A mechanism for moistening the lower stratosphere involving the Asian summer monsoon

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SUMMARY

This study employs European Centre for Medium-Range Weather Forecasts (ECMWF) re-analysis data and the contour advection technique to investigate the water vapour distribution in the upper troposphere and lower stratosphere. Water vapour is the primary greenhouse gas and understanding the processes which determine its distribution and transport is crucial. Of special interest is the exchange of water vapour across the tropopause. This study considers how the Asian summer monsoon affects the moisture budget of the upper troposphere and lower stratosphere. The region of the Asian summer monsoon is identified as a significant moisture source for the upper troposphere outside the deep tropics. Monsoon convection moistens the region of the upper-level monsoon anticyclone which is located close to the dynamical tropopause, where isentropes cross from the troposphere into the stratosphere. An isentropic analysis reveals that transport from the troposphere into the stratosphere in this region is normally prevented by the strong potential-vorticity gradients around the tropopause. However, midlatitude synoptic disturbances occasionally interact with the monsoon anticyclone and pull filaments of tropospheric air from its northern flank. These filaments, characterized by high values of humidity and low values of potential vorticity, can extend far north and transport moisture irreversibly into the northern hemisphere lower stratosphere. MOZAIC (Measurement of OZone by Airbus In-service airCraft) data are used as an independent data source to validate the results obtained from the ECMWF analyses.

KEYWORDS: Contour advection Troposphere–stratosphere exchange Water vapour

1. INTRODUCTION

The amount of water vapour in the upper troposphere and lower stratosphere is small. Nevertheless, water vapour in this region has an important influence on the climate system (Harries et al. 1996). First, upper-tropospheric humidity is radiatively very active; it dominates the radiative cooling of the earth in the far infrared in the absence of clouds and contributes strongly to the atmospheric greenhouse effect (Clough et al. 1992). Second, stratospheric water vapour is a very good tracer for transport processes in the middle atmosphere (Hintsa et al. 1994; Mote et al. 1996). Third, it plays an important part in the total hydrogen budget (mainly H₂O, CH₄, H₂) of the stratosphere (Dessler et al. 1994); studying the relationship between the different species of this budget can yield information about dynamical processes and physical conditions of the stratosphere. Finally, water vapour plays a crucial role in stratospheric chemistry; it is involved in the formation of polar stratospheric clouds (Solomon 1988) that are responsible for ozone destruction in polar regions in spring (Farman et al. 1985), known as the ozone hole. Water vapour is also chemically important as the main source for hydroxyl radicals in the stratosphere.

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(Brasseur and Solomon 1984). This very reactive species is involved in many photochemical reactions and can, for example, contribute to ozone depletion in middle latitudes. Consequently, knowledge of the distribution and time evolution of water vapour in the upper troposphere and lower stratosphere are of great interest, especially the transport of water vapour across the tropopause from the troposphere into the stratosphere.

The main sources of lower-stratospheric water vapour are believed to be the areas of deep convection in the tropics, where overshooting cumulonimbus clouds can mix tropospheric and stratospheric air. Brewer (1949) and Dobson (1956) were the first to postulate an exchange between the troposphere and stratosphere in the form of a circulation with mean large-scale ascent in the tropics, large-scale subsidence in the extratropics, poleward transport in the stratosphere and equatorward transport in the lower troposphere. Because the tropical tropopause is very cold, with zonal mean temperatures of $-70$ to $-80$ °C (Highwood and Hoskins 1998), air that crosses it in the ascending branch of the Brewer–Dobson circulation is freeze-dried (Newell and Gould-Stewart 1981; Johnston and Solomon 1979; Danielsen 1982). Consequently, the lower stratosphere is extremely dry with humidity concentrations around 2.5 mg kg$^{-1}$ or 4 p.p.m.v. (parts per million by volume) (Dessler et al. 1994). This is in strong contrast to the concentrations of upper-tropospheric humidity which can be of the order of several hundred p.p.m.v. (Dessler et al. 1995).

The basic idea of stratosphere–troposphere exchange as described by Brewer and Dobson is illustrated in Fig. 1(a). The location of the tropopause is represented by the thick line. It lies around 100 hPa in the tropics and slopes down to about 300 hPa in polar regions. The thin lines denote isentropic surfaces. Parcels rise diabatically in the tropics (indicated by A in Fig. 1(a)), crossing isentropic surfaces. This ascent is not zonally uniform in either space or time. It is concentrated in the regions of deep convection where strong updraughts transport water into the upper troposphere, with detrainment at the tops of the convection moistening the surrounding air. Overshooting cumulonimbus clouds can penetrate into the stratosphere, as illustrated in Fig. 1(a). Newell and Gould-Stewart (1981) called the convective regions where tropospheric air enters the stratosphere and is freeze-dried 'stratospheric fountains'. These include the Indonesian region between October and March, and the region of the Asian summer monsoon between May and September. The return flow from the stratosphere into the troposphere takes place in the extratropics (indicated by B in Fig. 1(a)), for example as a large-scale subsidence or in tropopause folds behind fronts.

Broadly speaking, the Brewer–Dobson circulation is a large-scale overturning circulation associated with diabatic heating and cooling, leading to a systematic crossing of isentropic surfaces by air parcels. Exchange between the troposphere and stratosphere potentially can also occur adiabatically, along isentropic surfaces, notably in the subtropics, where isentropes intersect the tropopause (indicated by C in Fig. 1(a)). This region is characterized by a strong potential-vorticity gradient, marking the transition zone between the troposphere and stratosphere. This gradient normally inhibits isentropic exchange between the two regions because it acts as a barrier to meridional displacements of air parcels (Juckes and McIntyre 1987). Exchange can only occur, as our paper will show, if this barrier is sufficiently distorted, e.g. by growing upper-tropospheric cyclones (see also Holton et al. 1995).

