

Report on the 22nd session of the Joint Scientific Committee (JSC) for the World Climate Research Programme (WCRP)

Boulder, USA, 19-23 March 2001

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The JSC provides overall scientific guidance for the WCRP and all the components of the Programme (the Global Energy and Water Cycle Experiment, GEWEX; Stratospheric Processes And their Role in Climate, SPARC; the World Ocean Circulation Experiment, WOCE; the Climate Variability and Predictability Study, CLIVAR; the Arctic Climate System Study/Climate and Cryosphere study, ACSYS/CLIC; the modelling activities conducted by the Working Group on Numerical Experimentation, WGNE, and the Working Group on Coupled Modelling, WGCM) were reviewed. The interactions and development of co-operation between WCRP and the other international global environmental change programmes, in particular the International Geosphere-Biosphere Programme (IGBP) and the International

Human Dimensions of Global Environmental Change Programme (IHDP) were also discussed.

Both the Co-chairs of the SPARC Scientific Steering Group, Marie-Lise Chanin and Marvin Geller participated in the meeting on behalf of SPARC. A presentation was provided to the JSC on the progress being made including particularly the main conclusions and results of the Second SPARC General Assembly ("SPARC 2000") in Mar del Plata, Argentina, 6-10 November 2000 (see report in SPARC Newsletter No. 16) and the recommendations from the following session of the SPARC Scientific Steering Group in Buenos Aires, 13-16 November (also reported in SPARC Newsletter No. 16). Attention was drawn to the future directions foreseen for SPARC and the approach

being planned for an integrated understanding of climate change in the stratosphere, and to a number of recent results relating to solar effects on climate.

The JSC expressed satisfaction at the work being carried out in SPARC and its continuing development. The good relationship between SPARC and WGNE in aiming to improve the representation of the stratosphere in models was particularly appreciated. The JSC especially noted the questions on the validity of the statistical analysis employed in suggesting the relationship between solar effects and land surface temperatures that were reported by M. Geller. Equally, there were questions concerning the analysis based on ISCCP data sets that had pointed to possible cloudiness

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variations associated with changes in solar output and cosmic rays. The JSC requested the GEWEX Scientific Steering Group to review and comment on these results. However, a more general proposal put forward by SPARC for a joint working group with SCOSTEP to assess solar influences on climate (that might also have included GEWEX and CLIVAR representatives) was not endorsed, being considered that a further broad evaluation of this topic was not necessary at present, so soon after the latest intensive IPCC discussions.

On other SPARC issues, the JSC voiced congratulations on the completion of the Water Vapour Assessment (WAVAS) (see SPARC Newsletter No. 16). The analyses in trends of ozone, temperature and water vapour carried out by SPARC had all been state-of-the-art summaries and had fed into the WMO/UNEP Ozone Assessment and into the IPCC Third Assessment. The anticipation of scientific needs in this way by SPARC was considered to be a good example for the whole WCRP. The JSC duly welcomed the study now being taken forward jointly by SPARC and SCOSTEP (together with the International Commission of the Middle Atmosphere) of upper stratospheric and mesospheric temperature trends and the SPARC participation in a workshop on ozone changes in the past decade leading up the next WMO/UNEP Assessment. The JSC also encouraged the integrated approach being developed by SPARC to see whether the different observed data variations provided a consistent picture of stratospheric climate changes (including possible trends) over the past decades (upon which shorter time-scale variations might be superimposed). Furthermore, the JSC recognised the importance of the SPARC study of gravity wave processes in the stratosphere and their parameterisation, including the construction of a stratospheric gravity wave climatology and the efforts devoted to the organisation of an international field experiment to examine the gravity-wave field forced by tropical convection (the "Effects of Tropical Convection Experiment, ECTE).

Another point raised in the SPARC report to the JSC was the dynamical coupling of the stratosphere and troposphere and the possible link between Arctic Oscillation (AO) and North Atlantic Oscillation (NAO), which appear to be different manifes-

tations of the same underlying dynamical phenomenon. The JSC urged SPARC, CLIVAR and ACSYS/CLIC to keep the questions involved under review, but did not recommend that any (joint) specific activity be organised in WCRP at present in view of the multiplicity of workshops/symposia on this subject being arranged by the scientific community at large.

In the field of co-operation between the global environmental change programmes, the Global Change Open Science Conference in Amsterdam (10-13 July 2001) is a particular highlight. All WCRP projects and groups were strongly encouraged to attend or be represented. A significant SPARC poster cluster has been arranged, and a new SPARC brochure especially prepared. More specifically, with respect to liaison between WCRP and IGBP, it was reported by Dr. B. Moore, Chair of the IGBP Scientific Advisory Committee that a number of IGBP core projects were expected to be significantly reorganised at the end of 2002, and it is to be determined how IGBP will treat the field of chemistry in the atmosphere in the future. In this context, consideration will be given to undertaking appropriate joint WCRP/IGBP activities in the area of "Atmospheric chemistry and climate", and Susan Solomon was asked to discuss the possibilities with Guy Brasseur (present Chairman of the IGAC Scientific Committee, and Chairman-designate of the IGBP Scientific Committee). In the meantime, the importance of continuing to carry forward the existing (successful) SPARC/IGAC co-operation to ensure that planned joint activities were carried forward in a timely and organised manner, was emphasised. The SPARC/IGAC joint activities would be a principal building block of an "Atmospheric Chemistry and Climate Initiative".

The relationship between WCRP and the space agencies and the Integrated Global Observing Strategy (IGOS) was discussed. It was noted that an Integrated Global Atmospheric Chemistry Observations (IGACO) theme was now being taken up. A report on a pilot project on ozone had earlier been agreed and published, and the JSC urged that implementation of proposals laid out in the ozone report that need not wait until the wider atmospheric chemistry theme had been fully defined should proceed as rapidly as possible.

As customary at its sessions, the JSC reviewed membership of WCRP Scientific Steering and Working Groups. With regard to the SPARC Scientific Steering Group, the JSC noted with the greatest regret that M.-L. Chanin has decided to relinquish her position as Co-chair which she had held since the inception of SPARC in 1992. Great appreciation and gratitude by all present were expressed to M.-L. Chanin for her outstanding contributions, during her period of service which had resulted in SPARC having become such a successful activity in WCRP and very much the focus for international stratospheric science. All were reassured that WCRP and SPARC would continue to benefit from her expertise and support for a further few years in her capacity as Director of the SPARC Office. The appointment of Alan O'Neill (University of Reading, UK) as the new Co-chair of the SPARC Scientific Steering Group was welcomed. Other new members nominated were P. Canziani (University of Buenos Aires, Argentina), A.R. Ravishankara (NOAA/ERL Aeronomy Laboratory, Boulder, USA) and V. Yushkov (Central Aerological Observatory, Moscow, Russian Federation). The membership of the SPARC Scientific Steering Group thus becomes:

Membership	Expiry of appointment
M. Geller, Co-chair	31 December 2002
A.J. O'Neill, Co-chair	31 December 2002
P. Canziani	31 December 2004
C. Granier	31 December 2002
K. Hamilton	31 December 2002
D. Karoly	31 December 2002
T. Peter	31 December 2002
A. Ravishankara	31 December 2004
U. Schmidt	31 December 2002
T. Shepherd	31 December 2002
S. Yoden	31 December 2002
V. Yushkov	31 December 2004



Report on the SPARC Tropopause Workshop

Bad Tölz, Germany, 17-21 April 2001

Organisers: Peter Haynes University of Cambridge, UK (p.h.haynes@damtp.cam.ac.uk)
Ted Shepherd University of Toronto, Canada (tgs@atmosp.physics.utoronto.ca) and
Volkmar Wirth University of Mainz, Germany (vwirth@mail.uni-mainz.de)

Introduction

The stratosphere is characterised by strong stability (to vertical displacements), high ozone and low water vapour, and the troposphere by weak stability, low ozone and high water vapour. The notional boundary between the two, the tropopause, has several definitions (WMO standard definition based on lapse rate, cold-point temperature, ozone tropopause, potential-vorticity definition, etc.), which are useful for different purposes. In reality, the transition between troposphere and stratosphere occurs over a layer of finite thickness and it is best to consider a tropopause region within which the different tropopauses corresponding to the different definitions might be expected to fall. Transport across the tropopause region is mainly upwards in the tropics and mainly downwards in the extratropics (the Brewer-Dobson circulation), but there is significant two-way transport, particularly in the extratropics.

The tropopause region is a critical region for climate. Chemistry relevant to climate is highly dependent on the OH radical, whose production is proportional to the product of ozone and water vapour and so is large in the tropopause region. Chemical and radiative timescales are relatively long (see Figure 1 p. I), which means that dynamical forcings (chemical transport and adiabatic warming or cooling) play a particularly strong role in controlling the structure of the region. But equally, it means that chemical concentrations are affected even by slow chemical processes, and temperature is highly sensitive to changes in radiative heating or cooling rates. Low temperatures imply the importance of condensed matter (liquid and solid clouds and aerosols) in this region, and therefore of heterogeneous and multiphase chemical reactions, as

well as associated radiative effects. Finally, the tropical tropopause temperature controls the amount of stratospheric water vapour through the freeze drying mechanism. These factors all make the tropopause region particularly crucial for climate sensitivity. It should be noted that current climate models, with vertical resolution of 1 km or coarser, cannot resolve this region because the transition between troposphere and stratosphere occurs over only a few kilometres.

shop in Cambridge in 1993 on Stratosphere-Troposphere Exchange, which led to the highly influential J. Holton *et al.* (1995) review paper, shifted the paradigm for exchange from local to global. However, the focus was still very much on dynamics and transport. Now there is a need to link the dynamical/transport picture to climate, which requires the inclusion of radiation and chemistry in the conceptual framework. Yet even some of



List of participants: M. Bourqui, B. Boville, J.-P. Cammas, G. Craig, V. Daniel, H. Davies, A. Dessler, G. Esler, V. Eyring, H. Fischer, I. Folkins, P. Forster, M. Fujiwara, A. Fusco, M. Geller, A. Gettelman, J. Gille, L. Goldfarb, V. Grewe, P. Haynes, J. Holton, P. James, D. Kley, M. Kramer, M. G. Lawrence, B. Legras, J. Logan, R. Mackenzie, C. Mari, C. Marquardt, P. Mascart, M. Milz, L. Moyer, R. Müller, W. Norton, L. Pan, T. Peter, D. Peters, L. Pfister, M. Ponater, A.R. Ravishankara, E. Ray, C. Reithmeier, G.-J. Roelofs, K. Rosenlof, K. Sato, C. Schiller, C. Schnadt, T. Schneider, U. Schumann, R. Scott, D. Seidel, T. Shepherd, S. Sherwood, M. Shiotani, E. Shuckburgh, P. Simon, L. Sparling, D. Stevenson, J. Thuburn, G. Vaughan, M. Volk, D. Waugh, P. Wennberg, H. Wernli, V. Wirth, D. Worsnop, A. Zahn.

Thus, the tropopause region exhibits a complex interplay between dynamics, transport, radiation, chemistry, and microphysics. This is particularly highlighted in the case of ozone and water vapour, which provide much of the climate sensitivity in this region.

Traditionally, atmospheric scientists have considered the tropopause region through a process-focused approach. The successful NATO/SPARC work-

the most basic questions about the tropopause region remain unanswered. This means that there is a need to go back to fundamentals.

To address this need, SPARC convened a workshop (co-sponsored by NASA) on the tropopause, which was held in Bad Tölz, Germany, from 17-21 April 2001. Its goal was to bring together a diverse group of scientists to consider various aspects of the tropopause region: what

it is, why it is what it is, how it affects climate, and how it might be expected to change in the future. The group included expertise in dynamics, radiation, chemistry, and microphysics, and likewise expertise in measurements, modelling, laboratory/field studies, and theory. The format was very open with a relatively small number of invited 'theme' talks, of a review/tutorial nature, and with plenty of time for open discussion led by a 'provocateur'. The meeting concluded with a half-day session led by 'rappor-teurs' who synthesised the salient points of the earlier sessions. The main conclusions drawn from the workshop are highlighted below.

Three other reports in this newsletter, written independently of the workshop (by **H. Fischer *et al.***, **A. Gettelman *et al.*** and **S. Sherwood**), are concerned with the tropopause region, and indeed the authors were participants at the Bad Tölz workshop.

General aspects

In a fundamental sense the troposphere is the part of the atmosphere in which the destabilising effect of solar heating of the surface leads to weak static stability and turbulent thermal energy transport. The stratosphere is the region above in which static stability is strong, thermal energy transport is relatively unimportant, and the dominant thermal balance is radiative. The tropopause is then the upper boundary to the turbulent energy transport (**T. Schneider**). In the tropics the turbulent energy transport is accomplished by moist convection. As

pointed out by **I. Held** (1982), in the extratropics it is more likely that the transport is accomplished by baroclinic instability.

If this basic picture is accepted then there are inevitable consequences for chemical distributions. From a chemical perspective, the stratosphere is a harsh UV environment, with relatively weak mixing and no rain-out, while the troposphere is a softer UV environment, with strong mixing and rain-out (**A.R. Ravishankara**). Within the troposphere the turbulence, be it convective or baroclinic, implies rapid contact with the surface. Chemical species which enter the atmosphere at the surface (this includes water vapour) are therefore relatively abundant. In the stratosphere, on the other hand, the chemical character of the air is likely to be very different, since the air has aged considerably since it left the surface and it may also have visited chemical source and depletion regions in the middle and upper stratosphere.

It used to be believed that the troposphere was an ozone sink, but it has become very clear in recent years that there is significant *in-situ* photochemical production (**A.R. Ravishankara**). In fact, typical estimates find that ozone transport from the stratosphere, while enough to refresh the tropospheric ozone burden every six months, is only about 10% of the photochemical production. (Of course some of the photochemical loss is closely coupled to photochemical production, so this figure could be misleading.) The upper troposphere in particular appears to be very chemically 'fertile' on account of

the presence of the OH radical, which is delivered to the upper troposphere by acetone (formaldehyde, methyl hydroperoxide and other species) in addition to the traditional ozone/UV/water source (**P. Wennberg**, **A.R. Ravishankara**). Ozone is produced in the so-called 'smog reactions' involving CO (hydrocarbon), OH, and NO.

The fact that some chemical species, particularly water vapour, are radiatively active implies that there is substantial coupling of dynamics, transport and chemistry. It is essential to understand this coupling thoroughly in order to make reliable predictions of future chemical-climate change (**V. Grewe**, **D. Stevenson**).

Tropics

Radiative-dynamical constraints

In the tropics, the cold-point tropopause (CPT) and the lapse-rate tropopause (LRT) are within 0.5km of each other at about 16-17km and have very similar temperatures, although the LRT is often ambiguous (**D. Seidel**). As there is probably not a useful distinction to be made between them, we can just refer to the tropical thermal tropopause (TTT). However, the TTT apparently has little significance in terms of the radiative-convective balance. Observations show that the vertical profile of temperature in the tropics matches that predicted by moist convective adjustment only up to about 11-12 km (which corresponds to the maximum Hadley-cell outflow). **E. Palmén** & **C. Newton** (1969, Fig.4.7) identified this level as the 'secondary tropical tropopause' (STT). Models of radiative-convective adjustment exhibit a similar structure, with CO₂ long-wave heating leading to a clear separation between the region of convective adjustment and the TTT (**J. Thuburn**). The notion of a 'tropical tropopause layer' (TTL) between the TTT and the STT was revived by **M. Atticks** & **G. Robinson** (1983) and, more recently, by **E. Highwood** & **B. Hoskins** (1998). Based on the overall radiative-convective balance, one might regard the TTL as being more stratospheric than tropospheric in character, and it has therefore also been dubbed the 'substratosphere' (**J. Thuburn**).

Tropopause acronyms

LRT: Lapse-rate tropopause: the conventional meteorological definition of the tropopause, in both tropics and extratropics, as the base of a layer at least 2km thick, in which the rate of decrease of temperature with height is less than 2K/km.

CPT: Cold-point tropopause: the level of minimum temperature. This is useful, and also significant, in the tropics.

TTT: Tropical thermal tropopause: since in the tropics the LRT and the CPT are usually less than 0.5km apart (with the LRT the lower) we ignore the distinction between them and refer to the TTT. The TTT is typically at 16-17km.

STT: Secondary tropical tropopause: the level of maximum convective outflow, above which the lapse rate departs from the moist adiabat. The STT is typically at 11-12km.

TTL: Tropical tropopause layer: the region between the STT and the TTT.

CSRT: Clear-sky radiative tropopause: the level at which the clear-sky heating is zero. Below the CSRT there is descent on average (outside convective clouds). Above the CSRT there is ascent on average. The CSRT is typically at 14-16km.

1 The altitude or potential-temperature values given here and below should be understood as rough values, which can be expected to vary with longitude and season.

On the other hand, the TTL is certainly penetrated by deep convection. There is some debate about the highest altitudes reached by convection, with evidence that it may go as high as 19km ($\theta=420\text{K}$), well above the TTT ($\theta=380\text{K}$) (S. Sherwood). However such events are very infrequent, and the frequency of convection drops rapidly with altitude above the STT (J. Holton). TRMM measurements suggest that only about 0.1% of the tropics has convection reaching the TTT at any given time. The distribution of convective penetration above the STT matches the distribution of equivalent potential temperature, θ_e , in the boundary layer (I. Folkins). The STT (11-12km, $\theta=345\text{K}$) corresponds roughly to the threshold value of θ_e above which convective available potential energy (CAPE) is positive and therefore deep convection is possible. The mode of the boundary-layer θ_e distribution falls at about the same value and the distribution is such that the frequency of convective penetration decreases rapidly as a function of θ for values greater than 345K. If one regards the stratosphere as being the region unreachable by convection, then the tropopause height would depend on the tail of the distribution of θ_e , but this would be an inherently fuzzy definition.

The upwelling Brewer-Dobson circulation in the tropical stratosphere is driven by a combination of effects including wave drag, transient radiative driving, and the stratospheric (i.e. non-convective) Hadley circulation. The relative role of these processes remains the subject of active research, although recent evidence suggests an important role for synoptic-scale wave drag in the subtropical lower stratosphere (W. Norton), rather than the extratropical planetary-wave 'pump' as had been previously proposed by J. Holton *et al.* (1995). This circulation certainly extends into the troposphere, cooling and elevating the TTT (see Figure 2). But the driving mechanisms do also, which makes any distinction between troposphere and stratosphere on this problematical basis.

Between $\theta_e=345\text{K}$ and about $\theta_e=355\text{K}$ (corresponding to roughly 13-14km), the rapidly decreasing likelihood of convection translates into a rapidly decreasing

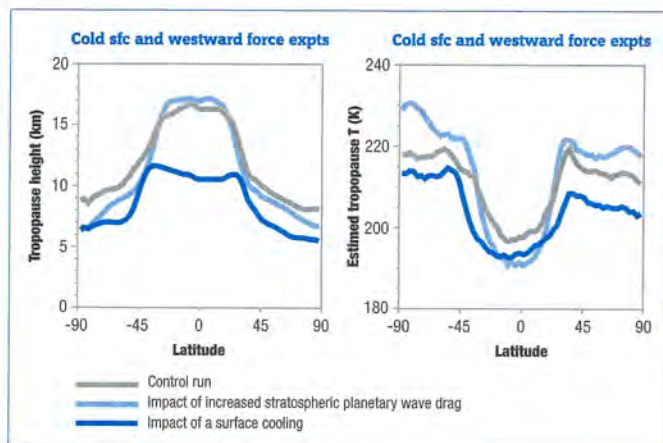


Figure 2: Sensitivity of the height and temperature of the tropopause to idealised changes in a general circulation model. The grey curves show the control case. The light-blue curves show the impact of increased stratospheric planetary wave drag (extending into low latitudes) which acts to raise and cool the tropopause in the tropics and to lower and warm it in the extratropics. The dark-blue curves show the impact of a surface cooling, which acts to lower and cool the tropopause everywhere, partly through the radiative effects of changes in stratospheric water vapour. The different character of the response to tropospheric and stratospheric changes is a useful fingerprint for attribution of observed changes. Courtesy of J. Thuburn, University of Reading.

convective outflow at the corresponding θ values, and the clear-sky radiative cooling appears to be essentially controlled by this outflow distribution (I. Folkins). This is the so-called 'mass flux scaling' (see Figure 3(a)). Of course some of the outflow also has to supply

another candidate tropopause, from a radiative-dynamical perspective, which we might call the clear-sky radiative tropopause (CSRT). From this point of view the TTT lies entirely within the stratosphere, and has no particular radiative-dynamical significance.

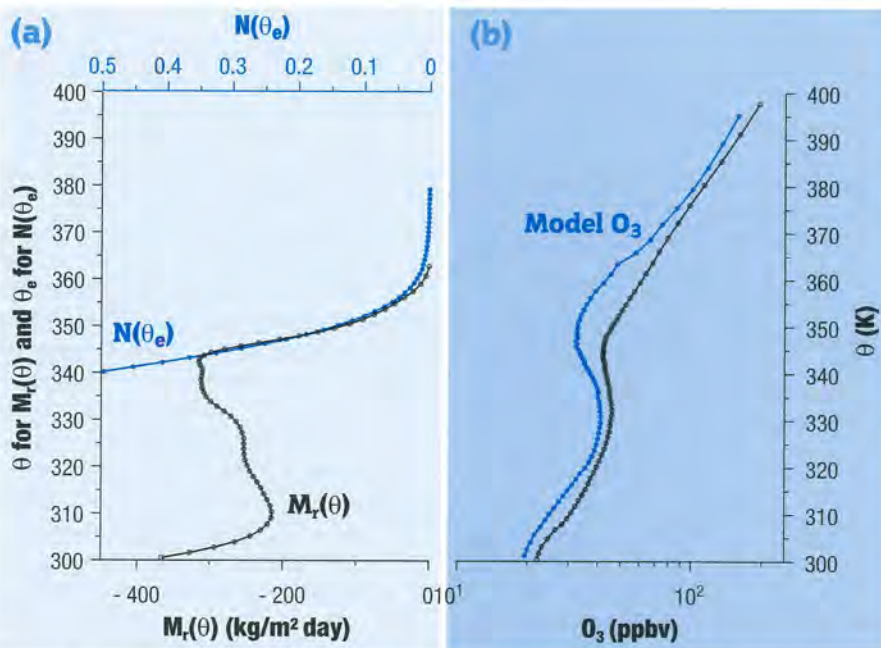


Figure 3: (a) Clear-sky radiative mass flux $M_r(\theta)$ and cumulative PDF of boundary-layer θ_e $N(\theta_e)$. M_r is evaluated between 30°S and 30°N . N is evaluated between 15°S and 15°N below 700 hPa, and is normalised to be unity at 310K. Below 345K, $M_r(\theta_e)$ is unrelated to $N(\theta_e)$; this is the region of moist convective adjustment. Between 345K and about 355K, they are seen to match closely; this is the region of so-called "mass flux scaling". (b) Ozone profile predicted from a one-dimensional model incorporating vertical transport in clear-sky regions, convective outflow (in which ozone mixing ratio is set to 25 ppbv) and chemical ageing, together with all tropical ozonesondes (within 15 degrees latitude of equator) from SHADOZ. Courtesy of C. Braun and I. Folkins, Dalhousie University.

Transport implications

None of the various levels described above are material surfaces under adiabatic conditions, which means that they are not transport barriers in any way (**J. Thuburn, B. Legras**). Thus the stratosphere and troposphere, however defined, cannot be regarded as distinct air masses on an instantaneous basis. In the case of the TTT, for example, the distribution of potential temperature θ (which does represent a material surface) at the TTT is a strong function of longitude (**A. Gettelman**), which means that air parcels close to the TTT move in and out of the stratosphere and troposphere, reversibly, as they move around the equatorial belt. Large-scale tropical waves, most notably Kelvin waves, further complicate this picture; the amplitude of Kelvin-wave effects on TTT parameters can be double that of the annual cycle. Thus, one must be careful of labelling air as 'tropospheric' or 'stratospheric' based on a sounding at a single location (**K. Sato**).

Below the STT, where moist convective adjustment holds, all air parcels have had recent contact with the boundary layer and chemical species are fairly well-mixed. Above this level, in the TTL, convective timescales are longer than radiative timescales (**J. Thuburn**) but all parcels, to the extent that the tropics are isolated, would have originated from the tropical boundary layer. Convective upward transport is thus the key to the delivery of short-lived chemically active species to the TTL. Chemical distributions in the TTL will then depend on the relative sizes of the convective timescales (i.e. the timescale since contact with the boundary layer), chemical timescales (since chemical concentrations may change through photochemical ageing), and timescales for lateral mixing with the extratropical lower stratosphere.

Above the CSRT, where the clear-sky mass flux is upwards, air parcels are on their way to the stratosphere (if they are not considered to be in it already), in some mean sense. However, convective overshoots and breaking waves can mix material downwards (**S. Sherwood, M. Fujiwara**), so one cannot say that air parcels are transported irreversibly into the stratosphere until they get well above the CSRT, if not above the TTT. It is worth noting that even if the region above the CSRT is 'strato-

spheric' from a radiative-dynamical perspective (because the net mass flux is upwards), for a tracer with a large contrast between tropospheric and stratospheric values, even a small mass exchange with air below the CSRT could be significant in which case the region might be regarded as 'tropospheric'.

