МІНІСТЕРСТВО ОСВІТИ І НАУКИ УКРАЇНИ
ОДЕСЬКИЙ ДЕРЖАВНИЙ ЕКОЛОГІЧНИЙ УНІВЕРСИТЕТ

Методичні вказівки
до виконання контрольної роботи
З АНГЛІЙСЬКОЇ МОВИ
для студентів ІІІ курсу
заочної форми навчання
Напрям підготовки – гідрометеорологія
Спеціальність - метеорологія

"Затверджено"
на засіданні робочої групи методичної
ради "Заочна та післядипломна освіта"

ОДЕСА – 2004

Методичні вказівки до виконання контрольної роботи з англійської мови для студентів III курсу заочної форми навчання.
Напрям підготовки – гідрометеорологія, спеціальність - метеорологія

Укладач: Куделіна О.Ю. Одеса - ОДЕКУ, 2004 р., 29 с.

Передмова

Практичне володіння англійською мовою при заочній формі навчання означає вміння самостійно за допомогою словника читати літературу за фахом англійською мовою, знаходити корисну для роботи інформацію, а також перекладати тексти за фахом рідною мовою.

Метою запропонованих методичних вказівок для самостійної роботи студентів (СРС) та навчального матеріалу з англійської мови для студентів III курсу заочної форми навчання, напрям підготовки – "гідрометеорологія" є:

– виробити у студентів навички читання та перекладу науково-технічної літератури англійською мовою за фахом "метеорологія";

– розвинути вміння розуміти зміст прочитаного;

– виробити навики постановки запитань до тексту англійською мовою;

– підготовити студентів до складання іспиту з англійської мови.

Навчальна програма для студентів III курсу заочної форми навчання розрахована на 144 годин СРС та на 22 години аудиторної роботи.

Програма з дисципліни англійська мова для студентів IIIк. заочної форми навчання

№ п.п	Назва теми заняття	Кількість СРС	Види контролю
1	Особливості перекладу видо-часових форм дієслів (Active Voice). Розмовна тема "About myself". Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	УО
2	Особливості перекладу видо-часових форм дієслів (Passive Voice). Розмовна тема "My future speciality". Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	УО
3	Особливості перекладу модальних дієслів. Розмовна тема "Ukraine". Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	УО
4	Особливості перекладу інфінитива. Розмовна тема "Kiev". Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	УО
5	Особливості перекладу дієприкметників. Розмовна тема "Odessa". Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	УО
6	Особливості перекладу герундія. Розмовна тема "Great Britain". Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	УО

№ п.п	Назва теми заняття	Кількість СРС	Види контролю
7	Особливості перекладу дієслів з післялогами. Особливості перекладу суспільно-політичного тексту. Розмовна тема “London”. Самостійний переклад суспільно-політичного тексту – 5 тис.др.зн.	13	УО
8	Особливості перекладу іменників з прийменниками. Особливості перекладу суспільно-політичного тексту. Розмовна тема “The political system of Ukraine”. Самостійний переклад суспільно-політичного тексту – 5 тис.др.зн.	13	КР№8
9	Особливості перекладу прикметників з прийменниками. Практика у перекладі. Розмовна тема “The political system of Great Britain”. Самостійний переклад суспільно-політичного тексту – 5 тис.др.зн.	13	КР№9
10	Практика у перекладі суспільно-політичного тексту та тексту за фахом. Самостійний переклад текстів за фахом – 5 тис.др.зн.	13	
11	Практика у перекладі суспільно-політичного тексту та тексту за фахом. Самостійний переклад текстів за фахом – 5 тис.др.зн.	14	

Контрольні роботи №8 та №9 передбачають контроль СРС. Кожний з трьох варіантів контрольних робіт №8 та №9 містять 7-8 тисяч друкованих знаків, що є контролем самостійної роботи студентів заочної форми навчання за ІІІ курс. Контрольна робота №10 передбачає письмовий переклад з англійської мови суспільно-політичного тексту на 10 тис.др. знаків за вільним вибором студента. Після виконання контрольних робіт студенти повинні скласти по п'ять спеціальних запитань англійською мовою до змісту тексту, визначити видочасові форми дієслів-присудків, знайти та визначити –ing форми дієслів (Participle I, Gerund). При виконанні контрольних робіт слід користуватися загальними та спеціалізованими перекладними словниками.

Критерії оцінки виконання контрольної роботи:

– **“зараховано”** – студент переклав не менш ніж 80% тексту без суттєвих граматичних помилок, склав запитання до змісту тексту та показав знання основних термінів щодо фаху “метеорологія”, які зустрічалися у контрольних роботах;

– **“незараховано”** – студент переклав менш ніж 80% текстів контрольних робіт, припустив більш 10 граматичних помилок при перекладі текстів та складанні запитань, та не показав знання основних термінів за фахом “метеорологія”, що зустрічалися у контрольних роботах.

КОНТРОЛЬНА РОБОТА №8
ВАРІАНТ №1

I. Зробіть письмовий переклад тексту за фахом:

Text COMPARISON BETWEEN THE STRATOSPHERE OF THE NORTHERN AND SOUTHERN HEMISPHERES

1. Introduction

Recently available data will be used to review differences between the extratropical circulation in the stratosphere of the two hemispheres. The approach will be to consider the stratosphere as an evolving dynamical system, rather than to consider time-averaged properties. Attention will be drawn to distinctive flow regimes in the two hemispheres, and to the extent of their interannual variability.