Some authors have suggested that upper-level monsoon anticyclones may play a part in isentropic exchange between the troposphere and stratosphere in the subtropics. This possibility is favoured by their location close to the potential-vorticity barrier. Figure 1(b) is a schematic illustration of the atmospheric structure (it is based on an actual cross-section, to be discussed later). It shows a convective tower in the subtropics which
Figure 1. (a) Schematic latitude–height cross-section showing the basic ideas of stratosphere–troposphere exchange. The thick line marks the tropopause, the stratosphere is lightly shaded, isentropes are represented by the thin lines, and the arrows indicate transport. (b) Schematic cross-section through the Asian monsoon anticyclone, isolines etc. as in (a), crosses depict easterly and dots westerly flow.
transports water vapour up to the sloping tropopause. The upper portion of this tower is
surrounded by an anticyclonic circulation at levels where isentropic surfaces intersect the
tropopause. Dunkerton (1995) argued that the circulation associated with a steady anticy-
cclone cannot by itself lead to stratosphere–troposphere exchange. A transient motion field
and irreversible mixing are required. He suggested, though he did not show examples, that
the transience could arise by the following means: changes in the intensity of the anti-
cclone, eastward propagating midlatitude synoptic disturbances, and westward propa-
gating tropical disturbances.

Chen (1995) presented evidence for transport from the stratosphere to the troposphere
associated with monsoon anticyclones. He showed an example of a tracer being drawn out
of the stratosphere into a filament that extended around the Asian monsoon anticyclone into
the troposphere. He also suggested that transport from the troposphere to the stratosphere
occurred on the west side of monsoon anticyclones, but presented no direct evidence for
this.

Although convincing evidence of transport from stratosphere to troposphere by mon-
soon anticyclones has been shown, specific examples of the converse process, troposphere
to stratosphere transport, have not been documented. Such transport could prove to be an
important path along which water vapour and other tracers of tropospheric origin could
enter the lower stratosphere and affect its chemical and radiative balance.

In this study the region of the Asian summer monsoon is identified as an important
moisture source for the northern hemisphere upper troposphere and lower stratosphere, and
a mechanism by which the Asian summer monsoon plays a role in moistening the extra-
tropical lower stratosphere is presented. This mechanism is based on interactions between
the upper-tropospheric monsoon anticyclone and midlatitude synoptic-scale tropospheric
cyclones. These systems draw moist air from the monsoon anticyclone, carrying it along
isentropic surfaces into the extratropical lower stratosphere, where it mixes irreversibly
with drier stratospheric air.

The re-analysis data set constructed by the European Centre for Medium-Range
Weather Forecasts (ECMWF) and the contour advection technique are used to identify
the mechanism for moistening; they are described in section 2. Section 3 presents back-
ground information about the Asian summer monsoon. The mechanism for moistening
is documented in section 4, and in section 5 the independent MOZAIC (Measurement of
OZone by Airbus In-service airCraft) aircraft data are used to validate the results obtained
from the ECMWF data. Conclusions are presented in section 6.

2. DESCRIPTION OF DATASETS AND TOOLS

(a) ECMWF re-analysis data

This study employs ECMWF re-analysis (ERA) data to investigate the upper-
tropospheric humidity distribution in the region of the Asian summer monsoon. The dataset
extends over fifteen years (1979–93), and was generated by assimilating data into a single
version of the ECMWF spectral model. The horizontal truncation of the model is T106
(equivalent to about 1.125° × 1.125°), and meteorological fields, including specific hu-
midity, were produced on 31 sigma levels in the vertical (Gibson et al. 1997). In this study,
fields are employed that have been interpolated from this dataset linearly in ln(p), where
p is pressure, onto pressure levels or isentropic surfaces.

Fields of water vapour for the ERA data were derived from radiosonde measurements
and from cloud-cleared radiances measured by the Tiros Operational Vertical Sounder
(TOVs). Radiosonde measurements of water vapour were used up to 275 hPa in the
assimilating model; above this level, the reliability was thought to be inadequate. Over sea,
humidity information from TOVS cloud-cleared radiances was incorporated at levels below 300 hPa (the weighting functions for humidity channels have their peaks below this level). Over land, only radiosonde information was used. An advantage of the assimilation system was that it used TOVS radiances from humidity and temperature channels directly, by using a one-dimensional variational (1D-VAR) analysis scheme (Gibson et al. 1997). The use of this technique led to improved results in both the horizontal and vertical distribution of specific humidity (McNally and Vesperini 1996). This improvement was largest in the upper troposphere.

In the stratosphere, because of the absence of reliable observational data, humidities in the ERA data were set to a climatological value of 2.5 mg kg⁻¹, or to the local saturation value if that was lower (Simmons et al. 1999). The temperature profile was used to determine which levels were stratospheric. Through advection by the model, errors introduced in this manner could affect humidity values at lower levels. Ovarlez and van Velthoven (1997) compared specific humidity from operational ECMWF analyses (for 1994 and 1995) with aircraft measurements from the POLINAT (Pollution from Aircraft Emissions in the North Atlantic flight corridor) experiment. They found that the fixing of the stratospheric humidity did not allow a realistic mixing-in of moist tropospheric air at the bottom of the stratosphere, so that the ECMWF analyses were drier than aircraft data at low humidities and in the stratosphere. Nevertheless, the model simulated well the contrast between moist tropospheric and dry stratospheric air at the tropopause, and was able to reproduce many of the small-scale structures seen in the observations.

A comparison of specific humidities from radiosonde ascents with those from profiles at grid points in the ERA data in the region of east Asia showed large discrepancies above 150 hPa, with ERA values being about an order of magnitude lower. Analysis of the temporal variability of upper-tropospheric humidity (not shown) also indicated that the imposition of a fixed humidity in the stratosphere affected the patterns of variability above about 150 hPa. Below this level the patterns were consistent throughout the troposphere. The region of the upper troposphere and lower stratosphere studied here, around 200 hPa, appears to be below the levels most strongly affected by the specification of stratospheric humidity, but above the levels where observational data directly influenced the analyses. It will be shown later that humidity fields at the levels of interest in this study are vertically consistent, because assimilating observations into a state-of-the-art forecast model also generates meteorologically realistic flow patterns above the levels of observational input. Studies with an independent dataset (section 5) will validate features seen in the ERA data.

(b) Contour advection

Other authors have shown that isentropic stratosphere–troposphere exchange can involve the development of filamentary intrusions (e.g. Yang and Pierrehumbert 1994). Chen (1995) also showed that filaments of stratospheric tracers were advected around the Asian monsoon anticyclone into the tropical troposphere. Contour advection is a technique which allows the study of the formation of small-scale structures in passive tracer fields in 2D isentropic flow (e.g. Norton 1994; Waugh and Plumb 1994). We employed this technique to study isentropic stratosphere–troposphere exchange in the vicinity of the monsoon region. The contour advection code used in this study was developed by Dritschel (1989) and modified by Norton (1994).