A basic physical constraint on vertical transport is that air parcels must be heated in order for their potential temperature to reach stratospheric values. This can be achieved by slow ascent and radiative heating (the Brewer-Dobson circulation perhaps augmented by local radiative effects), or by overshooting cumulus followed by immediate mixing with relatively warm stratospheric air (**J. Holton**). The importance of the latter is severely restricted by the constraints on deep convection mentioned earlier. In any case the evidence from water vapour and cirrus clouds, discussed below, appears to point to the dominant role of slow ascent.

Water vapour

It is when it comes to water vapour that the importance of the TTT manifests itself. A. Brewer (1949) proposed the 'freeze drying' of air passing through the tropical CPT as the explanation for extreme stratospheric dryness, and this hypothesis has stood the test of time remarkably well (**M. Geller**). (Strictly speaking the relevant level would be where the water vapour saturation mixing ratio (over ice) is a minimum, not the CPT, but again we can just refer to the TTT as a first approximation to this level.) Not only does the annually- and tropically-averaged minimum water vapour saturation mixing ratio of about 3.8ppmv appear to correspond to the annual mean entry-level water vapour for the stratosphere (**A. Dessler**), the annual cycle shows lowest values of water vapour occurring in NH winter when the mean TTT temperatures are lowest, as manifested in the water vapour 'tape recorder' of **P. Mote et al.** (1996). The strong nonlinearity of the Clausius-Clapeyron relation makes it difficult to understand how the mean values of water vapour and temperature can correspond so well, given the longitudinal and temporal variations in temperature. However, improving on the simple Brewer picture in a quantitative fashion has proven to be remarkably difficult.

Because the lowest TTT temperatures occur over Indonesia, attention has historically focused on this region as the controlling entry point to the stratosphere (**J. Holton, M. Geller**). The 'stratospheric fountain' of **R. Newell & S. Gould-Stewart** (1981) hypothesised that most of the transport from troposphere to stratosphere occurred in this region. However, this idea has now been largely superseded. The region immediately above the TTT over Indonesia exhibits subsidence, so is, if anything, a stratospheric 'drain' rather than a 'fountain' (**S. Sherwood**).

In the past, the lack of observed widespread cirrus clouds was taken as indirect evidence for the importance of condensation and dehydration on the convective scale. However, recent evidence has accumulated of the existence of extensive subvisible cirrus clouds in the tropics; **Figure 4 (p. 1)** shows an example of an 'ultrathin' cirrus cloud at 17km altitude observed in the APE-THESIO campaign, having an optical density 300 times lower than the threshold for visibility from the ground (**T. Peter**). Recent CHAMP (improved GPS-MET) measurements support this picture by confirming the existence of widespread regions of low temperatures (**C. Marquardt**). Of course, dehydration is a continuous process, with the relative importance of slow ascent versus convection varying with altitude through the TTL, but it now appears possible that the 'final' dehydration (controlling the amount of stratospheric water vapour) is accomplished in slow large-scale ascent. Such ascent combined with horizontal transport can, in principle, allow all air parcels to pass through the region of lowest temperatures (**J. Holton, S. Sherwood**). In this case the location of the lowest temperatures becomes largely irrelevant. Certainly single station measurements show no particular relation between temperature and water vapour (**M. Shiotani**). However, the observed horizontal inhomogeneity of the water vapour field above the TTT (**K. Rosenlof**) suggests that there are transport barriers associated with circulations within the tropics, so that not all air parcels entering the stratosphere experience the lowest temperatures. It is not until about $\theta=450\text{K}$, corresponding to an altitude of about 20km, that the water vapour field becomes horizontally homogeneous. Further progress on this issue will likely require accurate parcel trajectory calculations in the tropics, to

assess the temperature history of ensembles of air parcels. Only with such information will quantitative testing of Brewer's hypothesis be possible. The possible role of Kelvin waves, which are observed to affect cirrus formation (K. Sato), may need to be considered in such calculations.

But the assumption of 'freeze drying' occurring instantaneously and determined by local temperatures is itself also clearly an oversimplification. Unless ice particles fall out, they will evaporate and re-moisten the air once it has left the region of lowest temperatures. Thus the microphysics of dehydration is crucial to explain stratospheric dryness (T. Peter). It is fair to say that the fall out of ice particles has not been satisfactorily explained. If the air is clean (as one might imagine for slow ascent) then homogeneous nucleation is required, leading to large supersaturations. Roughly speaking an additional 4K cooling is required for homogeneous nucleation, which would produce another 1ppmv of water vapour compared to saturation values - a discrepancy which would need to be accounted for (T. Peter). Heterogeneous nucleation would reduce supersaturations, but is not well understood at this time. Heterogeneous nucleation is expected in convective overshoots, containing 'dirty' air from the boundary layer, but its extent in regions of slow ascent remains to be ascertained.

Chemistry

From a chemical point of view, also, the TTL is a transition region with characteristics of both troposphere and stratosphere. Indeed, measurements show that ozone increases with altitude throughout the TTL (M. Fujiwara), although there is a notable jump in the vertical gradient of ozone at the TTT (J. Logan). This picture is entirely consistent with the radiative-dynamical picture discussed above. Below the STT, air is well mixed, rain-out is effective, and ozone values are low (and roughly independent of altitude). This is the pure troposphere, chemically speaking. Deep convection brings low values of ozone from the lower troposphere into the TTL (M. Lawrence), together with short-lived species, but following detrainment into clear-sky regions the ozone levels increase through photolysis as the parcels age (I. Folkins). This effect together with the decreasing

mass outflow with altitude combine to produce an ozone profile remarkably like observations (see Figure 3(b)). Of course lateral mixing with ozone-rich air from the lowermost stratosphere must have an effect on these profiles too (D. Waugh), but this effect is not well estimated.

In addition to direct radiative effects through gas-phase species (principally ozone), chemistry affects the radiation budget indirectly through the growth of aerosols. Organic aerosols are especially important in the TTL, where they (and their precursors) are brought by convection and also produced *in-situ* by the smog reactions (D. Worsnop). The extent of the convective transport and the roles of both scavenging and rain-out in opposing such transport remain significant uncertainties (C. Mari). This applies to gas-phase chemicals as well as aerosols.

The preceding considerations would suggest that, from a chemical perspective, the atmosphere becomes increasingly 'stratospheric' as altitude increases above the STT. However, because water vapour is so important for chemistry (principally through its effect on OH), the TTT plays a distinguished role. To this one might add the chemical effects of cirrus clouds, which are confined to the TTT and below. Because air is pretty much irreversibly transported to the stratosphere once it gets 1km or so above the TTT, with no possibility of scavenging, the TTT essentially sets the chemical lower boundary condition for the tropical stratosphere (P. Wennberg). In this sense the TTT may be seen as a semi-permeable membrane, or trap door, that allows transport of air, but not water vapour, into the stratosphere (A.R. Ravishankara).

Variability and climate sensitivity

The temperature and altitude of the STT are controlled by moist convective adjustment, and are dependent on the moist static energy at the surface which is a strong function of temperature (B. Boville). Increasing the moist static energy should lead to a higher and warmer STT, independently of anything that might happen in the TTL and above. In contrast, the TTT reflects a balance of stratospheric and tropospheric radiative-dynamical processes, and its behaviour would

appear to be considerably more sensitive. Changes in stratospheric radiative properties should lead to a negative temperature-height (T-z) correlation, and this is indeed observed following volcanic eruptions when the TTT descends and warms (P. Forster). Upwelling from stratospheric wave drag effectively lowers the radiative equilibrium temperature at the TTT, also leading to a negative T-z correlation (J. Thuburn). This is seen in Figure 2 and appears to account for the negative T-z correlation seen in the annual cycle of the TTT as well as its QBO signal (D. Seidel).

In contrast, changes in surface temperatures are predicted to lead to a positive T-z correlation at the TTT (see Figure 2). If robust, this could provide a useful 'fingerprint' to distinguish stratospheric from tropospheric driving. For example, it points to the stratosphere being the cause of the annual cycle in TTT temperature, as argued originally by E. Yulaeva *et al.* (1994). It would also explain why it is that the coldest TTT occurs over Indonesia while the highest TTT occurs over South America and Africa (D. Seidel); land convection is much more energetic than ocean convection, leading (perhaps indirectly) to a warmer and higher tropopause in the latter regions (L. Pfister). (Note that this statement concerns only the radiosondes; the ERA-15 reanalyses have the coldest and highest TTT coinciding.) On the other hand, the ENSO signal, which is presumably tropospherically driven, has a negative T-z correlation (D. Seidel). This may point to the limitations of local considerations in trying to explain longitudinal differences; in principle one should consider a Gill-like solution to localised convective forcing, which would describe the spatial response. This might already help account for the fact that the lowest temperatures in Indonesia are not directly above the most vigorous convection (L. Pfister, S. Sherwood, A. Gettelman).

GCM experiments with enhanced greenhouse gas forcing produce a surface warming and therefore lead, by the above arguments, to significant warming of the STT. However whilst surface warming has been observed, the warming at the STT has not, which is something of a puzzle (B. Boville). The effects on the TTT from greenhouse gas forcing depend sensitively on the treatment of water vapour

radiative effects in the GCM (something which is typically not done very accurately); e.g. the NCAR CCM shows a warming, while most other GCMs show a cooling (**P. Forster**). Ozone changes are also expected to provide a significant climate feedback near the TTT, but the feedback is definitely two-way and requires a chemical climate model for its prediction. Increased emission of pollutants should increase upper tropospheric ozone, but greenhouse gas induced climate changes could well mitigate these changes through reduced PAN (which is highly sensitive to temperature) and reduced CH₄ (**D. Stevenson**). However, it must be said that there is a great deal of uncertainty attached to these predictions. Probably we need to be able to explain the past record before we can take future predictions seriously (**A.R. Ravishankara, J. Logan**).

Extratropics

Radiative-dynamical constraints

In the extratropics, as in the tropics, practical differentiation between troposphere and stratosphere may be on 'meteorological' or chemical grounds. The WMO lapse-rate definition of the tropopause was formulated on the basis of extratropical observations and rarely fails badly in the extratropics. The lower stratosphere tends to be rather isothermal and this guarantees a transition between tropospheric lapse rate and zero lapse rate, but makes a CPT ambiguous. The lapse-rate criterion does sometimes fail in Antarctic winter, where the very low temperatures in the lower stratosphere may imply that lapse rates that are tropospheric according to the WMO criterion continue well into the stratosphere. More generally, multiple LRTs are often observed due to tropopause folding. One of the primary disadvantages of the lapse rate definition is that through purely reversible dynamical changes air may change from 'tropospheric' to 'stratospheric' and back again. This has motivated the definition of a 'dynamical' tropopause, based on potential vorticity (PV) value, since PV is materially conserved on adiabatic timescales. Because a jump in lapse rate corresponds to a jump in PV there is a relation between the two definitions, although they differ significantly in cyclones and anticyclones (**V. Wirth**). Various threshold values of PV have been used to mark the

tropopause. This is a useful practical criterion and may be used as a basis for transport diagnostics (see later), but there is no reason why the tropopause should be marked by a single value of PV. **Figure 5 (p. I)** shows a typical height-latitude cross-section of PV at one longitude, showing the tropopause as a potential vorticity contour.

Simple radiative models allow the tropopause height to be predicted in terms of the tropospheric static stability, if the stratospheric radiatively determined temperature is known (**G. Craig**). However the static stability is unknown *a priori* and must be predicted as part of any complete radiative-dynamical model for the tropopause. Quantitative understanding of the extratropical tropopause remains inadequate, mainly because of this difficulty and its close relation to the classical problem of baroclinic eddy parameterisation, i.e. predicting baroclinic heat fluxes as a function of external parameters (**G. Craig**). Numerical experiments in simplified atmospheric models have shown that baroclinic instability alone can provide the necessary heat transport and lead to a realistic tropopause structure (**P. Haynes, T. Schneider**). Cleverly designed numerical simulations, e.g. in general circulation models, have provided substantial insight in a more realistic context. For example, recent experiments by **Barry, Craig & Thuburn** varied the strength of radiative heating terms and showed that in the transition from the strong heating limit (in which the static stability tends towards dry neutral) to the weak heating limit (in which the potential vorticity gradient on isentropic surfaces tends to zero), the corresponding change in tropospheric static stability leads only to a modest change in tropopause height (from 7.5km to 11km).

Transport implications

In the extratropics the tropopause may sensibly be regarded as a permeable transport barrier marking the boundary between different air masses. This is why a quasi-Lagrangian definition of the tropopause, e.g. based on a particular value of PV, is useful.

Exchange across the tropopause in the extratropics is accomplished by processes on a large range of scales. Whilst net exchange rates might best be considered as controlled by global-scale processes, as part of the Brewer-

Dobson circulation, two-way exchange occurs on synoptic and sub-synoptic scales and may have important consequences for chemical species. Synoptic-scale disturbances frequently bring air from the boundary layer to the upper troposphere, leading to strong chemical contrasts in the neighbourhood of the tropopause. Recent quantitative studies have reflected the fact that no single measure of exchange is likely to be adequate. One approach is to use a hybrid Eulerian-Lagrangian quantification in which fluxes across relevant control surfaces, such as the 1PV surface, are counted only if the corresponding air parcels then stay in the stratosphere or the troposphere for a minimum time (**H. Wernli**).

Ozone is a convenient species to use as a chemical basis for differentiation between tropospheric and stratospheric air, at least in winter when there is little photochemical production of ozone within the troposphere. Detailed analysis shows that sharp vertical gradients of ozone are often co-located with sharp vertical gradients of static stability. Ozone profiles often give a clear signature of the difference between tropospheric and stratospheric air in cases where the LRT is a poor indicator (e.g. in cyclones). The ozone-defined tropopause defined by **S. Bethan et al.** (1996), on the basis of observations primarily in winter and spring, appears to be located about 1km below the LRT (**G. Vaughan**).

Chemical measurements (e.g. of ozone and carbon monoxide) may be used to follow the transition across the tropopause from air with a tropospheric character to air with a stratospheric character. In winter there is little evidence for tropospheric air in the lowermost stratosphere (above 330K), suggesting that the extratropical tropopause is an effective barrier to transport. In summer, on the other hand, air in the lowermost stratosphere has substantial tropospheric chemical content indicating that the extratropical tropopause is a much less effective barrier to transport (although latitudinal gradients of chemical species persist in the lower stratosphere through the summer) (**H. Fischer**). (See **Figure 2** of **H. Fischer et al.** in this issue.) These winter-summer differences inferred from chemical evidence are consistent with passive tracer calculations using observed large-scale winds. They are also consistent with the observed

annual cycle of water vapour in the NH lowermost stratosphere, e.g. from POAM, which shows the widespread presence of high values in summer (C. Schiller). It has been suggested that this moistening arises through transport associated with the Asian summer monsoon. The possibility should not be ruled out that convective penetration into the summertime lowermost stratosphere, e.g. associated with deep convection over continents, may play an important role in the observed annual cycle of water vapour and other chemical species (S. Sherwood). However the annual cycle of downwelling is also relevant; in winter the stronger downwelling keeps the lowermost stratosphere dryer (C. Schiller).

Chemistry

As noted in the introduction, the upper troposphere is now understood to be far more chemically fertile than previously believed. Relatively short-lived species such as acetone play a significant role in the *in-situ* production of ozone in the upper troposphere. The possible impact of cirrus clouds on the ozone budget has yet to be determined. Nevertheless, if chlorine is present it must be activated on cirrus, so this activation could be a useful diagnostic of cirrus clouds (A.R. Ravishankara).

The seasonal cycle of ozone in the extratropics differs between lower stratosphere and upper troposphere. In the lower stratosphere the maxi-

mum ozone mixing ratios are observed in spring, as a result of downward advection of high mixing ratios by the winter Brewer-Dobson circulation. In the upper troposphere the maximum ozone mixing ratios are observed in summer, possibly as a result of *in-situ* photochemical production. The stratospheric signal is clear 2km above the thermal tropopause. The signal has a more transitional character 1km above and is tropospheric at the thermal tropopause itself (J. Logan); see Figure 6. At first sight, this is difficult to reconcile with the location of the ozone tropopause mentioned above, but the latter was determined without considering seasonal variation.

Variability and climate sensitivity

Column ozone decreases have been observed in mid-latitudes over the last few decades. The major part of the observed ozone trend occurs in the lowermost stratosphere, but in this region the budget of ozone is not well understood. Chemical and transport timescales are roughly comparable, making the problem especially difficult. Transport is obviously crucial to bring ozone to this region in the first place, but variations in transport are not well quantified. An important contribution to the ozone trend in this region may come from a trend in the tropopause height itself (J. Logan). It is notable in this respect that the ozone trends have significant longitudinal variations in this region.

Aircraft emissions of NO_x in the upper troposphere/lower stratosphere region have been studied in detail and are expected to lead to ozone increases in the lower stratosphere. Aircraft observations of chemical species show a great deal of small-scale spatial structure, with transition from tropospheric to stratospheric characteristics over horizontal scales of about 1km. There is almost certainly corresponding small-scale structure in water vapour. Reproducing this structure and, perhaps more importantly, estimating its effects on the large scale is an outstanding modelling problem (U. Schumann). Supersaturation with respect to ice is frequently observed in the extratropics (U. Schumann, C. Schiller), though is not presently represented in analyses. There may be local freeze drying in the extratropics, down to water vapour mixing ratios of less than 10 ppmv (C. Schiller).

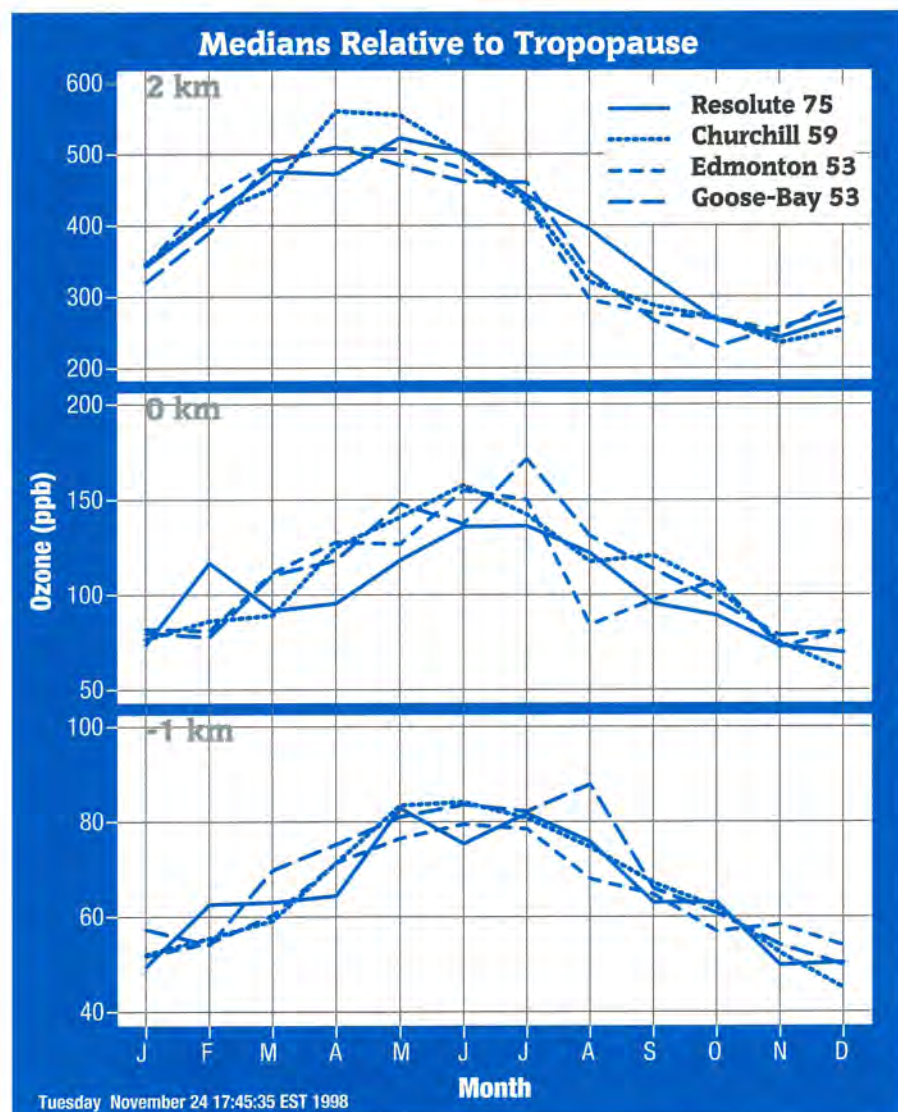


Figure 6: Annual cycle of ozone in the extratropics as a function of altitude relative to the LRT. The dramatic increase in ozone mixing ratio with altitude from troposphere to stratosphere is evident. The lower stratospheric annual cycle, seen 2km above the LRT, has a maximum in March/April, whilst the tropospheric annual cycle, seen 1km below the LRT, has a maximum in June/July. The annual cycle at the LRT has the same phase as the tropospheric cycle. Courtesy of J. Logan, Harvard University (see Logan (1999) for further details).

Conclusion

A greater proportion of the meeting was spent discussing tropical rather than extratropical issues. This may not seem surprising, given the crucial role that the tropics plays in determining stratospheric water vapour. (The dehydrating role of the Antarctic winter lower stratosphere was not discussed at this meeting.) But it also reflects the fact that, out of a set of investigations that began rather independently, there has been significant progress in our understanding of this region. As might be clear from the report, some of the discussion at the meeting concerned the advantages and disadvantages of different tropopause definitions. However, there is little to be gained, for example, from arguments over whether the TTL is more strato-spheric or tropospheric. What is more significant is that a coherent picture of the processes taking in the tropics is now beginning to emerge in which there is a great deal of consistency between different (radiative-dynamical, transport, chemical) viewpoints. This picture furthermore leads to the recognition of a region below the TTT where chemistry can be especially important for climate.

The picture in the extratropics is perhaps more complicated and certainly less coherent. Yet here too there is some encouraging consistency between consideration of dynamics

and transport on the one hand and chemical measurements on the other, e.g. with respect to the seasonal variation of the extratropical lowermost stratosphere.

The biggest remaining uncertainties in both tropics and extratropics, as might be expected, concern coupled processes. Our quantitative picture of tropical dehydration is far from complete and major uncertainties remain over microphysics and the role of convection. Coupled chemical-climate modelling of the tropopause region is also in its infancy, although this is precisely the region where one might expect the largest sensitivities.

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● Peter Haynes, Ted Shepherd

Report on the SPARC/IOC Workshop on Understanding Ozone Trends

Department of Meteorology,
University of Maryland, College Park, USA,
March 7-9, 2001

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Background

In the past few years, continued observations and recent analyses have raised a number of questions about how well past ozone changes can be explained in quantitative terms. In particular the importance of changes in

halogen loading, stratospheric aerosol loading, atmospheric dynamics radiation and temperature have been assessed in a number of ways. However there is no general consensus on the relative contributions of these factors to past ozone changes or on

how to assess them in a consistent manner. This is an important topic as the peak stratospheric halogen loading is passed. Such issues will be discussed in the next (i.e. 2002) WMO-UNEP Scientific Assessment of Ozone Depletion.

Accordingly, a workshop on the understanding of ozone trends was organised to stimulate discussion and relevant work ahead of the preparation of the 2002 assessment and to see what issues will require research over the longer term. This was done at the initiative of the SPARC Ozone Trend working group and the International Ozone Commission (IOC) in close conjunction with the co-chairs of the Montreal Protocol Scientific Assessment Panel and with WMO/GAW. Over 40 leading scientists in this field attended the workshop, including members of other SPARC working groups. The programme was balanced between presentations of new results and discussion of the issues raised.

Scientific Issues

There are a number of important scientific issues:

- The statistical model of ozone trends used in recent assessments has not allowed for long-term dynamical changes other than those related to the QBO and solar cycle. Nor has it yet incorporated the fact that the chlorine loading has stopped increasing and that the ozone-depleting efficiency of chlorine compounds is strongly modulated by stratospheric aerosols. The statistical approaches used to calculate ozone trends may need to be updated and developed in sophistication to address today's questions.

- There has been increasing evidence from published studies and work in progress that long-term dynamical changes (possibly related to climate change) may have been responsible for a significant fraction of the ozone trends derived from measurements at mid-latitudes. The studies vary from combined analyses of ozone and meteorological data through increasingly sophisticated 2D model studies using time-varying aerosols, etc., to 3D model runs using real winds. Estimates for the contribution for the dynamical contribution to the observed ozone trends have varied from zero to 100%. Simple-minded addition of all the current dynamical and chemical explanations of the mid-latitude trends would account for nearly twice the observed trend. Sound scientific approaches need to be developed so that these influences on ozone can be allowed for consistently in the statistical models used in the assessment.

- There are some signs of a change of behaviour, most obvious in the tropics, in the TOMS measurements of ozone since 1998 and a further shift in 2000. While discussion of issues related to the data quality of the various observing systems were not a major part of this workshop, attention was given to identifying whether such features in the observational record are real and to updating the estimates of the instrumental uncertainties.

