulate permanent tropical easterlies instead of a QBO. As the QBO is important for wave propagation and interaction with high latitudes, the latter CCMs therefore have a known deficit which would affect both the means and variabilities of trace gas distributions. Therefore, part of the community felt that QBO and solar variability should also be included in future years in the reference simulations to the year 2025 and, therefore, suggested using SCN2 as the reference simulation instead of REF2.

However, others have reservations about including a QBO and solar cycle in the future, since these are not anthropogenic forcings and, hence, cannot be predicted. In the case of the QBO, which is an internal mode of atmospheric variability and not a "forcing" at all, the amplitude and phase of the QBO will have no connection to the prognostic model variables or the SSTs and, of course, will not respond to climate changes.

The obvious way to address the opposing views is to encourage groups to run both simulations, REF2 and SCN2. However, due to limited time and computer resources it is not very likely that all, or even most, groups can afford to run both the REF2 and SCN2 simulations. Therefore, a decision was made to make REF1 and REF2 the highest priority and encourage groups to run SCN2 in addition if resources allow.

## Sea Surface Temperatures

One of the most critical issues in the discussion was the design of the future simulation REF2. The problem is how to extend SST observations into the future without introducing a discontinuity at the present-to-future transition.

One possibility would be to add time-evolving anomalies to the observed SSTs that are specified for REF1. However, the community sees at least two problems with this approach. First, the patterns and temporal variability are changed, depending on the shortcomings of the coupled system. Second, the ice distribution in the SST observational dataset is not the same as in the model. This is especially problematic in regions where the ice cover disagrees significantly between model and observations.

We can avoid these problems if the one-simulation-for-all design is abandoned in favour of a design including two separate simulations. The first would be for the observed period (REF1), for which we can assess the degree to which observed stratospheric dynamics and chemistry are reproduced. The second would be an internally consistent simulation from the past to the future (REF2). With this approach, fully coupled atmosphere-ocean CCMs with the atmosphere extending to the middle atmosphere and with coupled chemistry, will use their internally calculated SSTs in REF2, whereas all other CCMs will use modelled SSTs from a coupled atmosphere-ocean simulation for the full time period (1980 to 2025 or longer). One constraint is to make the external SST dataset consistent with the GHG scenario A1B.

There has also been a debate on whether or not the model simulations should use the same set of SSTs for future years in REF2. Obviously, if different SSTs are used, the forced low frequency variability could be quite different between the simulations. One of the biggest uncertainties is the predictability of the decadal timescale and the separation of internal from externally-forced variability in the models. However, the focus of the future simulation is not a model-model intercomparison. Rather we would like to provide the best available prediction of the future. REF2 is a simulation that focuses on consistency and that follows the IPCC simulations. Essentially we are asking that modelling groups make their best prediction. Therefore, it is not necessary to have consistent SSTs. In fact, by applying different SSTs, the change in climate and its variability are effectively included in the simulation. (To use a common set of SSTs would certainly underestimate the uncertainty in future climate predictions, and any error in those SST predictions would lead to a bias in the model predictions.)

Finally, an agreement was reached that at least a subset of groups will run with the same SST forcings, whereas others will use internally calculated SSTs or model-consistent SSTs. This will allow us to address both views.

## Summary and Outlook

CCM Modelling groups are encouraged to run the proposed reference simulations with the same forcings. In order to facilitate the set-up of the reference simulations, CCMVal has established a website where the forcings for the simulations can be downloaded (http://www.pa.op.dlr.de/CCMVal/Forcings/CCMVal_Forcings.html). This web site was developed to serve the needs of the CCM community, and encourage consistency of anthropogenic and natural forcings in future model-model and model-observation intercomparisons. Any updates as well as detailed explanation and further discussion will be placed on this website.

We encourage the groups to run both simulations, REF1 and REF2. If a model only provides REF2 it will be more difficult to assess the model's ability to simulate realistic trends and variability. Changes on the decadal timescale are not necessarily part of the secular trend. It is quite probable that some of the changes are due to low frequency variability that is likely to be unpredictable if the source is internal. It is then possible that some of the differences between the deterministic model predictions will be attributed to unpredictability and not to differences in the fundamental forcings and responses of the models. For these reasons, we encourage groups to run ensembles. Depending on computer resources, a subset of groups might also be able to carry out sensitivity simulations. Especially if the prediction simulation only covers the short-term prediction (e.g. until 2025), it would be very useful to see how the prediction changes if a solar cycle and the QBO are included. If you are interested in this topic, please run the sensitivity simulation SCN2.

In agreement with the experts in this field, it has been decided that the enhanced stratospheric bromine scenario should not be included in the reference simulation (REF1 and REF2). Enhancing the inorganic bromine reservoir increases BrO, a reaction partner for anthropogenically derived ClO, above that found in the standard simulation in the first few kilometres of the stratosphere. The sensitivity of ozone to enhanced bromine in the lowermost stratosphere will likely depend on details of the model simulation of ClO just above the tropopause. Due to the inherent three-dimensional nature of accurately simulating ClO and BrO near the tropopause, it is hoped that one or more of the CCMs will carry out simulation SCN1. It is also expected that 2D and 3D chemical transport model (CTM) simulations will be relied upon to further assess the sensitivity of ozone trends to bromine in the lowermost stratosphere for the next UNEP/WMO ozone assessment.

CCMVal will provide a list of model recommendations that will be placed on the website. We encourage groups to check the CCMVal forcing website for recommendations concerning the model set-up and the variables that should be stored in order to allow for sophisticated intercomparisons of chemistry, transport, dynamics and radiation within the CCM.

A detailed intercomparison of CCM results and observations has successfully started. Model results from 10 European model groups that are participating in the European Integrated Project SCOUT-O3 and one model group from outside Europe (CCSR/NIES) have been obtained. The first phase of the intercomparison will be based on existing runs. With the exception of total column ozone, only transient model simulations for the time period 1980 to 1999 will be compared (no time slice experiments). We would like to encourage other model groups to join in the intercomparison and to send data from existing runs. As soon as the results of the CCMVal simulations with equal forcings become available, the intercomparisons and analyses will be repeated. It will be interesting to see how the results and interpretation change when runs with equal forcings are compared. CCMVal is still looking for volunteers from around the world to assist with the intercomparison. If you are interested in a certain diagnostic or scientific topics, please contact us.

A second CCMVal workshop will be held from October 17 to 19, 2005 at the National Center for Atmospheric Research in Boulder, Colorado (http://www.pa.op.dlr.de/workshops/CCMVal2005/). The 2005 Chemistry-Climate Modelling Workshop will focus on progress in chemistry-climate modelling and process-oriented validation of CCMs. The aims of the workshop are to identify near-term and long-term goals within the validation architecture and to coordinate activities among the participating modelling groups. In addition we will discuss how CCM results can support the WMO/UNEP Scientific Assessment of Ozone Depletion 2006 and other upcoming assessments. We encourage the participation of global modellers as well as scientists who make atmospheric observations that are relevant for model evaluation.

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# Report on the 1st WG 1 Expert Meeting on the LAUTLOS Campaign at Lindenberg August 24-27, 2004.

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The LAUTLOS field campaign was hosted by the FMI Arctic Research Centre, Sodankylä, and was assisted by Vaisala. It was successfully conducted in January and February of 2004. The purpose of LAUTLOS-WAVVAP (LAPBIAT Upper Troposphere Lower Stratosphere Water Vapour Validation Project) is the comparison and validation of the world's best hygrometers that are usable as research-type radiosondes for precise water vapour measurements in the troposphere and stratosphere (up to 10 hPa), and to improve and validate these hygrometers and radiosondes. The instruments used in the study included the Meteolabor Snow White hygrometer, the NOAA frostpoint hygrometer, the CAO Flash Lyman-alpha hygrometer, the Lindenberg FN sonde, and Vaisala's latest RS92 GPS-version. The aim is to define an optimal working range in temperature, water vapour mixing ratio, relative humidity, and pressure for each of the participating hygrometers/radiosondes. In addition to the balloon-borne instruments, the University of Bern operated its ground based 22 GHz microwave instrument, MIAWARA, at Sodankylä to obtain water vapour profiles from approximately 25 to 70 km. A further microwave radiometer was operated from a LearJet of the Swiss Air Force to obtain water vapour profiles close to the balloon locations. The field campaign consisted of 30 balloon flights carrying integrated payloads.

The WG1 expert meeting was sponsored by COST 723 (see home page **http://www.cost723.org/meetings/wg1_2)** and co-sponsored by SPARC. At the meeting, the principal investigators reported on the status of the data archiving and the evaluation process. The participants discussed the results obtained during the campaign, the details of the comparison procedures, and possible future publications. The LAUTLOS data archive consists of vertical profiles from 9 different systems.

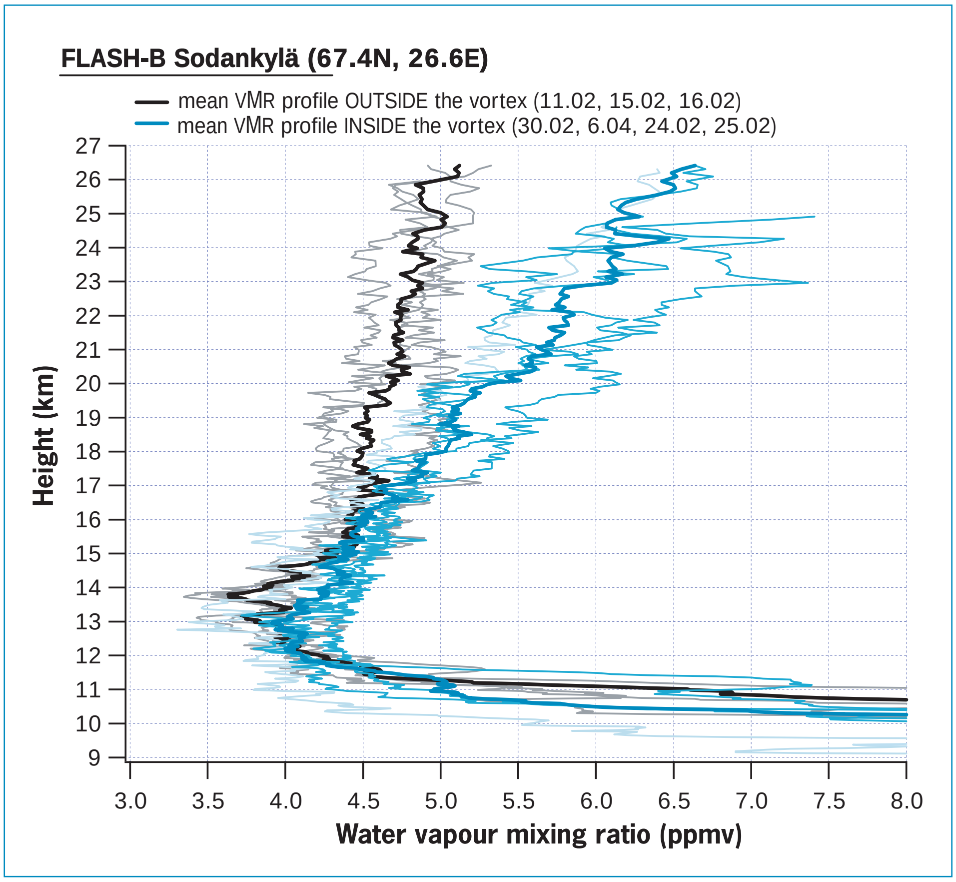


*Figure 2: Mean water vapour profiles inside (thick blue) and outside (thick black) the polar vortex calculated using the measurements made by the FLASH-B hygrometer during the LAUTLOS campaign (January 29 – February 27). Thin blue and black lines on the plot indicate the profiles obtained inside the vortex (flights on 30.01, 6.02, 24.02, 25.02.2004) and outside the vortex (flights on 11.02, 15.02, 16.02.2004).*

Figure 1 (see colour insert I) shows a comparison of relative humidity measured by different humidity sondes during balloon ascent. As seen from the plot, the sondes are generally in good agreement with each other and do not reveal significant biases. The temperature profile taken by the Vaisala RS-80 temperature sensor is also shown on the plot in order to demonstrate the sondes relative performance at different temperatures.

During the LAUTLOS campaign, 11 water vapour profiles were obtained using the

FLASH-B instrument; among them four inside the polar vortex, four on the edge of the vortex, and three outside the vortex. The polar vortex during this time period (February 2004) appeared to be relatively weak, with generally warm temperatures. Therefore it is assumed that no PSCs formed for the duration of the campaign.

Figure 2 shows the calculated mean water vapour profile both inside and outside the polar vortex using the FLASH-B measurements, which appear to be in a good agreement with those obtained by the NOAA frost point hygrometer (see Figure 3). The profiles obtained inside the vortex clearly show higher water vapour values than those taken outside the vortex. This is caused by the fact that the water vapour mixing ratio in the stratosphere increases with height through the oxidation of methane. Thus, downwelling air motion in the vortex (diabatic descent) leads to higher water vapour content inside the vortex, compared to that outside the vortex at the same altitude. The difference reaches 1.4 ppmv at 25 km altitude. Since the FLASH-B water vapour measurements are in good agreement with the other independent sensors, the calculated mean water vapour profiles can be considered as a reference for the Arctic stratosphere over Sodankylä in February with the moderate-strength vortex.

The water vapour profiles obtained by the FLASH-B and NOAA frostpoint hygrometers at the edge of polar vortex on 17 February are shown in Figure 3. The profiles indicate some filamentary structure, which can be explained by differential advection of air masses originating from inside and outside the vortex. This is confirmed by a calculation of potential vorticity from a back-trajectory analysis performed using ECMWF operational data. The laminae are captured well by both hygrometers.