The code was used to study tracer transport on the 340 and 350 K isentropic surfaces, which lie in the upper troposphere at low latitudes and in the lower stratosphere at high latitudes. Material contours of a passive tracer were initialized on an isentropic surface on a 3.75° longitude by 2.5° latitude grid, and isentropic winds from the ERA data were used to advect the contours in time. Winds were updated every 6 hours, with linear interpolation
between the wind records. Because additional particles were inserted along contours in regions of high curvature, high accuracy of the contours was maintained. This allowed the investigation of small-scale features typical of chaotic transport, as the contours were, for example, stretched into long fine-scale filaments.

Potential-vorticity and specific-humidity fields were used as initial tracer fields, which were then advected as passive tracers. In the real atmosphere, potential vorticity is not necessarily conserved because of diabatic processes and the effects of unresolved small-scale processes. However, as will be shown, the flow associated with the features of interest here is approximately isentropic away from the monsoon region where the diabatic heating is strong. Specific humidity is not materially conserved on isentropic surfaces if saturation is reached. Condensation is a significant limitation on the use of water vapour as a passive tracer. However, in this study, advected water vapour fields are used only qualitatively to identify the transport processes that moisten the lower stratosphere. A quantitative analysis would require a calculation that explicitly included sources and sinks of water vapour.

3. ASIAN SUMMER MONSOON: BACKGROUND

The Asian summer monsoon circulation is basically thermally driven. Differential heating between the hot land masses and the colder oceans is responsible for the formation of a shallow heat-low over northern India, the Tibetan plateau and especially central Asia during northern summer. The inflow into these areas converges, and the resulting convection and latent-heat release enhance the thermal contrast. The climatological low-level wind field for July over India is shown in Fig. 2(a). A westerly flow brings moist air into the Indian sub-continent. The winds converge over the Bay of Bengal and Bangladesh, where the main areas of monsoon convection, precipitation (contours in Fig. 2(a)), and latent-heat release can be found.

The upper-level flow (Fig. 2(b)) is dominated by the monsoon anticyclone that is located to the north-west of the areas of latent heating. It builds as a Gill-type Rossby-wave response (Gill 1980). The climatological monsoon anticyclone is a very elongated feature, extending from Iran over northern India and the Tibetan plateau into China. It has a deep vertical structure and extends into the lower stratosphere (Dunkerton 1995). The anticyclone is flanked by the equatorial easterly jet to the south and the subtropical westerly jet to the north. Immediately to the north of the anticyclone lies a zone where the gradient of potential vorticity is strong, corresponding to the transition zone from tropospheric to stratospheric air. (The heavy line in Fig. 2(b) is an isopleth of potential vorticity in this zone, given in PVU, potential-vorticity units; 1 PVU = 10^{-6} m^{-2} K s^{-1} kg^{-1}). This strong gradient acts as a dynamical barrier to quasi-horizontal transport between the stratosphere and the troposphere (e.g. Juckes and McIntyre 1987).

Figure 3 shows the distribution of specific humidity at 200 hPa for July, averaged over the fifteen years of the ERA data set. Values are high in the tropics along the intertropical convergence zone (ITCZ), where deep convection moistens the upper troposphere (e.g. Soden and Fu 1995). Maximum values are, however, not found along the ITCZ but away from the equator in the subtropics in association with convection in the Asian summer monsoon. Monsoon convection moistens the air in the upper troposphere close to the sloping tropopause and the associated potential-vorticity barrier between the troposphere and stratosphere. The region of high humidity correlates roughly with the location of the upper-level monsoon anticyclone. The elongated shape of the humidity maximum to the north-east and, to a lesser extent, to the south-west, suggests that, in the upper troposphere, moisture is exported from the monsoon region by the subtropical westerly and tropical easterly jets. This will be addressed later.
Figure 2. (a) Climatological (1979–93) 850 hPa wind field for July from the ERA data (see text) and climatological precipitation for July from the Xie–Arkin precipitation data set (Arkin and Xie 1994; Xie and Arkin 1997). Contour interval for precipitation is 2 mm day\(^{-1}\) and values in excess of 2 mm day\(^{-1}\) are shaded. Winds are shown as vectors, the reference vector is 10 m s\(^{-1}\). (b) Climatological (1979–93) 200 hPa wind field for July from the ERA data, the reference vector is 20 m s\(^{-1}\). The thick line shows the location of the mean 3 PVU (see text) potential-vorticity contour, an isopleth in the transition zone from tropospheric to stratospheric air that marks the dynamical tropopause.
As noted earlier, freeze-drying of air ascending into the stratosphere above regions of intense convection in the tropics has been the favoured explanation of the dryness of the stratosphere. Some authors have also proposed that a similar freeze-drying mechanism operates above strong convection associated with the Asian summer monsoon (Johnston and Solomon 1979; Newell and Gould-Stewart 1981; Mote et al. 1996). However, the atmospheric structure is quite different near the monsoon convection compared with convective regions in the deep tropics. A cross-section through the monsoon anticyclone for July is shown in Fig. 4, which served as a basis for the schematic in Fig. 1(b). The ERA data were averaged over fifteen years (1979–93) and over the longitudinal extent of the anticyclone (40°–100°E) to give a representative picture. The northern flank of the anticyclone is marked by the subtropical westerly jet (arrow A, Fig. 4), and the southern flank by the tropical easterly jet (arrow B, Fig. 4). The zonal wind contours show that the anticyclone extends upwards into the lower stratosphere, as previously noted by Dunkerton (1995).

The heavily shaded band in Fig. 4 is the region of the atmosphere where values of potential vorticity increase rapidly from characteristically tropospheric to stratospheric values; it serves to define a dynamical tropopause, which coincides with the temperature tropopause (at about 380 K) at low latitudes. For studies of exchange between the stratosphere and troposphere in midlatitudes, it is natural to focus on the dynamical tropopause because this structure is relevant to the meridional displacement of air parcels.