Aims

The main aims of the workshop were to:

- Identify the major current issues concerning ozone trends,
- Improve quantification of the contributions and uncertainties of the chemical and dynamical mechanisms to observed ozone trends, particularly at mid-latitudes,
- Identify how to assess the consistency of these proposed contributions.

Success in meeting these aims has resulted in significant progress in the preparations for the forthcoming 2002 WMO-UNEP assessment. Steps that will assist the assessment process (further work, comparisons, publication, etc.) were explicitly discussed at the workshop.

In addition, although not a primary aim, there was some discussion on how the approaches that can be developed to improve our understanding of past changes could be used to identify the recovery of the ozone layer as a result of actions taken under the Montreal Protocol.

Discussion and findings

A. Data quality and availability

The overall quality and availability of ozone data are in reasonably good shape. Clarification is needed however on the possible satellite bias between the Northern and Southern hemispheres and some apparent degradation in EP-TOMS in 2000. Several composite data sets for total ozone, in which the instrument to instrument biases have been addressed, have been prepared. The aim is to inter-compare these composite data sets and to have identified and understood discrepancies by December 2001, in time for the assessment. The need for continued high quality ozone data from ground and space was clearly recognised.

B. Chemical Influence

Current estimates of chemical loss of ozone, which come principally from 2D models, have changed little since the 1998 WMO-UNEP assessment, the biggest change resulting from improved reaction rates for NO_x chemistry. The 2D models can reproduce most of the northern mid-latitude ozone trends. In particular, several 2D models have simulated the observed drop in ozone following the Mt Pinatubo eruption in 1991, as well as the more recent higher values, supporting the link between chlorine chemistry and aerosol loading. The agreement between measurements and models does not appear to be as good for the Southern Hemisphere. It has proven hard to accurately quantify the impact of polar chemical processes at mid-latitudes (a certain influence on the mid-latitude trends) from measurements and from models. The high altitude loss should be the simplest to model.

C. Dynamical Influence

There was much discussion on the influence of dynamics on ozone abundances and trends. Ozone trends in the lowermost stratosphere are particularly sensitive to changes in dynamics, whether changes in Rossby wave events, in the frequency of low ozone events (mini-holes and laminae) or changes in principally tropospheric phenomena such as the North Atlantic Oscillation / Arctic Oscillation. Diagnosing the links between these undoubtedly coupled phenomena is hard, but it is necessary in order to avoid 'double counting' when calculating ozone trends.

In the long term, the magnitude of the dynamical influence on ozone trends depends critically on whether the short term relations between dynamics and ozone stay constant (as currently assumed) or vary with time as the overall atmospheric system responds to the original forcing (whatever its origin). As a result, the dynamical influence depends on the time period considered. The current estimates of the influence are likely to be upper limits. It is important not to simply add ozone trends derived from individual statistical studies looking at the influence of one particular dynamical process, as the various dynamical processes in the atmosphere are clearly related.

It is unclear which dynamical proxies should be used in statistical models. A better understanding of which dynamical processes affect the ozone is clearly needed, as well as any radiative feedbacks or radiative forcings from the changes in stratospheric water vapour or greenhouse gases. Simple 3D CTMs, driven by assimilated winds, show a large inter-annual dynamical influence on ozone trends and this will affect the estimate of ozone trends. It is hard to attribute causality in models; i.e. what is a feedback and what is a driving force? Improved understanding of ozone-climate interactions and past temperature trends is needed for improved characterisation of past ozone changes.

D. Regimes

Another approach, which takes account of atmospheric motions, is to consider the atmosphere as a number of regimes with boundaries chosen according to dynamic criteria rather than simple geographic criteria, which in ozone trend studies have typically been latitude bands. New work was presented that uses the meteorological tropopause fronts to separate total ozone into regimes. Narrow distributions of total ozone were found if the regions close to the fronts were excluded. Within the regimes, small trends in total ozone were observed. Changes in the relative contributions of the tropical, mid-latitude and polar regimes to the overall trends at mid-latitudes were also reported. Trends in the lowermost stratosphere, calculated relative to the tropopause height, have been used to distinguish stratospheric and tropospheric effects. In addition, polar ozone trends have been investigated using equivalent latitudes on a number of isentropic surfaces in the lower stratosphere.

E. Use of "Fingerprints"

There is a need to identify, consistently, the spatial (latitude, longitude and altitude) and temporal signatures expected in ozone as a result of the various chemical and dynamical processes. Attempts should be made to quantify ozone loss, temperature, chlorine, etc. at 40 km using adapted climate 'fingerprinting' techniques. At lower altitudes one can probably only make qualitative estimates at present. Wherever possible, external variables, such as water vapour, methane, hydrogen chloride, temperature, etc. should be used.

F. Statistical models

Much of the discussion on this subject centred on which proxy variables should be used to describe the dynamical influence and the halogen-induced ozone loss. Ideally such proxies should be independent (orthogonal), but this may not be possible given that the atmospheric processes involved are coupled. This issue remains an important one to resolve.

The development of tools to investigate the turnaround resulting from decreased halogen loading is progressing, and it currently seems that any turnaround will not become observable until at least 2010. An improved understanding of the processes affecting ozone trends is an essential element of this work. Most of the current statistical effort is being devoted to the further development of time series approaches rather than the introduction of different statistical techniques.

The way forward

The main areas of activity in the coming year will be the continuation of the on-going research projects and the preparation of the 2002 WMO-UNEP report. In order to facilitate general discussion and information flow on this issue, the IOC will host a joint IOC/SPARC web site on understanding ozone trends. It is anticipated that a further joint initiative will be undertaken in the second half of 2002 which will promote progress on the issues which are still outstanding after the WMO-UNEP report's considerations.

As stated above, the understanding of the chemical depletion of ozone has not changed much since the 1998 WMO-UNEP report, and those advances which have occurred have done so principally as a result of revisions of laboratory data. Recently there has been more work using chemical composition measurements to provide constraints on the chemical loss mechanisms in the lower stratosphere, and this work needs to continue. A thorough evaluation of high altitude ozone trends taking into account all the known or likely changes in chemical composition, temperature and possibly large-scale circulation in that region is still needed.

It is now clear that the effect on ozone of the decadal changes in dynamics

needs to be considered in the calculation of ozone trends. What is currently less clear is how to evaluate the magnitude of their effect on ozone quantitatively and how to separate purely dynamical changes from each other and from those induced by the ozone changes themselves. A more comprehensive account of all the dynamical processes needs to be made simultaneously on a hemispheric basis. The very recent work done where total ozone is separated according to meteorological regimes at the tropopause provoked a great deal of interest, but how the results relate to the previous view of mid-latitude ozone depletion remains a puzzle.

Developments in 3D models have meant that these have started to be used in serious trend studies and they offer a new way to assess the relative importance of chemistry and dynamics. However there is still a lot of work to be done in this area in order to get realistic descriptions of chemical processes, to develop appropriate diagnostics to distinguish cause from effect and to evaluate the consistency with the current 2D models.

Overall, further developments are likely to be made along the general principles governing the evolution of the overall SPARC programme. These will involve progress toward a more integrated understanding of all the changes which have occurred in the stratosphere (water, temperature, radiation, dynamics as well as chemical composition including halogen and aerosol loading). However a word of caution is needed here as the uncertainties associated with the observed changes may preclude an exact interpretation.

Acknowledgements

We wish to thank the participants of the meeting for their enthusiasm during the workshop and for their assistance in the preparation of this summary. We thank SPARC, the University of Maryland, the EC Research DG (through the CRUSOE concerted action – EVK2-1999-00252) and the UK DETR Global Atmospheres Division for their support in the organisation of the workshop, as well as the many other funding agencies who helped support individuals to attend.

Attendees

A-L. Ajavon, D. Albritton, D. Balis, G. Bodeker, R. Bojkov, D. Cunnold, V. Fioletov, E. Fleming, L. Flynn, P. Forster, S. Godin, N.R.P. Harris, S. Hollandsworth, L. Hood, R.D. Hudson,

I. Isaksen, J. Kerr, J. Logan, R. McPeters, G. Mégie, A.J. Miller, H. Nakane, M. Newchurch, P. Newman, S. Oltmans, R. Portmann, M. Proffitt, J.A. Pyle, V. Ramaswamy, W. Randel, S. Reid, G. Reinsel, N. Sabogal, S. Solomon,

J. Staehelin, R. Stolarski, O. Uchino, B. Weatherhead, M. Weber, D. Wuebbles, J. Zawodny, C. Zerefos.



Report on the Joint US-Japanese Workshop on "Coupling of the Troposphere and Stratosphere by Dynamical, Radiative and Chemical Processes"

Kyoto, Japan, 13-17 March, 2001

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This workshop was jointly sponsored by the Japanese Society for the Promotion of Science and the US National Science Foundation, with additional support from the Japanese Ministry of Education, Culture, Sports, Science and Technology (Grant-in-Aid for Scientific Research on 'Stratospheric Variations and Climate') and the Kyoto University Foundation. The workshop was arranged on the occasion of the retirement of Professor Isamu Hirota from Kyoto University, and jointly organised by Shigeo Yoden (Kyoto University) and William Randel (National Center for Atmospheric Research). A web page discussing the workshop program can be found at <http://www-mete.kugi.kyoto-u.ac.jp/Kyoto2001/index.htm>

Dynamical links between the troposphere and stratosphere are inevitably two-way. Several kinds of waves generated in the troposphere propagate into the stratosphere and interact with the mean flow, while the stratospheric circulation also has influences on the large-scale flow in the troposphere. Radiative, chemical and transport processes (such as ozone transported across the mid-latitude tropopause) are also important for coupling these regions. Although mean structure and variability have sometimes been investigated separately in the troposphere and middle atmosphere research communities, the importance of coupling is becoming more evident in data analysis and modelling studies.

The main objective of the workshop was to foster continued collaborations

between Japanese and US scientists involved in stratospheric and tropospheric research, with a focus on studying mechanisms which couple the two regions over a broad range of time scales. The participants consisted of a number of Japanese and US researchers (approximately 15 from each country), plus a few scientists from Canada, Europe, Asia and Australia. Also included were a number of younger (postdoctoral-level) Japanese and US scientists. The workshop was organised according to the themes of:

- Overviews and General Circulation Model studies,
- Transport and chemistry,
- Wave processes,
- Intraseasonal and interannual variability.

In between four full days of invited lectures covering the preceding topics, one day was devoted to a morning of presentations by young scientists combined with an afternoon celebrating 'Professor Hirota and his old friends'. The morning lectures were arranged in the form of alternating Japanese and US talks, with friendly competitive 'matches'. The afternoon featured talks by Professors Taroh Matsuno, Mike Wallace, Hiroshi Yanaka, Jim Holton, Susumu Kato, Alan Plumb, Saburo Miyahara and Matt Hitchman reminiscing on the career of Professor Hirota, including highlighting his pioneering research and original contributions on the general circulation

and waves in the middle atmosphere. Later, Professor Hirota gave a keynote presentation on the 'Early days of equatorial middle atmosphere studies: personal retrospective', providing a unique perspective on the leading scientists and important early discoveries in middle atmospheric science. Finally the entire conference shared a traditional Japanese dinner at a Zen-style Tofu restaurant.

A written proceedings of this workshop will take the form of a special issue of the Journal of the Meteorological Society of Japan, containing review papers or original research articles from each of the workshop participants, with a publication date of early 2002. Overall the workshop exceeded expectations in terms of combining high quality science with many opportunities for interactions, all in the historic setting of Kyoto. This workshop will serve as a lasting tribute to the distinguished career of Professor Hirota (see **photography p. II**).



Northern Hemisphere Winter Climate Response to Greenhouse Gas, Ozone, Solar, and Volcanic Forcing

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Introduction

Observations show large increases in surface temperatures over the Northern Hemisphere (NH) continents during winter over the past few decades (Figure 1, bottom p. II). Large areas have warmed at a rate more than an order of magnitude larger than the global annual average rate. These large changes appear to result largely from an increase in the westerly air flow around the NH, which brings warmer, wetter oceanic air over the continents and cooler interior air to the eastern coasts and the oceans. This westerly flow is associated with the leading variability pattern of NH cold season sea level pressure (SLP), called the "Arctic Oscillation" (AO) (Figure 1, middle p. II). This oscillation is a hemispheric scale pattern which contains the North Atlantic Oscillation (NAO) in its Atlantic sector. Observations show an apparent upward trend in the amplitude of the AO pattern, or equivalently, a bias toward the positive phase of the pattern, in recent decades [Thompson and Wallace, 1998; Thompson *et al.*, 2000]. The variability associated with this pattern extends from the surface up into the stratosphere, and in fact the variability pattern can be equivalently defined as being composed of variability at all levels from the surface through the lower stratosphere [Baldwin and Dunkerton, 1999]. The Goddard Institute for Space Studies (GISS) climate model is able to reproduce the observed trend in the AO in response to increasing greenhouse gases [Shindell *et al.*, 1999a] only with the inclusion of a well-resolved stratosphere. Consistent with those results, observations indicate that changes in the stratospheric circulation typically precede the changes in the lower atmosphere [Kodera and Koide, 1997; Baldwin and Dunkerton, 1999]. As a result, the stratospheric climate model was better able to reproduce the large continental wintertime warming trends seen in the NH than a similar "tropospheric model" (Figure 1, top p. II).

It is interesting to explore the various ways that stratospheric changes can influence tropospheric circulation. For example, it has been shown that the circulation in the lower atmosphere can be affected by stratospheric changes such as large volcanic eruptions [Rind *et al.*, 1992; Graf *et al.*, 1993; Kodera, 1994] or solar forcing [Haigh, 1999; Shindell *et al.*, 1999b]. But what is the mechanism by which stratospheric perturbations affect surface changes? Greenhouse gas increases, volcanic aerosols, ozone depletion, or solar forcing can all induce meridional temperature gradients in the stratosphere during wintertime. GCM experiments reveal a positive feedback whereby zonal wind anomalies induced by these thermal gradients near the tropopause are amplified through planetary wave refraction. Deflection of upward propagating tropospheric waves at lower and lower altitudes and the resulting changes in angular momentum transport in effect carry the anomaly steadily down from the stratosphere, allowing stratospheric changes to affect surface climate by altering tropospheric energy flow.

It is also useful to examine the extent to which each individual stratospheric forcing excites the natural patterns of variability. This is accomplished by projecting the induced changes onto empirical orthogonal functions (EOFs), fixed spatial patterns of variability ranked by the amount of variability accounted for by each pattern [e.g., Kutzbach, 1970], derived from a control run with no external forcings. In the observations, the recent trend in SLP has occurred primarily through enhancement of the AO (EOF 1) [Thompson *et al.*, 2000]. In order to reproduce the observations, it is therefore necessary that a model yield a trend in the amplitude of the AO pattern similar to the observations, while at the same time keeping the amplitudes of changes in the other variability patterns small. The analysis of our model simulations shows that while greenhouse gas increases, volcanic

aerosols, or Arctic ozone depletion can excite primarily this leading mode, only increasing greenhouse gases can excite an AO trend comparable to the observed value.

Simulated AO response

Several forcings may have contributed to surface climate change via the stratosphere. Increases have been observed in the abundance of atmospheric greenhouse gases, Arctic winter-spring ozone loss, and solar irradiance over decadal and longer time scales. While there is not thought to have been a long-term increase in volcanic eruptions, they can periodically cause very large perturbations to the stratosphere, providing an important test of the stratosphere's influence on surface climate. A systematic comparison of the mechanism by which perturbations to the stratosphere affect surface climate has been performed with the GISS GCM. Solar cycle variability is used here as a proxy for longer-term solar variations, which are not well constrained.

Empirical orthogonal functions (EOFs) were calculated from the NH cold season (November–April) sea-level pressure (SLP) time series of the control run. The trend from the transient greenhouse gas runs was then projected upon those EOFs. Other forcings were not run as transients, so differences between two states are used. For the solar cycle variability simulations, the difference between the runs with solar maximum and solar minimum conditions was used. For the volcanic forcing, the difference between the years when stratospheric aerosol loading was large (primarily the few years immediately following eruptions) and years with background levels was used. Results for the ozone hole simulations are differences between the simulation with polar heterogeneous chemistry included and a control run without the chemistry parameterisation. For the transient simulations we define the time series corresponding to the leading EOF as

the “index” of the model’s AO. The spatial pattern has been normalised so that the value of the AO index is equal to the opposite of the SLP anomaly averaged poleward of 60°N, and is given in hPa. For the transient runs the AO index trend is given over a 30-year period to match the observational record. The three-decade value we present here was consistent for the entire five decades during which the AO increased steadily between the initial spin-up and eventual saturation. For the tropospheric runs the averaging period has a minimal effect on the results, as the trends were so small that no saturation was seen. For the solar, volcanic, and ozone hole simulations, whose results are differences, we obtain a single value for the change in the AO index based upon the component of the total SLP difference associated with the AO.

While the leading EOF is well separated from subsequent patterns, the higher EOFs are not adequately distinct from one another. In the stratospheric model forced with increasing greenhouse gases, EOF 1 accounts for 13.3% of the variance, but subsequent EOFs account for 5.9, 4.8, 4.1, and 3.5 percent, respectively, which are not statistically different within the calculated variance range of about 2% (calculations based on the entire 110-year run). An EOF analysis of the observations yields similar results, with 21.3, 11.7, 11.1, 8.2, and 7.0 percent variance explained by the first five EOFs (calculations based on 1947–1997 data). Again, only the leading pattern is well separated from the others. Since the higher modes are mathematical constructs which are forced to be orthogonal to the leading mode, not well separated, and lacking an obvious physical interpretation, we restrict our analysis to the leading EOF (the AO).

As discussed by Shindell *et al.* [1999a], the version of the GISS model containing a realistic representation of stratospheric processes produces an increasing AO trend in response to greenhouse gas increases of ~0.8 hPa/decade during the several decades between spin-up and saturation during which the AO index is increasing steadily. This value is comparable to the roughly 1 hPa/decade suggested by observations in recent decades. In contrast, model versions lacking a realistic stratosphere, “tropospheric models” with only one or two layers in the stratosphere, produce a leading EOF that also looks very much like the AO, but its index shows

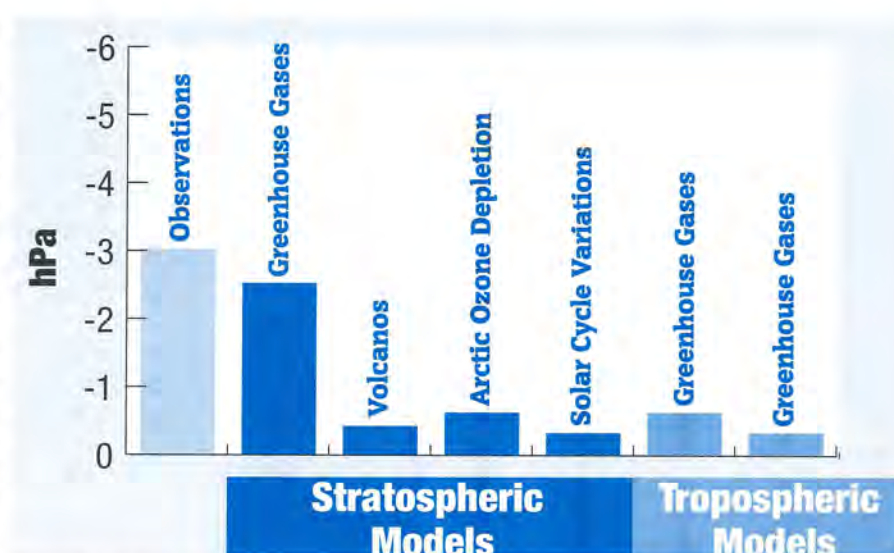


Figure 2. The magnitude of the trend or change in the AO index induced by each forcing. The magnitude of the AO is defined as the SLP anomaly averaged northward of 60 degrees. Values are calculated using a linear least squares regression through the observations or model output. Observations are updated from Trenberth and Paolino [1980].

only a weak increasing trend (+0.1–0.3 hPa/decade) which is not statistically significant (Figure 2). It is interesting that when stratospheric model forced with increasing water vapour is allowed to increase steadily, as a result of tropopause warming and methane oxidation [Shindell, 2001], this distinction between stratospheric and tropospheric models is slightly reduced. When water vapour and greenhouse gases both increase, the AO trend is reduced to ~0.6 hPa/decade, outside the range of the trend in the other three increasing greenhouse gas simulations (0.83, 0.78, and 0.79). This is due to the slight decrease in the meridional temperature gradient that results from mid-latitude cooling that ensues from water-vapour-induced ozone destruction in the lower stratosphere [Shindell, 2001]. Those simulations do not include the potential effect of increased water vapour on heterogeneous ozone destruction in the polar region, however, which would oppose the mid-latitude ozone-induced cooling and could potentially cause a significant increase in the meridional temperature gradient in the spring. Water vapour on its own, as a greenhouse gas, increases the bias toward the high phase of the AO.

We find that the inclusion of an ozone hole in the model, another candidate for driving the observed increases [e.g., Graf *et al.*, 1998; Volodin and Galin, 1999], does cause an increase in the model’s AO index, but of only ~0.6 hPa (1990s ozone depletion relative to the 1970s and earlier, when there was no depletion, or approximately 0.3 hPa/decade over the past

two decades). Though this has likely contributed to the measured trend, it seems insufficient to account for a great deal of the observed increase of several hPa over the past three decades. This is perhaps not surprising, since the trend in the AO has been observed throughout the November-to-April cold season, but severe ozone depletion in the Arctic does not take place in general until the spring, when sunlight falls on cold vortex air. This conclusion is in agreement with the mild increase in the wintertime meridional temperature gradient at northern high latitudes calculated using TOMS ozone trends [e.g., Ramaswamy *et al.*, 1996]. While the TOMS observations were extrapolated to high latitudes during the polar night, the use of *in-situ* ozonesonde data [Randel and Wu, 1999a] does not change this conclusion [Rosier and Shine, 2000]. This contrasts with the situation in the Antarctic, where ozone depletion seems to account for nearly the entirety of the observed temperature trends [Randel and Wu, 1999b].

Volcanic forcing can be strong in the years immediately following an eruption, leading to an AO difference of 0.4 hPa between volcanic and non volcanic years, but since it is an intermittent forcing that decays rapidly, it also seems unlikely to have contributed greatly to the long-term observed trend. Solar cycle variability causes an increase in the AO index of 0.3 hPa between solar maximum and minimum. Since the longer-term trend in irradiance over the past 30 years has been of comparable size to solar cycle variability, it also seems unlikely that

solar variability has been responsible for much of the observed trend. Furthermore, estimated solar irradiance increased as much in the first half of the twentieth century as in the second [Lean *et al.*, 1997], but the AO index showed no increase during the former. However, the long-term response would include a response of the ocean to solar forcing, while the shorter-term solar cycle does not, so that the response may in fact be somewhat different over longer time scales.

The stratosphere model is distinct from the tropospheric models not only for exhibiting a trend of magnitude comparable to the observed value, but because the spatial distribution of its SLP trend closely resembles the AO spatial pattern, as observed [Thompson and Wallace, 1998]. The resemblance of the SLP trend to the AO spatial pattern can be quantified by decomposing the trend into the EOFs of SLP and measuring the contribution of each EOF to the trend variance (Figure 3, p. II), following Fyfe *et al.* [1999]. The trend in observations of cold season SLP between 1967 and 1997 [Trenberth and Paolino, 1980, updated to 1997] is dominated by a change in the AO (the leading EOF), which contributes 56% of the trend. The GISS stratospheric model with increasing greenhouse gases puts 64% of the trend in the AO (the average value of the ensemble of four simulations). The stratospheric model with either an ozone hole or volcanic forcing also projects primarily onto the first EOF; however, solar cycle variability does not. The stratospheric model with increasing water vapour projects less strongly onto the AO than do the other increasing greenhouse gas runs, but its trend is still dominated by the AO (and actually looks most like the observed trend projections).

In contrast, GISS model versions lacking a detailed stratosphere ("tropospheric models") forced with increasing greenhouse gases produce trends that project only 5-20% onto the AO (Figure 3). The GISS tropospheric model including sulphate aerosols puts just 6% of its trend into the leading EOF, with most of its trend spread out over a large number of patterns. Transient greenhouse gas experiments by other modelling groups seem to be consistent with this distinction (Figure 3). The Canadian Climate Center CCCma model [Fyfe *et al.*, 1999], which shows a positive AO trend in response to increasing greenhouse gases despite the absence of

detailed stratospheric dynamics, similarly fails to reproduce the observed dominance of the AO pattern, putting less than 30% of its total trend into its leading EOF. A Geophysical Fluid Dynamics Laboratory (GFDL) climate model version lacking a detailed stratosphere also finds that the majority of its SLP trend in response to increasing greenhouse gases is not in the AO (P. Kushner, personal communication, 1999). Several other groups have presented the simulated AO response to increasing greenhouse gases. The ECHAM3 model from the Max Planck Institute for Meteorology, without a full representation of the stratosphere, shows a systematic increase in the AO index in response to increasing greenhouse gases [Graf *et al.*, 1995, 1998; Perlwitz *et al.*, 2000], but its amplitude is stated to be weaker than observed. Paeth *et al.* [1999] present results of simulations with both the ECHAM3 and ECHAM4 models (showing NAO trends, but these are likely very similar to AO trends). Zorita and Gonzalez-Rouco [2000] present the AO trends from the same simulations, adding in results from the Hadley Center model 2 as well. Along with Fyfe *et al.* [1999], all of these groups use normalised units to present their trends, making it impossible to compare the magnitude of the increase in those simulations to the observed value. We suggest that documentation of a model's AO trend should include, at minimum (1) the trend's magnitude in hPa and its statistical significance, with a description of the area weighting used, and (2) the decomposition of the trend into its component EOFs. This information would facilitate a comparison across models to document more clearly the role of stratospheric dynamics. As yet, only models that realistically simulate the stratosphere have quantitatively demonstrated an AO trend of roughly the right magnitude and with the same predominance within the SLP trend as the observations.