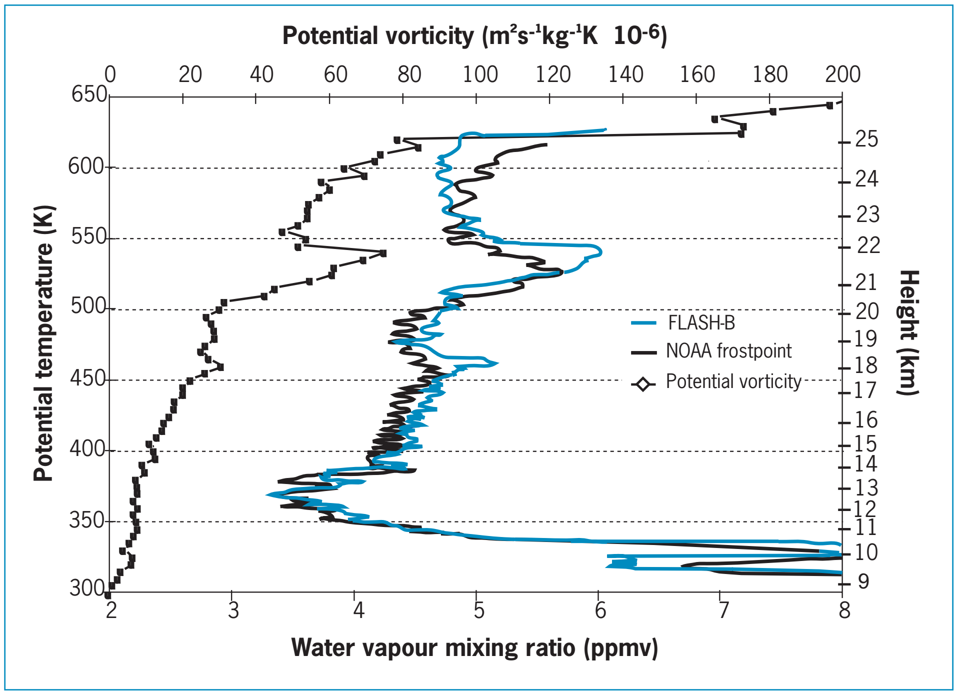


*Figure 3: Vertical water vapour profiles measured by FLASH-B (blue) and NOAA (black) hygrometers at the edge of the polar vortex on 17 February, and the potential vorticity profile obtained from back-trajectory analysis. Potential temperature and estimated height are shown on the vertical axis. Laminae at approximately 22 km and 18 km seen in the potential vorticity profile are also captured by both the FLASH-B and NOAA frostpoint hygrometers.*

The stratospheric water vapour mixing ratio inside, outside, and at the edge of the polar vortex has been accurately measured. The large dry bias in Arctic stratospheric water vapour typically found in models implies the need for future regular measurements of water vapour in the polar stratosphere to allow for validation and improvement of climate models. The numerous comparisons are now in progress and results are due by May 2005. All data are available in FMIARC ftp-server for the participants. The LAUTLOS database will be free for all interested scientists after May 31, 2005.

It was proposed that the 2nd LAUTLOS data evaluation meeting be held in Finland from August 29 to September 03, 2005.

# Tropical Troposphere-to-Stratosphere Transport: A Lagrangian Perspective

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## Introduction

A growing body of observations and theoretical considerations suggests that the transition from the troposphere to the stratosphere is gradual, rather than a relatively sharp discontinuity at the tropopause. In the tropics, it has been suggested that a tropical tropopause layer (TTL) spans the transition region from the convectively dominated overturning circulation of the Hadley cell to the region of slow upwelling (primarily wave-driven) of the lower stratospheric Brewer-Dobson circulation. Also see the article by Ian Folkins in this issue.

The transport of chemical tracers through the TTL is an important part of the global climate system. Air enters the stratospheric 'overworld' primarily in the tropics (Holton *et al.*,1995), and therefore processes in the TTL play a significant role in determining timescales for transport into the stratosphere of chemical species with tropospheric sources. For various reasons, among them the problem that current global-scale models cannot resolve individual convective cells penetrating the TTL, the modelling of chemical tracer transport in this region provides a formidable challenge. We have recently carried out a number of experiments with trajectory calculations based on 3-dimensional wind fields from global-scale atmospheric models and meteorological analysis datasets in order to characterize the circulation in the TTL as represented in these models and datasets. In particular, we focused on identifying typical pathways of tropical troposphere-to-stratosphere transport (TST), and their implications for stratospheric water vapour concentrations.

Here, we provide an overview of some of the results reported by Hatsushika and Yamazaki (2003) (henceforth HY03), Bonazzola and Haynes (2004) (henceforth BH04), Fueglistaler *et al.* (2004, 2005) (henceforth FWP04, FBHP05) and Fueglistaler and Haynes (2005) (henceforth FH05). HY03 used an atmospheric general circulation model with a resolution of T42L50, while the other experiments used data from the European Center for Medium-range Weather Forecasts (ECMWF). BH04 and FWP04 used operational analyses with resolutions of T106/L31 and T106/L50, and T511L60, respectively, and FBHP05 and FH05 used the ERA-40 reanalysis data with resolution T159/L60.



## Circulation in the TTL and tropical TST

An intriguing result found in all experiments is that TST-trajectories enter the TTL predominantly over the western Pacific. This reflects the models' enhanced vertical transport due to convection over the western Pacific warm pool, and may have important implications, particularly for the transport of short-lived species into the stratosphere. We note, however, that this result needs careful interpretation given that none of the models explicitly resolves individual convective cells. An integration of these results from a global perspective with results obtained from mesoscale cloud resolving simulations is therefore an important and challenging task. It remains to be seen whether the latter lead to substantial changes of the picture obtained from the global-scale models.

Within the TTL itself, horizontal advection by the upper-level monsoonal anticyclones and the equatorial easterlies plays an important role in determining the path of TST trajectories. A small fraction of TST trajectories could be identified as travelling with the northern subtropical jet around the globe (the southern subtropical jet appears to be more detached from the circulation governing TST). Typically, TST trajectories were found to travel several thousand kilometers horizontally from their point of entry into the TTL to the point where they encountered their minimum temperatures. Experiments to determine average residence times in the TTL show a peak of about 13 days for a change of 10 K in potential temperature at 360 K, which is approximately the level of zero net radiative heating derived from radiative transfer calculations (Gettelman *et al.*, 2004).

The results of BH04 and FWP04 indicate that during boreal summer the distance travelled by the particles is shorter than during boreal winter, but that their ascent rate is faster during boreal winter, as expected from the annual cycle of the strength of the stratospheric Brewer-Dobson circulation. These results support the notion of Holton and Gettelman (2001) that horizontal advection is a crucial aspect of tropical TST. Our results, however, also emphasize the role of zonal inhomogeneities of vertical transport. Firstly, the convection over the western Pacific warm pool is the predominant source for air in the TTL, and eventually the stratosphere. Secondly, within the TTL this region is characterized by enhanced upwelling and an associated cold temperature anomaly. Matsuno (1966) and Gill (1980) showed that these planetary-scale features can be understood as a stationary, planetary-scale wave response due to localized heating in the troposphere as a result of convection over the western Pacific warm pool.

Figure 1 (see colour insert I) provides a schematic of tropical TST, emphasizing the role of the Western Pacific warm pool region and the upper-level anticyclones. During boreal summer, the Indian/Southeast-Asian monsoon is found to play a similar role, although it is much less symmetric around the equator as the convection, which plays an important role in determining the flux into the TTL and the pattern of horizontal circulation within the TTL itself, is shifted significantly to the north of the equator.

Experiments addressing interannual variability show that the largest modifications of the circulation patterns of tropical TST occur due to El-Nino/Southern Oscillation (ENSO) events. In particular, during El-Nino, convection is more homogeneous over the entire tropical Pacific, which in turn induces a homogenization of temperatures and circulation within the TTL.

Figure 2 (see colour insert II) shows the horizontal wind and temperature fields at 90 hPa from ERA-40 data during boreal winter for a strong La-Nina, a strong El-Nino situation, and the climatological mean state for the period 1979-2001. In addition, the figure shows the spatial density distribution where trajectories rising from the troposphere to the stratosphere are found to enter the TTL (red contour lines).

## Stratospheric water vapour

Although it has been widely accepted since the seminal work of Brewer (1949) that the extremely low temperatures at the tropical tropopause constrain the water vapour flux into the stratosphere to a few parts per million by volume (ppmv), the details of the dehydration processes of tropical TST remain controversial. We therefore pursue two goals with the trajectory calculations. Firstly, we want to quantify the relevance of the previously discussed large-scale spatial structure of temperature and circulation in the TTL for water vapour mixing ratio of air entering the stratosphere ($H_2O$). Secondly, we want to quantify the degree to which observations of stratospheric water vapour concentrations can be explained by model calculations that greatly simplify cloud microphysics and mesoscale processes, but take into account the 3-dimensional temperature history of tropical TST as resolved by global scale models. Of particular interest in this context is the minimum temperature experienced by an air parcel as it ascends from the troposphere to the stratosphere. FBHP05 termed this point the 'Lagrangian cold point', to be contrasted with the traditional cold point tropopause.

In all experiments we found that the previously discussed circulation in the TTL is such that tropical TST trajectories efficiently sample the regions of lowest temperatures. Correspondingly, the spatial density distribution of the Lagrangian cold points as shown in Figure 2 (black contour lines) is highest in regions of lowest temperatures. These zonal temperature anomalies substantially lower $H_2O$ compared to what might be expected from zonal mean tropical tropopause temperatures. Therefore we conclude that the large-scale 4-dimensional structure of temperatures and circulation in the TTL is crucial for understanding $H_2O$. FBHP05 showed that predictions of $H_2O$ based on the saturation mixing ratio at the Lagrangian cold point of TST-trajectories agree very well with a broad range of observations, with a remaining moist bias of about 0.2 ppmv, or about 5% of $H_2O$.

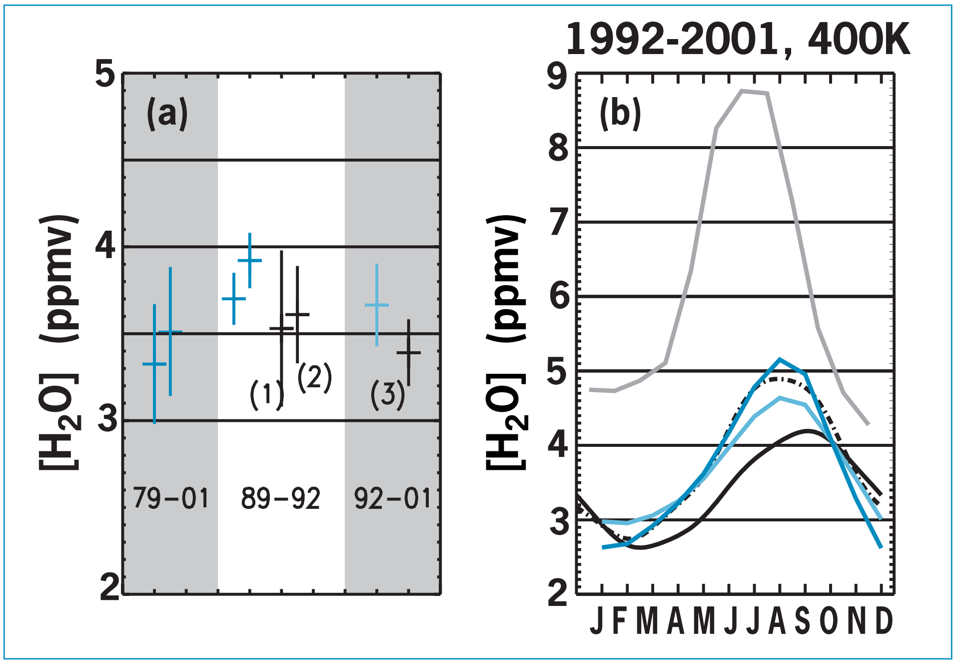


*Figure 3: Comparison of model predictions and observations of stratospheric water vapour, taken from FBHP05. (a): Model predictions for concentration of water vapour $[H_2O]_e$ (blue, with (left) and without (right) taking into account the seasonal cycle of the strength of the stratospheric Brewer-Dobson circulation) and tropical (30°S-30°N) mean water vapour mixing ratio at 400 K potential temperature (cyan) for the periods 1979-2001, 1989-1992 and 1992-2001. Observations (black) of $[H_2O]_e$ are from Michelsen et al. (2000) (1) and Engel et al. (1996) (2), and (3) is the tropical mean water vapour mixing ratio at 400 K from HALOE. (b): Climatological mean annual cycle of model prediction for $[H_2O]_e$ (blue) and tropical mean at 400 K (cyan), HALOE at 380 K (black, dash-dot) and at 400 K (black, solid), and, for reference, the saturation mixing ratio of zonal mean (10°S-10°N) cold point tropopause temperatures based on ERA-40 data.*

Figure 3 reproduces some results shown by FBHP05 for mean entry mixing ratios of $H_2O$ over selected periods, and the climatological mean annual cycle for the period 1992-2002. Note, for example, that the annual cycle of the model calculations based on the 3-dimensional temperature history of TST (Figure 3, cyan line) significantly improves the agreement with observations (Figure 3, black line) compared with a prediction based on tropical zonal mean cold point temperatures (Figure 3b, grey line). Remaining differences are discussed in detail by FBHP05, note for example that the obvious phase shift of approximately 1 month between the model prediction (cyan) and observation (black) is likely due to known problems of the stratospheric circulation in ERA-40 data.