The tropopause is higher in the tropics than in the extratropics. It slopes downward with increasing latitude in the westerly jet at the northern flank of the anticyclone (arrow A, Fig. 4). Isentropic surfaces (quasi-horizontal thin lines in Fig. 4) intersect this sloping tropopause in the westerly jet. This configuration opens the possibility that exchange of air between the troposphere and stratosphere might take place in this region through nearly horizontal, adiabatic motion, without air being freeze-dried by ascent through regions of very low temperatures. Monsoon convection moistens the upper troposphere close to the sloping tropopause, and Fig. 4 illustrates the close proximity of moist tropospheric air within the anticyclone to dry stratospheric air to the north of it. (The elevated tropopause
above the monsoon region, marked by arrow C, was noted by Highwood and Hoskins (1998) and is due to diabatic heating in the monsoon anticyclone. The monsoon anticyclone is thus a potential moisture source for the extratropical lower stratosphere, and transport from the troposphere into the stratosphere along quasi-horizontal isentropes can lead, as we will show, to a moistening of the lower stratosphere.

Additional observational evidence that moist air from the troposphere enters the stratosphere in the subtropics along isentropic surfaces was obtained by Dessler et al. (1995). They detected air in the extratropical lower stratosphere (below the 380 K isentropic surface) with moisture values that were too high to be consistent with ascent of air through the cold tropical tropopause. They suggested that the observed values could be explained by isentropic transport of air from the troposphere to the stratosphere across the subtropical tropopause (pathway C in Fig. 1(a)).

Further observational support for a moistening of the lower stratosphere by the Asian summer monsoon was presented by Jackson et al. (1998), who used data from the HALOE (Halogen Occultation Experiment) instrument on board the Upper Atmosphere Research Satellite. They found humidity maxima in the lower stratosphere in the region of the Asian summer monsoon which they linked to the strong monsoon convection. Harries (1997) showed that a moistening of the lower stratosphere at the latitudes of the Asian summer monsoon occurred every summer, and that there was no equivalent moistening in the lower stratosphere of the southern hemisphere. Consistent with these findings, Rosenlof
et al. (1997) and Pan et al. (1997) noted that the extratropical lower stratosphere in the northern hemisphere was moister than that in the southern hemisphere. They proposed that the interhemispheric asymmetry in summer could be partly accounted for by moisture transport during the Asian summer monsoon, though they did not propose a mechanism for this transport.

4. THE ROLE OF THE MONSOON IN MOISTENING THE SUBTROPICAL UPPER TROPOSPHERE AND EXTRATROPICAL LOWER STRATOSPHERE

In this section the distributions of moisture, wind and potential vorticity associated with the monsoon anticyclone in the upper troposphere are described, considering individual days to point out some specific features of the transport of water vapour in the upper troposphere. Subsection (a) describes the role of the monsoon in moistening the upper troposphere in the subtropics near the dynamical tropopause. Subsection (b) presents a mechanism for the quasi-horizontal transport of moisture from the troposphere to the stratosphere, and subsection (c) looks at the seasonal evolution of the moisture transport into the lower stratosphere of the northern hemisphere.

(a) The moistening of the subtropical upper troposphere

Trajectory studies using a program developed by Methven (1997) were carried out to investigate the upper-level circulation and the transport associated with the Asian summer monsoon. A cluster of particles was initialized on 1 July 1988 over India (every 2°, in a box 70°–100°E, 20°–40°N) on the 360 K isentropic surface, and winds from the ERA data were used to calculate forward trajectories for the eight subsequent days. The results of the trajectory study are presented in Fig. 5. For clarity, only every tenth trajectory is plotted. The trajectory study confirms the existence of a barrier to northward meridional displacements.

![Figure 5. Trajectories (3-dimensional) of particles initialized on the 360 K surface over India (in the marked box) on 1 July 1988 and integrated forward for 8 days.](image-url)
in the region between 40° and 120°E, which coincides with the potential-vorticity gradient associated with the dynamical tropopause. Under quasi-steady conditions, transport is around the anticyclone, rather than to the north, in agreement with the findings of Dunkerton (1995).

A small anticyclonic circulation is centred over Iran in which some of the particles are caught. Synoptic charts show that the monsoon anticyclone often splits into two parts, with one centre over Iran and the other over south China, as for example in this case (Fig. 5) on 3 July 1988. Other studies (He et al. 1987; Dunkerton 1995) showed that these split anticyclones, usually having one centre over Iran and another over Tibet, are quite common features. As a result, humidity that is transported into the upper troposphere in the main areas of monsoon convection (e.g. the Bay of Bengal) can be redistributed in the upper troposphere, and such moisture transport can lead to humidity maxima remote from the source regions.

Such upper-level transport processes can explain the upper-tropospheric humidity distribution seen in HALOE data (Jackson et al. 1998, their Figs. 1(c) and 2(c)). These satellite data show a humidity maximum over south Asia in the seasonal mean over June, July and August, which agrees well with the location of the monsoon anticyclone. The largest humidity values, however, are located over Iran and Saudi Arabia, slightly displaced to the north-west of the main areas of monsoon convection, and may be a result of moisture redistribution in the upper troposphere away from the source regions.

Figure 5 shows that usually particles either stay in the region of the upper-level anticyclone or leave it on the north-eastern or south-western sides, i.e. in the subtropical westerly and tropical easterly jets. This fanning out in the regions of divergence is found frequently and leads to the elongated shape of the humidity maximum in Fig. 3. It can also lead to the formation of elongated streamers of moist air that are being drawn from the monsoon anticyclone. An example of such a streamer is presented in Fig. 6, which shows the distributions of specific humidity and winds on the 340 K isentropic surface on 10 July
1988 in the vicinity of the monsoon anticyclone. A long streamer of air with relatively high values of humidity has been drawn out of the monsoon anticyclone and stretches eastward along the subtropical westerly jet. Similar streamers of moist air develop regularly in the upper troposphere during the monsoon season and can extend far over the Pacific Ocean, contributing to the climatological eastward extension of the moisture maximum shown in Fig. 3. Through the formation of such streamers the monsoon anticyclone serves as a moisture source for much of the subtropical upper troposphere near the dynamical tropopause.

(b) The moistening of the extratropical lower stratosphere

As noted earlier, the strong potential-vorticity gradients associated with the subtropical westerly jet normally prevent northward transport out of the monsoon anticyclone along sloping isentropes. However, the monsoon anticyclone is a very deep structure and, being located next to the steep dynamical tropopause (Fig. 4), it seems likely to have some potential for troposphere to stratosphere transport if a distortion of the dynamical tropopause and irreversible mixing occurs.