Physical Mechanism

The upper troposphere in the tropics warms significantly in response to increases in greenhouse gases. This leads to a sharp contrast with latitude in the temperature response in the tropopause region. While the upper troposphere is closely connected to the surface by moist convective processes at low latitudes, and therefore warms, the same altitudes lie within the stratosphere at higher latitudes, as the

height of the tropopause decreases abruptly poleward of the mid-latitude jet. Increasing greenhouse gases cool the stratosphere, and therefore enhance the meridional temperature gradient across the jet core. This gradient increase, which is largely zonally symmetric, is associated with strengthened westerlies within the jet. The strengthened mid-latitude jet alters the propagation of planetary waves, which can potentially feed back on the jet through eddy forcing.

Volcanic forcing is similar, as aerosol heating in the sunlit portion of the winter atmosphere warms that region relative to the part of the atmosphere in polar darkness, enhancing the latitudinal temperature gradient. Arctic ozone depletion does the same thing too, as less ozone at high latitudes leads to a cooling there due to the reduction in short-wave absorption by ozone. This happens much later than the others, however, as sunlight returns to high latitudes and ozone depletion takes place only in the spring around March and April, so that this has a relatively small impact over the entire November-to-April cold-season, and no real effect during December-to-February.

Amplification by planetary waves is largest in NH winter, when planetary wave activity is greatest. While we used the November-April cold season for SLP, to facilitate comparison with other analyses of the AO, we return now to the conventional December-February winter period for easier comparison with model output and with observations of more standard meteorological parameters. Planetary wave refraction is governed by wind shear, among other factors, so that enhanced wave refraction occurs as the waves approach the area of increased wind speed. In this case, since planetary waves are propagating up from the surface, they are refracted by the increased vertical shear below the enhanced zonal wind. Equatorward refraction of planetary waves at the lower edge of the wind anomaly (Figure 4, p. III) leads to wave divergence and hence an acceleration of the zonal wind in that region. Over time, the wind anomaly itself thus propagates downward [Haynes *et al.*, 1991] from the lower stratosphere to the surface.

The precise location of the enhanced westerlies shown in Figure 4 therefore depends upon the interaction between planetary waves, wind shears, eddy fluxes and angular momentum fluxes, and is not a simple function of the location of the strongest temperature

contrast. This is likely the cause of the difference in location between the strongest enhancement of the temperature gradient, which is at the latitude of the jet stream, and the location of the strongest zonal wind enhancement, which is at the latitude of the polar night jet. Further work will be required to fully elucidate the link between the meridional structures of the temperature and wind responses to external forcing.

The overall equatorward refraction of planetary waves and reduced upward propagation at high latitudes shown in Figure 4 affect the location of wave dissipation. The net result is a reduced ability of individual waves to abruptly enhance the residual circulation and create sudden warmings. In fact, there is a strong anticorrelation between the frequency of sudden stratospheric warmings and the AO index strength [Hartmann *et al.*, 2000].

The behaviour of the troposphere is in accord with theory stating that angular momentum is in general transported in the opposite meridional direction to planetary wave energy [Andrews *et al.*, 1987]. In this case, angular momentum is then preferentially transported poleward, enhancing westerlies, as waves are refracted towards the equator in the upper troposphere. Geostrophic and hydrostatic balance in the atmosphere is maintained by generation of a vertical circulation cell consistent with the northerly angular momentum transfer. Increased westerlies are transformed into northerly flow by surface friction, leading to a cell with rising air in the polar region, and descending air at middle latitudes from about 40° to 55°N throughout the troposphere and the lower stratosphere. The effects are clearly visible in the SLP field as a decrease in the Arctic concurrent with increased mid-latitude pressure. Adiabatic expansion of rising air at high latitudes must be balanced by radiative heating, which occurs as a result of further cooling of the air below its radiative equilibrium temperature. Conversely, sinking air at mid-latitudes warms. The increased latitudinal temperature gradient that results is consistent with the increased westerly zonal wind around 55°N seen in the model and in observations [Baldwin and Dunkerton, 1999; Thompson *et al.*, 2000]. The surface wind anomaly is therefore created by the effect of the increased lower stratospheric zonal wind on planetary waves via linked changes in planetary wave and angular momentum fluxes. It is this increase in

surface wind that leads to greater advection of warm oceanic air over the downstream continents. The role of the stratosphere in influencing the behaviour of surface meteorology in the model is in agreement with the observed co-variability between the lower stratosphere and the surface [e.g., Perlwitz and Graf, 1995; Graf *et al.*, 1995; Thompson and Wallace, 1998]. It is also consistent with the large role played by planetary waves in creating the AO pattern itself during midwinter in the NH, as evidenced by the fact that the AO structure amplifies with height up into the stratosphere only during this season, when the zonal flow is conducive to strong wave-mean flow interaction [Baldwin and Dunkerton, 1999; Thompson and Wallace, 2000]. This would also explain why there has only been an upward trend in the AO during NH winter.

Conclusions

The model simulations have shown that a surface climate response is induced by increasing greenhouse gases, ozone depletion, volcanic eruptions, and solar variability. Except for solar variability, the others all affect surface climate at NH middle and high latitudes during winter primarily by favouring the positive phase of the AO, the dominant natural variability pattern, while concurrently strengthening the polar vortex aloft. Nonetheless, only increasing greenhouse gases seem capable of causing an AO trend in our model as large as the observed trend.

The AO sensitivity of models developed by several groups, including GISS, which lack realistic representations of the stratosphere, is much weaker than the GISS stratospheric model. It seems that models with an overly simplistic stratosphere fail to adequately capture the positive feedback of planetary wave flux changes on circulation anomalies seen in the stratospheric model. Furthermore, the simulated trends in those models are not primarily composed of the AO pattern, as are the observations, while the stratospheric model does reproduce this behaviour.

The surface climate response of the GISS stratospheric model to the injection of volcanic aerosols into the stratosphere corresponds well with observations. This provides important evidence that the modelled linkage between forcings and the AO does indeed occur in the atmosphere. While longer-term AO changes will need to continue for many more years to be fully accepted as a distinct trend, the AO enhancement after erup-

tions is already quite clear in the observational record. Since the AO responses to greenhouse gas increases, ozone depletion, and volcanic aerosols all take place via the same mechanism, the clearly observed response to large volcanic eruptions suggests that the same linkage is likely at work in the case of the slower, longer-term greenhouse gas and ozone depletion forcings as well.

An obvious question is how would the response vary from model to model, using models with a realistic stratosphere? There are likely to be considerable intermodel variations, as one of the dominant factors controlling the sensitivity of the AO to forcings is the surface temperature response to forcings, which differs considerably between models [IPCC, 1995]. The magnitude of the tropical surface warming seems to play a major role in changes in the latitudinal temperature gradient via its control over warming in the tropical upper troposphere. The degree of this warming is probably the largest uncertainty in the overall magnitude of the change in the latitudinal temperature gradient, as the stratospheric cooling is purely radiative, and similar in most models [e.g., WMO, 1999]. Thus the tropical surface warming directly governs a significant part of the AO in response to climate forcings. This implies that the surface response is crucial to the stratospheric response, an interesting variation on the stratosphere-climate linkages that are the focus of SPARC. These interactions clearly go in both directions.

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Seasonal Variation of Extratropical Cross-tropopause Transport Inferred from Chemical Tracer Measurements

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Introduction

The stratosphere can be subdivided into the "overworld" and the "lowermost stratosphere" (Holton *et al.*, 1995). The lowermost stratosphere is defined as the volume of air between the local tropopause and a potential temperature (Θ) level (isentropes) of approximately 380 K, corresponding to the mean height of the tropical tropopause. The overworld represents

the volume of stratospheric air above this isentropic surface. In the large scale Brewer-Dobson circulation, tropospheric air enters the stratosphere predominantly in the tropics, while stratospheric air returns to the troposphere at high latitudes. Although the net flux in the extratropics is from the stratosphere into the troposphere, the exchange can take place in both directions. In general, the chemical trace gas composition of the lowermost

stratosphere is therefore a mixture of aged stratospheric air descending from the overworld through the 380 K isentropes and recent injections of tropospheric air (here the "age" of stratospheric air refers to the time passed after last contact with tropospheric air). The latter injections can be related to either diabatic processes, e.g. via mid- and high-latitude convection, or adiabatic along isentropes across the tropopause, e.g. in the

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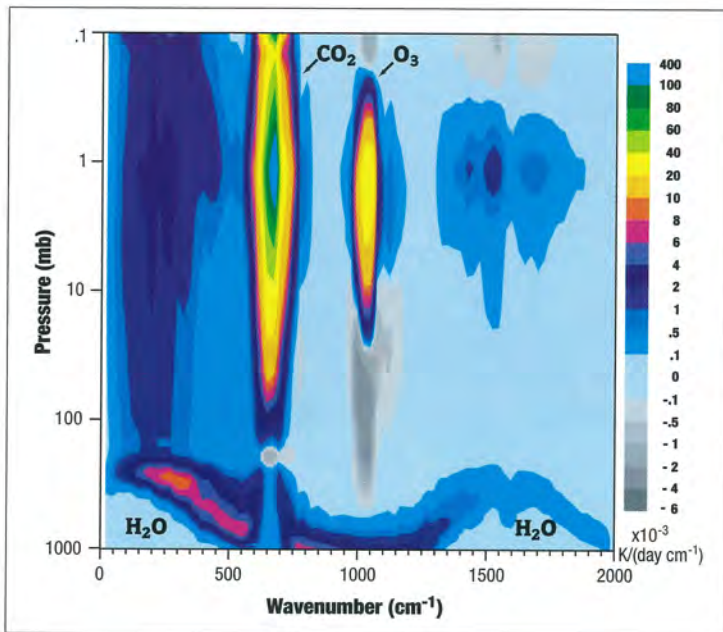


Figure 1

Thermal infrared cooling rates for H₂O, CO₂ (355 ppm), and O₃ as a function of wavenumber and pressure for mid-latitude summer conditions. Colour scale × 10⁻³ is in units of K d⁻¹ (cm⁻¹)⁻¹. The figure demonstrates the special nature of the UT/LS region in the atmosphere radiative balance. The band centred at 667 cm⁻¹ is dominated by CO₂, and exhibits strong stratospheric cooling (explaining why a CO₂ increase leads to stratospheric cooling) but a local warming at the tropopause. The band around 1043 cm⁻¹ is dominated by O₃ and exhibits warming throughout the lower stratosphere. Water vapour makes a strong contribution to upper tropospheric cooling near 300 cm⁻¹, although its effects are moderated by CO₂. In the UT/LS region the thermal infrared warming and cooling effects tend to largely cancel, implying a strong radiative sensitivity to greenhouse gas changes. Courtesy of M. Iacono, AER.

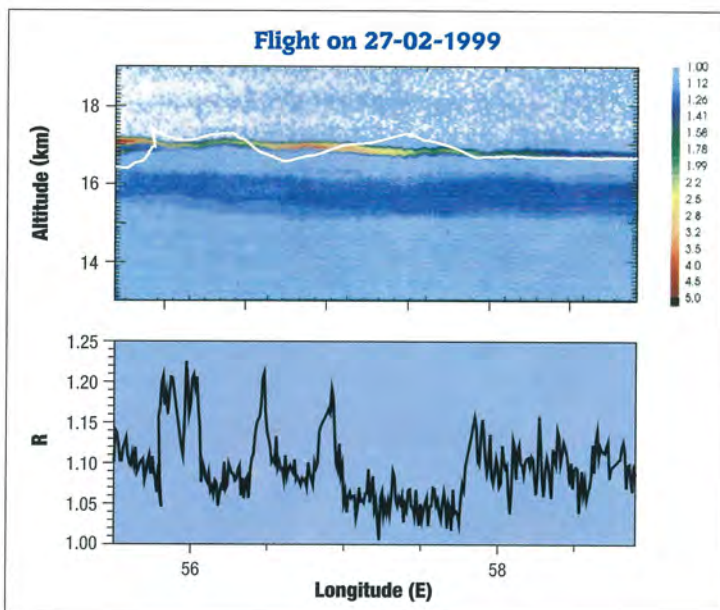


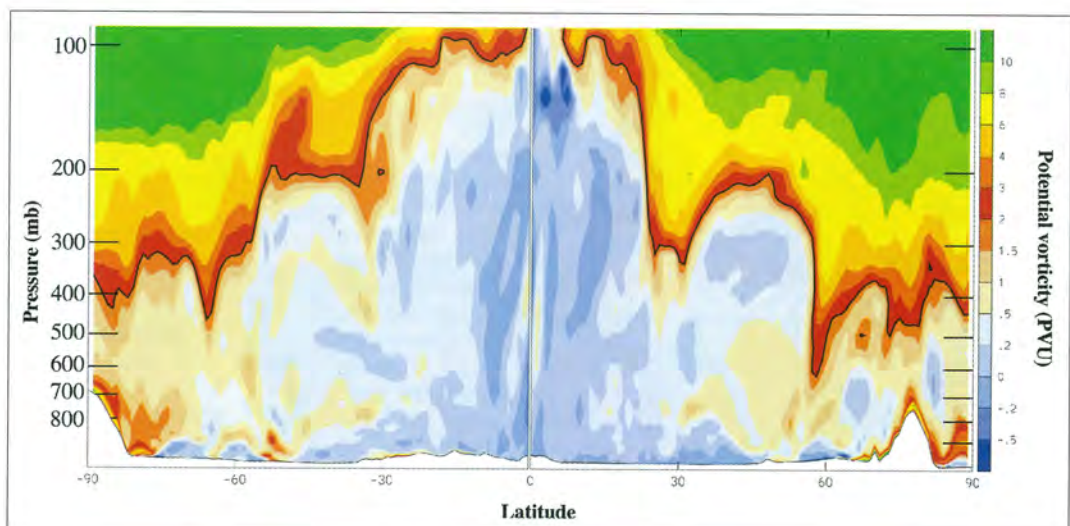
Figure 4 - (top panel) Ultrathin tropical cirrus cloud at 10°S and 17km altitude, located about 300m below the CPT, detected through backscatter ratio at 1064nm measured by OLEX on the German research aircraft Falcon during the APE-THESIO experiment in February 1999.

(bottom panel) *In-situ* measurements of backscatter ratio at 532nm measured by the aerosol sonde MAS on board the Russian high-altitude research aircraft Geophysica, whose flight track is indicated by the white line in the top panel. The cloud is only about 200-300m thick and has an optical density of about 10⁻⁴, 300 times lower than that required for visibility from the ground.

These results suggest that thin, subvisible, cirrus clouds may be far more prevalent than previously imagined. Such clouds might be important for chemistry via heterogeneous reactions and possibly also for scavenging (e.g. of HNO₃). The cloud at 17km carries very low concentrations of water and therefore does not by itself indicate that significant dehydration is occurring. However, modelling indicates that upon cooling the particles in the cloud can grow to much enhanced sizes. The dehydration effect would then be much more significant. Courtesy of DLR Oberpfaffenhofen, CNR Rome and the APE/THESIO community.

Figure 5

Instantaneous height-latitude cross section of potential vorticity along a single longitude (55W), with the tropopause marked (in black) as the 2PVU contour. The complex geometry of the tropopause associated with day-to-day variations in the circulation is evident, including upward doming associated with a subtropical anticyclone and downward intrusion and folding at higher latitudes. Courtesy of H. Wernli, ETH Zurich.



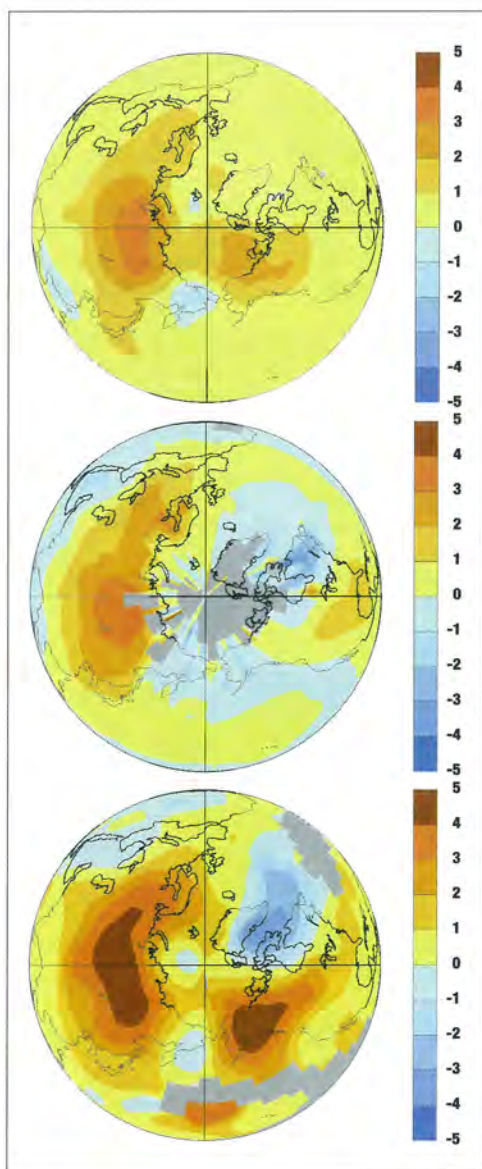
Report on the Joint US-Japanese Workshop on « Coupling of the Troposphere and Stratosphere by Dynamical, Radiative and Chemical Processes », Kyoto, Japan, 13-17 March, 2001

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Northern Hemisphere Winter Climate Response to Greenhouse Gas, Ozone, Solar, and Volcanic Forcing



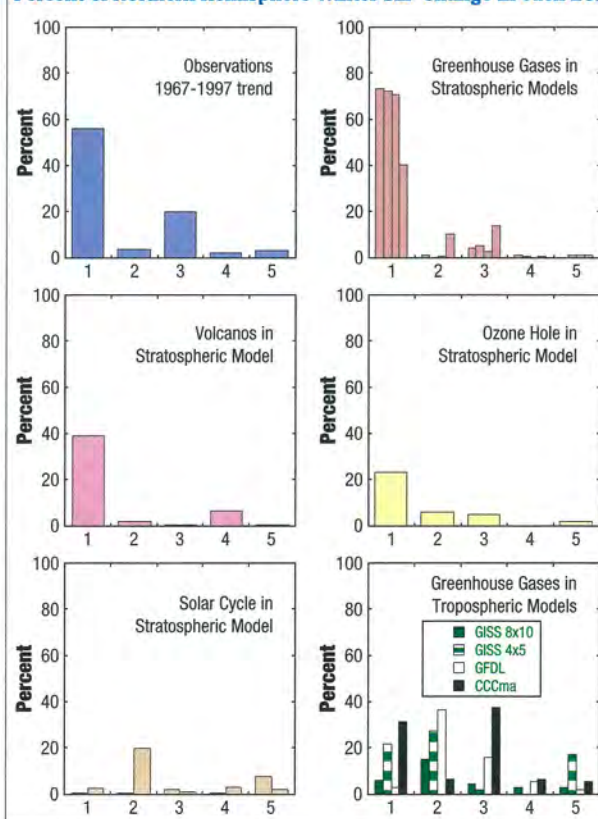
◀ Figure 1

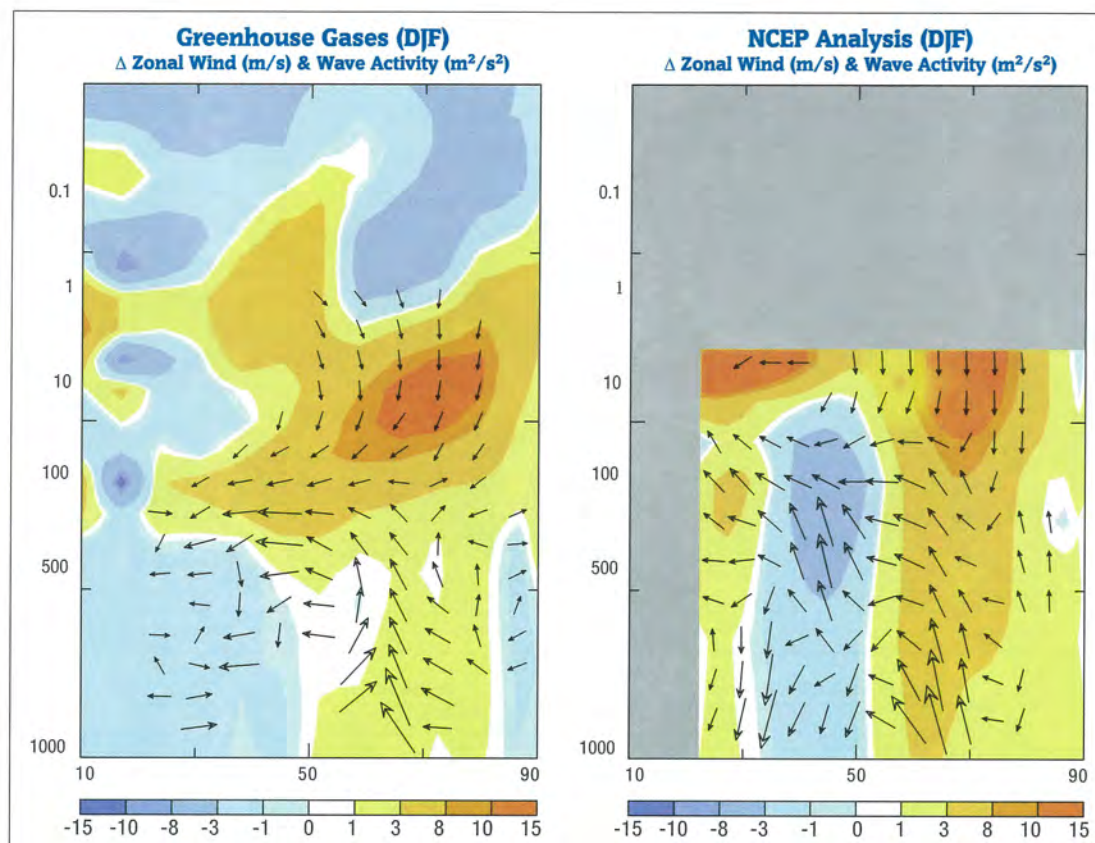
Surface temperature trends ($^{\circ}\text{C}$). (bottom) Observed surface temperature change from 1965 to 1995, December to February average [Hansen *et al.*, 1999]. (middle) The Arctic Oscillation (AO) contribution to the total trend, obtained by regressing the temperature trend onto the AO spatial pattern (as in the work by Thompson *et al.* [2000]). (top) The AO component of the total trend in the GISS stratospheric model, similarly obtained by regressing the model's temperature trend onto its AO spatial pattern. Grey areas are those where no data is present. Note that the difference in grey areas between the bottom and middle maps results from the use of slightly different temperature data sets, which do not match exactly in areas with few observations (Arctic Ocean and low-latitude Atlantic and Pacific).

► Figure 3

Percentage of the total Northern Hemisphere November-April sea level pressure change in each empirical orthogonal function (EOF). Trends from each indicated model simulation were projected upon EOFs taken from the control run. Observations are updated from Trenberth and Paolino [1980], as in Figure 2.