FH05 further showed that for the period 1995-2002 their model predictions can explain observed interannual variability of water vapour concentrations in the tropical lower stratosphere to within measurement uncertainties. Largest modulations of $H_2O$ are found to result from the temperature perturbation around the tropical tropopause due to the Quasi-Biennial Oscillation (QBO) and due to strong El-Nino situations. Based on simulations of idealized ENSO situations, HY03 predicted that La-Nina situations should be drier than El-Nino. This was partly confirmed by FH05 in their reconstruction of $H_2O$ for the period 1979-2001. However, the interference of ENSO with other processes that affect tropical tropopause temperatures, most notably the QBO, makes it difficult to classify unambiguously the effect of ENSO, particularly La-Nina situations, on $H_2O$.

## Summary

The good agreement of model predictions with observations of stratospheric water vapour suggests that the trajectory calculations, and the assumption of dehydration to the saturation mixing ratio of the Lagrangian cold point, capture the main processes controlling $H_2O$. Thus, the synoptic-scale temperature history of tropical TST apparently can explain $H_2O$ very well, and smaller scale effects such as overshooting convection apparently need not be invoked at first order. Consequently, it appears that current global-scale models may do a better job in representing tropical TST than might have been expected. That notwithstanding, a comprehensive theory explaining the spatio-temporal circulation and temperature structure of the TTL that encompasses all scales, from individual convective cells to the planetary-scale stationary wave pattern, and the role of the stratospheric Brewer-Dobson circulation, remains an important and ambitious task.



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# Temperatures, Transport, and Chemistry in the TTL

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The Tropical Tropopause Layer (TTL) has been recently introduced (Highwood and Hoskins, 1998, Folkins *et al.*, 1999) as a layer whose properties are intermediate between the troposphere and stratosphere. If examined with a fine enough spatial scale, any atmospheric property will vary continuously in going from one atmospheric layer to the next. One could therefore invoke the existence of a transition layer between any two adjacent atmospheric layers. This article will attempt to explain what is unique about the TTL, and why it provides conceptual advantages in thinking about some atmospheric science problems.

The issue of how to define the TTL is not settled. Here, the base of the TTL will be defined as the Level of Zero clear sky radiative Heating (LZH), which occurs near 15 km. The transition from radiative cooling (below 15 km) to radiative heating (above 15 km) is driven by the combination of a rapid decrease in water vapour mixing rations (such that longwave cooling from water vapour is negligible above 15 km), and also by the onset of extremely cold temperatures, which suppress longwave emission from $CO_2$ and $O_3$. From radiative considerations, an air parcel which detrains from a deep convective clouds above the LZH will rise upward across isentropic surfaces into the stratosphere. This definition of the base of the TTL therefore seems to be the definition most relevant to stratosphere-troposphere exchange (STE). Air parcels that detrain into the TTL have some probability of influencing the chemical composition of the stratosphere, while those which detrain below the TTL have very little. It should be kept in mind, however, that air parcels moving horizontally near 15 km will roughly follow an isentropic surface that can undulate above and below 15 km, so that the base of the TTL is not a material surface. Perhaps the most reasonable definition of the base of the TTL is the height at which an air parcel detraining from a deep convective cloud first attains a 50% likelihood of ascending into the stratosphere. This height probably occurs near the LZH because radiative heating is the most dominant diabatic process in the clear sky atmosphere.

It is more difficult to define the top of the TTL. A useful conceptual definition is that it is the height at which the upward convective mass flux becomes small in comparison to the Brewer Dobson mass flux. That is, that the convective outflow that feeds the base of the Brewer Dobson circulation has essentially become exhausted. Unfortunately, it is intrinsically difficult to diagnose the high altitude tail of the convective detrainment profile from observations. Measurements of ozone and other chemical species suggest that undiluted convective outflow can occur as high as 17 km (Folkins *et al.*, 2002a, Tuck *et al.*, 2004), which would place the top of the TTL near the climatological height of the cold point tropopause. This is also a convenient definition for TTL dehydration. However, deep convective overshooting through the tropical tropopause, followed by mixing with ambient stratospheric air, provides a mechanism for convective detrainment above 17 km whose existence is difficult to detect from chemical tracers.

It has been argued that the input of significant amounts of overshooting tropospheric air above the tropical tropopause is unlikely, since it would undermine the seasonal variation in $CO_2$ propagating upward from the tropopause (Boering *et al.*, 1995). Model simulations suggest, however, that this is not true in all circumstances (Sherwood and Dessler, 2003).

As defined here, the TTL is a consequence of the geometry of the flow in the upper tropical troposphere. In order to feed the Brewer-Dobson circulation, there must be some convective detrainment above the altitude at which the background mean flow changes direction. Since the Brewer-Dobson circulation is about 100 times smaller than the Hadley circulation, one would anticipate that roughly 1% of the mass flux from tropical deep convection detrains into the TTL.

Figure 1 gives an overview of the mass flux divergences in the upper tropical troposphere associated with convective outflow ($\delta_c$), radiative cooling ($\delta_r$), and evaporative cooling ($\delta_e$). These were obtained using a simple one-dimensional model of the tropical atmosphere constrained by observed temperature and moisture profiles (Folkins and Martin, 2005). To first order, radiative convergence balances convective divergence. In other words, convective outflow is the mass source required to supply a downwardly increasing subsiding radiative mass flux (or to supply an upwardly increasing ascending radiative mass flux in the TTL).

The magnitude of the evaporative driven downward mass flux increases below 13 km, giving rise to a convergence, which partially balances the divergence from convection. Due to the extremely cold temperatures, saturated water vapour mixing ratios in the TTL are very low, typically less that 10 ppmv. At such low mixing ratios, heat releases associated with changes in phase of water are typically smaller than those due to radiative heating or cooling, so that from a thermodynamic perspective, air parcels in the TTL can be effectively treated as dry. In the TTL, one should not ordinarily have to worry about the role of evaporating ice crystal in driving vertical motions.

Figure 1 shows that the rate at which deep convection injects air into the upper tropical troposphere reaches a maximum of 0.4/day near 12.5 km. The convective replacement time $\tau_r$ can be defined as $\tau_r=1/\delta_c$. This replacement time varies from 2.5 days at the peak of deep outflow mode, to about two weeks at the base of the TTL. Within the TTL, estimations of $\delta_c$ and $\delta_r$ using simple models become highly uncertain. This is due to a variety of reasons. The divergences shown in Figure 1 were calculated by assuming that the mass budget of the tropics (here defined as 20°S-20°N) could be treated in isolation from the extratropics. This assumption is probably valid at most heights. In the TTL, however, it is likely that the rate of mass transfer between the tropics and extratropics is comparable with the convective divergence. The curve labelled $\delta_{ex}$ in Figure 1 is an estimate of the rate at which tropical air is exported from the tropics to the extratropics, using assimilated meteorological data from the Goddard Earth Observing System (GEOS) of the NASA Data Assimilation Office. In principle, diagnostic approaches to estimating the strength of the convective divergence in the TTL should take this mass transport pathway into account. In addition, the radiative divergence shown in Figure 1 was calculated by assuming clear sky conditions. Clear sky heating rates in the TTL are very small. The relative importance of cloud radiative effects can therefore be expected to be much higher in the TTL than in the rest of the tropical troposphere.

The ideas behind the TTL are not entirely new. Until recently, however, there had been a widely held implicit assumption that much of the air rising in deep convective clouds detrained near, or just below, the tropical tropopause. This view partly arose from the fact that stratiform anvil clouds from deep convection frequently extend to the cold point tropopause (~17 km). Inferring mass outflow from the presence of ice crystals is, however, problematic. Ice crystals may arise from vertical motions associated with gravity waves generated by convection, rather than by the updrafts themselves. In addition, stratiform anvils can extend from 10 km to 17 km. There is no reason to assume a priori that outflow is preferentially distributed near cloud top.

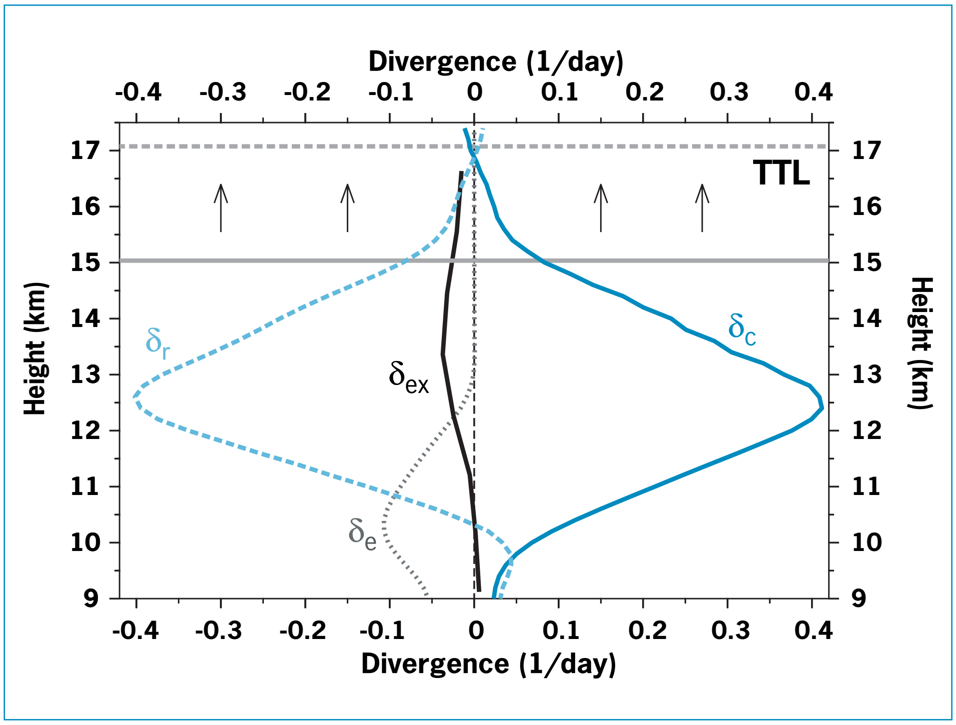


*Figure 1: Tropical mean (20°S – 20°N) mass flux divergences calculated from a one-dimensional model of the tropical atmosphere (Folkins and Martin, 2005). $\delta_c$ refers to the convective divergence, $\delta_r$ to the radiative divergence, $\delta_e$ to the evaporative divergence, and $\delta_{ex}$ to the divergence associated with export of tropical air into the extratropics. The vertical arrows indicate that the clear sky mass flux is upward above 15 km. The top of the TTL, near 17 km, has been drawn with a dashed line to indicate that its height is uncertain.*



Temperature profiles from radiosondes also frequently give the impression of a strong convective influence extending up to the cold point tropopause. Deep convection is often associated with a cold and well-defined cold point tropopause, with layers near the tropopause where the lapse rate approaches the dry adiabatic value. The strong influence of deep convection of TTL temperatures does not necessarily imply rapid mass outflow rates in the TTL, however, since the weakness of radiative heating rates in the TTL implies long radiative damping timescales (Hartmann *et al.*, 2001). Temperature anomalies in the TTL arising from deep convection can be therefore expected to be very persistent.

The notion that most of the air transported upward in deep convective clouds reaches the tropical tropopause also gives rise to an apparent thermodynamic paradox. In the absence of mixing, the level of neutral buoyancy of an air parcel rising upward inside a convective updraft occurs at the height at which the equivalent potential temperature of the air parcel is equal to the potential temperature of the background atmosphere. (This statement makes a number of assumptions, among them that water vapour concentrations are sufficiently low that their effect on density can be ignored, and that the presence of condensate does not have an appreciable effect on the density of an air parcel.) While the potential temperature at the cold point tropopause is rarely lower than 365 K, near surface air parcels with $\theta_e > 365$ K are extremely rare (Folkins and Martin, 2005). On thermodynamic grounds, one would therefore also expect convective detrainment near the cold point tropopause to be extremely rare.

There have been a number of attempts to reconcile the idea of significant convective detrainment near the tropical tropopause with its apparent thermodynamic implausibility. One mechanism that would allow air parcels to detrain at a higher potential temperature surface is convective overshooting and irreversible mixing. It has also been pointed out that the potential temperature of the cold point tropopause roughly corresponds to the saturated equivalent potential temperature of the warmest tropical sea surface temperatures, so that air parcels with $\theta_e \sim 365$ K can in fact be generated near the surface (Chimonas and Rossi, 1987). Another suggestion is that the freezing of lofted ice could act as a source of heating that would allow air parcels to detrain near the tropical tropopause (Zipser, 2003). Such estimates are, however, quite sensitive to assumptions on the amount of lofted ice, and the efficiency with which rising air parcels retain this additional heat.

Both simple diagnostic models, as well as *in situ* chemical evidence, indicate that outflow from deep convective clouds ranges from 10 km to 17 km. This suggests that approaches that attempt to relate a mean detrainment potential temperature to a mean boundary layer $\theta_e$ may be misguided. Instead, it may be more appropriate to attempt to find some relationship between the shape of the deep convective detrainment profile in the upper troposphere with the probability distribution of boundary layer $\theta_e$ in actively convecting regions (Folkins and Martin, 2005).

The mean profile of any chemical tracer in the tropics should reflect an approximate balance between the competing tendencies of *in situ* chemistry, vertical advection, convective detrainment, as well as other possible influences such as STE. In the vicinity of the 12.5 km $\theta_c$ maximum, chemical tracers with upper tropospheric lifetime longer than a week should be maintained near their boundary layer values by the strength of deep convective detrainment. However, as the timescale for convective replacement increases toward the cold-point tropopause, the mean concentration of a chemical tracer can be expected to become increasingly determined by *in situ* chemistry.