The key result of this research suggests that from time to time midlatitude synoptic-scale tropospheric cyclones, which move eastwards north of the Tibetan plateau, interact with the monsoon anticyclone during northern summer. During these interactions, filaments of air that are characterized by high values of humidity and low values of potential vorticity are drawn off the monsoon anticyclone. These filaments are transported into the extratropics where they are eventually mixed irreversibly with the surrounding stratospheric air. Although this study focuses on transport of moisture into the stratosphere in the vicinity of the monsoon anticyclone, such events do not occur exclusively in this region. A similar mechanism for the transport of moisture from the troposphere to the stratosphere operates along the storm tracks at other locations near the dynamical tropopause, although the local moisture concentrations are typically not as high there as they are near the monsoon anticyclone.

The mechanism for moistening the lower stratosphere involves an irreversible buckling of the dynamical tropopause by cyclones that extend through the troposphere into the lower stratosphere. This buckling leads to a two-way exchange of air: moist filaments are transported north out of the monsoon anticyclone and mixed irreversibly with stratospheric air; dry, high-potential-vorticity air is subsequently mixed into the troposphere around the eastern side of the anticyclone. We illustrate the mechanism by presenting an example of an event during July 1988 when an eastward travelling tropospheric cyclone led to the irreversible transport of moisture from the monsoon anticyclone into the lower stratosphere. Later in the paper another example is presented (in less detail) for July 1995, for which MOZAIC data are available for the validation of the results.

The synoptic development of the July 1988 event is depicted in Fig. 7, which shows the evolution of the geopotential height field at 200 hPa from 22 to 30 July 1988. On 22 July, the geopotential height field shows an elongated monsoon anticyclone, the subtropical westerly jet to the north, and a cyclone centred at about 70°E to the north of the subtropical jet. During the following week, this cyclone moves eastwards, and an associated ridge develops downstream, involving a strong distortion and elongation of the monsoon anticyclone (26 July). The monsoon anticyclone eventually splits (30 July).

The transport associated with the deformation of the monsoon anticyclone by the eastward travelling cyclone can be inferred from Fig. 8, which shows the humidity field on 26 July 1988 on the 350 K isentropic surface, and also the corresponding potential-vorticity field. Moist air is being advected northwards in the geopotential ridge ahead of the cyclone. The moisture streamer extends far into the extratropics on an isentropic surface
Figure 7. Geopotential height in decametres (dm) of the 200 hPa surface on (a) 22, (b) 26, and (c) 30 July 1988 from the ERA data (see text). Contours are drawn every 16 dm and values greater than 1252 dm are shaded.
which, at these latitudes, lies mainly in the stratosphere. Contour advection studies will later show that irreversible mixing of moist air occurs in the stratosphere during this event, so that at least some of the moisture is transported irreversibly out of the troposphere. The extratropical stratosphere is close to radiative equilibrium in summer, so that the vertical transport of moisture entering into that region would be expected to be small.

The potential-vorticity field on the 350 K surface (Fig. 8) on 26 July closely resembles the humidity field. A filament of tropospheric air (potential vorticity less than 2 PVU) extends into the extratropical lower stratosphere. The dynamical tropopause is in the pro-
cess of being irreversibly distorted, as can be seen from the deformation of the region of strong horizontal potential-vorticity gradients. Downstream of this low-potential-vorticity filament, a filament of high-potential-vorticity air of stratospheric origin is being advected southwards into the subtropics along the eastern flank of the monsoon anticyclone. This feature coincides with a dry filament in the humidity picture, supporting the notion that this air is of stratospheric origin.

The tropospheric cyclones, which deform the dynamical tropopause, extend through the depth of the troposphere into the stratosphere, where they fade with increasing height (Fig. 9). They are dynamically active features in the so-called 'middleworld' (Hoskins 1991), where isentropic surfaces intersect the sloping tropopause. Figure 9 is a latitude–height cross-section of zonal wind, potential temperature, and potential vorticity through the monsoon anticyclone and the eastward travelling cyclone. Arrow A marks the subtropical westerly jet lying between the monsoon anticyclone and the eastward travelling cyclone. Arrow B marks the local easterly jet north of the cyclone. The jet structure shows that the cyclone extends well above 100 hPa into the lower stratosphere. Westward advection of low-potential-vorticity air on the northern flank of the cyclone (see Fig. 8) leads to a raised tropopause at B in Fig. 9 near 60°N, so that air of tropospheric origin can be found at high latitudes at altitudes normally occupied by stratospheric air.

To test the hypothesis that the filament of high humidity/low-potential-vorticity air extending into the stratosphere, as depicted in Fig. 8, is largely the result of quasi-isentropic, horizontal advection out of the monsoon region, we calculated 3D air parcel trajectories. These trajectories are not isentropic. They were computed using 3D velocity fields from the ERA data. The ECMWF data analysis system has a representation of atmospheric diabatic processes, and the velocity fields will be affected by such processes. To determine the origin of the moist filament shown in Fig. 8, we calculated backward trajectories from

Figure 9. Latitude–height cross section along 90°E on 26 July 1988. Details as in Fig. 4, but arrow A marks the subtropical westerly jet, and arrow B the local extratropical easterly jet associated with the cyclone.
the filament as follows. Backward trajectories were calculated for all parcels arriving on 26 July on the grid points of a 1° by 1° grid on the 350 K isentropic surface in box A (Fig. 10(a)). The trajectories plotted in Fig. 10 are those that had humidity and potential-vorticity values characteristic of the filament (humidity greater than 0.05 g kg⁻¹, and potential-vorticity values between 0 and 1.5 PVU). Those trajectories marked by solid lines are more or less isentropic, whereas the dashed ones originated within the monsoon anticyclone and rose diabatically (Fig. 10(b)).

The trajectories confirm that the majority of particles in the filament were pulled off the northern flank of the monsoon anticyclone along isentropic surfaces (marked by B). These calculations imply that the moist filament that extended into the stratosphere was formed largely by quasi-isentropic transport of air from the monsoon anticyclone, but that humidity values were augmented to some extent by cross-isentropic convection.

To investigate further the isentropic aspect of the transport during the interaction and to examine smaller-scale features, the contour advection technique was used. This technique gives a revealing fluid-dynamical picture of the isentropic exchange of air between the troposphere and the stratosphere in the vicinity of the dynamical tropopause. In the following, we shall focus on stratosphere–troposphere exchange in the vicinity of the Asian summer monsoon, but other examples of exchange between the stratosphere and the troposphere near the dynamical tropopause will also be noted.