Percent of Northern Hemisphere Winter SLP Change in each EOF



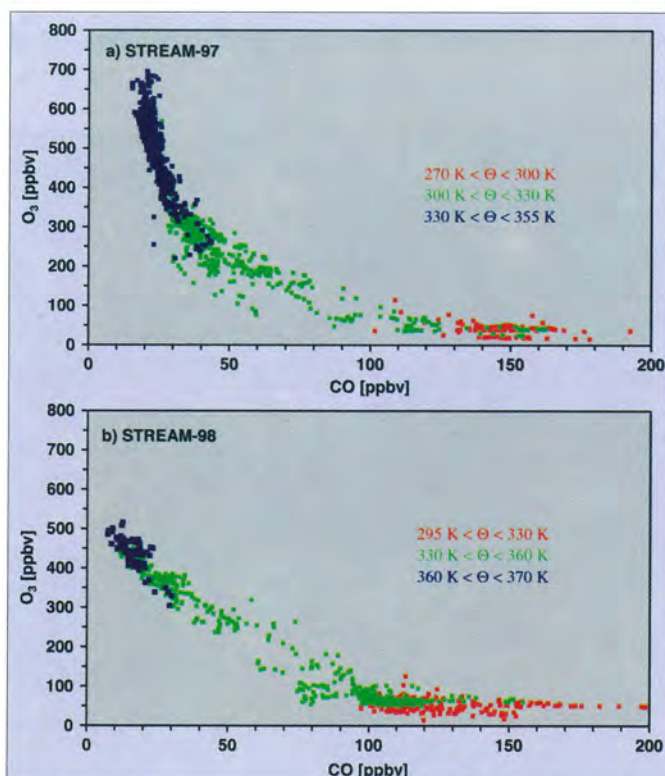


◀ **Figure 4**
Trends in zonal mean zonal wind (m/s) and planetary wave activity (m^2/s^2) (left) in the greenhouse gas simulations and (right) in analysed observations. Model trends are shown for the five decades during which the AO increased steadily between the initial spin-up and eventual saturation (1980-2030). Observed trends are differences between the high-AO years 1981, 1983, and 1984 and the low-AO years 1979, 1982, and 1992 using NCEP analyses of observational data. Values are the sum of the total trend components on the first two EOFs of these trends, as given by Ohhashi and

Yamazaki [1999, Figures 7 and 8] (Hartmann *et al.* [2000] show similar results). Since the first two EOFs contain nearly all the variability, these should be comparable to the model's total trend values. Vertical fluxes have been scaled by the inverse of pressure for convenience in all plots. The maximum vector length corresponds to $7 \text{ m}^2/\text{s}^2$. The total AO index change in the model over this five-decade period was about 4 hPa, somewhat larger than the 1979 to 1992 variability of about 2.5 hPa.

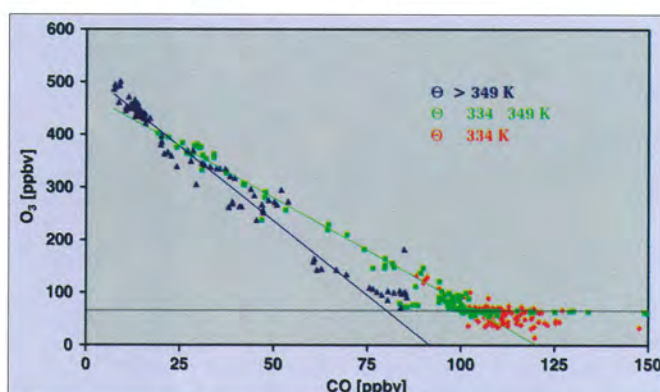
III

Seasonal Variation of Extratropical Cross-tropopause Transport Inferred from Chemical Tracer Measurements



◀ **Figure 2**
Scatter plots of O_3 versus CO obtained during the STREAM measurement campaigns in early March 1997 (a) and July 1998 (b). The colour coding refers to different potential temperature intervals.

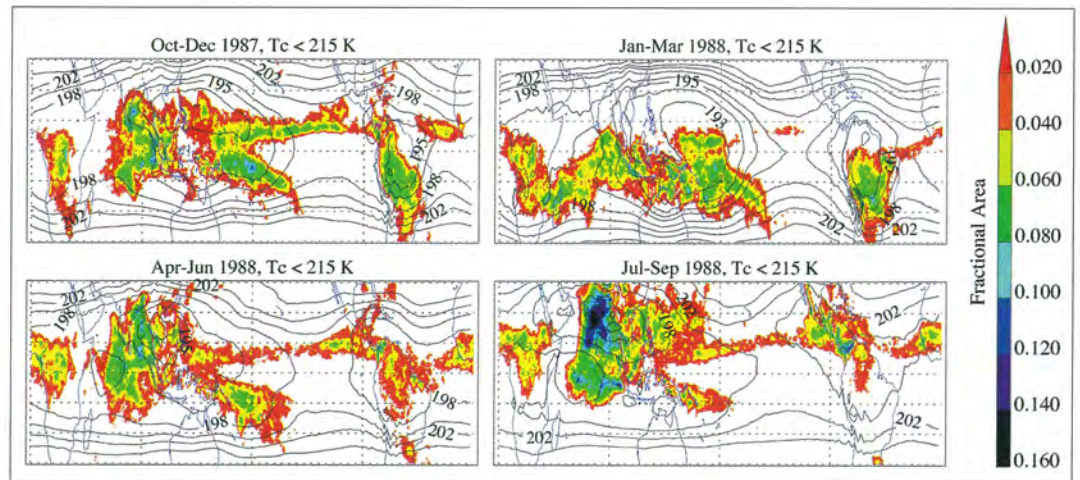
▼ **Figure 3**
Scatter plot of CO and O_3 for the flights on July 1 and 12, 1998. Colours represent potential temperature. Note the different slopes of the stratospheric mixing lines for different Θ intervals, suggestive of different tropospheric reservoirs.



Convection in the Tropical Tropopause Region and Stratosphere-Troposphere Exchange

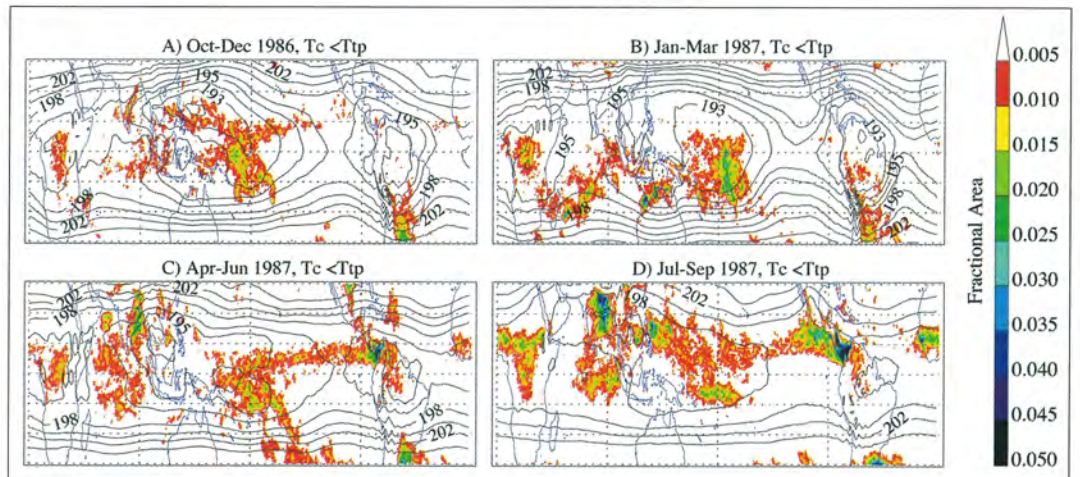
► **Figure 1**

Fractional area with brightness temperature colder than 215° K in GCI data (η_{215}) for A) October-December 1986, B) January-March 1987, C) April-June 1987 and D) July-September 1987. Adapted from Gettelman *et al.* (2001).

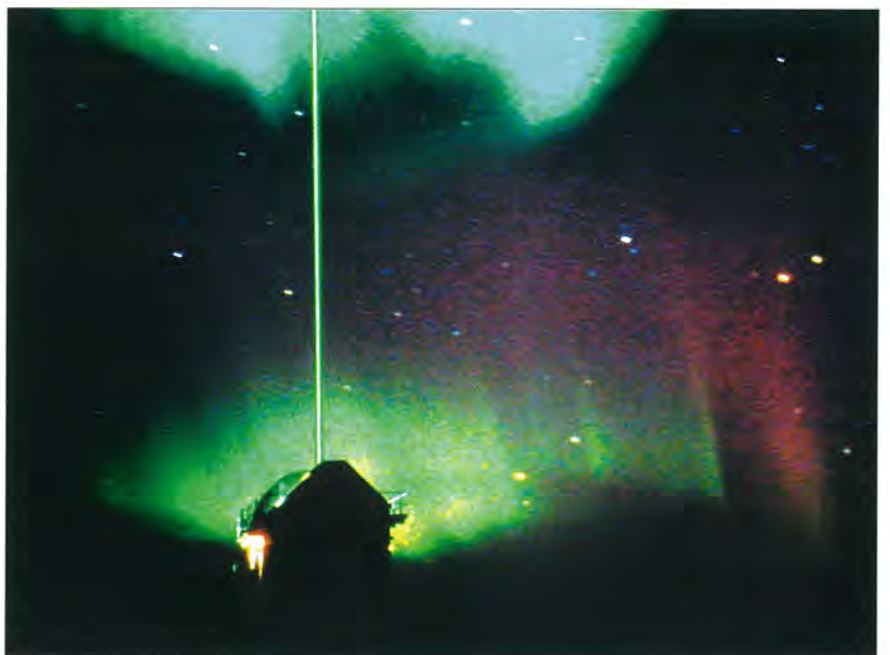


► **Figure 3**

Fractional area with brightness temperature colder than the tropopause in GCI data (η_{TP}) for A) October-December 1986, B) January-March 1987, C) April-June 1987 and D) July-September 1987. Contours are seasonal mean tropopause temperature. Adapted from Gettelman *et al.* (2001).



Australian Activities Relevant to Sparc



► **Figure 3**

Lidar and aurora (Courtesy of David Correll, Australian Antarctic Division, 2001).

vicinity of tropopause breaks associated with the subtropical and the polar front jet streams (Danielsen, 1968; Hoerling *et al.*, 1993). This results in the formation of a mixing layer above the local tropopause (Danielsen, 1968; Kritz *et al.*, 1991; Hoerling *et al.*, 1993; Ray *et al.*, 1999; Fischer *et al.*, 2000; Hoor *et al.*, 2001). Model studies indicate that bi-directional cross-tropopause transport at potential temperature levels of 340–360 K in the vicinity of the subtropical jet is strongly inhibited in winter, while substantial exchange occurs at these altitudes in summer (Chen, 1995; Haynes and Shuckburgh, 2000). These studies also indicate strong exchange below the 330 K isentropic surface during all seasons, associated with synoptic disturbances at the polar front.

Chemical composition of the mixing layer

While chemical signatures from transport of stratospheric air into the upper troposphere are difficult to detect due to the rapid vertical mixing in the troposphere, the transport of tropospheric air into the lowermost stratosphere can be more easily identified from *in-situ* measurements of a large number of trace gases that exhibit a strong concentration gradient across the tropopause (e.g. H_2O , N_2O , CO, O_3 and NOy, the sum of all oxidised nitrogen species except N_2O). In particular, the slope of chemical tracer-tracer-relations in the lowermost stratosphere can be strongly affected by extratropical cross-tropopause transport (Hipskind *et al.*, 1987; Fischer *et al.*, 2000). This is illustrated in Figure 1 for the O_3 -CO relation. Hipskind *et al.* (1987) first used correlation of O_3 and CO in the

lower stratosphere and the marine troposphere to quantitatively describe the mixing between these two reservoirs in a tropopause fold. The mixing ratios of both species exhibit strong gradients at the tropopause, with high CO (low O_3) values in the troposphere and low CO (high O_3) values in the stratosphere. In general, the concentration of O_3 strongly increases in the stratosphere, while its mixing ratio in the troposphere is relatively constant at levels smaller than approximately 100 ppbv. Sources of CO are primarily associated with combustion processes and hydrocarbon oxidation in the troposphere, leading to rather high concentrations of the order of 100 ppbv in the free troposphere of the Northern Hemisphere. Apart from extratropical stratosphere-troposphere exchange, the only source of CO in the stratosphere is the OH initiated oxidation of CH_4 . The produced CO is further oxidised to CO_2 by reaction with OH, resulting in a rather constant mixing ratio of 10 to 20 ppbv in the overworld. Therefore, in the absence of cross-tropopause mixing, it is expected that simultaneous CO and O_3 measurements should follow the L-shape indicated in Figure 1 (black lines). Mixing of tropospheric air into the lowermost stratosphere will result in deviations from this L-shape in the form of mixing lines connecting the chemical reservoirs of stratospheric and tropospheric air involved in the mixing process (dashed and dotted blue lines in Figure 1). The slope of the mixing lines is a function of the concentrations of CO and O_3 in the tropospheric and stratospheric reservoirs. In general, the CO concentration in the Northern Hemisphere Upper Troposphere is a function of latitude, with higher concentrations at high latitudes. This is due to the large sources at high latitudes, and the strong sink

in the tropics associated with high OH levels. Therefore we expect that mixing at the subtropical jet (dotted blue line) will result in a different mixing line compared to the polar jet (dashed blue line). In addition, the slope will significantly depend on the O_3 concentration of the stratospheric reservoir. Fresh mixing events will produce a linear mixing line, while chemical ageing of mixed airmasses in the lowermost stratosphere will result in a non-linear mixing line due to the shorter photochemical lifetime of CO (~ 3 month) compared to O_3 (~ 6 month) in this part of the atmosphere.

STREAM measurements

Figure 2 shows scatter plots of O_3 versus CO obtained during two STREAM (Stratosphere Troposphere Experiments by Aircraft Measurements) campaigns by Aircraft Measurements) campaigns from Kiruna, Sweden (68°N , 20°E) in early March 1997 (Figure 2a (p. III)) and Timmins, Ontario Canada (48.2°N , 79.3°W) in July 1998 (Figure 2b (p. III)). The winter data were obtained during four measurement flights covering an area extending from 2°E to 20°E and 68°N to 78°N up to a maximum altitude of 13 km. The summer campaign (8 flights) probed a region between 44° and 56°N and 71° to 90°W at similar altitudes. Both campaigns were performed in close proximity to the polar jet. The data in Figure 2 have been colour-coded with respect to potential temperature. In the troposphere at Θ -levels less than approximately 300 K in winter and less than approximately 330 K in summer, CO mixing ratios in excess of 100 ppbv are found while the O_3 mixing ratio is nearly constant at approximately 60 ppbv (red data points). As expected, the CO- O_3 regression shows no correlation at all in the troposphere. For the Arctic STREAM campaign in March 1997 a strong negative relation between CO and O_3 is observed for CO mixing ratios between 15 ppbv and approximately 30 ppbv (blue data points above the $\Theta = 330$ K level in Figure 2a), mixing ratios typical for aged stratospheric air. Cross-tropopause mixing can be identified by deviations from the L-shape for CO mixing ratios between about 30 and 100 ppbv (green data points in Figure 2a). During the summer campaign the purely stratospheric branch of the O_3 -CO correlation is almost missing (Figure 2b). Instead, the mixing line seems to extend to the highest O_3 mixing ratios and to rather small values of CO, of the order of 10 to 15 ppbv. Hoor *et al.* (2001) have shown

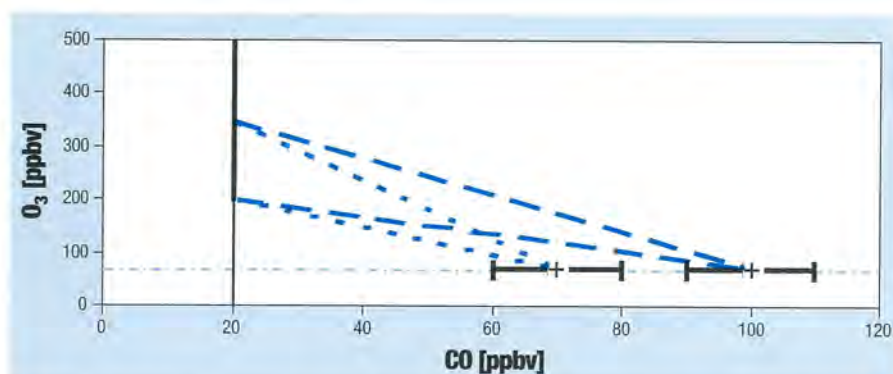


Figure 1: Schematic representation of the influence of stratosphere-troposphere exchange on the O_3 -CO relation. The vertical line represents the stratospheric (overworld) reservoir, while the horizontal lines show the ranges for two different tropospheric reservoirs. Blue dashed and dotted lines indicate mixing between these reservoirs.

in more detail that in the winter season (Figure 2a) the mixing zone is restricted to potential temperature levels below approximately 330 K and does not extend beyond 20 K above the local tropopause. On the other hand, the summer measurements show that the mixing line reaches up to at least the 360 K potential temperature level, approximately 40 K above the local tropopause.

Air mass origin in the mixing layer

From Figure 2 we can estimate the CO and O₃ concentrations for the stratospheric (transition from green to blue data points) and tropospheric (green to red data points) reservoirs involved in the mixing process. In the winter mixing occurs between a tropospheric reservoir of approximately ~120 ppbv CO and ~50 ppbv O₃ and a stratospheric reservoir of ~30 ppbv CO and ~350 ppbv O₃ (Figure 2a). During summer the tropospheric reservoir appears to contain between 100 and 120 ppbv CO at O₃ levels of 60 ppbv. The stratospheric reservoir, representing descending air masses from the overworld, is characterised by CO less than 20 ppbv and ~400 ppbv of O₃. In particular the lower CO concentration indicates that the stratospheric reservoir in the summer is composed of older air masses compared to the winter season. This is confirmed by CO₂ measurements indicating an air mass age of approximately 2 years for the summer measurements (Hoor *et al.*, 2001). In general the mixing line for the summer season is very compact and linear, contrary to the mixing line observed for the winter season, which is slightly curved and exhibits more scatter. This indicates photochemical ageing of the air in the mixing layer during winter, while during summer the linear shape points to a rather continuous mixing process. Despite its compactness the mixing line for the summer season can be separated into two different lines. This is mainly due to data points obtained during the STREAM measurement flights on July 1 and 12, 1998, which have been performed on the cyclonic and anticyclonic side of the polar jet stream (Figure 3 p III). The green data points, for a potential temperature interval between 334 and 349 K, seem to arise from mixing with a tropospheric reservoir characterised by ~110 ppbv CO and ~70 ppbv O₃. These values agree very well with upper tropospheric measurements during STREAM 98 for air masses

with a polar background, as indicated by 5-day backward trajectories. The blue data points in Figure 3 ($\Theta > 349$ K) seem to result from mixing with a tropospheric reservoir with significantly less CO (~75 ppbv), indicating a more southerly subtropical/tropical origin of these air masses.

This finding is consistent with a seasonal variation of cross-tropopause exchange at the subtropical jet as predicted by model studies (Chen, 1995; Haynes and Shuckburgh, 2000). Although the two STREAM campaigns have been performed in different locations, they both took place in close proximity to the polar jet stream. It seems that transport of tropospheric air into the lowermost stratosphere at potential temperature levels less than 330 K, associated with exchange across the mid-latitude polar jet, is found in both seasons. In the summer season, additional exchange occurs at the subtropical jet, mixing low latitude air masses from the troposphere into the lowermost stratosphere up to potential temperature levels of approximately 360 K.

Excess NO_y in the lowermost stratosphere

The identification of the mixing layer is not restricted to investigations of the O₃-CO relationship. Other studies used *in-situ* measurements of radioactive tracers (Kritz *et al.*, 1991), H₂O (Dessler *et al.*, 1995, Hintsa *et al.*, 1998), CFCs, SF₆ and CO₂ (Ray *et al.*, 1999) and the NO_y-N₂O relation (Fischer *et al.* 2000). In particular the ratio between total reactive nitrogen (NO_y = NO + NO₂ + NO₃ + 2 N₂O₅ + HONO + HNO₃ + PAN + particulate

nitrate) and NO_y*, a proxy for NO_y produced in the overworld by photochemical destruction of N₂O, is very sensitive to additional NO_y mixed into the lowermost stratosphere during cross-tropopause exchange processes (Fischer *et al.*, 2000). This is illustrated in Figure 4. In the overworld NO_y has been found to be anticorrelated with N₂O following the linear relation $\text{NO}_y = (312.7 - \text{N}_2\text{O}) \times 0.073$ ($50 < \text{N}_2\text{O} < 310$ ppbv; dashed line in Figure 4) (Keim *et al.*, 1997). Therefore, N₂O measurements can be used to calculate the amount of NO_y produced in the stratosphere descending from the overworld into the lowermost stratosphere. Mixing between stratospheric air and tropospheric air can result in deviations from this linear behaviour (dotted line in Figure 4), in particular if the NO_y concentration in the upper troposphere is high, e.g. by convective transport from the polluted continental boundary layer, lightning or aircraft emissions of NO_x. Figure 5 shows data obtained during the two STREAM campaigns discussed in the previous section. The difference between the *in-situ* measurements of NO_y and NO_y* has been plotted against N₂O, which is used here as an altitude proxy above the tropopause. An N₂O value of ~312 ppbv is representative for the well-mixed troposphere. Average concentrations of NO_y in the upper troposphere were of the order of 1 ppbv during STREAM 97, and approximately 1.5 ppbv during STREAM 98 (data points below the dotted horizontal line in Figure 4). Cross-tropopause mixing results in an enhancement of NO_y in the lowermost stratosphere. In the winter

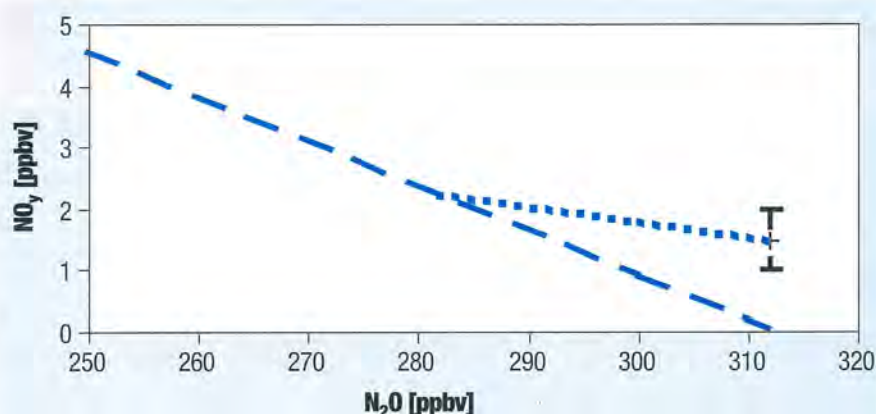


Figure 4: Schematic representation of the influence of stratosphere-troposphere exchange on the NO_y-N₂O relation. The dashed line has been calculated from ER-2 measurements (Keim *et al.*, 1997) and represents overworld conditions. In particular during summer upper tropospheric NO_y can be strongly enhanced (vertical bar at N₂O = 312 ppbv). Cross-tropopause transport of tropospheric air into the lowermost stratosphere can cause excess NO_y within the mixing region (blue dotted line).

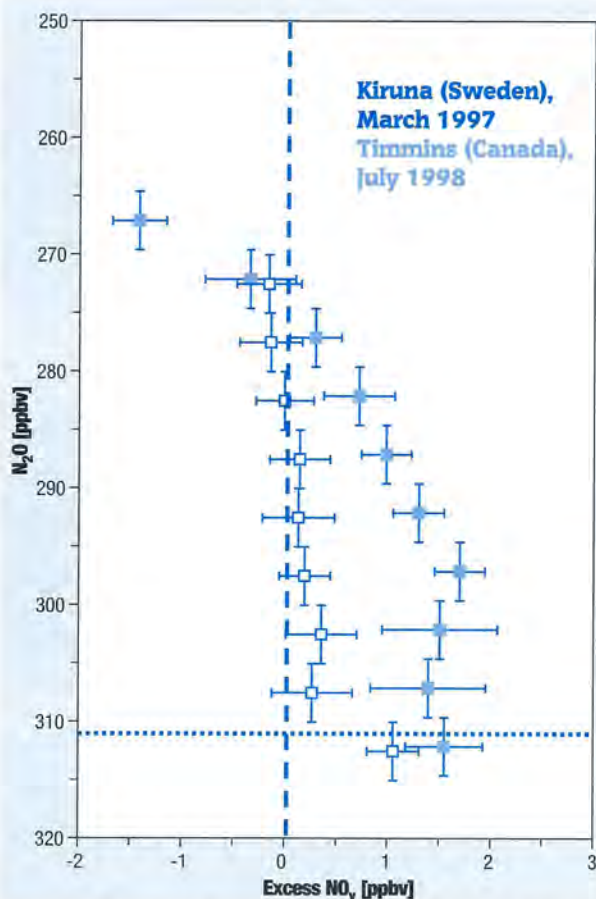


Figure 5: Average excess NO_y ($= \text{NO}_{y\text{measured}} - \text{NO}_{y*}$) versus N_2O . The vertical error bars represent the N_2O ranges used to bin the data. The horizontal bars indicate the 1σ -standard deviation. Note that the two data sets have been vertically shifted by 2 ppbv to avoid overlap of the error bars. The horizontal dotted line indicates the tropospheric N_2O mixing ratio of 312 ppbv. NO_y values below this line represent upper tropospheric measurements.

season this enhancement is small and restricted to N_2O mixing ratios less than 300 ppbv, corresponding to Θ -levels of less than 330 K. In agreement with the findings of Figure 2, excess NO_y is found up to much higher levels in the lowermost stratosphere during the summer campaign ($\text{N}_2\text{O} < 280$ ppbv corresponding to Θ -levels of ~ 360 K). Also the absolute amount of the excess NO_y is much higher in summer compared to the winter season. This is due to the higher tropospheric NO_y concentrations observed over Canada in July 1998, reflecting strong upward NO_x transport and lightning production in thunderstorms.

Conclusions and outlook

Trace gas measurements of CO , N_2O , NO_y , and O_3 in the lowermost stratosphere in March 1997 and July 1998 confirm the existence of a mixing layer above the local tropopause up to potential temperature levels of ~ 330 K

latitudes, so that a latitudinal dependency of cross tropopause exchange can not be excluded as the cause of the observed differences between Stream 97 and STREAM 98.