Figure 2 (see colour insert II) shows ozone climatologies at ten tropical locations from the SHADOZ campaign (Thompson *et al.*, 2003). Most of these profiles exhibit the classic "S-shape". The minimum near 12.5 km is probably due to the strong convective outflow at this altitude. Ozone mixing ratios at locations characterized by maritime deep convection are smaller than those with continental deep convection. The two dashed lines are ozone profiles calculated from a model forced by the convective divergence profile shown in Figure 1 (Folkins and Martin, 2005). In one of the profiles, the mean ozone mixing ratio in air detraining from convective clouds is assumed to be 20 ppbv, while in the other, it is assumed to be 30 ppbv. In the model, the increase in ozone mixing ratios above 13 km is due to the increase in the convective replacement timescale, which allows ozone to approach its steady state mixing ratio. The increase in ozone above 13 km is not driven by an increase in the rate of *in situ* chemical production from $O_2$ photolysis. This source of ozone is insignificant below 16.5 km.

Figure 3 show profiles of CO in the tropics obtained from various field campaigns. Unfortunately, there are far fewer measurements of CO than $O_3$ in the tropics, especially between DC8 (~11 km) and ER-2 flight altitudes (15 km to 20 km). None of the CO profiles shown in Figure 3 contain enough data to constitute a climatology. CO is primarily destroyed in the upper troposphere by OH attack. Since the sources of CO are mainly at the surface, one would expect deep convection to maintain high concentrations of CO near the 12.5 km convective outflow maximum, and decrease toward the tropopause as the convective source weakened. This expectation is broadly consistent with the CO profiles shown in Figure 3.

One would also expect tracers, which are influenced by STE to exhibit a vertical gradient in the upper tropical troposphere. It has been noted that the onset of a decrease in CFC-11 sometimes occurs below the tropical tropopause (Tuck, 1997). This was attributed to the existence of a "standing reserve" of stratospheric air in the upper tropical troposphere. The existence of such a reservoir of stratospheric air in the TTL is consistent with Figure 1, which suggests that the timescales for sideways tropical-extratropical exchange and deep convection in the TTL may be comparable. In these cases, the vertical gradient in the tracer concentration is not necessarily caused by a change in the magnitude of STE with height, but by a decrease in convective outflow, which gives any stratospheric input a longer chemical signature.

The base of the TTL denotes a change in the nature of the clear sky water vapour budget. In most of the clear sky tropical troposphere, the occurrence of supersaturation with respect to ice is inhibited by large-scale descent. In the TTL, however, air parcels experience colder temperatures as they ascend toward the cold point tropopause. Supersaturation with respect to ice should be quite frequent, provided ice deposition nuclei are sufficiently rare. Measurements have, in fact, suggested an increase in the frequency of clear sky supersaturation at the base of the TTL (Folkins *et al.*, 2002b). These measurements would tend to support the view that dehydration occurs throughout the TTL rather than simply in the vicinity of the cold point tropopause. Unambiguous detection of clear sky supersaturation with respect to ice remains a significant experimental challenge, however, because in regions where the relative humidity approaches 100%, it is possible to observe supersaturation with respect to ice simply due to the existence of random (and possibly bias) errors in measurements of water vapour mixing ratio and temperature.

The concept of the TTL was partly motivated by modelling studies indicating that convective control of tropical temperatures did not extend as high as the cold point tropopause (Thuburn and Craig, 2002). It is clear that the base of the TTL is associated with a change in the nature of radiative-convective equilibrium. Figure 4 (see colour insert III) shows the difference in December to February (DJF) temperatures at 14 tropical radiosondes locations from a DJF climatology obtained by averaging over all 14 locations. Below 14 -15 km, locations characterized by persistent deep convection tend to be anomalously warm, while those with less convection tend to be anomalously cold.

One of the ways in which convection influences the large scale temperature profile is by the emission of gravity waves. These gravity waves give rise to irreversible vertical motions in the background atmosphere, which diminish the contrast in density between the atmosphere and the convective plumes 25

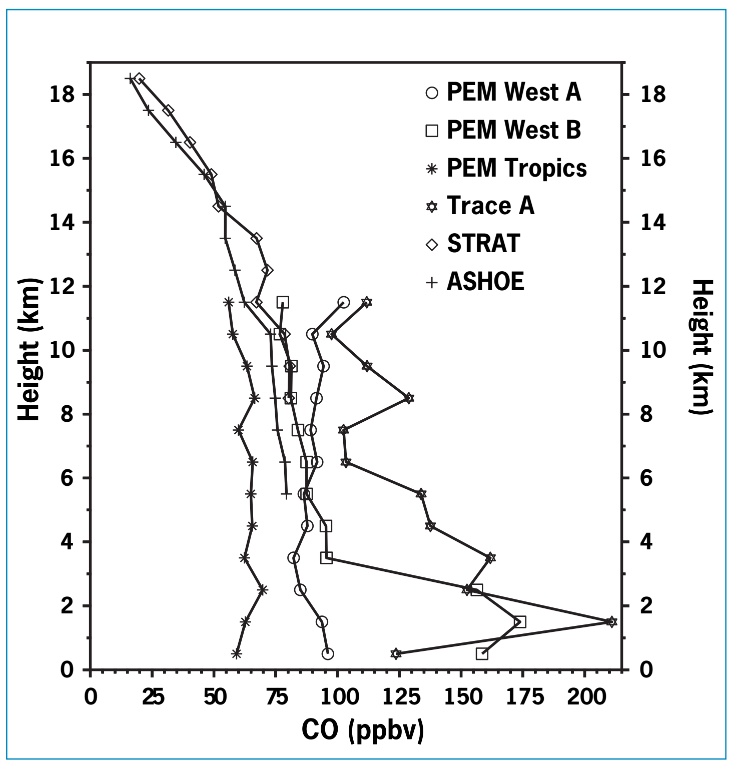


*Figure 3: CO climatologies obtained by averaging over CO measurements (15°S – 15°N) from a variety of aircraft field campaigns.*

(Bretherton and Smolarkiewicz, 1989). A warm positively buoyant plume would be expected to give rise to downward motions in the background atmosphere, while a negatively buoyant plume (overshooting updraft or downdraft) would give rise to upward motions. The association between convection and warm anomalies below the TTL would suggest that convective plumes are on average positively buoyant below 15 km, while the association between convection and cold anomalies in the TTL would suggest that convective plumes in this region are negatively buoyant.

The coincidence of the LZH with the height at which the relationship between convective frequency and temperature changes sign is coincident with the view that deep convection exerts a dominant control on temperature at these altitudes. Below the TTL, the induced downward motions from deep convection increase tropical temperatures away from radiative equilibrium, and give rise to radiative cooling, which within the TTL, the induced upward motions from deep convection decrease tropical temperatures away from radiative equilibrium, and give rise to radiative heating.

There have been other explanations put forward for the association between cold temperatures and deep convection in the TTL. One possibility is that it is due to the deep convection injection of cold, negatively buoyant air which irreversibly mixes with the ambient warmer air (Sherwood *et al.*, 2003, Kuang and Bretherton, 2004).

In the lower tropical stratosphere, upwelling is strongest during the December to Febraury (DJF) season and weakest during the June to August (JJA) season. This is associated with a seasonal temperature cycle, in which temperatures are coldest during the DJF season and warmest during the JJA season. Figure 5 (see colour insert III) shows the difference between the JJA and DJF seasonal temperature climatologies at 14 radiosonde locations. The amplitude of the lower stratospheric seasonal cycle peaks near 18 km. At most locations, the onset of this seasonal cycle occurs near 15 km, so that the stratospheric seasonal cycle extends to the base of the TTL. Below 15 km, most tropical locations are warmer during the DJF season.

Predictions of the future evolution of tropical cold point temperatures are made difficult both by uncertainty in the mechanism by which tropical deep convection influences TTL temperatures, and also because the relative roles played by the Hadley and Brewer-Dobson circulations in controlling TTL temperatures have not yet been established. It is important to determine the nature of this control, however, because the Hadley and Brewer Dobson circulations can be expected to change in response to future changes in $CO_2$. By influencing tropical cold point temperatures, these changes can be expected to influence stratospheric water vapour.

Ozone plays an important role in the radiative budget of the TTL. Its climatological profile will be influenced by changes in both the shape of the convective detrainment profile, and by changes in stratospheric upwelling. It would therefore be desirable to investigate the future evolution of TTL temperatures using models in which the dynamics is self-consistently coupled with radiation and chemistry.

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# Stratosphere-Troposphere Dynamical Coupling

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## Introduction

It is widely accepted that the troposphere has a strong dynamical effect on the stratosphere, primarily through the upward propagation of waves, both low-frequency large-scale Rossby waves ('planetary waves') and high-frequency inertia-gravity waves. Understanding of this effect is based on simple theories of wave propagation (including the well-known Charney-Drazin criterion for vertical Rossby wave propagation), experiments in many different types of numerical models, plus observational indicators such as differences in stratospheric circulation between summer and winter, and between the hemispheres. An important aspect of this effect is that in the stratosphere there is a two-way interaction between waves and mean flow. Breaking or dissipating waves exert a systematic mean force $\mathcal{G}$ that changes the mean flow. The mean flow, on the other hand, affects the propagation, breaking and dissipation of waves and hence itself affects $\mathcal{G}$. The two-way interaction can lead, for example, to sensitivity to initial conditions, or to internal dynamical variability. Yoden *et al.* (2002) and Haynes (2005) review some these issues.

Nonetheless, it is still not yet the case that our understanding of the dynamical effect of the troposphere on the stratosphere can be said to be complete. For example, for events such as stratospheric sudden warmings (including the unexpected Southern Hemisphere sudden warming of September 2002), in which the stratospheric circulation is highly perturbed, it does not seem possible to identify unambiguously anomalously large tropospheric wave forcing as the cause. (See for example papers in the recent special issue of Journal of Atmospheric Sciences, volume 62, number 3.) One of the reasons may be that the view of the troposphere having a one-way dynamical effect on the stratosphere is seriously limited. There are plenty of reasons why the coupling should be two-way. The large-scale extratropical dynamics of the atmosphere is inherently non-local in both horizontal and vertical. Changes in the stratosphere must inevitably affect the troposphere and vice versa – the key question, of course, is how much?

## Stratospheric cause and tropospheric effect?

The dynamical effect of the stratosphere on the troposphere is now a major research activity. (See previous SPARC newsletter articles by Gillett *et al.* 2003, Hartmann 2004.) Whilst there has for some time been significant evidence from numerical model studies, early examples being Boville (1984) and Kodera *et al.* (1990), that imposed perturbations to the stratosphere lead to changes in the tropospheric circulation and that the mechanism for communication of these changes is dynamical, much of the recent heightened interest in this topic has arisen from studies of the Northern Hemisphere (NH) and Southern Hemisphere (SH) 'annular modes' (hereafter NAM and SAM). Methods such as EOF analysis have identified these as dominant signals in variability in both troposphere (e.g. Thompson and Wallace 2000 and references therein) and stratosphere (e.g. Baldwin and Dunkerton 1999). In both troposphere and stratosphere, the annular modes are associated with variation in the strength and position of the jet (the mid-latitude jet in the troposphere, the polar night jet in the stratosphere). However the underlying dynamical character of the modes is somewhat different in the two cases.

In the troposphere the annular mode variability is believed to arise from two-way interaction between baroclinic eddies and the tropospheric mid-latitude jet (e.g. Robinson 1991, Feldstein and Lee 1998, Hartmann and Lo 1998). In the stratosphere, on the other hand, the annular variability is a manifestation of the variation in the strength of the polar-night jet, driven by the wave force $\mathcal{G}$. A significant ingredient is the variability in tropospheric wave forcing, although as noted earlier, there is an important role for two-way interaction between waves and mean flow here too. 

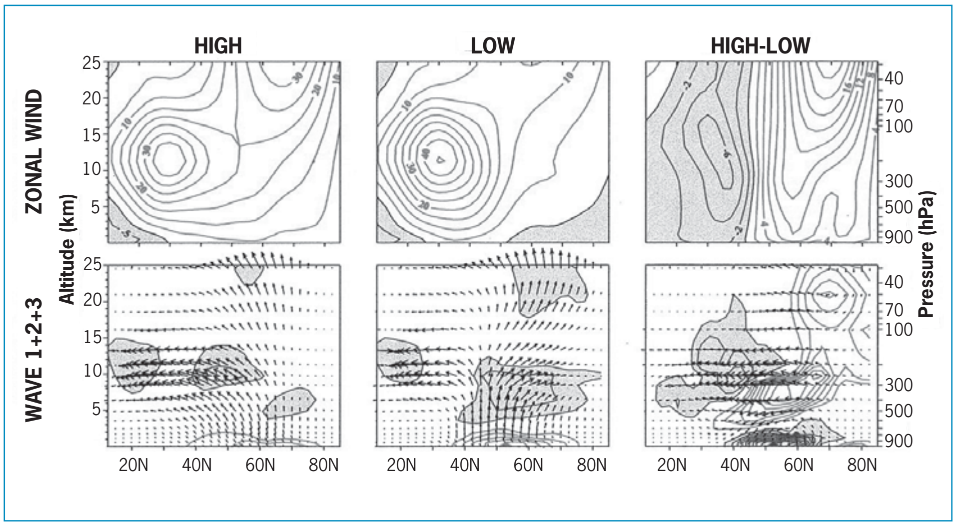


*Figure 1: (From Hartmann et al. 2000.) Composites for periods of high and low NAM index and their difference (left, centre and right, respectively) in longitudinal wind (top) and Eliassen-Palm flux, which indicates wave propagation and transport of westward momentum, and its divergence, which indicates eastward wave force. Positive contours are grey, negative contours black, with negative regions shaded. The Eliassen-Palm flux is calculated only for longitudinal wavenumbers 1, 2 and 3. In the 'high' phase wave fluxes tend to be directed equatorward within the troposphere and to converge in the subtropical troposphere, whereas in the 'low' phase wave fluxes tend to be directed upward from troposphere to stratosphere and to converge, implying an anomalous westward wave force, in the mid- and high latitude stratosphere. (Copyright (2000) National Academy of Sciences, USA. Reproduced with permission.)*