Figure 11(a) to (c) shows potential-vorticity fields, and Fig. 11(d) to (f) specific-humidity fields, for 24, 26 and 28 July 1988, on the 350 K isentropic surface. Contours for potential vorticity are: 1–2 PVU (dark blue), 2–3 PVU (light blue), 3–4 PVU (green), 4–5 PVU (orange), greater than 5 PVU (red). Contours for specific humidity are 0.025–0.05 g kg⁻¹ (dark blue), 0.05–0.075 g kg⁻¹ (light blue), 0.075–0.1 g kg⁻¹ (green), 0.1–0.5 g kg⁻¹ (orange), 0.5–1 g kg⁻¹ (red), 1–5 g kg⁻¹ (brown). The unshaded areas in (d) to (f), therefore, mark very low humidities, including the stratospheric areas with fixed climatological humidities of 2.5 mg kg⁻¹.

The potential-vorticity field on the 350 K isentropic surface on 24 July 1988 from the ERA data was used as an initial tracer distribution (Fig. 11(a)). The contour advection technique was employed to advect isentropically a set of isolines of potential vorticity as material lines, by using winds from the ERA data. Results of a 5-day integration are shown in Fig. 11(a) to (c) (90°E is at the bottom of these figures). The initial potential-vorticity field on 24 July 1988 (day 0) identifies the dynamical tropopause as a region of strong potential-vorticity gradients (rapid colour change) at the outer boundary of a region of potential vorticity with high values, characteristic of stratospheric air (red contours are values greater than 5 PVU). The sequence of figures shows that the dynamical tropopause is strongly distorted by synoptic systems. We note in particular the filament of low potential vorticity on 26 July (day 2) near 115°E (Fig. 11(b)), which extends from the north-eastern part of the Asian monsoon region into high latitudes. The filament subsequently rolls up anticyclonically (day 4, Fig. 11(c)) to form a cut-off region of air (north of 60°N at 120°E) with low values of potential vorticity, characteristic of the troposphere, surrounded by air with high values of potential vorticity, characteristic of the stratosphere.

During the same period, stratospheric air with high values of potential vorticity (red colours) is being advected southward along the east Asian coast (Fig. 11(b)). This filament subsequently rolls up into a cyclonic vortex over south China (Fig. 11(c)) to form a region of air of stratospheric origin surrounded by tropospheric air. This illustrates the two-way nature of the interaction. Another example of apparent exchange between the troposphere and stratosphere, distant from the Asian monsoon region, can be seen on day 4 (Fig. 11(c)) near 90°W. Here tropospheric and stratospheric air are being mixed in a developing baroclinic system.
Figure 10. (a) 3-dimensional backward trajectories of parcels initialized on the 350 K surface on 26 July 1988 in box A. The particles are followed back in time for seven days to 20 July 1988. Depicted are all those trajectories that have initial specific humidity greater than 0.05 g kg\(^{-1}\) and initial potential vorticity between 0 and 1.5 PVU (see text) in box A, i.e. those that contribute to the filament in Fig. 8. Dashed lines depict particles that rise adiabatically in the monsoon anticyclone. (b) Potential temperature in K along the trajectories (40 time steps per day, time step 0 is the initial time in box A).
Figure 11. (a) to (c) Potential-vorticity fields on 24, 26 and 28 July 1988, respectively, from contour advection calculations, initialized using the contours in (a); and (d) to (f) specific-humidity fields on 24, 26 and 28 July 1988, respectively, initialized using the contours in (d). 90°E is at the bottom of the figures. See text for key and discussion.
The contour advection calculations indicate that the effect of irreversibly deforming the dynamical tropopause by synoptic systems is to sharpen the potential-vorticity gradients that define the dynamical tropopause. Locally, colour contrasts in Fig. 11(c) are somewhat sharper than in the initial low-resolution field in Fig. 11(a). Since such mixing events must be continually occurring in association with synoptic-scale system in the troposphere, this sharpening must be offset by other processes, such as diabatic heating and cooling.

The previous calculations were repeated with specific humidity on the 350 K isentropic surface as the initial tracer field for 24 July 1988. These results are shown in Fig. 11(d) to (f). The initial specific-humidity field on 24 July (day 0) shows the largest values (red and brown contours) in the tropical troposphere associated with the Asian summer monsoon (Fig. 11(d)). Unshaded regions mark relatively dry air. Corresponding to the potential-vorticity feature near 115°E on day 2 (in Fig. 11(b)) is an intrusion of very moist air northward into the dry air of the stratosphere (Fig. 11(e)). This leads to a moisture input into the stratosphere around 120°E north of 60°N on day 4 (Fig. 11(f)).

The eastward advection of moist air out of the monsoon region along the subtropical westerly jet (previously referred to in Fig. 6) is also evident in Fig. 11(d) to (f). An elongated streamer of moist air is being drawn from the monsoon region and advected eastwards over the Pacific. Considerable filamentation occurs here, as the moist air is being mixed with the drier local air. Notice that moisture from the Asian monsoon also spreads westwards in the tropical easterly jet. This air is either transported around the anticyclone (red contours in Fig. 11(f)), exported into the western hemisphere (orange contours in Fig. 11(e) and (f)), or transported across the equator into the southern hemisphere over Africa (day 5 of the calculations, not shown). The results from the contour advection calculations hence confirm the importance of the Asian summer monsoon region as a moisture source for much of the subtropics.

It has to be kept in mind that these contour advection calculations do not give an accurate quantitative picture of the isentropic water vapour transport, because they do not take account of water vapour sources and sinks, such as condensation and small-scale vertical mixing. We also note that our calculations do not include a moisture source from convection in the monsoon region, and therefore moisture removed from this region by advection is not replenished. Furthermore, water vapour gradients will be incorrect near the tropopause because of the fixed humidity values in the stratosphere. This can lead to errors in the details of the filament structure, but does not alter the overall idea of moistening by rapid, small-scale irreversible transport. Nevertheless, the contour advection calculations give a revealing, qualitative picture of the transport of air out of the Asian summer monsoon region, which can serve as a basis for more detailed quantitative studies.

The calculations capture the following features: (1) the anticyclonic circulation of moist air in the region of the upper-level anticyclone, (2) the eastward export of moist air in the subtropical westerly jet from the monsoon region in form of streamers, (3) the export of moist air to the west by the tropical easterly jet, (4) the moistening of the lower stratosphere to the north of the monsoon region by interaction with synoptic disturbances, transport across the dynamical tropopause, and irreversible mixing into the lower stratosphere, and (5) the southward transport and mixing of dry stratospheric air into the subtropical troposphere.