Additional information will be gained from a forthcoming project (SPURT), supported by the German Federal Ministry for Education and Research (BMBF), in the next 3 years. Within SPURT a total of 8 measurement campaigns will be performed in the lowermost stratosphere (up to 14 km), covering the latitude range from $28^\circ - 78^\circ\text{N}$ along the western border of Europe, during all seasons. On board a Learjet 35A, high precision measurement of CO , N_2O , CH_4 , CO_2 (MPI-Mainz, PI: H. Fischer), H_2O , O_3 (FZ Jülich, PI: C. Schiller), F11 , F12 , SF_6 (University of Frankfurt, PI: A. Engel, U. Schmidt), NO and NO_y (ETH Zürich, PI: T. Peter) will be performed. For the data interpretation and evaluation an idealised mechanistic tropopause model (University

and ~ 360 K, respectively. Above this layer, levels of trace gases and correlation have signatures typical for the overworld, notably very low CO mixing ratios of about 20 ppbv. Adiabatic transport from mid-latitudes, in particular isentropic troposphere to stratosphere mixing, is restricted to the $\Theta < 330$ K layer in the winter season, associated with exchange near the polar jet. During summer additional exchange across the subtropical jet is most probably responsible for the extension of the mixing layer to higher potential temperature surfaces (~ 360 K). This picture is consistent with recent model studies (Chen, 1995; Haynes and Shuckburgh, 2000), indicating a strong seasonal variation of cross tropopause exchange above 330 - 340 K. Unfortunately, the STREAM measurements have been performed at different

of Mainz, PI: V. Wirth), a lagrangian chemistry transport model (FZ Jülich, PI: C. Schiller) and a 3D chemistry-transport model (MPI-C, PI: J. Lelieveld) will be used. It is expected that this project will further improve our understanding of stratosphere-troposphere-exchange at mid-latitudes.

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Convection in the Tropical Tropopause Region and Stratosphere-Troposphere Exchange

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Introduction

It is widely accepted that there are several processes that contribute to the exchange of air between the stratosphere and the troposphere in the tropics, chief among them deep convection (a fast vertical process) and the stratospheric circulation (a slow vertical process). Yet, there is still a troubling uncertainty regarding the balance of processes that regulate the motion of air across the tropical tropopause. This issue is highlighted in the recent SPARC Water Vapour Assessment (SPARC 2000).

There has been a reported increase in stratospheric water vapour (SPARC 2000), which cannot be explained by increases in methane or trends in tropical tropopause temperatures (SPARC 2000). Because of the importance of water vapour for the hydrogen budget of the stratosphere, there is a need to understand the balance of processes that govern the exchange of air across the tropical tropopause. The roles of convection and stratospheric driven upwelling in stratosphere-troposphere exchange are of interest chiefly because rapid convective injection may yield air with a much different chemical signature than slow ascent, and may localise the injection of air into the stratosphere. In addition, given significant anthropogenic radiative forcing, slow ascent (driven by extratropical wave driving) may change over time in a much different way than convection might change (driven by low level tropical convergence and tropical near surface temperatures).

The concept of the tropical tropopause as a 'layer' rather than a single surface was discussed by Atticks & Robinson

(1983) and more recently by Highwood and Hoskins (1998). The tropical tropopause region or tropical tropopause layer (TTL) is the region between the convective equilibrium of the troposphere, and the radiative equilibrium of the stratosphere. It is a transition zone where both stratospheric and tropospheric processes interact. We will focus on the analysis of convection in this transition layer. A convenient definition of this layer is the region between the main convective outflow (10-12km) and the cold point tropopause.

It has long been known that deep convective clouds may penetrate the tropopause in mid-latitudes. In the tropics, there is anecdotal evidence of even deeper convection penetrating the tropical tropopause. These convective events have been invoked by Danielsen (1982) to explain the dehydration of the stratosphere by convective overshooting. More recently Sherwood and Dessler (2001) illustrated how convection can dehydrate the tropopause region.

But how frequent is this deep and very cold convection in the tropopause region and above the tropopause? Drawing on recent work using satellite data with global coverage (Gettelman *et al.*, 2001) we will attempt to illustrate that convective penetration of the tropopause does occur, but it is confined to specific regions and seasons. We also detail the penetration of convection into the tropopause region. This leads to an integrated picture of how convection and subsequent horizontal transport of air in the tropopause region set the "boundary condition" for air that enters the stratosphere.

Convection in the Tropopause Region

To determine the position of deep convection in the tropopause region, cloud brightness temperatures observed from satellites have been analysed by Gettelman *et al.* (2001). Satellite brightness temperature is represented at high space-time resolution in Global Cloud Imagery (GCI), described by Salby *et al.* [1991]. This data set has a spatial resolution of ~0.5° (lat and lon), every 3 hours. Pixels are derived from satellite infrared brightness temperature measurements with a footprint of 8km. These data provide a qualitative description of optically thick cloud in the tropopause region. The uncertainty in actual height for each observation is easily ± 500m. The averaging of 8km observations does not appear to introduce a bias (see Gettelman *et al.* 2001 for a more complete discussion). The advantage to the GCI data is to be able to cover the whole tropics, and to look at the relative frequency of cloud top temperatures observed in different regions at different times of the year.

To examine cloud temperatures relative to the tropopause, the tropopause is defined using the WMO lapse rate criteria, and data from the Reanalyses developed by the National Centers for Environmental Prediction (NCEP) and the National Center for Atmospheric Research (NCAR). Comparing clouds to the position of the tropopause introduces uncertainties associated with the tropopause temperature [Randel *et al.*, 2000] and the modification and deformation of the local tropopause by the overshooting clouds themselves. These clouds are colder than their environment. Thus these estimates of clouds in

the tropopause region do not represent permanent transport into the stratosphere, but rather the locations of a mixing process. In this process convective air overshooting its level of neutral buoyancy will mix with some tropopause region air and sink back down to a new equilibrium level.

Given these caveats, we present the following qualitative view of clouds reaching into the tropical tropopause region. Deep convection occurs throughout the tropics, but in preferred regions. The analysis of brightness temperature from GCI enables the complete convective pattern to be observed with global coverage and sufficient space-time resolution to capture the dominant scales of organised convection. The analysis focuses on clouds in the tropopause region, that is, those clouds above 150 hPa. The 150 hPa level is climatologically at a temperature of about 215° K and the 340 K potential temperature surface.

We define the fractional area with brightness temperature below some temperature in the GCI data as η_T , the fractional area coverage at some temperature T . The fractional area with brightness temperatures colder than 215° K in the GCI data (η_{215}) is illustrated in **Figure 1 (P. IV)**. Seasonally averaged η_{215} has maxima over the continents and the Inter-Tropical Convergence Zone (ITCZ) during local summer. In general, the summer hemisphere has far more convective activity reaching 215°K than the winter hemisphere, and the maxima in fractional area coverage are typically centred off the equator. The highest frequency of cold brightness temperatures at this level are associated with the South Asian Monsoon in July-September, with values nearly as high over Central America during these same months.

The zonal mean fractional area in the GCI data covered by clouds below various threshold brightness temperatures (η_T) is illustrated in **Figure 2** for two months as a function of latitude and temperature. The zonal mean tropopause temperatures are also plotted in **Figure 2** (note that temperature decreases upward).

During both months, the fractional coverage at any temperature is a maximum in the summer hemisphere about 10° from the equator. The edge of winter storm activity in mid-latitudes is visible at 25–30° latitude in the winter hemisphere. The tropopause is ~3°C colder during January (Figure 2a) than during July (Figure 2b). In each month, a small fraction of cloud (~0.5%) does appear colder than the mean tropopause. The maxi-

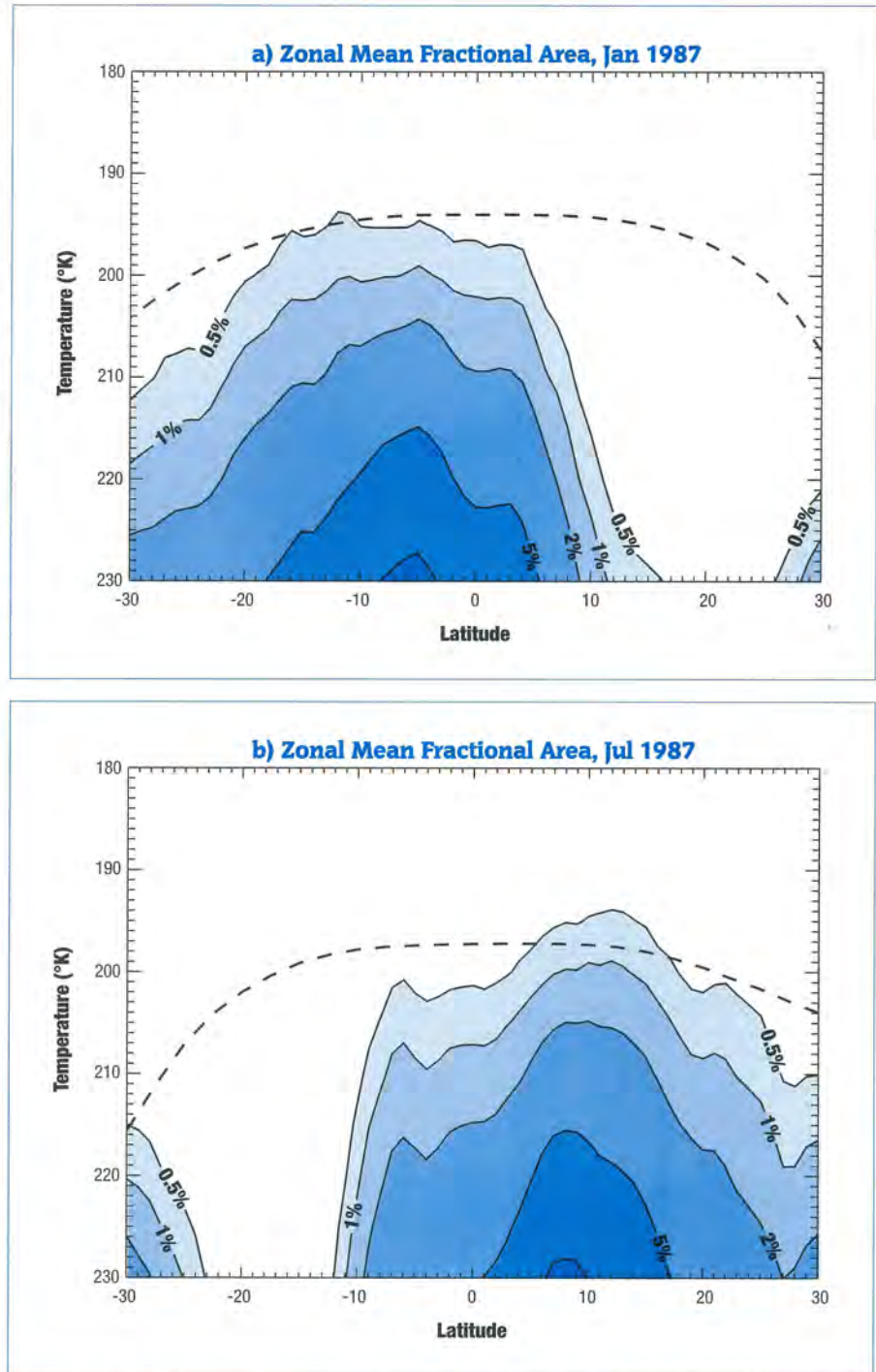


Figure 2: Zonal mean η_T (in percent) as a function of Latitude and Temperature for a) January 1987 and b) July 1987. Dashed line is the zonal mean thermal tropopause. Adapted from Gettelman *et al.* (2001).

mum fractional coverage above the tropopause is larger in July (at ~12°N in Figure 2b) than in January (at ~15°S in Figure 2a), which reflects the warmer tropopause temperatures, since the peak zonal mean fractional coverage is about the same temperature in January and July. The fractional coverage above the tropopause would be smaller if the (4°C) warm bias to the NCEP thermal tropopause is taken into account (Randel *et al.* 2000), or if the cold point tropopause definition is used.

The distribution of clouds whose brightness temperatures are colder than the analysed tropopause is presented in **Figure 3 p. IV**. As discussed, this does not imply irreversible transport of air above the tropopause, only that clouds overshooting their level of neutral buoyancy reach this level, where they may mix with some quantity of air before returning to a lower equilibrium level. A clear annual cycle is visible in Figure 3. Convective events colder than the tropopause are

broadly found in the same regions as deep convective activity in Figure 1. The highest frequencies of brightness temperatures colder than the tropopause temperature in Figure 3, as for those at lower levels in Figure 1, are found over the Indian subcontinent and over Central America in July-September (Figure 3d). The region of North-East India was studied and noted for its convective penetration events by Cornford and Spavins [1973]. Oceanic convection penetrates the tropopause significantly in the central Pacific in January-March of this year (Figure 3b), which was an El Niño warm event. High frequencies of clouds colder than the tropopause are found over the inter-tropical convergence zones, and also over the convective regions of central Africa.

The distribution of convective events into the tropopause region falls off rapidly above the average level of neutral buoyancy, which can be derived from the surface equivalent potential temperature and is approximately the 345 K potential temperature level (Folkins *et al.* 2000). A very small fraction of this convection reaches tropopause levels, as expected from a small proportion of locations with surface equivalent potential temperatures of 370-390 K. Very few individual locations see clouds colder than the tropopause even 3-4% of the time (Figure 3). Even in regions of the most active convection, clouds are colder than the tropopause less than 0.5% of

the time (Figure 2). These observations are consistent with rapid dehydration by a spectrum of convection penetrating into the tropopause region. Sherwood and Dessler (2001) have shown that the required amount of convection necessary to dehydrate air to stratospheric levels (less than 10 ppmv) is at least within the range of the frequency of convection reaching tropopause levels implied in the GCI data.

In general, the maximum occurrence of brightness temperatures colder than the local tropopause temperature occurs in the summer hemisphere, consistent with the annual cycle of convection. Higher frequencies of cloud colder than the tropopause occur in July-September when tropopause temperatures are warmer than in January-March.

Transport and Implications for Water Vapour

Given the distribution of convection in time and space it is natural to ask how it is related to the transport of air into the tropopause region and into the stratosphere. Note that in Figure 2 and Figure 3 that convection colder than the tropopause occurs mostly in the summer hemisphere, and more occurs in August than in February. If convection was the major process governing the entry of water vapour into the stratosphere, then we might expect the water vapour minimum to be in July-September rather than in January-March as observed.

To investigate this discrepancy further, we look at the relationship between convection, temperature and water vapour in the tropopause region. We choose January for this comparison as the month with the coldest averaged tropopause temperatures and lowest averaged water vapour. Figure 4 links the frequency of convection averaged zonally over the western Pacific with cold point temperatures below 188° K (shaded region), regionally averaged tropopause temperatures (dashed line), NCEP/NCAR reanalysis winds (vectors), and water vapour observations (contours) derived from the HALOE instrument on the UARS satellite (Randel *et al.*, 2001). There is a spatial separation in Figure 4 between the deepest convection and the lowest water vapour concentrations. While the deepest convection is south of the equator in the summer hemisphere, the lowest concentrations of water vapour are centred about 15° north of the equator in the winter hemisphere. The minimum descends slightly in the subtropics, consistent with the wind vectors indicating the northward flowing upper branch of the Hadley circulation. These winds imply transport along isentropic surfaces (which slope downward and poleward in this region). The coldest temperatures, unlike the convection, are nearly symmetric around the equator. A picture similar to Figure 4 (though with higher absolute concentrations of water vapour) can also be generated for July, with a water vapour minimum south of the equator, and more convection north of the equator (Figure 2).

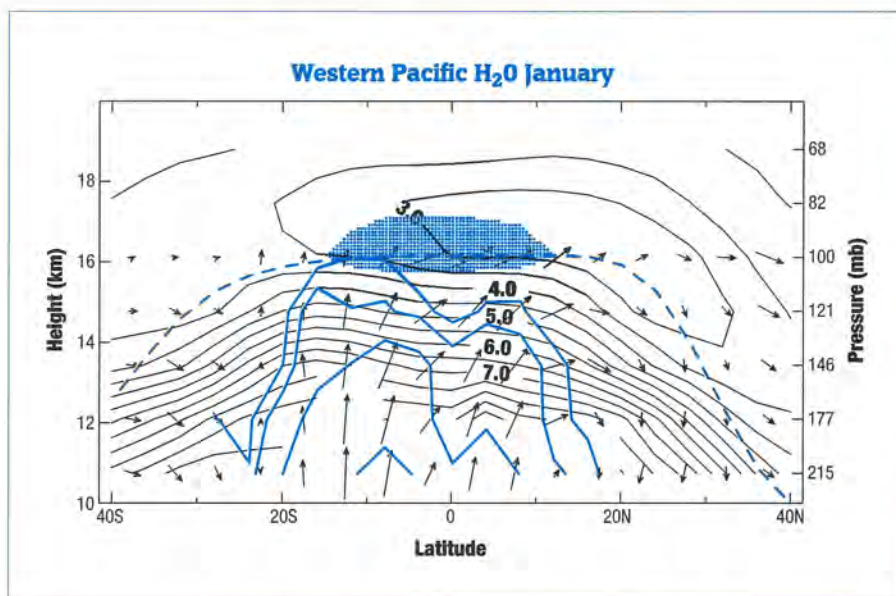


Figure 4: Western Pacific (100-180 longitude) mean: fractional area coverage of clouds colder than a given brightness temperature (thick contours at 10%, 5%, 2%, 1%), water vapour observed by HALOE in ppmv (thin contours), and NCEP/NCAR areal averaged kinematic velocities (vectors, scaled to motion in one day). Shaded region indicates regional monthly mean temperatures below 188° K. Adapted from Randel *et al.* (2001).

Figure 4 hints at some interesting relationships between these fields. This integrated picture implies that both vertical and horizontal motion are important for understanding the distribution of water vapour, since the minimum in water vapour just above the tropopause is well away from convection. One explanation for the water vapour minimum north of the equator is that the rapid horizontal transport of air through very cold regions dehydrates air further as it slowly ascends. This slow ascent is due to radiative heating in the tropopause region and the stratospheric circulation, and a characteristic vertical velocity is on the order of mm/s, in contrast to m/s horizontal velocities. Such a process would help explain the presence of widespread thin sub-visible cirrus clouds (Wang *et al.* 1996). Space based LIDAR measurements suggest that in many cases these cirrus are not

obviously related to convection (Winker and Trepte 1998). The generation and maintenance of these cirrus are not well characterised, but Holton and Gettelman (2001) have recently shown using a simple model that horizontal motions in the tropopause region can dehydrate air and generate cirrus clouds, even in the presence of the reported subsidence in the cold regions of the western Pacific (Sherwood, this issue). In addition to the large-scale temperature fields, small-scale temperature perturbations due to waves (Boehm and Verlinde 2000) may also be important.

The question of what causes the regional distribution of tropopause temperatures is still uncertain. While the seasonal variation of zonal mean tropopause temperatures should be set by the residual circulation of the stratosphere (Holton *et al.* 1995), Johnson and Kriete (1982) demonstrated using radiosondes during MONEX that convection itself has a significant impact on tropopause temperatures. Cooling following mixing with convective air is also consistent with observed and modelled downward motions in the tropopause region (see Sherwood, this issue).

Summary and Future Observations

Satellite brightness temperatures locate convection in the Tropical Tropopause region, and indicate the frequency of convection colder than the local tropopause. These results are qualitatively consistent with the few available observations. The coupling of convection with observations of water vapour, temperature and wind fields allows us to better understand the role of convection in the tropopause region. Convection supplies much of the air into the tropical tropopause region, but subsequent horizontal transport of air is also important for understanding how air enters the stratosphere, and with what water vapour concentration. Observations imply that there is dehydration in the tropopause region away from convection.

What does this integrated picture mean for trends in water vapour? Tropopause level water vapour appears determined (to first order) by the tropopause temperature. This temperature is regulated by convection and the uplift and heating associated with the stratospheric circulation. The

details of the water vapour distribution might be sensitive to horizontal motion in the tropopause region. Furthermore, the magnitude, level and temperature of the convective influx into the tropopause region also affect the supply of water vapour and perhaps constrain its final value. Mixing of convective air is also likely to affect the tropopause region. Two other important processes affecting the entry of air into the stratosphere that have not been discussed here are isentropic mixing from the stratosphere into the tropical tropopause region, and changes in the circulation in the lower stratosphere above the tropopause.

These results can be tested with more advanced instruments than the infrared measurements on operational polar orbiters and geosynchronous platforms (which comprise the GCI data). These instruments include the radiometers and radars on the TRMM satellite (currently flying), and chemical and cloud sensors like HIRDLS and MLS on EOS-AURA (launch in 2003) and the LIDAR on PICASSO-CENA (also launch in 2003). In addition to these new sensors, interpretation of these observations would be aided by a better understanding of the radiative environment in the tropopause region, and the physics surrounding clouds that overshoot their equilibrium level. While the annual cycle of tropopause temperature is broadly understood, and the basic large-scale response of the tropopause to convective heating in the troposphere is fairly well described, the effect of deep convection on the local radiation balance in the tropopause region and above is not well constrained. In addition, the cloud scale dynamics and radiation will likely interact to help determine how the deepest convection evolves, and how it affects the thermodynamics of the tropopause region. Detailed observations from a field campaign in the tropics can be used in conjunction with this comprehensive suite of satellite observations to answer both the small and large-scale questions raised by this coarse picture of the tropopause region.

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Recent Insights into the Regulation of Water Vapour Entering the Stratosphere

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Introduction

Water vapour in the upper troposphere and lower stratosphere has come into increasing focus in recent years. Low-latitude water vapour at these levels is important not only for its greenhouse forcing, but also because the air here is advected into the stratosphere where water vapour plays an important role in regulating stratospheric chemistry and temperatures. The chemical role involves both the production of important OH hydroxyl radicals, and the formation of polar stratospheric clouds which subsequently help destroy ozone. The cooling role exacerbates ozone destruction as well.

As air passes the tropical tropopause it is dehydrated to a value approximately equal to the saturation vapour mixing ratio there, but the processes that decide exactly how much vapour stays in the air are not well understood. A recent SPARC assessment (SPARC, 2000) has thoroughly examined the observations of water vapour available in the stratosphere and concluded that previously reported, and unexplained, upward trends of about 1% per year are consistent over a number of available instrumental records spanning nearly four decades (Rosenlof *et al.*, 2001). This SPARC effort has exposed both the inability of trends in tropopause temperature or chemical production to explain those of water vapour, and the acute need for more and better observations in the tropopause region. The failure of known influences on water vapour to account properly for these important trends reveals a genuine need to understand the stratospheric dehydration processes better than they are now.

Until recently, efforts to deduce further details of stratospheric dehydration have concentrated on precise quantitative comparisons of the mean mixing ratio found in the stratosphere to that expected from calculation of the saturation value at the tropical tropopause with respect to the ice phase. Comparison of these quantities based on data from the 1970's using seasonal tropopause climatologies led to the conclusion (Newell and Gould-Stewart, 1982) that dehydration had to

be limited to the coldest regions of the tropopause (particularly over Indonesia and the western Pacific Ocean), since the observed stratospheric mixing ratios were lower than expected based on the wider tropics. They assumed further that air must be entering the stratosphere in these regions, dubbing this a "stratospheric fountain."

Recent work casts doubt on this entry conclusion, since observations show air to be sinking out of the stratosphere in the regions where the tropopause is coldest, as would be expected in a thermally direct baroclinic circulation (Sherwood, 2000), rather than ascending into it. Though contradicting the "fountain" image, this finding does not rule out localisation of dehydration in the indicated region since (contrary to common assumption) dehydration and upward motion need not have anything to do with one another, at least at large scales. More will be said about this important point below. Though not requiring a "fountain," the old argument in favour of regional dehydration has itself been found lacking since it rests on climatological tropopause values; if instantaneous water-vapour saturation minima from individual soundings are averaged rather than averaging temperature first on pressure surfaces, a much lower mean value results that is no longer inconsistent with stratospheric mixing ratios at least during the UARS era (Dessler, 1999). Thus the question of where air is dehydrated is not really constrained by the early observational methodologies; temporal fluctuations of temperature and moisture must be reckoned with as well as spatial ones.

Dehydration mechanisms

Any mechanism that can effectively dehydrate the stratosphere requires attainment of cold temperatures and removal of enough of the condensed water to reach stratospheric mixing ratios. Two mechanisms currently enjoy favour as candidates for doing this: first, dehydration within very cold convective overshoots as suggested by Johnston and Solomon (1979) and Danielsen (1982), and second, dehydration within large, stably-

stratified air masses near the tropopause lifted by Kelvin waves or other transient motions (e.g., Potter and Holton, 1995). Both mechanisms involve temporary, local cooling of air below the climatological mean value. As it happens, either mechanism would probably be confined to more or less those cold-tropopause regions identified by Newell and Gould-Stewart. This is principally because both mechanisms are either directly or indirectly associated with intense convection, which is in turn highly correlated with cold tropopause temperatures within the tropics. Another factor is that dehydration would be easier in areas of lower mean temperature even if transient fluctuations were not exclusive to those areas.