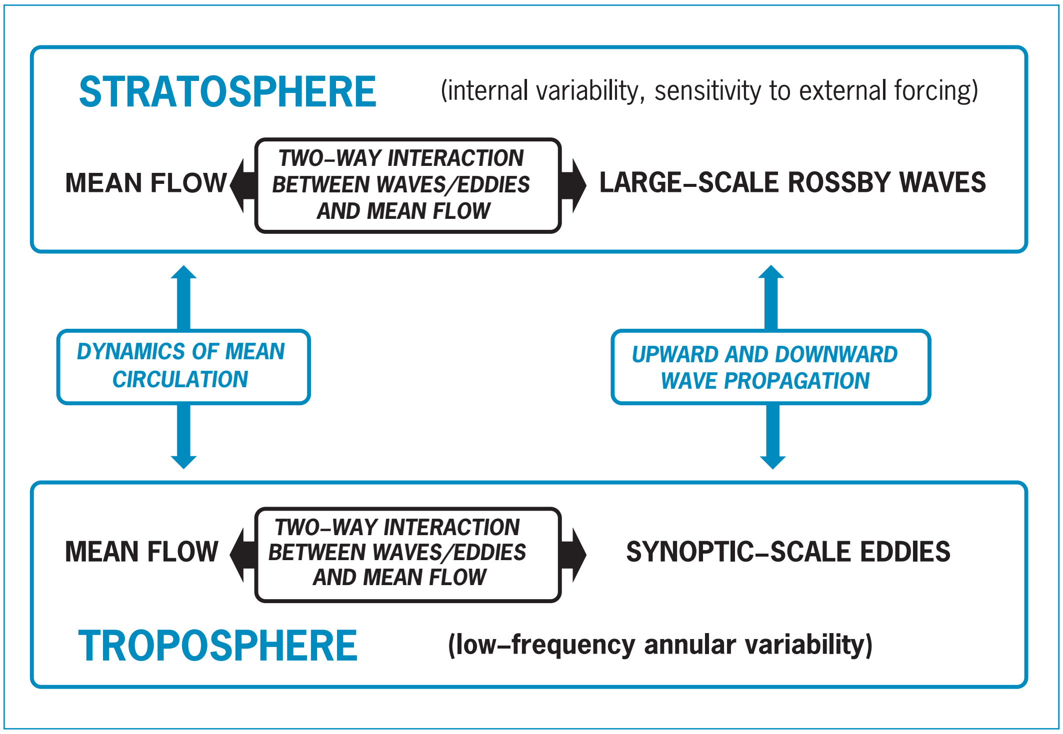


*Figure 2: Schematic diagram indicating the role of different aspects of the dynamics in the dynamical mechanisms discussed in the text. Note that 'dynamics of mean circulation' includes non-local PV inversion (or equivalently the short-time effect of the meridional circulation) and the effect of the meridional circulation on longer time scales, including the 'downward control' limit.*



What has caused widespread interest is that there seem to be strong correlations between the annular mode variability in the troposphere and that in the stratosphere. For example, in NH winter there is significant correlation between the annular mode index defined on the basis of the surface pressure field and the stratospheric circulation, so that when there is a strong pole-to-equator pressure gradient at the surface, indicating strong eastward surface flow, there are also strong eastward winds throughout the mid-latitude troposphere and in the mid-to-high latitude stratosphere (Thompson and Wallace, 2000). There is corresponding organisation in the wave fluxes indicating propagation and momentum transport, both in the troposphere (Limpasuvan and Hartmann 2000) (as is expected from the accepted mechanism for the variability) and also in the stratosphere (Hartmann *et al.* 2000) (see Figure 1) implying corresponding differences in the wave force $\mathcal{G}$ .

Baldwin and Dunkerton (1999, 2001) have shown that the vertical structure of NAM variation in NH winter typically shows a downward propagation from middle stratosphere to troposphere and the corresponding figure from their 2001 paper is now (quite justifiably) de rigeur in any scientific talk in this area. Does this picture imply a direct effect, with some delay, of anomalies in the stratospheric circulation in mid-stratosphere on the troposphere?

The answer is 'not necessarily'. The equatorial quasi-biennial oscillation (QBO) is an educational example. The QBO is manifested by downward propagating anomalies in zonal wind, however it was pointed out by Plumb (1977) that, at least in simple dynamical models of the QBO, there is in fact no downward propagation of information. The evolution of the flow below some given level is completely independent of any changes that are made above that level. The distinction in elementary wave theory between phase propagation on the one hand, and group propagation on the other, with only the latter unambiguously associated with propagation of information, is well-known. The downward propagation of wind anomalies in the QBO can be seen as a sort of phase propagation.

Simple dynamical models of the equatorial QBO of the type studied by Plumb (1977) are not necessarily relevant to the extratropical stratosphere, in particular because rotation is omitted and because a WKBJ approximation is used in calculating the structure of the waves for given mean flow. However Plumb and Semeniuk (2003) have shown in a simple wave-mean model of the extratropical stratosphere (that does incorporate rotation and uses a weaker approximation for the spatial structure of the waves) that forcing at low levels can give rise in the stratosphere above to downward propagating structures similar to those observed by Baldwin and Dunkerton. Therefore, such structures do not necessarily imply downward propagation of information.

There is similar uncertainty in interpretation of the correlations between anomalies of zonal velocities and wave fluxes of the type shown by Hartmann *et al.* (2000), (Figure 1), for example. While these correlations strongly suggest dynamical connections between troposphere and stratosphere involving two-way interactions between mean flow and waves, it is impossible to tell from the correlations alone whether there is an effect of the stratosphere on the troposphere, or the troposphere on the stratosphere (or indeed whether it makes sense only to think of the troposphere-stratosphere system as intrinsically coupled, with the whole idea of the effect of one component on the other as intrinsically flawed).

Notwithstanding the above, there are now many examples of simulations in numerical models where changes in the troposphere have been shown to result from imposed changes in the stratosphere. In these cases downward propagation of information is difficult to discount. Some relevant examples include imposed perturbations to the upper stratosphere (Kodera *et al.* 1990, Gray 2003) as a simple representation of solar cycle effects, changes to stratospheric radiative equilibrium temperature profiles in a simplified general circulation model (Polvani and Kushner 2002, Kushner and Polvani 2004), changes to Southern Hemisphere stratospheric ozone in a general circulation model (Gillett and Thompson 2003) and changes to stratospheric initial conditions in a numerical weather prediction model (Charlton *et al.* 2004). The results of Scott and Polvani (2004), obtained in a simplied dynamical model with a damped troposphere, show nicely that the wave flux from troposphere to stratosphere cannot be considered to be set by the tropospheric wave forcing alone and that the stratosphere can to some extent determine how much wave flux it accepts.

## Dynamical mechanisms

There are several possible dynamical mechanisms by which the stratosphere might affect the troposphere. There are two aspects to this: (a) how information might be communicated in the vertical and (b) why the tropospheric response might be larger than expected. Figure 2 is a schematic diagram indicating some of the points discussed below.

Taking (a) first, one mechanism is via the non-local inversion operator that determines quantities such as velocity or temperature from the potential vorticity (PV) distribution. (The non-locality arises from the rapid propagation of inertio-gravity waves required to maintain a state of geostrophic and hydrostatic balance.) Any change in the PV distribution in the lower stratosphere will inevitably give rise to changes in wind and temperature in the troposphere. Hartley *et al.* (1998), Ambaum and Hoskins (2002) and Black (2002) show explicit calculations to illustrate this point. This might account for some of the lower part of the height-time plots shown by Baldwin and Dunkerton (1999, 2001). If the restriction is made to zonal mean fields, then the vertical non-locality of PV inversion is precisely equivalent to a statement that a wave-induced force localised to the stratosphere will, through the instantaneous induced meridional circulation, give rise to an acceleration in the troposphere below and many of the papers on this topic have chosen the latter description. An important refinement is that on longer timescales, in the presence of radiative damping, the meridional circulation tends to be narrower and deeper below (Haynes *et al.* 1991, Holton *et al.* 1995) (tending to a 'downward control' response in the steady state limit) potentially allowing an enhanced tropospheric response to a stratospheric wave force.

A second distinct mechanism for communication in the vertical is via Rossby wave propagation. The propagation of Rossby waves out of the troposphere might be sensitive to by variation in the 'refractive' properties of the lower stratospheric flow (Hartmann *et al.*, 2000, see also Limpasuvan and Hartman 2000), or indeed there might be downward reflection of Rossby waves from higher in the stratosphere (e.g. Perlwitz and Harnik 2003, 2004).

A third mechanism for downward propagation of information might be through a two-way interaction between waves and mean flow. The results of Plumb and Semeniuk (2003) discussed earlier do not completely rule out the possibility that there can be real downward propagation of information through this interaction. Recent investigation (Steven Hardiman, personal communication) using the same model as Plumb and Semeniuk has shown no evidence of any distinct downward propagation of the effect of imposed upper level stratospheric perturbations through a such a mechanism when stratospheric Rossby wave amplitudes are weak. When stratospheric wave amplitudes are large, however, the dynamics of waves and mean flow is highly nonlinear and imposed upper level perturbations can have significant effects at lower levels.

Turning to (b), the two-way interaction between baroclinic waves and mean flow in the troposphere that gives rise to annular mode variability may also serve as an 'amplifier' for external forcing (including dynamical forcing from the stratosphere) (e.g. Hartmann *et al.* 2000). This may allow the tropospheric response to be significantly larger than might be expected from zonal-mean dynamics, for example.

Putting (a) and (b) together Song and Robinson (2004) report numerical experiments showing the effect of imposed stratospheric perturbations on the troposphere and the effect arises through a downward penetrating response in the mean circulation communicates the effect of stratospheric wave forcing to the troposphere, where the response is amplified by the eddy (i.e. wave) feedbacks associated with annular variability. They name their mechanism 'downward control with eddy feedback'. However, having proposed this mechanism (which to this author at least seemed interesting and plausible) Song and Robinson then present further numerical experiments that show that it cannot be the full explanation for the effect of stratospheric perturbations on tropospheric circulation that they observe. In particular they show that the effect on the tropospheric circulation is much weaker when the Rossby waves in the stratosphere are artificially damped. The conclusion is therefore that the Rossby waves play a significant role in downward communication of information.

Other strong evidence that the stratosphere plays an active, rather than a passive, role in tropospheric variations associated with the NAM comes from observed and modelled changes on the time scale of the NAM variations. Baldwin *et al.* (2003) have shown that this time scale is significantly longer at times of the year (NH winter, SH spring) when there is strong Rossby wave propagation into the stratosphere. Correspondingly, artificial suppression of stratospheric variability in model simulations reduces the time scale of the tropospheric NAM (Norton 2003). The dynamical mechanism here is likely to be that, when there is significant flux of Rossby waves into the stratosphere, the flow in the stratosphere acts as an integrator (and hence low-pass-filter) of this flux (or rather the variability in this flux), since stratospheric damping times are relatively long. Any stratospheric effect on the troposphere will therefore tend to increase the time scales of the variability. When there is little flux of Rossby waves into the stratosphere (in summer, or in SH midwinter) the effect is absent.

## Conclusion

The possibility of significant stratospheric effects on the troposphere has implications for many aspects of month-to-month and year-to-year variability and systematic change in the tropospheric circulation. It suggests possible mechanisms for explaining apparent signals in the tropospheric circulation of the solar cycle, inputs of volcanic aerosol to the stratosphere, and the equatorial QBO. It also strengthens the link between between possible climate change in the troposphere and changes in the stratosphere, due to ozone depletion or increasing greenhouse gases. Whilst care is needed in interpreting observations as implying real downward influence or downward propagation, there is convincing evidence from numerical model simulations that changes in the stratosphere can sometimes lead to significant effects in the troposphere.

The dynamical mechanisms required to explain these effects are still being investigated. It seems clear that the two-way interaction between synoptic-scale waves/eddies and mean flow in the troposphere that gives rise to 'annular variability' is important, notwithstanding the fact that the nature of annular variability is still being vigorously debated (e.g. Cash *et al*. 2005). The two-way interaction between waves and mean flow in the stratosphere also seems likely to be relevant, though the relative roles of waves, mean flow, and coupling between them is not yet clear.

Further clarification of these dynamical mechanisms and their role in the real atmosphere will most likely come from careful studies in a sequence of numerical models. Much has already been learned from simplified models that include the large-scale dynamics plus highly simplified representations of processes such as radiation, and more work with these models, as well as with sophisticated general circulation models, is surely needed to resolve some of the remaining uncertainties.

An important general point that has been revived by the recent interest in troposphere-stratosphere coupling is that, whether or not one is interested in the dynamical details, the fact is that the coupled system exhibits strong (dynamical) internal variability and that any attempt to explore correlations between one part of the atmosphere and the other, or to predict future changes, needs to take this into account. Such studies therefore need to use long integrations or large ensembles, requiring significant computational resources. There is understandable pressure to make as rapid progress as possible with coupled chemical-climate simulations, which also requires significant computational resources, but there is still much to learn about the variability and predictability of the coupled physics and dynamics of troposphere-stratosphere system without coupling to chemistry, and this should not be overlooked.



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# Assimilation of Stratospheric Meteorological and Constituent Observations: A Review

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## Introduction

The assimilation of stratospheric observations has been the focus of several research groups in the past fifteen years. The use of products from the assimilation of meteorological data is now widespread. As new data types become available researchers are anxious to try assimilation experiments. This article will review progress in the last 3-5 years and evaluate that progress in terms of the underlying geophysical robustness of data assimilation systems.