(c) Seasonal evolution of moisture transport into the lower stratosphere of the northern hemisphere

We have noted that the Asian summer monsoon can serve as an upper-tropospheric moisture source not only for the Asian region but for much of the subtropical zone that lies near the dynamical tropopause. Other monsoon circulations (e.g. the Mexican monsoon)
are likely to contribute in a similar manner, though these other systems are less intense and their upper-tropospheric moisture source is therefore weaker (see Fig. 3).

We have used the whole ERA dataset from 1979 to 1993 to examine the seasonality of events that transport moist air northwards into the extratropics in the region of the middleworld. Results are shown for the year 1988 only; other years exhibit similar seasonality, though the strength of the monsoon as a moisture source for the upper troposphere varies from year to year. Figure 12 shows the seasonal evolution of specific humidity between 0 and 180°E at a notional latitude of 50°N on the 345 K isentropic surface in the middleworld. Humidity values at this notional latitude were computed as meridional averages of humidity values in the latitude band between 45 and 60°N. The figure shows the strong seasonality of the moisture input into the upper troposphere and lower stratosphere. Moisture values are largest at the longitudinal position of the Asian summer monsoon. High values occur during the monsoon season from July to October and are absent outside that season. During the first half of the monsoon season, the monsoon anticyclone migrates from its position over south-east Asia to its established location over northern India (Ju and Slingo 1995), and the area of strongest convection also migrates north-westwards towards
the Bay of Bengal (Lau and Yang 1996). This movement is reflected in Fig. 12 by the westward movement of high moisture values during June and July. During the second half of the monsoon season, the area of strongest convection retreats south-eastwards towards Indonesia, and this movement is reflected by the eastward movement of high moisture values between August and October.

5. INDEPENDENT VALIDATION USING MOZAIC DATA

The results obtained from the ERA data, trajectory studies and contour advection calculations give a consistent fluid-dynamical picture of water vapour transport in the vicinity of the Asian summer monsoon. Nevertheless, we are concerned about the lack of observational input into the upper troposphere in the ERA-assimilating model. Unfortunately, we do not have an independent dataset to validate the findings obtained from the ERA data. However, MOZAIC data have recently become available and allow comparisons with later operational ECMWF analyses. We compare ECMWF operational analyses and MOZAIC data for an event in July 1995 that followed a similar pattern to that described in section 4(b). The daily operational analyses between April 1995 and February 1996 used the same assimilating model as the ERA data (Gibson et al. 1997), although the analyses were performed at a higher horizontal resolution (triangular truncation T213). In this study we use the operational analyses for 1995 diagnosed at the same horizontal resolution as the ERA data (see section 2(a)), so that validation of these later ECMWF analyses will give credibility to the findings obtained from the ERA data.

(a) Datasets

(i) MOZAIC data. The MOZAIC project is a collaboration of European scientists, aircraft manufacturers and airlines with the aim of measuring ozone and water vapour in the atmosphere (Marenco et al. 1998). Five Airbus 340 commercial aircraft were equipped with sensors to measure water vapour and ozone during routine flights. The project began in January 1993, and aircraft data for scientific evaluation have been available from September 1994 onwards. Most of the measurements are taken at cruise level (9–12 km, corresponding to about 250–150 hPa), the altitude range of interest in this study.

The MOZAIC water vapour sensor is described by Helten et al. (1998). This sensor has uncertainties of ±7% in relative humidity between 9 and 12 km. The response time of the sensor is 1–3 minutes at 10–12 km, corresponding to a horizontal resolution of between 15 and 50 km at aircraft speeds of about 250 m s⁻¹. At high altitudes, in the stratosphere, the relative errors are large.

The ozone sensor as described by Thourret et al. (1998) has an overall precision of ±2 parts per billion by volume (p.p.b.v.) (i.e. ±2% of typical upper-tropospheric ozone values). Our study does not require high precision of the MOZAIC sensors, but uses their ability to distinguish between tropospheric and stratospheric air.

(ii) Interpolation of ECMWF analyses onto MOZAIC flightpaths. ECMWF operational analyses for July 1995 were interpolated onto the MOZAIC flightpaths for comparison with this independent dataset. Six-hourly data at truncation T106 on 14 pressure levels between 1000 and 50 hPa were interpolated in time and space to the aircrafts' positions. Area weighting was used in the horizontal, whilst the vertical interpolation was linear in ln(p) for relative and specific humidity, and linear in pressure for potential vorticity. Time interpolation was also linear.
Relative humidities (the field that is measured by the MOZAIC humidity device) from MOZAIC and ECMWF data were compared along several flight tracks, to investigate how well the ECMWF data captured the overall humidity distribution and to compare the absolute values. Ozone measurements from the MOZAIC data were used to distinguish between air originating from the troposphere and from the stratosphere. In the free troposphere, ozone concentrations are usually low, with values below 100 p.p.b.v. Concentrations increase in the lower stratosphere, where they are of the order of several hundred p.p.b.v. ECMWF data do not provide ozone concentrations, but fields of potential vorticity can be calculated, and they can also be used to distinguish between air originating from the troposphere and stratosphere (Hoskins et al. 1985). Accordingly, derived values of potential vorticity from the ECMWF analyses were interpolated onto several flightpaths for comparison with measured ozone concentrations.

(b) Results from MOZAIC data

To set the scene, we show results from contour advection calculations for an interaction event in July 1995, when there were MOZAIC data available to validate findings. Potential-vorticity and specific-humidity fields on the 340 K isentropic surface on 19 July 1995 from the ECMWF analyses were used as initial tracer distributions for the contour advection calculations. The results on day 4 (23 July 1995) of a 5-day integration are shown in Fig. 13.

Superimposed is a MOZAIC flight track between Europe and Asia. A filament of low potential vorticity develops (between day 1 and day 3 of the calculations) from the northeastern part of the Asian monsoon region (near 100°E) and extends into high latitudes. The filament subsequently rolls up anticyclonically to form a cut-off region of air (north of 60°N at 90°E) with low values of potential vorticity and locally high values of specific humidity on 23 July 1995 (Fig. 13). This cut-off feature retains its identity for about a week in the potential-vorticity field, before dissipating either through diabatic processes or mixing.