The main objection to the convective mechanism is that convection rarely overshoots the tropical tropopause, while the minimum in water vapour is at least as high or higher than the mean tropopause thus requiring a dehydration mechanism active at that height. Most convective outflows occur between 150 and 200 hPa in the regions where intense convection occurs (outflows are even lower in places such as the ITCZ). However, some clouds do penetrate the tropopause. Quantifying the vertical extent of convective penetration and mixing in a way that is relevant to stratospheric dehydration (see Gettelman, this issue) is rather difficult.

The main hurdle faced by the second or "*in-situ*" mechanism is the difficulty of nucleating and removing ice crystals. Very few ice nuclei are present in the ambient air at these levels, so homogeneous nucleation is required to initiate freezing; this occurs only at supersaturations of ~60%. Once crystal formation begins, the water vapour mixing ratios quickly fall below the threshold for homogeneous nucleation, shutting off further crystal formation unless temperatures continue to fall. With the right ascent rates, however, reasonably effective dehydration can occur over time, but temperatures must fall farther than would be necessary with equilibrium thermodynamics (Jensen *et al.*, 2001)

and must remain low for at least a day or more. The cirrus clouds that would accompany this process are indeed widely observed, particularly over convective regions where temperatures are coldest and where wave activity is greatest, although they contain little ice. Current observations cannot yet tell us how much actual dehydration occurs in these clouds.

Recent support for, and clarification of, a role by the "overshoot" mechanism

Two recent studies (Sherwood and Dessler, 2000, 2001) have provided further support for the idea that a substantial contribution to the required dehydration comes directly from drying within convective cells. These studies reconsider the behaviour of convective overshoots in light of current understanding of the physics of convection and clouds, and use more recent data.

The earlier of these two studies used correlations found among water vapour, ozone, and other tracer data obtained by the ER-2 at low latitudes (see Figure 1) to argue that the youngest air crossing any given theta surface near the tropopause is also the driest, implying a rapid dehydration process followed by a slower ascent into the stratosphere. In interpreting the relationship, the authors assumed that (outside tropical convection) air can reach theta surfaces near the tropical tropopause only from below in the tropics or from the "overworld" (above 380 K) at mid-latitudes, in accord with prevailing views (Holton *et al.*, 1995). *In-situ* drying of air during uniform ascent in the tropics should not produce such correlations because slow lifting can take many weeks to lift air from the peak outflow level of convection (13–14 km) up to the hygropause (16–18 km), which should allow plenty of time for horizontal mixing processes within the tropics to give each parcel the same dehydration opportunities

regardless of its initial ozone or ascent rate. Nor can the relationship be caused by mixing with overworld air, because the latter's ratio of ozone to water is far too large. It would be worth examining these issues in global models with prognostic water vapour and ozone as they become available.

Probably the most important new point emphasised by the recent work relates to the time-scale argument above but is independent of the dehydration mechanism. This point is that the slowness of the residual, diabatic vertical motion means that dehydration can easily be localised in cold-tropopause regions regardless of

residual circulation details. Both candidate dehydration mechanisms operate within an extended (at least a couple of km) vertical space. This means air is mainly processed through the dehydration regions horizontally rather than vertically. In other words, even air that crosses the tropopause in its warmer sectors will usually, at some point, have passed through the cold sectors where dehydration is occurring. Therefore, variations in how the ascent is *horizontally* distributed would not particularly be expected to cause variations in entry moisture, even if they meant that uplift occurred on average at warmer or colder tropopause temperatures.

Instead, variations of (1) the temperature in the dehydration regions specifically, (2) the rate of horizontal processing of air through these regions, or (3) the efficiency of drying in these regions, would be required to change stratospheric moisture.

Earlier, Sherwood (2000) concluded that the observed sinking through the tropopause over Indonesia required a heat sink or diabatic cooling effect and suggested cooling by convective overshoots as the only apparently workable explanation for this. Building on the earlier studies, Sherwood and Dessler (2001) quantified the drying effect of overshoots by constraining a convective model to produce the correct cooling effect. Getting the right cooling required clouds overshooting the tropopause to cover about 0.5% of the tropics. With this number of clouds and CAPE (convective available potential energy) values consistent with observations, they were also able to obtain realistic profiles of water vapour (see Figure 2), ozone, and cloud ice including a water vapour minimum located just above the mean tropopause. The results were quite sensitive to assumptions about cloud mixing, and did not include a seasonal cycle, so they do not provide firm support for single-handed dehydration by overshoots. However, they go a long way toward deactivating criticisms of the possible importance of overshoots that are based on

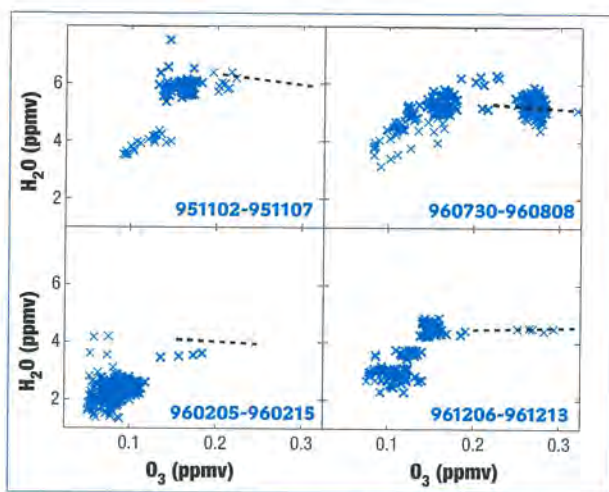


Figure 1. Water vapour vs. ozone from ER-2 data collected near the 380K potential temperature surface during four phases of the STRAT campaign, excluding air of apparent overworld origin. Dashed lines show regression to the overworld data. From Sherwood and Dessler (2000).

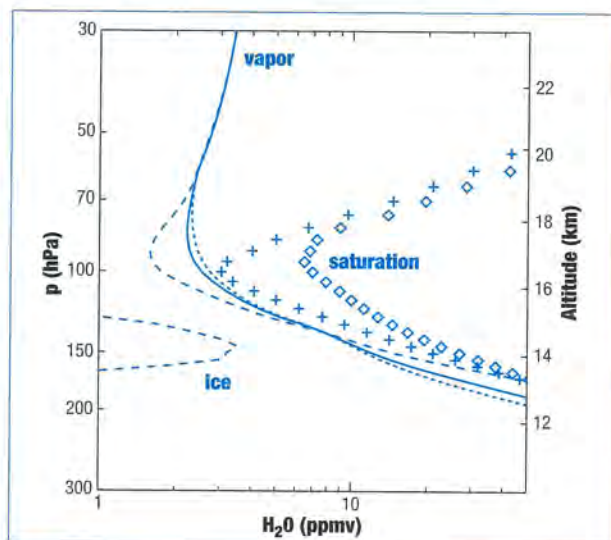


Figure 2. Water vapour (heavy-dashed/dotted lines) respectively (near to / far from) overshooting convection and ice profiles near convection (light dashed), simulated by a radiative-convective model with dehydration by overshoots and horizontal mixing given the environmental saturation curves (plus signs/diamonds). Thick line shows simulated tropical mean vapour. From Sherwood and Dessler (2001).

observed quantities of overshoots being too small, observed cloud-top heights being too low, or ozone below the tropopause being too great. In their simulations, dehydration in a limited region was easily able to control moisture entering the stratosphere throughout the tropics, through horizontal mixing.

It is important to emphasise what is different between the model of Sherwood and Dessler (2001) and common perceptions of overshooting behaviour. It is not physically possible for a convective updraft to shoot past its level of neutral buoyancy (LNB, which is always below the tropopause) and remain there without mixing with the environment. This point was recognised by earlier authors. But if such mixing does occur, it will cool the environment diabatically and thereby pull air gravitationally downward across theta surfaces until final detrainment occurs. Though this downward-moving air mass moves through less distance than did the overshooting updraft, it possesses greater mass since it is a mixture of overshoot plus environment. Thus, the net upward mass flux created by overshoots must be negative for some distance below the highest levels reached, if there is mixing. The sign of this flux at any given level (e.g. the tropopause) depends on overshooting and mixing details. But the only way to achieve permanent upward motion through the highest levels at which mixing occurs is by steady lofting outside of the overshoots, in balance with radiative heating.

Thus it is nonsense to ask what fraction of air "enters the stratosphere" by overshooting versus slow lofting, since the existence of overshooting/mixing only increases the necessary lofting rate at the key levels! The processes do not compete, they conspire. Our best estimates of the amount of ambient lofting through the tropopause (balanced by radiative heating) exceed the Brewer-Dobson mass flux in the lower stratosphere by a factor of several, which requires either flux out of the tropics just above the tropopause, the downward pulling process described above, or both.

What if overshoots occur but do not mix with their environment? If the convective elements were all of uniform LNB, the overshooting ones would sink back into the troposphere and would be much drier but otherwise

no different from non-overshooting ones. Such behaviour would accomplish dehydration. But it would not explain the water-vapour/ozone correlations, would not provide drying at a level higher than about 150 hPa, and would not account for the downward motion at the Indonesian tropopause. It would also not be consistent with the behaviour of other convective boundary layers including those in the laboratory, although quantitative comparison between these analogs and the tropopause region is not straightforward.

On the other hand, if non-mixing clouds had significant variations in LNB, this would produce the same net result as mixing except that the mixing effect would in this case be confined to levels at and below the maximum LNB rather than above, thus failing to reach the tropopause. In this case the H₂O-ozone tracer correlations would still be explained, but not the sinking and drying effects inferred above the highest LNB. This option is possible if the height of the sinking and hygro-pause are exaggerated in the observations or if these features are caused by something other than convective processing of air.

The existence of a robust seasonal cycle in CO₂ and other constituents is sometimes cited as evidence against significant overshooting, but these observations do not rule out an important impact of overshoots on constituents with shorter lifetimes. In either overshooting scenario described above, the overwhelming majority of tracer observations in the lower stratosphere (especially if they are collected far from the most vigorous convection) will inevitably be consistent with the existence of slow lofting of air since, at and above the tropopause, that is what over 99% of the air is doing. Discrepancies will be easy to find only when two species each having concentrations that can change significantly in a few weeks (e.g. water vapour and ozone) are examined together, in which case the lag between convective mixing and lofting can emerge (Figure 1). On a cautionary note, however, there must be a limit as to how deeply in altitude significant mixing or convective detrainment height variation can spread, beyond which damping of the seasonal cycles of all trace constituents entering the lower stratosphere would be implied that is inconsistent with observations. It is not clear yet what

that limit is, especially since the diabatic ascent rate itself varies seasonally, but the matter should be considered carefully in future modelling work.

An issue with both mechanisms is the ability to remove frozen condensate. This issue is less problematic with overshoots, since these will be full of large particles on which vapour can condense and be removed in short order, but the issue must still be considered before the mechanism can be accepted as important. Danielsen's (1982) argument about anvil destabilisation is untenable in light of current understanding of radiative transfer, but may not be necessary. The question has already been explored to some degree for *in-situ* dehydration, but in this case progress is limited by our knowledge of statistics of the large-scale conditions.

Why do we care about the details of which of these processes are dominant? This question goes back to the observed long-term trend in water vapour. For this trend to be explained, some variable needs to be found that can change the effectiveness of an important dehydration process. From our arguments here, it does not appear that changes in the residual circulation would be very effective in doing this, although changes in the horizontal circulation could conceivably have an impact. If convective dehydration is important, then (as noted by Sherwood and Dessler, 2001) changing convective meteorology or ice microphysics should be able to change the moisture entering the stratosphere without any changes in temperature or circulation. Conversely, if *in-situ* removal were responsible, then one would look for changes in wave intensity or available ice nuclei near the tropopause to explain long-term changes in the drying effectiveness.

Conclusion

To summarise, two basic dehydration hypotheses (dehydration within energetic convective updrafts or by dynamical lifting at larger scales) have been discussed. Many conclusions can be drawn that would apply to either mechanism. Both mechanisms would act mainly in limited regions where energetic deep convection is prevalent and the tropopause is coldest. Though such localisation is not supported by the line of reasoning first used to argue it (based on mean mixing ratios in the stratosphere), it is supported both by the

links between either mechanism and convection, and by time-scale arguments that tend to couple horizontal variations in dehydration to those of temperature rather than the residual circulation.

Either mechanism would operate over an extended vertical layer rather than on a surface such as the tropopause. Thus strict passage of air across some infinitesimal surface, regardless of how that surface is defined, should not be envisaged as a physically relevant event at least regarding stratospheric dehydration. Air passing into the stratosphere must run a gantlet of one or more dehydration mechanisms for at least a few weeks before it has safely ascended into the stratospheric overworld. During this time the air should have a chance to sample nearly all longitudes multiple times, which means that even a dehydration mechanism that is confined to a limited horizontal region can operate on nearly all air parcels that enter the stratosphere. Temperatures outside dehydration regions are irrelevant to stratospheric moisture.

The focus here has been on recent work concerning the first, or "overshooting," mechanism. This mechanism fell out of favour after data collected during the 1990's seemed to support only the alternatives, but recent work shows that the mechanism is still viable. Variables concerning this mechanism are how high overshoots reach, and to what extent they mix with their surroundings. Currently these variables are not well constrained. The results of Sherwood and Dessler (2000) are consistent with

either variable energy or variable mixing of overshoots, but in any case argue in favour of rapid dehydration before air begins a lofting process of substantial duration, as opposed to dehydration during or after the lofting. The model of Sherwood and Dessler (2001) assumes overshoots with limited variability of LNB but robust mixing with the near-tropopause environment, and obtains a reasonable result, but this does not prove that mixing of overshoots is important or occurs as high as in their model.

Convective dehydration undoubtedly plays a dominant role in dehydrating tropospheric air from near-surface values of 30,000 ppmv to values of 10-20 ppmv at the main outflow levels. The most important question really amounts to whether convection finishes the job (down to ~4 ppmv) by dehydration in unusually strong updrafts, whether the job is finished by a post-convective "*in-situ*" removal process, or whether there is a combination "one-two punch" involving both. Following up on that question, we must then ask, what controls the effectiveness of these processes?

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Australian activities relevant to Sparc

David Karoly, Monash University, Melbourne, Australia (d.karoly@sci.monash.edu.au)

The major areas of research and operational activity in atmospheric science are concentrated in two government organisations: the Bureau of Meteorology and CSIRO Atmospheric Research, both based in Melbourne. However, these two organisations have relatively little involvement in stratospheric research, and the activities related to SPARC are spread over a number of

small groups in different organisations. From 1993, the Co-operative Research Centre (CRC) for Southern Hemisphere (S.H.) Meteorology, based at Monash University in Melbourne, provided a focus for stratospheric research in Australia and it hosted the first SPARC General Assembly in 1996. However, funding for this Centre ceased and it closed in June 2000.

Observational activities

Ozone monitoring.

Art Downey, Bureau of Meteorology (BM), Victoria

Dobson spectrophotometers are used to estimate total column ozone at five stations (Melbourne, Perth, Brisbane, Macquarie Island, and Darwin). Figure 1 shows column ozone time series from Melbourne. At

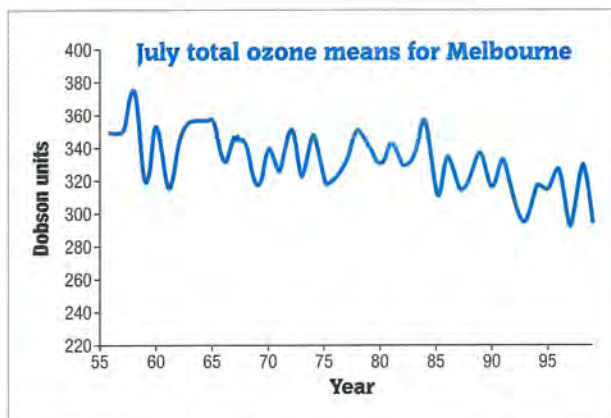


Figure 1. July mean total column ozone at Melbourne from Dobson observations for 1955-99.

non-ozonesonde stations, the Dobson spectrophotometers are also used to perform Umkehr observations, to estimate the vertical profile of ozone in the stratosphere. Regular weekly ozonesonde flights are carried out at Melbourne. In addition, a research ozonesonde program, of one flight per week, was established at Macquarie Island at the end of 1996; the BM plans to continue to operate this important station until at least the middle of 2001 while funding is available.

The data are sent to the WMO World Ozone and UV Data Centre in Toronto, Canada, for archival and international perusal 3 to 4 months after collection. For Macquarie Island, the data cycle is somewhat extended because of logistic problems due to the remoteness of the site.

Ground-based remote sensing of trace gases by solar spectroscopy.

David Griffith, University of Wollongong, NSW. <http://www.uow.edu.au/science/research/acrg/>

A complementary site in the Network for Detection of Stratospheric Change (NDSC) using ground-based FTIR solar spectroscopy has operated at the University of Wollongong, NSW since 1996. High-resolution solar FTIR measurements are made automatically on all days when weather permits, and the spectra are analysed for total vertical columns and partially resolved vertical profiles of over 30 trace gases. The system is augmented with UV/Vis measurements made along the same optical path. Figure 2 illustrates typical measurements (Griffith *et al.*, 1998).

UV radiation monitoring.

Peter Gies, Australian Radiation Protection and Nuclear Safety Agency (ARPANSA), Victoria (Peter.Gies@health.gov.au)

Melbourne, Sydney, Newcastle, Brisbane, Townsville, Cairns, Darwin, Alice Springs, Adelaide, Perth, Macquarie Island, and Mawson, Davis, and Casey in the Antarctic.

In addition, a weather-proof spectral system is running almost continuously in Melbourne. There are also some spectral measurements being done at Davis. Several small portable spectral units are being built to do more regular calibrations at sites other than Melbourne.

Proposed studies of the Antarctic stratosphere using a high spectral resolution Lidar.

Andrew Klekociuk, Ray Morris, Australian Antarctic Division (AAD), Tasmania (andrew.klekociuk@aad.gov.au)

The AAD and the University of Adelaide have developed an atmospheric lidar for investigating climate change in the middle atmosphere (Morris *et al.* (1995), Klekociuk *et al.* (1994). Figure 3 (p. IV) shows the lidar

The UV group in ARPANSA (formerly the Australian Radiation Laboratory) is slowly converting their measurement data-logger sites to two unit systems, a Solar Light UV Biometer 501 (biologically-effective UVR) and Eppley TUVs (total UVR 290 to 400 nm). The IL actinic and UVB detectors are being phased out as they are getting old. Currently, they are running sites at

in operation. From 2001, the lidar will be located at Davis station to profile temperature, wind velocity and aerosol loading in the Antarctic stratosphere and mesosphere over a full solar cycle.

The lidar altitude range is 10km to between about 65km and 90km during both day and night. The upper altitude limit depends on the type of observation being conducted, the time of day, and the time resolution required. The maximum vertical resolution will be about 20m. The basic data set will comprise profiles of molecular density, temperature, wind velocity and aerosol loading as a function of altitude and time.

The instrument can be configured in either a monostatic arrangement which allows coaxial viewing to a maximum zenith angle of 45°, or a higher power biaxial arrangement which is restricted to observing the zenith. The laser wavelength is 532 nm, and the maximum power-aperture product of the system is 24Wm² (see <http://www.antdiv.gov.au/aad/sci/atmos/lidar/lidar.html>).

At Davis, intensive Rayleigh lidar observations will be conducted using both coaxial and biaxial optical configurations. It is also planned to fly radiosondes to altitudes of up to 35km, which when combined with the expected lower ceiling of the stratospheric aerosol layer, should allow the density retrieval method to be calibrated with *in-situ* measurements. In addition, temperature profiles will also be obtained from incoherent Doppler and Raman scatter measurements, which will provide for important

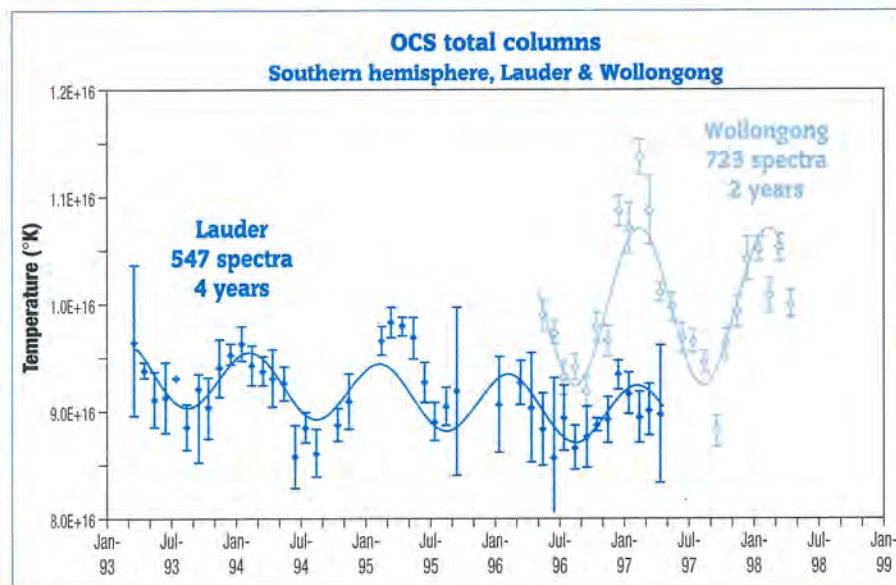


Figure 2. Typical FTIR solar spectroscopy measurements showing a monthly mean time series of total column OCS above Wollongong

cross-comparisons potentially extending down through the aerosol layer.

Ozone-depleting chemicals.

Paul Fraser, CSIRO Atmos. Res., Victoria

In-situ analyses of air at Cape Grim, in north-west Tasmania, and from the Cape Grim air archive, with sensitive GC-ECD and GC-MS systems, are being used to investigate trends in the background atmosphere of ozone-depleting substances (ODS) and their replacements.

The data show that the combined accumulation of ODS has peaked and is declining slowly, due largely to the rapid loss of methyl chloroform from the atmosphere. The on-going decline of total ODS is threatened by the continuing growth of some CFCs (chlorofluorocarbons) and halons, which are still being emitted from their large *ébanksí* in the developed world (old refrigeration and fire-fighting systems) and by their expanding use in the developing world. The rapid accumulation of CFC replacements, such as the HCFCs (hydrochlorofluorocarbons) and HFCs (hydrofluorocarbons), in the background atmosphere is in accord with expectations based on the phase-out of CFCs and other ODS as mandated by the Montreal Protocol.

Gravity waves and radar observations of the middle atmosphere.

Robert Vincent, University of Adelaide, SA (rvincent@physics.adelaide.edu.au)

Observational characteristics of GW in the stratosphere are being determined

from high-resolution radiosonde data. This study is being co-ordinated with the SPARC gravity wave initiative. Radar studies are being used to investigate the variability of winds, GW and turbulence in the middle atmosphere.

Modelling activities

UV radiation.

Lilia Lemus-Deschamps, Lawrie Rikus, BM Research Centre, Victoria

In Australia, operational UV forecasts have been issued by the BM to the public since September 1996. The UV forecast system was developed by the CRC for the S.H. Meteorology in collaboration with the BM (Lemus-Deschamps *et al.*, 1998). The biologically effective radiation dose is calculated from the product of the UV radiation incident at the surface and the erythmal action-spectrum (CIE) integrated over wavelength and divided by 25 mW/m^2 , giving a unitless number defined as the UV Index. The clear sky UV Index is forecast daily for the next day at local solar noon. The forecast system is based on a two stream delta Eddington radiation model, a deterministic total ozone scheme, and the meteorological fields from the global numerical weather forecast model of the BM (GASP). Cloud effects have been included by using a very simple approach; the clear sky UV Index is multiplied by a cloud factor, which is prescribed for different cloud conditions, and which has been determined from the relationship between the erythmal dose and cloud cover.

In an attempt to derive cloudy UV Index forecasts for Australia from meteorological variables and the standard radiative transfer formulation, two approaches have been investigated to derive cloud transmission factors from:

(i) an empirical algorithm between the observed UV Index (Australian Radiation Laboratory), the model clear sky UV Index and the cloud cover (BM), and

(ii) including detailed interactive cloud optical properties in the UV radiation scheme.

Stratospheric circulation modelling.

Greg Roff, BM Research Centre, Victoria

A stratospheric version of the BM global spectral atmospheric circulation model (BAM) has been developed. This is being used as part of GRIPS Task 1 to examine how well the various GRIPS models can capture the initiation, evolution and eventual decay of the polar vortices in both hemispheres. To this end, elliptical diagnostics are being used to compare the simulated vortex 4D lifecycle from several years of daily model and climatology data. Daily elliptical diagnostics are calculated on individual isentropic surfaces by fitting an ellipse to the polar vortex, which is defined as the potential vorticity (PV) contour with the maximum latitudinal PV gradient. The full model and climatology data sets are being processed at present.

Stratospheric chemical transport modelling.

Ian Plumb, CSIRO Telecommunications and Industrial Physics, NSW

This group maintains a suite of atmospheric models, comprising a box model and 2D and 3D CTMs. The models share common chemical solvers and radiative transfer routines. The transport component of the 3D CTM is based on the NCAR MATCH model.