A dictionary definition of assimilation is: to incorporate or absorb, for instance, into the mind or the prevailing culture. For Earth science, assimilation is the incorporation of observational information into a physical model. Or more specifically, data assimilation is the objective melding of observed information with model-predicted information.

Assimilation rigorously combines statistical modelling with physical modelling. Daley (1991) is the standard text on data assimilation. Cohn (1997) explores the theory of data assimilation, and its foundation in estimation theory. Swinbank *et al.* (2003) is a collection of tutorial lectures on data assimilation. Assimilation is difficult to do well, easy to do poorly, and its role in Earth science is expanding and sometimes controversial.

In the atmosphere the predominant observations that are assimilated are temperatures, or radiances that represent the temperature. These observations come from radiosondes, aircraft and satellites. Wind data are also assimilated, and are of crucial importance to the quality of the assimilation. As a function of altitude the amount of wind data available for assimilation drops dramatically above the tropopause. Hence, the state of the stratosphere is determined primarily by temperature observations and information propagated from the surface and the troposphere. A good tropospheric assimilation is required for a good stratospheric assimilation.

A number of constituents are also of interest. Water vapour has been routinely assimilated in the troposphere for many years and is important to the specification of tropospheric physics. More recently, several centres and groups have undertaken the assimilation of ozone. There have also been a number of experiments addressing interacting chemical species. Both ozone assimilation and chemical assimilation are reviewed in Swinbank *et al.* (2003).

There are a number of factors that motivate the desire to assimilate geophysical parameters. The list below is specifically for ozone, but it is representative of the motivation of other geophysical parameters as well.

## Motivations for Ozone Assimilation

- **Mapping**: There are spatial and temporal gaps in the ozone observing system. A basic goal of ozone assimilation is to provide vertically resolved global maps of ozone.
- **Short-term ozone forecasting**: There is interest in providing operational ozone forecasts in order to predict the fluctuations of ultraviolet radiation at the surface of the earth (Long *et al.*, 1996).
- **Chemical constraints**: Ozone is important in many chemical cycles. Assimilation of ozone into a chemistry model provides constraints on other observed constituents and helps to provide estimates of unobserved constituents.

| | | |
|---|---|---|
| Boundary Conditions | Emissions, SST, … | **e** |
| Representative Equations | $DA/Dt=P-LA-(n/H)A+q/H$ | e |
| Discrete/Parametrize | $(A_{n+\Delta t}-A_n)/\Delta t = ….$ | $(\mathbf{e}_d, \mathbf{e}_p)$ |
| Theory/Constraints | $\partial u_g/\partial z= -(\partial T/\partial y)R/(Hf_0)$ | Scale Analysis |
| Primary Products (i.e. A) | $T, u, v, F, H_2O, O_3$ | $(\mathbf{e}_b, \mathbf{e}_v)$ |
| Derived Products (F(A)) | Potential Vorticity, v*,w*, … | Consistent |

*Table 1: Simulation Framework: General Circulation Model, or forecast model is associated with errors: ($e_d$, $e_p$) = (discretization error, parametrization error) and ($e_b$, $e_v$) = (bias error, variability error). The size of 'e' in each case represents the size of the error associated with the process in each row. The derived products are likely to be physically consistent, but have significant errors;* i.e. *the theory-based constraints are met.*

- **Unified ozone data sets**: There are several sources of ozone data with significant differences in spatial and temporal characteristics as well as their expected error characteristics. Data assimilation provides a potential strategy for combining these data sets into a unified data set.
- **Tropospheric ozone**: Most of the ozone is in the stratosphere, and tropospheric ozone is sensitive to surface emission of pollutants. Therefore, the challenges of obtaining accurate tropospheric ozone measurements from space are significant. The combination of observations with the meteorological information provided by the model offers one of the better approaches available to obtain global estimates of tropospheric ozone.
- **Improvement of wind analysis**: The photochemical time scale for ozone is long compared with transport timescales in the lower stratosphere and upper troposphere. Therefore ozone measurements contain information about the wind field that might be obtained in multi-variate assimilation.
- **Radiative transfer**: Ozone is important in both longwave and shortwave radiative transfer. Therefore accurate representation of ozone is important in the radiative transfer calculations needed to extract (retrieve) information from many satellite instruments. In addition, accurate representation of ozone has the potential to impact the quality of the temperature analysis in multi-variate assimilation.
- **Observing system monitoring**: Ozone assimilation offers an effective way to characterize instrument performance relative to other sources of ozone observations as well as the stability of measurements over the lifetime of an instrument.
- **Retrieval of ozone**: Ozone assimilation offers the possibility of providing more accurate initial guesses for ozone retrieval algorithms than are currently available.
- **Assimilation research**: Ozone (constituent) assimilation can be productively approached as a univariate linear problem. Therefore it is a good framework for investigating assimilation science; for example, the impact of flow dependent covariance functions.
- **Model and observation validation**: Ozone assimilation provides several approaches to contribute to the validation of models and observations.

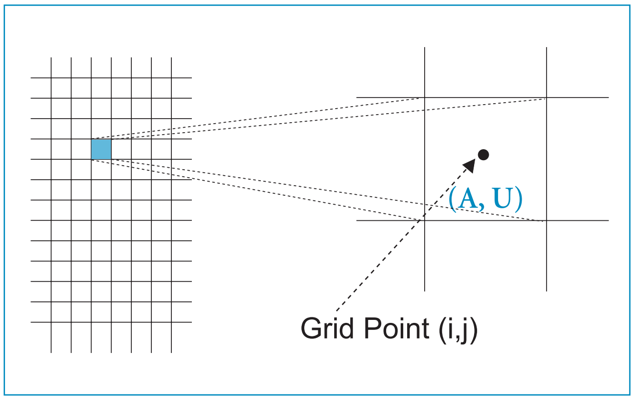


*Figure 1: Discretization of resolved transport and the choice of where to represent information impacts the physics, such as conservation, scale analysis limits, and stability (Rood, 1987). Grids may be orthogonal, uniform area, adaptive, or unstructured.*

Some of the goals mentioned above can be meaningfully addressed with the current state of the art, others cannot. It is straightforward to produce global maps of total column ozone which can be used in, for instance, radiative transfer calculations. The use of ozone measurements to provide constraints on other reactive species is an application that has been explored since the 1980's (see Jackman *et al.*, 1987) and modern data assimilation techniques potentially advance this field. The impact of ozone assimilation on the meteorological analysis of temperature and wind, and hence improvement of the weather forecast, is also possible. The most straightforward impact would be on the temperature analysis in the stratosphere. The improvement of the wind analysis is a more difficult challenge and confounded by the fact that where improvements in the wind analyses are most needed, the tropics, the ozone gradients are relatively weak. The use of ozone assimilation to monitor instrument performance and to characterize new observing systems is currently possible and productive (see Stajner *et al.*, 2004). The improvement of retrievals using assimilation techniques to provide ozone first guess fields that are representative of the specific environmental conditions is also an active research topic. The goal of producing unified ozone data sets from several instruments is of little value until bias can be correctly accommodated in data assimilation. This final topic will be discussed more fully below.

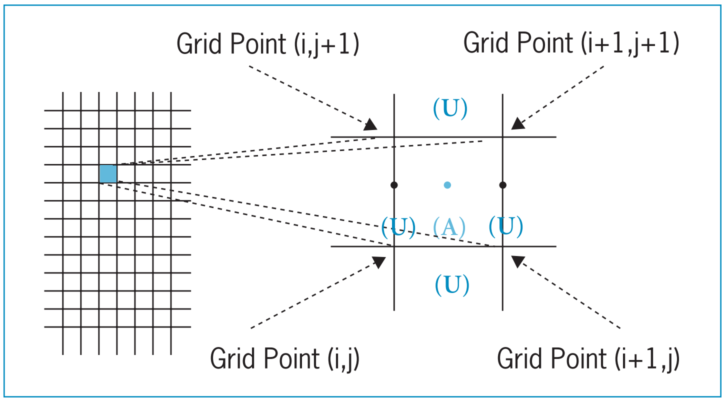


*Figure 2: Another choice of where to represent information on the grid.*

In order to discuss why assimilation is successful in addressing some of these goals and less successful in others, the underlying physical foundation of assimilation will be considered. This will be based on the discussion of modelling and assimilation and comparison of the attributes of the concepts of performing simulation (modelling) and assimilation. Following this, expository examples will be presented. An underlying tenet is that by comparing results from a simulation to results from an assimilation using the same model, the researcher is investigating cause and effect in a controlled experiment.

## Simulation and Assimilation

In order to provide an overarching background for thinking about model simulation, it is useful to consider the elements of the modelling, or simulation, framework described in Table 1. In this framework are six major ingredients. The first is the specification of boundary and initial conditions (*e.g.* topography and sea surface temperatures for an atmospheric model). Boundary conditions are generally prescribed from external sources of information.

The next three items in the table are intimately related. They are the representative equations, the discrete or parameterized equations, and the constraints drawn from theory. The representative equations are the analytic form of forces or factors that are considered important in the representation of the evolution of a set of parameters. All of the analytic expressions used in atmospheric modelling are approximations; therefore, even the equations the modeller is trying to solve have *a priori* errors. That is, the equations are scaled from some more complete representation, and only terms that are expected to be important are included in the analytic expressions (see Holton, 2004). The solution is, therefore, a balance amongst competing forces and tendencies. Most commonly, the analytic equations are a set of nonlinear partial differential equations.

The discrete or parameterized equations arise because it is generally not possible to solve the analytic equations in closed form. The strategy used by scientists is to develop a discrete representation of the equations which are then solved using numerical techniques. These solutions are, at best, discrete estimates to solutions of the analytic equations. The discretization and parameterization of the analytic equations introduce a large source of error. This introduces another level of balancing in the model; namely, these errors are generally managed through a subjective balancing process that keeps the numerical solution from running away to obviously incorrect estimates.

While all of the terms in the analytic equation are potentially important, there are conditions or times when there is a dominant balance between, for instance, two terms. An example of this is thermal wind balance in the middle latitudes of the stratosphere (see Holton, 2004). It is these balances, generally at the extremes of spatial and temporal scales, which provide the constraints drawn from theory. If the modeller implements discrete methods which consistently represent the relationship between the analytic equations and the constraints drawn from theory, then the modeller maintains a substantive scientific basis for the interpretation of model results.

The last two items in Table 1 represent the products that are drawn from the model. These are divided into two types: primary products and derived products. The primary products are variables such as wind, temperature, water, ozone – parameters that are most often, explicitly modelled; that is, an equation is written for them. The derived products are often functional relationships between the primary products; for instance, potential vorticity (Holton, 2004). A common derived product is the balance, or the budget, of the different terms of the discretized equations. The budget is then studied, explicitly, on how the balance is maintained and how this compares with budgets derived directly from observations. In general, the primary products can be directly evaluated with observations and errors of bias and variability estimated. If attention has been paid in the discretization of the analytic equations to honour the theoretical constraints, then the derived products will behave consistently with the primary products and theory. They will have errors of bias and variability, but they will behave in a way that supports scientific scrutiny.

In order to explore the elements of the modelling framework described above, the following continuity equation for a constituent, A, will be posed as the representative equation. The continuity equation represents the conservation of mass for a constituent and is an archetypical equation of geophysical models. Brasseur and Solomon (1986) and Dessler (2000) provide good backgrounds for understanding atmospheric chemistry and transport. The continuity equation for A is:

$$\frac{\partial A}{\partial t} = -\nabla \bullet \mathbf{U}A + M + P - LA - \frac{n}{H}A - \frac{q}{H} \quad (1)$$

where A is some constituent, **U** is velocity (“resolved” transport, “advection”), M is “Mixing” (“unresolved” transport, parameterization), P is production, L is loss, n is “deposition velocity”, q is emission, H is representative length scale for n and q, t is time, and ∇ is the gradient operator.

Attention will be focused on the discretization of the resolved advective transport. Figures 1 and 2 illustrate the basic concepts. On the left of the figure a mesh has been laid down to cover the spatial domain of interest. In this case it is a rectangular mesh. The mesh does not have to be rectangular, uniform, or orthogonal. In fact the mesh can be unstructured or can be built to adapt to the features that are being modelled. The choice of the mesh is determined by the modeller and depends upon the diagnostic and prognostic applications of the model (see Randall, 2000). The choice of mesh can also be determined by the computational advantages that might be realized.

Points can be prescribed to determine location with the mesh. In Figure 1 both the advective velocity and the constituent are prescribed at the centre of the cell. In Figure 2, the velocities are prescribed at the centre of the cell edges, and the constituent is prescribed in the centre of the cell. There are no hard and fast rules about where the parameters are prescribed, but small differences in their prescription can have a huge impact on the quality of the estimated solution to the equation, *i.e.* the simulation. The prescription directly impacts the ability of the model to represent conservation properties and to provide the link between the analytic equations and the theoretical constraints (see Rood, 1987; Lin and Rood, 1996; Lin, 2004). In addition, the prescription is strongly related to the stability of the numerical method; that is, the ability to represent any credible estimate at all. 