Between 19 and 25 July 1995 several MOZAIC aircraft flew between Asia and western Europe, across the filament of high humidity and low potential vorticity drawn from the region of the Asian monsoon. We show one example (a MOZAIC flight from Vienna to Tokyo on 23 July 1995) in Fig. 14, at a time when the filament was well developed and in the process of rolling up. Figure 14(a) compares relative humidities measured by the aircraft instrument, and relative humidities from the ECMWF data interpolated onto the flight track. Figure 14(b) compares ozone measured by the aircraft sensor and potential vorticity from the ECMWF data interpolated onto the flight track.

The overall agreement between the measurements and the derived ECMWF fields is generally good along the flight track, though the variations of the ECMWF data are smoother than the aircraft measurements, as might be expected. Figures 13 and 14 show that the flight track intersected the rolled up tip of the filament on 23 July between 15 hours and 17 hours. During this period, the aircraft measured high values of relative humidity, detecting the relatively moist air in the filament (Fig. 14(a)). Values are in reasonably good agreement with the interpolated ECMWF data. The aircraft instruments also measured low values of ozone during this time (Fig. 14(b)), coinciding with low values of potential vorticity in the interpolated ECMWF data. Before and after the aircraft detected this filament it was flying in stratospheric air, as evidenced by the low values of relative humidity and the high values of ozone.

These results were supported by similar comparisons along the other relevant MOZAIC flight tracks during the period of interest (not shown). We therefore conclude that the ECMWF operational analyses and also the ERA data can be used with a measure of confidence to study the transport of moisture from the troposphere into the stratosphere.
at altitudes considered in this study, notwithstanding the serious shortcomings of the humidity data at higher altitudes. On this basis, our mechanism for moistening the lower stratosphere has credibility.

6. Conclusions

Water vapour in the upper troposphere and lower stratosphere has an important influence on the climate system (Harries et al. 1996), and plays a crucial role in stratospheric chemistry (Solomon 1988). Therefore it is important to understand the processes which determine its transport and distribution in these regions of the atmosphere, including the processes by which moisture can be exchanged between the troposphere and the lower stratosphere.
Figure 14. Results from MOZAIC (see text) flight from Vienna to Tokyo on 23 July 1995. (a) Relative humidity in per cent as measured by the MOZAIC sensor (solid line), relative humidity from ECMWF data interpolated onto the flightpath (dotted line), and flight level in hPa (thick solid line). Times are UTC. (b) Ozone in parts per billion by volume (ppb) from MOZAIC data (solid line), and potential vorticity in PVU (see text) from ECMWF data interpolated onto the flightpath (dotted line).

In this paper, a mechanism has been identified for moistening the extratropical lower stratosphere in the northern hemisphere in summer. This mechanism involves the transport of moisture from the upper tropospheric moisture source associated with the Asian summer monsoon. Interactions occur between midlatitude tropospheric cyclones and the upper-level monsoon anticyclone, during which filaments of moist tropospheric air are stripped from the anticyclone. These filaments are carried northwards across the dynamical tropopause, which normally acts as a barrier to meridional transport, and some of the moisture is irreversibly mixed into the lower stratosphere.

An example of the mechanism was presented for July 1988, based on ERA data, and interpreted by using trajectory calculations and the technique of contour advection. Similar cases were identified throughout the northern summer months of 1988 and other years. The essential features of the mechanism are illustrated schematically in Fig. 15.
Figure 15. Schematic depicting the interactions between midlatitude synoptic disturbances and the monsoon anticyclone (shaded ellipse). The lines denote the flow or the potential-vorticity gradient. (a) Undisturbed case, (b) approaching synoptic disturbance, (c) interaction, and (d) final stage.

Figure 15(a) shows the unperturbed case; the upper-level monsoon anticyclone is depicted as the shaded ellipse. As a synoptic-scale tropospheric cyclone moves eastwards north of the Tibetan plateau (Fig. 15(b)) an associated downstream ridge develops. In this ridge, air out of the monsoon anticyclone is advected to the north (Fig. 15(c)). This air is eventually pulled quasi-isentropically off the source region and some of it is irreversibly mixed with the surrounding stratospheric air. In the final state (Fig. 15(d)), the upper-level trough often splits the monsoon anticyclone into two parts, and the moisture supply stops locally.

Contour advection calculations showed that the exchange of moisture from the upper troposphere into the lower stratosphere across the dynamical tropopause did not occur exclusively near the region of the Asian summer monsoon. It also occurred at other longitudes, mainly in association with extratropical cyclones that developed along the north Pacific storm track. Some of the moisture for these events was supplied by transport of moist air from the monsoon anticyclone along upper-tropospheric jet streams that lie near the dynamical tropopause. Hence, the role of the Asian monsoon in moistening the upper troposphere and lower stratosphere is not just a local one. To a lesser degree, the Mexican monsoon may serve as a moisture source for the Atlantic storm track.

The southern hemisphere does not have such an intense monsoon system in summer. Therefore, if this mechanism proves to be quantitatively important, it may explain the finding that the extratropical lower stratosphere in the summer hemisphere is moister during northern summer than it is during southern summer (Rosenlof et al. 1997; Pan et al. 1997).
In this study, the mechanism for moistening the lower stratosphere was elucidated by using meteorological and humidity data from the ERA dataset. Since no direct humidity measurements went into the ECMWF assimilating model at the altitudes that are of interest in this study, independent MOZAIC data were used to validate the findings. These comparisons showed a good overall agreement between MOZAIC and ECMWF data, though the ECMWF data were spatially smoother. Instances were shown of flight tracks that intersected filaments of moist, low-potential-vorticity air revealed by the ECMWF data. Fluctuations of relative humidity and ozone measured along MOZAIC flight tracks supported inferences made from the ECMWF data.

Further research is required to determine whether this mechanism for moistening the extratropical lower stratosphere in northern summer can account quantitatively for the observed seasonal evolution of moisture values in that region, for interhemispheric differences, and for year-to-year variability. Although the ERA dataset suffices to reveal the qualitative features of the mechanism, an accurate, quantitative analysis requires higher-fidelity humidity fields in the upper troposphere and lower stratosphere than are currently available, as well as an explicit calculation of sources and sinks of water vapour along air parcel trajectories. Current plans for a 40-year re-analysis by ECMWF include an increased emphasis on a more accurate representation of water vapour in the upper troposphere and lower stratosphere. This dataset should allow the importance of the transport processes highlighted in this paper to be quantified.

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