Current research activities include modelling long-term trends in ozone, with particular emphasis on the role of temperature changes resulting from greenhouse gases (see **Figure 4**) and on increasing concentrations of N_2O and CH_4 . These factors are important in determining the evolution of stratospheric ozone and therefore the amount of UV radiation reaching the ground (Plumb and Ryan, 1998), as halogen levels in the atmosphere decline. Recent laboratory measurements of key reactions involving nitrogen species have resulted in increased calculated levels of active nitrogen in the middle stratosphere (Randeniya *et al.*, 1999) and inclusion of these kinetic data in the model also has a major effect on calculated ozone trends.

In other studies, the role of transport variations in determining long-term ozone trends and the influence of polar processes on mid-latitude trends are being investigated. Polar processes are also being studied with the 3D CTM, using both GCM and analysed winds. The aim of this work is to simulate the interannual variations in Arctic and

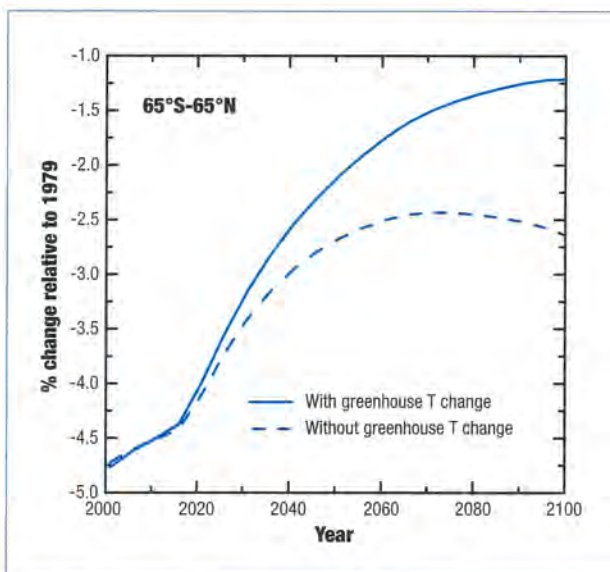


Figure 4. 2D ctm STYDY of impact of greenhouse-gas-induced temperature changes on the future recovery of global column ozone amounts.

Antarctic ozone depletion. Comparison of the concentrations and correlation (Plumb *et al.*, 1999) of long-lived tracer fields generated using the 2D and 3D CTMs provides additional insight into the role of transport processes in the stratosphere. The impact of increases in air traffic, including both subsonic aircraft and proposed fleets of supersonic aircraft, is also being studied.

Interannual variability and trends in the Southern Hemisphere stratospheric circulation.

David Karoly, Monash University, Victoria

In collaboration with the stratospheric modelling groups at BMRC in Melbourne and CSIRO TIP in Sydney mentioned above, a diagnostic and modelling study is being undertaken to investigate the causes of the interannual variations and

long-term trends in the S. H. stratospheric circulation and ozone. This project is an extension of research activities started in the CRC for S. H. Meteorology.

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EGS-2001 General Assembly

Nice, France, 24-29 March 2001

OA.21 - Tropospheric aerosols: formation and heterogeneous chemistry

Conveners : Markus Ammann (PSI, Villigen, Switzerland, markus.ammann@psi.ch), Christian George (CNRS-LACE, France, Christian.George@univ-lyon1.fr), Thorsten Hoffmann (ISAS, Germany, hoffmann@isas-dortmund.de)

In an overview lecture, **A.R. Ravishankara** (Aeronomy Lab., NOAA, Boulder, USA) discussed the role of gas phase and heterogeneous/multiphase reactions in the atmosphere, with particular emphasis on the interplay between the reactions in these media. Following talks dealt with the atmospheric processing (or ageing) of aerosols in the atmosphere as observed in the field, or simulated in the laboratory or in numerical models.

J. Abbatt (Univ. of Toronto) presented new studies of tropospheric ice chemistry in which the interactions of a number of trace gases with laboratory ice surfaces have been observed.

It appears that sulphur dioxide and hydrogen peroxide interact poorly but react together on ice to form sulphuric acid. These reactions may be non-negligible in the presence of moderately thick ice clouds. **P. Behr** *et al.* (Univ. of Essen and Wisconsin) also showed molecular beam studies directly useful for a better understanding of the aerosol-mediated decomposition of HCl and HBr on sulphuric acid aerosols. Such studies have the potential to bring the capture of many molecules. Ion induced aerosol formation in the upper troposphere (UT) was reviewed by **F. Arnold** *et al.* (MPI Heidelberg) with an emphasis placed upon findings obtained in aircraft based and laboratory measurements. They measured the mass distribution and composition of ions in different UT environments including cloud free air masses, clouds and exhaust plumes of jet aircraft. These measurements clearly show cases of rapid ion growth leading to large ions, which represent already stable aerosol particles.

C. O'Dowd (Univ. of Ireland and Helsinki) discussed recent advances in elucidating new particle formation in the marine and coastal boundary layers. It is speculated that iodine oxide over the open ocean may also provide the material for the production of new particles of detectable sizes. **T. Koop** *et al.* (ETH Zurich) also discussed the microphysics of sea-salt aerosols at low temperature suggesting that they are most likely liquid most of the time under polar marine boundary conditions. **R. Sander** *et al.* (MPI Mainz) then discussed the existing published and unpublished data sets showing widespread halogen activation. They compiled measurements of bromine in aerosols particles and in the gas phase at marine locations at low and mid-latitudes.

These presentations were followed by a short but lively poster session.

It was decided to hold a similar session at the EGS assembly next year with more time devoted to posters.

OA.22 - Chemistry of the upper troposphere and lower stratosphere

Conveners: **Th. Peter**, Institute for Atmospheric and Climate Science, ETH Zürich, Switzerland, Thomas.Peter@ethz.ch, <http://www.iac.ethz.ch>, **K. Law**, Department of Chemistry, University of Cambridge, U.K., Kathy.Law@atm.ch.cam.ac.uk, <http://www.atm.ch.cam.ac.uk/>, **A. Bregman**, IMAU, Utrecht University, The Netherlands, A.Bregman@phys.uu.nl, <http://www.fys.ruu.nl/~wwwimau/>

The session was supported by NERC (UK). It included three sub-sessions:

1. Photochemical processes,
2. Chemical observations and mixing processes,
3. Heterogeneous chemistry, aerosol and cloud processes.

The first sub-session, focussing on processes studied by laboratory measurements and by chemistry-transport modelling, was started by **R. Zellner**, who stressed the general importance of acetone photolysis for the HO_x budget in the tropopause region. Besides the commonly assumed products CH_3CO and CH_3 , a second pathway is possible from the excited CH_3CO , leading to 2 CH_3 radicals and one CO molecule. Although the quantum yield of this process is almost unity below about 230 nm, it almost drops to zero beyond 270 nm, so that this photolysis pathway gives only a minor contribution. **N.M. Donahue** presented laboratory studies on the reaction $\text{NO}_2 + \text{OH}$ and showed that the formation of several metastable products is possible, with HONO as the most favourable. However, no evidence of these products was found in the experiments, basically corroborating the JPL recommendations. Balloonborne observations of IO at Kiruna and Gap in 1999 were presented by **H. Bösch**. Taking proper account of solar centre-limb darkening corrections reduces the fraction attributed to IO absorption significantly. They deduced an upper limit in the lower stratosphere of 0.06–0.99 pptv at Kiruna and 0.14 pptv at Gap. **Th. Röckmann** investigated N_2O isotope fractionation. Photochemical processes lead to altitude dependent isotopic enrichment of heavy isotopes that provides information on the relative contribution of photochemical and transport processes in the stratosphere. Comparisons between balloon-

borne observations and results from the Mainz global 2D chemistry-transport model showed an underestimation of the isotopic enrichment by the model possibly caused by model transport deficiencies. **M. Dameris** presented model results of the chemical effect of decreasing the cruising altitude of the subsonic air traffic by 1 km. In the tropopause region this resulted in a decrease by 30 % in NO_x , and in the free troposphere in a 10 % increase.

The second sub-session was devoted to measurements in the tropopause region. **J. Baehr** showed NO , NO_y , O_3 , and CO measurements both in the northern and southern hemisphere (SH and NH) as part of the INCA campaign. She showed concentrations in the SH to be on average a factor of 3–5 lower than in the NH. Enhanced levels were observed in the free troposphere in the ITCZ. **D.E. Oram** discussed canister samples in the free troposphere of a variety of halons from aircraft flights between Frankfurt and Johannesburg as part of the CARIBIC program. He demonstrated that several CFC substitutes increased between 1998 and 2000. The total bromine budget was estimated as 22–24 pptv. Evidence was further given of biomass burning from the correlation between CO and CH_3I . **M.O. Koehler** presented an ozone climatology based on MOZAIC between 1994 and 2000. Especially in the relatively warm winters 1997/98 and 1998/99 there was more ozone in the mid-latitudes at the 340 K isentropic level. A comparison with SLIMCAT revealed the model to underestimate ozone during these winters due to ozone loss calculated at higher latitudes. **W.J. Collins et al.** showed result from a Lagrangian model experiment including 10^5 air parcels and 70 chemical species. Their approach allows to distinguish shallow stratosphere-troposphere exchange processes from deep exchange. **S.A. Penkett** gave an overview of the results from the ACTO and EXPORT aircraft measurement campaigns. The observations included a large set of different species and particles, yielding evidence of anthropogenic combustion products in the free troposphere, transported upwards by convection. The discussion of ACTO was deepened by **J. Methven** illustrating how these observations can be used to identify the origin of the air masses. **A. Zahn** introduced the

chemopause as a measure of the tropopause, derived from the O_3 -CO relation observed on the CARIBIC flights. He showed that the chemopause exhibited a significant seasonality with altitudes varying between 80–120 ppbv. **E. Pavelin** showed results from LIDAR and airborne tracer measurements in the tropopause region above Wales. They traced a significant variation in ozone correlating with wind gusts. From the observations they could deduce an eddy diffusivity of $1.6 \text{ m}^2 \text{ s}^{-1}$ in gravity waves and 0.05 – $1.1 \text{ m}^2 \text{ s}^{-1}$ in the jet stream.

The final sub-session covered heterogeneous chemistry in observations and modelling. **R. Salawitch** highlighted the conflicting interpretation of lower stratospheric HCl measurements during the last decade. The observations seem to indicate a recovery of HCl after Mount Pinatubo eruption in 1991, while another interpretation of the same data set concluded no change in HCl. He also demonstrated an imbalance in inorganic chlorine (Cl_y), based on recent airborne campaigns (SPADE, SOLVE). **J. Hendricks** showed box model results of heterogeneous chemistry on cirrus cloud particles in the upper troposphere. By introducing periodic cirrus events he concluded that the fate of HNO_3 taken up on the ice particles is central for the calculated ozone loss. **G. Pitari** presented model intercomparisons of the effect of heterogeneous chemistry on ozone by assuming a double aerosol density produced by increasing subsonic air traffic. A substantial decrease of the current radiative forcing estimate must be expected, resulting from heterogeneous chemistry. **H. Ziereis** discussed NO_y uptake on cirrus cloud particles, based on the INCA observations. For surface areas of 100 – $1000 \mu\text{m}^2/\text{cm}^3$ more than 95% of the NO_y resided in the gas phase, whereas with larger surface areas ($> 5000 \mu\text{m}^2/\text{cm}^3$) 20–30 % uptake was observed. The estimated fractional coverage was less than one monolayer. Finally, **A. Thomas** showed particle profiles in the tropical region, observed during APE-THESIO, February 1999. Ultrathin high level clouds were detected at 17–18 km altitude, close to the tropical tropopause, with particle diameters around $10 \mu\text{m}$. The effects of these clouds, e.g. on dehydration, are not yet well understood.

The three sub-sessions were corroborated by a common poster session featuring 14 contributions, spanning the full spectrum of investigations in the UTLS, from 3D modelling work, via lab studies to field observations.

ST2.01 - Middle atmospheric dynamics

Convener: Martin Juckes, Clarendon Laboratory, Oxford, UK (juckes@atm.ox.ac.uk)

The 1.5-day session began with several talks on the issues surrounding the Arctic Oscillation: is it a 'physical' mode, in the sense of having an intrinsic dynamical identity, or is it just a statistical construction? Does the downward phase propagation seen in several papers represent downward influence or is it purely an artefact of differing response times? Studies using a variety of approaches, from simple 1D models through to comprehensive statistical studies of NCEP and ECMWF data showed many new insights, but a comprehensive consensus remains elusive. Two presentations showed that the downward propagating signal was associated with strongly disturbed winters. Another study showed the characteristic downward phase propagation to be associated with a coherent 4-month cycle in EOF's 1 and 2 of the zonal mean zonal wind.

Studies of transport emphasised the importance of chemistry-dynamics feedbacks in the dispersion of volcanic aerosols and the importance of regional distribution of troposphere to stratosphere exchange in determining the water vapour budget of the stratosphere.

Work on the influences of solar variability emphasised the importance of resolving the differences in variability in different spectral bands. One study predicted that the Antarctic ozone hole would persist beyond 2050.

A mechanistic study of polar stratospheric warmings found surprisingly strong sensitivity to the zonal wind in the upper (40-50km) tropical stratosphere. The modulation of stratospheric mixing associated with inter-annual variability in the troposphere was described.

In the sub-session on gravity waves (GW) new results suggesting the exis-

tence of a significant source of GW in the southern winter polar stratosphere were presented. The observed peak in gravity wave energy in the tropics was discussed. Another study showed that GW can force planetary waves in the mesosphere. Measurements by the Egrett aircraft over Wales gave very high-resolution information over the structure of mountain waves and the turbulence caused by their breaking. A parameterisation of GW which reproduces the k^{-3} saturated spectrum was described.

The poster session described sensitivity of the Brewer-Dobson circulation to CO_2 and sea surface temperature changes, analysed discrepancies between ground based and space based water vapour measurements, described a modelling study of the QBO and gave an outline of the resources provided by the British Atmospheric Data Centre.

A more detailed report is available at http://www.atm.ox.ac.uk/user/juckes/egs01_st201_rep.html

ST2.02 - Middle atmosphere chemistry

Convener: Rolf Müller, Forschungszentrum Jülich, Germany (ro.mueller@fz-juelich.de)

Owing to the nature of the session, a wide variety of topics were covered. T. Neubert *et al.* presented results on the first European sprite campaign; there were also presentations on the importance of CH_3CN and HCN in the middle atmosphere. R. Salawitch (and colleagues) from JPL discussed the importance of HNO_4 for HOx chemistry. Two presentations on both theoretical analysis and on measurements of isotopic enrichment of stratospheric N_2O were made by a group of scientists from the Max-Planck Institute for Chemistry in Mainz. Several satellite measurements were discussed: M. Riese *et al.* discussed the modelling of nitrogen species from CRISTA, S. Tilmes *et al.* showed an analysis of chemical ozone loss in the Arctic in 1996-97 from ILAS and HALOE, and very recent results from the winter 2000-2001 from GOME were presented by M. Weber of the GOME team in Bremen. GCM results on the coupling between climate and ozone chemistry (C. Brühl *et al.*) and CTM results on the interaction on mixing

and chemistry were presented (Y. Orsolini *et al.*). Further foci were PSC and H_2O observations as well as several papers on ozone measurements from mid-latitudes, and from Norway (ALOMAR, Andenes).

ST.8 - Atmospheric ozone (co-sponsored by OA)

ST8.01 - Global ozone

Conveners: Michaela M. Hirschberg, Technical University, München, Germany, (hirschberg@met.forst.tu-muenchen.de), Peter Fabian, University of Munich, Germany (fabian@met.forst.uni-muenchen.de)

This year sub-session on 'Global Ozone' attracted 13 oral contributions and 15 poster presentations. The audience of about 50 scientists listened to an interesting program with lively discussions.

Five papers were given on model calculations to investigate the feedback of dynamical, chemical and radiative processes. A. Jrrar (Cambridge Univ., UK) studied the dynamical contribution of the downward trend in northern mid-latitudinal ozone with the SLIMCAT model and by a statistical trend analysis. M. Dameris (DLR, Germany) presented an interactive coupled chemistry-climate model to estimate the development of dynamical and chemical key parameters. Another Cambridge University study of P. Braesicke and J. Pyle was assessed to how the tropospheric forcing is related to the total ozone winter increase in high latitudes. G. Pitari (Univ. of L'Aquila, Italy) investigated the impact of stratospheric sulphate aerosols in relation to the ozone recovery rate. C. Brühl (MPCH Mainz, Germany) gave an overview on 20 years 'time-slice' experiments with a fully coupled CCM to investigate the future of the Arctic ozone behaviour.

Two papers showed measurements of total ozone and its interannual variations in high latitudes (C. Basdevant and H. Teitelbaum, LMD-IPSL, Paris, France) and over Tibet (Zou *et al.*, Chinese Academy of Science, Beijing, China).

Two papers dealing with GOME total ozone observations were presented. H. Shets (Royal Met. Inst., Belgium) discussed a comparison with ground-based observations, and A. Ortenzi

(Univ. of Tor Vergata Rome, Italy) analysed the November 1999 ozone hole over Europe. Another November 1999 mini hole investigation based on POAM III data was given by **L. Hood** (Univ. of Arizona, USA).

A statistical approach to ozone was introduced by **R. Toumi** (Imperial College, London, UK), who discussed in his presentation the non-Gaussian or Gaussian statistics of time series of ozone and temperature.

The last two presentations showed measurements of tropospheric ozone. **R. Zbinden** (Observatoire Midi-Pyrénées, France) gave an interesting overview on the measurements made on 15,000 flights of normal A340 aircraft between 1994-2000 in the frame of MOZAIC. **G. Laneve** (Univ. of Rome, Italy) discussed the importance of ozone data from the African continent which can contribute to our knowledge of the tropical atmosphere.

The poster session was very well attended. 15 posters were presented both on ground based measurements and on model based studies.

ST.23 - Results from the APE-GAIA campaign in the Antarctic region

Conveners: Bruno Carli, IROE-CNR, Italy (carli@iroe.fi.cnr.it), Cornelis Blom, IMK-FZK, Germany (cornelis.blom@imk.fzk.de), Gianluca Redaelli, University of L'Aquila, Italy (gianluca.redaelli@aquila.infn.it)

This session aimed at the presentation of first consolidated results from the APE-GAIA (Airborne Polar Experiment-Geophysica Aircraft In Antarctica) field campaign carried out over the Antarctic peninsula, to investigate lower stratospheric chemistry and transport at the boundary of the southern polar vortex. The campaign was based in Ushuaia (Argentina) and performed in September-October 1999.

An overview was given by **U. Cortesi** (IROE-CNR, Italy). He recalled the scientific objectives and the main components of the mission funded by the Italian National Programme for Antarctic Research (PNRA) with contributions from nine countries involved in aircraft measurements, meteorological and modelling support and other co-ordinated activities. A

synopsis of the five Antarctic flights was provided, along with a summary of results from both airborne observations and ground-based or balloon-borne correlative measurements.

The subsequent presentations focussed on the findings from individual instruments and on specific events. **M. Hoepfner** (Forschungszentrum Karlsruhe, Germany) reported about limb-sounding emission measurements of ClONO_2 and HNO_3 performed by the Mid-infrared spectrometer MIPAS-STR during three flights at the boundary and inside the Antarctic stratospheric vortex. During the 990923 (23rd September 1999) and 991012 flight the inner part of the vortex was reached, with clear evidence of denitrification. On 991002, the aircraft entered a region with very low temperatures due to mountain wave activity over the Antarctic peninsula at the vortex edge where dense PSC were observed. By correlation of MIPAS HNO_3 and ClONO_2 data with tracers, the amount of de-nitrification and its vertical extent below the aircraft could be estimated. On 990923 the denitrification was observed above 15km and on 991012 above 14km.

An important case study, aiming at cross-validation of lower stratospheric composition measurements performed by the two limb sounders onboard the M-55, was presented by **C. Blom** (Forschungszentrum Karlsruhe, Germany). Intercomparison of 2D cross sections for ozone measured by SAFIRE-A and MIPAS-STR during the flight of 990923 highlighted, within a good overall agreement, specific discrepancies that require further analysis.

Results obtained by the GASCOD-A4 π UV/Vis spectroradiometer were discussed by **I. Kostadinov** (ISAO-CNR, Italy). The instrument, operating in the 280-850 nm spectral range, performed remote-sensing measurements of O_3 , NO_2 and BrO slant column by applying the DOAS methodology. Quasi *in-situ* values for the same gases were calculated by using a new retrieval code developed at ISAO. GASCOD-A4 π observed also 2 π upwelling and down-welling solar irradiation, thus measuring the actinic flux and calculating the photolysis rate of NO_2 along the flight.

G. Redaelli (Aquila, Italy) discussed the modelling tools used to identify transport related tracer variation during the flights. Based on such pro-

ducts, a filament of vortex air had been correctly forecasted for the 991008 and then sampled with great detail at different altitudes, by *in-situ* and remote sensing instruments onboard the M-55. He presented also a work by **M. Volk** (J.W. Goethe-Univ. Frankfurt, Germany), on the ozone loss and mixing, as derived from M-55 *in-situ* observations of tracers and ozone. While nearly complete ozone loss was found only inside the vortex, ozone loss up to 60% was found routinely even outside the region delineated by the sharp tracer gradients. **F. Cairo** (IFA-CNR, Italy) discussed the principal results on PSC observed by the M-55 aerosol payload, in comparison with synoptic and mesoscale modelling.

Posters addressed both technical and scientific APE-GAIA issues. One poster focussed on the stabilisation of the line of sight implemented for the MIPAS-STR and on its validation from atmospheric spectra measured during the 990923 flight. An estimate of the total inorganic chlorine budget and partitioning, based on simultaneous measurements of ClONO_2 by MIPAS-STR and of HCl and ClO by SAFIRE-A, was shown. Finally, *in-situ* measurements of ozone by the fast chemiluminescence sensor FOZAN, and water vapour by the fluorescent Hygrometer FLASH, were described. These data were used for the detection of the chemical boundary of the ozone-depleted region and a filament out of the vortex edge.

Future SPARC and SPARC-related Meetings

- 18-30 August 2001 : IAGA-IASEI 2001 Joint scientific Assembly, Hanoi, Vietnam**
(<http://www.IAGAandIASPEL.org.vn/pta.htm>).
- **G2.04:** Solar Activity Effects on the Middle and Lower Atmosphere. Convener: E.S. Kazimirovsky, Institute of Solar Terrestrial Physics, Russian Academy of Sciences, PO BOX 4026, 664033, Irkutsk, Russia. Fax: +7 3952 462557, email: edkaz@iszf.irk.ru. Co-convener: S. Sabatini (sofia@astro.yale.edu).
 - **G2.05:** Propagation of Tidal and Gravity Waves and Planetary Signature Into and their Effects on the Mesosphere, Thermosphere, and Ionosphere. Convener: J. M. Forbes, Department of Aerospace Engineering Sciences, Campus Box 429, University of Colorado, Boulder, CO 80309-0429, USA. Fax: +1 303 492 7881, email: forbes@zeke.colorado.edu. Co-conveners: K.P. Hamilton (USA), A.D. Aylward (UK), D. Pancheva (Bulgaria).
 - **G2.09:** Long-term Trends in the Mesosphere, Thermosphere, and Ionosphere Systems. Convener: G. Beig, Indian Institute of Tropical Meteorology, Dr. Homi Bhabha Road, Pashan, Pune-411 088, India. Fax: +91 20 5893825, email: beig@tropmet.ernet.in. Co-convener: J. Lastovicka (Czech Republic).
- 19 August - 01 September 2001: Cargese International School, Cargèse, Corse, France**
(<http://cargese.univ-corse.fr>)
- Dynamical barriers, stirring and mixing in geophysical flows - Mathematical model and applications
- 20-24 August 2001: Workshop on Atmospheric Chemistry and Climate Interaction during the Climate Conference 2001, Utrecht, The Netherlands**
(<http://www.fys.ruu.nl/~wwwimau/cc2001.html>)
- 24-27 September 2001: Network for the Detection of Stratospheric Change: NDSC 2001 Symposium «Celebrating 10 years of Atmospheric Research», Arcachon, France**
(<http://www.observ.u-bordeaux.fr/public/ndsc.symp>)
- 25 September - 5 October 2001: School on the Physics of the Equatorial Atmosphere, ICTP, Trieste, Italy**
(<http://www.ictp.trieste.it/cgi-bin/ICTPsmr/mklinks/mklist?smr1328>)
- 4-7 December 2001: SPARC Scientific Steering Group, Ninth Session, TUPC, Honolulu, Hawaii. Convener: K. Hamilton (kph@phoenix.princeton.edu)**
- 22-26 April 2002: EGS XXVII General Assembly, Nice, France**
(<http://www.copernicus.org/EGS/egsga/nice02/programme/overview.htm>)
- 10-19 October 2002: 34th COSPAR Scientific Assembly and 2nd World Space Congress, Houston, TX, USA**
(<http://www.copernicus.org/COSPAR/COSPAR.html>)
- **A1.1/C2.7:** Climate Change Processes in the Stratosphere and at the Tropopause. Convener: J. Gille.
 - **C2.3:** Long-term Trends in Stratosphere-Mesosphere-Thermosphere Coupling. Convener: D.K. Chakrabarty

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