The use of numerical techniques to represent the partial differential equations that represent the model physics is a straightforward way to develop a model. There are many approaches to discretization of the dynamical equations that govern geophysical processes (Jacobson, 1998; Randall, 2000). One approach that has been recently adopted by several modelling centres is

| | | |
|---|---|---|
| Emissions, SST, … | **e** | Boundary Conditions |
| $DA/Dt=P-LA-(n/H)A+q/H$ | e | $(\mathbf{O}P_f\mathbf{O}^T+R)x=A_0-\mathbf{O}A_f$ |
| $(A_{n+\Delta t}-A_n)/\Delta t=\ldots$ | **e** | Discrete/Error Modelling |
| $\partial u_g/\partial z=-(\partial T/\partial y)R/(Hf_0)$ | Scale Analysis | Constraints on Increments |
| $A_i$=T, u, v, F, $H_2O$, $O_3$ | ($\mathbf{e}_b$, $\mathbf{e}_v$) | ($\mathbf{e}_b$, $\mathbf{e}_v$) reduced |
| Pot. Vorticity, v*, w*, … | Consistent | Inconsistent |

*Table 2: Assimilation Framework:* **O** *is the “observation” operator, $P_f$ is forecast model error covariance, R is the observation error covariance, and x is the innovation. Generally resolved, predicted variables are assimilated into the models.*

described in Lin (2004). In this approach the cells are treated as finite volumes and piecewise continuous functions are fit locally to the cells. These piecewise continuous functions are then integrated around the volume to yield the forces acting on the volume. This method, which was derived with physical consistency as a requirement for the scheme, has proven to have numerous scientific advantages. The scheme uses the philosophy that if the physics are properly represented, then the accuracy of the scheme can be robustly built on a physical foundation.

Table 2 shows elements of an assimilation framework that parallels the modelling elements in Table 1. The concept of boundary conditions remains the same; that is, some specified information at the spatial and temporal domain edges. Of particular note, the motivation for doing data assimilation is often to provide the initial conditions for predictive forecasts.

Data assimilation adds an additional forcing to the representative equations of the physical model; namely, information from the observations. This forcing is formally added through a correction to the model that is calculated, for example, by (see Stajner *et al.*, 2001):



$$(\boldsymbol{O}P_f\boldsymbol{O}^T + R)x = A_o - \boldsymbol{O}A_f \qquad (2)$$

The terms in the equation are as follows: $A_o$ are observations of the constituent, $A_f$ are model forecast (simulated estimates of the constituent, often called the first guess), $\boldsymbol{O}$ is the observation operator, $P_f$ is the error covariance function of the forecast, R is the error covariance function of the observations, x is the innovation that represents the observation-based correction to the model, and $(\,)^T$ is the matrix transform operation. The observation operator, $\boldsymbol{O}$, is a function that maps the parameter to be assimilated into observation space.

The error covariance functions $P_f$ and R represent the errors of the information from the forecast model and the information from the observations, respectively. This explicitly shows that data assimilation is the error-weighted combination of information from two primary sources. From first principles, the error covariance functions are prohibitive to calculate. Stajner *et al.* (2001) show a method for estimating the error covariances in an ozone assimilation system.

Parallel to the elements in the simulation framework (Table 1), discrete numerical methods are needed to estimate the errors as well as to solve the matrix equations in Equation (2). Addressing physical constraints from theory is a matter of both importance and difficulty. Often, for example, it is assumed that the increments of different parameters that are used to correct the model are in some sort of physical balance. For instance, wind and temperature increments might be expected to be in geostrophic balance. However, in general, the data insertion process acts like an additional forcing term in the equation, and contributes a significant portion of the budget. This explicitly alters the physical balance defined by the representative equations of the model. Therefore, there is no reason to expect that the correct geophysical balances are represented in an assimilated data product. This is contrary to the prevailing notion that the model and observations are 'consistent' with one another after assimilation.

The final two elements in Table 2 are, again, the products. In a good assimilation the primary products, most often the prognostic variables, are well estimated. That is, both the bias errors and the variance errors are reduced relative to the model simulation. However, the derived products are likely to be physically inconsistent because of the nature of the corrective forcing added by the observations. These errors are often found to be larger than those in self-determining model simulations. This is of great consequence as many users look to data assimilation to provide estimates of unobserved or derived quantities. Molod *et al.* (1996) and Kistler *et al.* (2001) provide discussions on the characteristics of the errors associated with primary and derived products in data assimilation systems.

As suggested earlier, the specification of forecast and model error covariances and their evolution with time is a difficult problem. In order to get a handle on these problems it is generally assumed that the observational errors and model errors are unbiased over some suitable period of time, *e.g.* the length of the forecast between times of data insertion. It is also assumed that the errors are in a Gaussian distribution. The majority of assimilation theory is developed based on these assumptions, which are, in fact, not valid assumptions. In particular, when the observations are biased, there would be the expectation that the actual balance of geophysical terms is different from the balance determined by the assimilation. Furthermore, since the biases will have spatial and temporal variability, the balances determined by the assimilation are quite complex. Aside from biases between the observations and the model prediction, there are biases between different observation systems for the same parameters. These biases are potentially correctible if there is a known standard of accuracy defined by a particular observing system. However, the problem of bias is a difficult one to address and perhaps the greatest challenge facing assimilation (see, Dee and da Silva, 1998).

As a final general consideration, there are many time scales represented by the representative equations of the model. Some of these time scales represent balances that are achieved almost instantly between different variables. Other time scales are long (*e.g.* the general circulation), which will determine the distribution of long-lived trace constituents. It is possible in assimilation to produce a very accurate representation of the observed state variables, and those variables which are balanced on fast time scales. On the other hand, improved estimates in the state variables are found, at least sometimes, to be associated with degraded estimates of those features determined by long time scales. Conceptually, this can be thought of as the impact of bias propagating through the physical model. With the assumption that the observations are fundamentally accurate, this indicates errors in the specification of the physics that demand further research.

## Transport applications: What have we learned?

Rood *et al.* (1989) first used winds and temperatures from a meteorological assimilation to study stratospheric transport. Since that time there have been productive studies of both tropospheric and stratospheric transport. However, a number of barriers have been met in recent years, and the question arises - has a wall been reached where foundational elements of data assimilation are limiting the ability to do quantitative transport applications? Stohl *et al.* (2004; and references therein) provide an overview of some of the limits that need to be considered in transport applications.

In transport applications, winds and temperatures are taken from a meteorological assimilation and used as input to a constituent transport model (CTM). The resultant distributions of trace constituents are then compared with observations. The constituent observations are telling indicators of atmospheric motions on all time scales. Further, there is a wealth of very high quality constituent observations from many observational platforms. Rigorous quantitative Earth science has been significantly advanced by comparison of constituent observations and model estimates. Overall, it is found that the meteorological analyses do a very good job of representing variability associated with synoptic and planetary waves. This has been invaluable in accounting for dynamical variability, and allowing the evaluation of constituents from multiple observational platforms. On the other hand, those geophysical parameters that rely on the representation of the general circulation, for instance the lifetime of long-lived constituents, are poorly represented.

Returning to the concepts introduced in the previous section, when the primary products of assimilation (*e.g.* winds and temperature) dominate the variability, then the variability is well represented. When the derived products (*e.g.* the residual circulation) are responsible for the variability, then the variability is poorly represented. This is found to be consistently true. Transport calculations from all assimilation systems are found to have too much mixing between the tropics and the middle latitudes. In the tropics, where the temperature observations do not strongly determine the winds, the quality of the transport degrades relative to the middle latitudes. In the troposphere, if the transport features are associated with convection or surface exchanges, *i.e.* the quantities derived from the constrained physical parameterizations, then they are not robustly represented.

By the nature of data assimilation, improvements to the system provide better estimates to the variables that are being assimilated. Simmons *et al.* (2003) show impressive improvement of the representation of winds in the tropics in the products from the European Centre for Medium-range Weather Forecasts. However, a number of conference papers have shown that these improvements have not extended to the representation of transport features associated with the general circulation; the derived parameters associated with long time scales for adjustment are degraded. Ruhnke *et al.* (2003) were amongst the first to demonstrate this through the transport of ozone, where overestimates of lower stratospheric ozone increased with newer versions of the data assimilation system.

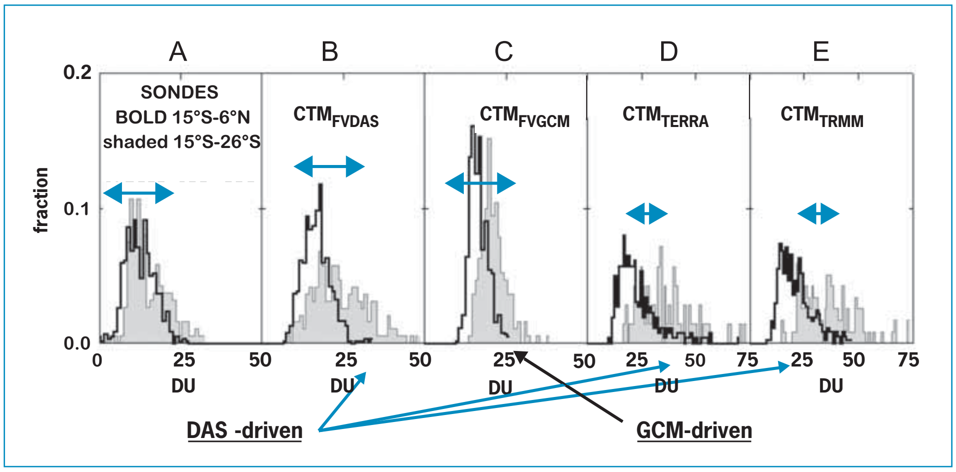


*Figure 3: PDFs of total ozone, observations and chemical transport model (CTM). PDFs from DAS-driven show displaced means and spreads that are too wide, whereas GCM-driven PDFs have displaced means with a better half-width. This shows that there is too much tropical-extratropical mixing in DAS. From Douglass et al. (JGR, 2003).*

Douglass *et al.* (2003) and Schoeberl *et al.* (2003) each provide detailed studies that expose some of the foundational shortcomings of the physical consistency of data assimilation. In their studies they investigate the transport and mixing of atmospheric constituents in the upper troposphere and the lower stratosphere. Figure 3 is from Douglass *et al.* (2003) and shows ozone probability distribution functions in two latitude bands from four experiments using a constituent transport model. In three of these experiments (Panels B, D, and E), winds and temperatures are taken from an assimilation system. In panel C, results are from an experiment using winds from a general circulation model; that is, a free-running model without assimilation. Panel A shows ozonesonde observations; the sondes reflect similar distributions in the two latitude bands. In all of the numerical experiments, the means in the two latitude bands are displaced from each other, unlike the observations. In the three experiments using winds from different data assimilation systems (DAS), the half-width of the distributions is much too wide.

There are a number of points to be made in this figure. First, the winds from the assimilation system in Panel B and the model in Panel C both use the finite-volume dynamics of Lin (2004). Therefore, these experiments are side-by-side comparisons that show the impact of inserting data into the model. Aside from developing a bias, the assimilation system shows much more mixing. As Douglass *et al.* show, the instantaneous representation of the wind is better in the assimilation, but the transport is worse. This is attributed to the fact that there are consistent biases in the model prediction of the tropical winds and the correction added by the data insertion causes spurious mixing. Tan *et al.* [2004] investigate the dynamical mechanisms of the mixing in the tropics and the subtropics. Second, the assimilation systems used for Panels D and E have a different assimilation model, and their representation of transport is worse than that from the finite-volume model. This improvement is attributed to the fact that the finite-volume model represents the physics of the atmosphere better, in particular, the representation of the vertical velocity. Third, the results in Panel B show significant improvement compared to the older assimilation systems used in Panels D and E. Older assimilation systems have enough deficiencies that scientists have shied away from doing tropical transport studies. This example demonstrates both the improvements that have been gained in recent years, and indicates that the use of winds from assimilation in transport studies might have fundamental limitations.

Figure 4 (see colour insert IV) is from Schoeberl *et al.* (2003). The Schoeberl *et al.* (2003) study is similar in spirit to the Douglass *et al.* study, but uses Lagrangian trajectories instead of Eulerian advection schemes. This allows Schoeberl *et al.* (2003) to address, directly, whether or not the spurious mixing revealed in the Douglass *et al.* (2003) paper is related to the advection scheme. In this figure the results from two completely independent assimilation systems are used; UKMO (United Kingdom Met Office) and DAO (Data Assimilation Office). The DAO system uses the finite-volume dynamical core (labelled DAO) and the finite-volume GCM (labelled GCM). Vertical winds are also calculated two ways, diabatically using the heating rate information from the assimilation system, kinematically, through continuity, using the horizontal winds from the assimilation.



The figure shows the impact of the method of calculating the vertical wind using the diabatic information. When the diabatic information is used there is much less transport in the vertical. While this is generally in better agreement with observations and theory, the diabatic winds no longer satisfy mass continuity with the horizontal winds. This points to a self-limiting aspect of using diabatic winds in Eulerian calculations such as the ones of Douglass *et al.* (2003). The Schoeberl *et al.* calculations also show that even with the diabatic vertical winds, there remains significant horizontal mixing, which is compressed along isentropic surfaces. The final panel shows that for the simulation, the free-running model, there is much less dispersion, which is in better agreement with both observations and theory. Schoeberl *et al.* attribute the excess dispersion in the assimilation systems to noise that is introduced by data insertion. (They also note that the finite-volume dynamics is much improved relative to previous generation models.)

These two studies point to the fact that data insertion impacts the physics that maintains the balances in the conservation equations of momentum, heat, and mass. Both bias and the generation of noise have an impact. Both problems are difficult to address, with the problem of bias having fundamental issues of tractability. Again, while the data assimilation system does indeed provide better estimates of the primary variables, as the impact of data insertion is adjusted through the physics represented in the model, the derived parameters are often degraded. (Lait (2002) provides an interesting exposition of subtle artifacts related to biases between different radiosonde instruments.) One conclusion is that while there may be greater discrepancies in the absolute, day-to-day representation of constituents with free-running models, the consistent representation of the underlying physics allows more robust study of transport mechanisms and those features in the constituent data which are directly related to dynamics.

## Ozone Assimilation

The last five years have seen the publication of a number of papers on the assimilation of ozone and other constituents. These studies show constant improvement in the state of the art. They demonstrate the importance of having both total column observations as well as high-resolution profile information. Because of this progress, products from ozone assimilation are on the verge of being geophysically interesting. Applications, for example, include improvement of radiative transfer, monitoring of instruments, and providing information useful for estimating tropospheric ozone.

Figure 5 (see colour insert IV) shows an example from an assimilation of ozone data (Wargan *et al.* 2005). In this example, there are two satellite instruments, the Solar Backscattered Ultraviolet/2 (SBUV) and the Michelson Interferometer for Passive Atmospheric Sounding (MIPAS). SBUV is a nadir sounder and measures very thick layers with the vertical information in the middle and upper stratosphere. MIPAS is a limb sounder with much finer vertical resolution and measurements extending into the lower stratosphere. SBUV also measures total column ozone, which is assimilated in all experiments. The results from three assimilation experiments are shown through comparison with an ozonesonde profile. Ozonesondes were not assimilated into the systems, therefore, these data provide an independent measure of performance.

There are several attributes to be noted in Figure 5. The quality of the MIPAS-only (+ SBUV total column) assimilation is the best of those presented. This suggests that the vertical resolution of MIPAS instrument has a large impact. Even though the MIPAS observations are assimilated only above 70 hPa, the assimilation captures the essence of the structure of the ozone profile down to 300 hPa. This indicates that the model information in the lower stratosphere and upper troposphere is geophysically meaningful. Further, the model is effectively distributing information in the horizontal between the satellite profiles. The comparison with the SBUV-only (+ SBUV total column) assimilation shows that the thick-layered information of the SBUV observations, even in combination with the model information, does not represent the ozone peak very well. This impacts the quality of the lower stratospheric analysis as the column is adjusted to represent the constraints of the total ozone observations. Finally, from first principles, the combined SBUV and MIPAS assimilation might be expected to be the best since it has the maximum amount of information. This is not found to be the case, and suggests that the weights of the various error covariances and the use of the observations can be improved. The optimal balance of nadir and limb observations is not straightforward, and these experiments reveal the challenges that need to be addressed when multiple types of instruments are used in data assimilation.

## Summary

In the past 3-5 years there has been notable improvement in the representation by data assimilation of the primary parameters that describe the physical state of the stratosphere, *i.e.* wind and temperature. There has also been a significant improvement in the state of the art of ozone assimilation. Derived parameters, such as the residual circulation, have not seen similar improvements. This is attributed to insertion of data into the model during the data assimilation cycle. Both bias and noise will have a response that is realized through changes in the physical balance of the model being used for assimilation. As primary variables are pushed closer and closer to observations, this physical response often pushes the derived parameters further from reality. A problem that has not been discussed is the difficult problem of gravity waves and gravity wave dissipation. The data insertion process acts as an additional source of gravity waves, and hence, there are both near field and far field impacts.

The question is raised as to whether or not we have reached a limit in the transport and climate problems that can be addressed with assimilated data. Clearly, we have reached a point where the lack of physical consistency is impacting the ability to study the problems at hand. This is true not only in transport studies, but in any studies that require a consistent, closed budget. Furthermore, because of the exquisite sensitivity of data assimilation systems to the input observations, long-term trends strongly reflect changes in the observing system. Improvement in the quality of assimilated data systems is most likely to follow the use of new observations, the reduction of bias, and the development of bias correction techniques. The reduction of bias, and the key to development of assimilated data sets for use in chemistry and climate, will depend on the development of improved physical parameterizations. These improvements should include the ability to directly predict and constrain new observations of the physical processes. Based on recent experience, the details of the assimilation method, *i.e.* the statistical analysis algorithm, and the improvement of the representation of error covariance are secondary to addressing the problems of bias and physics.

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Update

# The SPARC Polar Stratospheric Cloud Assessment

The SPARC Polar Stratospheric Cloud Assessment (SPA) has kicked off with a chapter scoping meeting at the Coolfont (Spa) resort in West Virginia, USA from 12-13 May. The meeting was kindly hosted by our local organiser Mike Fromm of the US Naval Research Laboratory (NRL).

The aim of SPA is to assess our understanding of polar stratospheric clouds (PSCs). The motivation for the assessment is that there remain genuine gaps in our understanding of PSC distribution, formation and long term change that are important to stratospheric chemistry. Our understanding is patchy and specific rather than global and integrated and there has been a tendency to undertake detailed process studies rather than integrative studies. There is also no consensus on how to describe PSCs and denitrification in global models, which means we are unable to reliably predict changes that might occur in a future stratosphere. An important factor that limits our understanding is the lack of a large-scale consistent and evaluated dataset for model testing, the creation of which is one aim of SPA.

The purpose of the meeting was to agree on the organisation of the chapters and distribute writing tasks. We also managed to create an additional chapter on meteorological processes and agreed to produce a "Twenty Questions and Answers About PSCs" document for the stratospheric community, along the lines of David Fahey's excellent WMO Ozone Assessment pamphlet. If written well, we believe that such a document will allow the stratospheric community to better comprehend the rather acronym burdened world of PSCs.

The chapters are as follows:

1. **PSC processes (Niels Larsen)**
   This chapter will set the scene and also serve to pull the whole assessment together.
2. **Temperatures and meteorological diagnostics for PSC studies (Gloria Manney and Steve Eckermann)**
   This chapter was seen as a very important addition to SPA and was created during the meeting. Much of the analysis of PSC occurrence and the interpretation of datasets in terms of microphysical quantities requires accurate meteorological analyses. We also know that mesoscale processes play an important role in forming solid PSCs. The aim of this chapter will be to assess how accurate the meteorological analyses need to be and what effect uncertainties have on our interpretation of PSC measurements.
3. **PSC detection and discrimination (Beiping Luo and Mike Fromm)**
   The overall aim of this chapter is to define the range of PSC properties that can be detected by different instruments. Statements have previously been made about the frequency of occurrence of various cloud types that may have more to do with instrument detection thresholds than with actual PSC existence, particularly when it comes to the existence of a few large solid particles. This chapter will provide a consistent way of converting PSC signatures into ranges of microphysical properties, where possible.
4. **PSC observations and their interpretation (Terry Deshler and Lamont Poole)**
   This is the chapter that most closely resembles a PSC climatology. Climatologies have been produced in the past. The aim here is to refine them to provide information about different types of PSC.
5. **Denitrification and dehydration observations (Michelle Santee and Gerald Nedoluha)**
   The occurrence of denitrification and dehydration is closely related to that of PSCs. This chapter will mirror chapter 4 by examining the satellite and *in situ* record of denitrification and dehydration.
6. **Using models to assess our understanding of PSCs (Katja Drdla and Ken Carslaw)**
   How good are the different PSC models that are used? Models range from those that include, as far as possible, all the microphysical processes to those that parametrise PSC processes for use in large scale models. This chapter will assess these models, including denitrification and dehydration, for a few well defined cases based on the data from chapters 4 and 5. An important aspect will be the extent to which laboratory measurements of PSC formation, models and field observations are consistent.
7. **PSCs in a changing stratosphere (Markus Rex and Richard Bevilacqua)**
   The aim of this chapter is to understand what controls long term changes and interannual variability in PSCs. It will also explore simple ways for global models such as CCMs to test whether they have the right physics to correctly predict long term changes that are relevant to ozone depletion studies.

SPA has ambitious aims that go beyond a straightforward review of understanding. A 10 month period was felt to be essential for all participants to make substantial progress, particularly in cases where large datasets need to be assembled and analysed. Our next meeting is planned for early spring 2006 in Europe.

**Co-chairs: Ken Carslaw, University of Leeds, UK (carslaw@env.leeds.ac.uk)**
**Katja Drdla, Ames Research Centre, USA (Katja.Drdla-1@nasa.gov)**

**Meeting participants: Chapter lead authors (Niels Larsen, Gloria Manney, Terry Deshler, Lamont Poole, Beiping Luo, Mike Fromm, Michelle Santee, Gerald Nedoluha, Richard Bevilacqua) and coauthors (Harald Flentje, John Remedios, Christiane Voigt, Hideaki Nakajima, Jerome Alfred, Tony Strawa)**

# In Memory of Gérard Mégie 1946-2004

Gérard Mégie died at the age of 58, following an extended illness which did not prevent him from assuming his duties as Chairman of CNRS until the last minute. He left a feeling of emptiness in the numerous national, European and international organisations where he was active.

Gérard Mégie was well known in the SPARC community. The bulk of his research dealt with the development of original methods for measuring atmospheric variables by lidar, and the modelling of the natural variability of ozone and how it is influenced by human activities. He participated in the implementation of many different means for observing the atmosphere from the ground and from other platforms (airplane, balloon, satellite) and played a leading role in setting up the Network for Detection of Stratospheric Changes (NDSC).

Besides his increasing responsibilities in the national research community, he chaired the International Ozone Committee of the International Council for Science (ICSU) from 1988 to 1996, the Earth Observation Committee of the European Space Agency (ESA) from 1994 to 1999, and the Stratosphere Scientific Committee of the European Commission (1989-2004). He co-chaired the International Scientific Committee of the Montreal Protocol for the protection of the ozone layer and the WMO-UNEP Scientific Assessment of Ozone Depletion. He became a member of the European Research Advisory Board (EURAB) of the European Commission in 2001, and a member of the European Research Council Expert Group (ERCEG) of the European Union in 2003.

Gérard Mégie was the author of more than 240 scientific publications and two books on stratospheric ozone: '*Stratosphère et Couche d'Ozone*' (Editions Masson) in 1991 and '*Ozone, l'Equilibre Rompu*' (Presses du CNRS) in 1989. He was also the scientific coordinator of two reports of the French Academy of Sciences: 'Ozone et Propriétés Oxydantes de la Troposphère' (1993), and 'Ozone Stratosphérique' (1998) published by Editions Lavoisier.

Gérard Mégie was an internationally recognized scientist as well as a highly cultivated person, a humanist and a teacher, appreciated by everyone for his open-mindedness, his intellectual discipline and his keen analytical ability. We miss him greatly.

**Marie-Lise Chanin**



# Announcement

## The New SPARC Data Center

Marvin Geller, Stony Brook, New York, USA (marvin.geller@sunysb.edu)

Dr. Stefan Liess from Stony Brook University has been serving as the SPARC Data Center Scientist at http://www.sparc.sunysb.edu since December 2004. Stefan Liess will continue the work of his predecessors Petra Udelhofen and Xuelong Zhou in working with the SPARC projects to acquire newly available data as needed, as well as to maintain archives of data that have been used in past SPARC projects so that they may be acquired as needed by the world community, and also be available for updating should the SPARC projects desire to do so. Subject to funding, plans for the near future include a modernization of hardware facilities in order to accommodate the growing demand for data storage. This demand arises from several sources, e.g. from an ongoing collection of high-resolution radiosonde data from 93 U.S. stations, a planned chemistry-climate model intercomparison project (CCMVal), and to provide storage space for SPARC-IPY data. The implementation of online plotting in order to ease data access and decrease data transfer is also being planned. Please contact Stefan Liess at stefan.liess@stonybrook.edu for any questions regarding the SPARC Data Center. The SPARC Data Center has been funded by NASA grants to Stony Brook University with Marvin Geller as Principal Investigator. A new proposal is being prepared for submission to NASA for continued support.

**SPARC Data Center Website: http://www.sparc.sunysb.edu**

# Future SPARC and SPARC-related Meetings

## 2005

**2-11 August:** **9th Scientific Assembly of IAMAS**, Beijing, China (http://iamas2005.com)

**22-26 August:** **IAG/IAPSO with IABO Joint Scientific Assembly**, Cairns, Australia (http://www.gfy.ku.dk/~iag, http://www.iugg.org/iapso)

**12-16 September:** **Third Stratospheric Processes And their Role in Climate (SPARC) Data Assimilation (SPARC-DA3) workshop**, Banff, Canada
Sept 12-14: Third Stratospheric Processes And their Role in Climate (SPARC) Data Assimilation (SPARC-DA3) Workshop
Sept 14-16: SPARC Workshop on Stratospheric Winds

**12-16 September:** **5th Annual Meeting of the European Meteorological Society**, Utrecht, Netherlands

**3-15 October:** **Cargese International School COST ACTION 723 Upper Troposphere and Lower Stratosphere**, Cargese, Corsica, France (http://www.cost723.org/school)

**17-19 October:** **Chemistry-Climate Workshop CCMVal**, Boulder, Colorado, USA

**17-22 October:** **ICSU 28th General Assembly**, Beijing, China www.icsu.org

**5-9 December:** **AGU Fall Meeting**, San Francisco, California, USA.

## 2006

**29 January- 02 February: 86th AMS Annual Meeting**, Atlanta, Georgia
The Jim Holton Symposium
18th Conference on Climate Variability and Change,
Tenth Symposium on Integrated Observing and Assimilation Systems for Atmosphere, Oceans, and Land Surface (IOAS-AOLS),
The Eighth Conference on Atmospheric Chemistry





### Themes and Group Leaders

- **Climate-Chemistry Interactions:** C. Granier (France), T. Peter (Switzerland), A. R. Ravishankara (USA)
- **Detection, Attribution, and Prediction of Stratospheric Change:** W. Randel (USA), T.G. Shepherd (Canada)
- Stratosphere-Troposphere Dynamical Coupling: M. Baldwin (USA), S. Yoden (Japan)
- **Gravitiy Waves:** K. Hamilton (USA), R. Vincent (New Zealand)
- **Data Assimilation:** S. Polavarapu (Canada)
- **GRIPS:** S. Pawson (USA)
- **CCM Validation:** V. Eyring (Germany), N. Harris (UK), T.G. Shepherd (Canada)
- **Laboratory Studies joint with IGAC:** A. R. Ravishankara (USA), R. A. Cox (IGAC)
- **PSC Climatology:** K. Carslaw (UK), K. Drdla (USA)
- **Solar Influences joint with SCOSTEP:** K. Kodera (Japan)


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