2. Data

The data used in this study derive from two sources: the UK Meteorological Office (UKMO) and the European Centre for Medium Range Weather Forecasts (ECMWF). The UKMO data are global meteorological analyses of geopotential height, wind and temperature from the surface to 0,3 hPa in the lower mesosphere. These analyses were produced by the technique of data assimilation, as described by Swinbank and O'Neill (1994). The ECMWF data cover the period 1979 to the present. They extend to 10 hPa in the middle stratosphere. The data up until 1993 form the ECMWF Re-Analysis (ERA) data set.

3. Climatology and Seasonal Evolution of the Stratosphere

The circulation in the winter stratosphere is dominated in both hemispheres by a strong circumpolar cyclonic vortex, within which wind speeds can be well in excess of 100 ms^{-1}. The vortex is created by radiative cooling as the zone of total darkness spreads out from the pole as winter approaches. The evolution of the polar vortex is strongly affected by planetary-scale disturbances originating in the troposphere. This evolution is reflected in that of zonal-mean winds.

Figure 1 shows the variation of zonal-mean winds in the stratosphere for the period September 1992 to February 1994. Zonal-mean westerly (west to east) winds are generally weaker in the stratosphere of the northern hemisphere (NH) than in the southern hemisphere (SH) and exhibit greater variability during the seasonal evolution. In mid and late winter, sudden warmings may occur which reverse the zonal-mean winds from westerly to easterly. These wind reversals correspond to a weakening of the stratospheric polar vortex in association with the growth of anticyclonic vortices. Zonal-mean wind reversals are not found in the stratosphere of the SH, except during the seasonal breakdown of the polar vortex, the so-called final warming. Zonal-mean winds in the stratosphere exhibit much greater interannual variability in the NH than in the southern hemisphere. Correspondingly, temperatures exhibit greater interannual variability in the NH than in the SH. This includes the lower stratosphere, where anthropogenic ozone losses are strongly sensitive to temperature.

4. Dynamical Phenomena in the Extratropics

The mid-winter sudden warmings referred to above affect the stratosphere much more strongly in the NH than in the southern hemisphere. These warmings can be divided into three classes, each with distinct features, and each exhibiting strong interhemispheric differences.

4.1. Early Winter Warmings

These warmings are distinctive both in structure and evolution. In both hemispheres, warmings of this class occur about a month before mid winter. (This regularity of timing must shed some light on the dynamics and origin of the phenomenon.) Such warmings do not occur every year, and are much rarer in the SH than in the NH. In the NH, they are referred to as Canadian warmings, and involve the sudden eastward translation of a previously stationary Aleutian High (AH), from its climatological position over the dateline eastwards the Greenwich Meridian (passing over Canada). The warming affects the lower and middle stratosphere, but not the upper stratosphere. The circulation has a barotropic structure, associated with large meridional displacements of air parcels, but temperature changes are small compared with warmings that occur later in the year. The polar vortex, though strongly distorted and displaced from the pole, is not broken down, but continues to intensify by radioative cooling after the passage of the disturbance. An event of similar structure and evolution in the SH was recorded by Mechoso et al. (1992), who referred to it as a South Pacific warming, on account of the eastward translation of an anticyclone around the polar vortex over the South Pacific.

4.2. Mid-Winter Warmings

Strong mid-winter warmings are common in the stratosphere of the NH. While events that can be classified as warmings are witnessed in the SH, they are much less disruptive to the stratospheric circulation than their counterparts in the NH.

In the NH, two types of warming are found. The more common type, sometimes called a wave-1 warming, corresponds to an intensification of the AH near the dateline. Such warmings are often preceded by or involve the merger of an eastward-travelling anticyclone with the AH. A recent example occurred in January 1992. Although warmings are generally regarded as being triggered by the amplification of planetary waves in the troposphere, it is, in fact, often impossible to tie such warmings to particular transient events in the troposphere. It seems that transience in the tropospheric circulation is not actually necessary; stratospheric transience can be a nonlinear response even to steady wave forcing in the troposphere.

The second, less common, type is sometimes called a wave-2 warming. This has a much greater effect on the circulation. It corresponds to the development of a second anticyclone about 180^{o} in longitude from the AH (i.e. near the Greenwich Meridian). In extreme events, the polar vortex splits between the "pincers" formed by the pair of anticyclones. The anticyclones may then merge over the polar cap, and the remnants of the polar vortex, now two smaller cyclonic vortices, may be sheared out around the dominant anticyclone. A recent example of a wave-2 warming occurred in the NH during January/February 1995. Compared with other examples that have been seen, this was a somewhat weak event, leading only to a partial split of the polar vortex. Events of this class, unlike warmings of the wave-1 class, can often be related to a corresponding transient planetaty-scale events in the troposphere.

In the SH, where tropospheric planetary wave amplitudes are weaker than those of the NH for much of winter, the polar vortex has never been seen to break down to the extent that zonal-mean easterly winds replace westerly winds. The most dynamically circulation regime in the SH is found during spring.

4.3. Springtime Final Warmings

The final warming in the stratosphere is the irregular transition in spring from westerly winds to easterly winds. The irregularity arises from a sequence of minor warmings that culminate in the final breakdown of the polar vortex.

In the NH, final warmings can be classified loosely into two types: dynamically active and quiescent. The former evolve in a similar manner to mid-winter warmings of the wave-1 type. The AH intensifies, displacing the polar vortex from the pole. The AH systematically erodes the polar vortex by drawing off air with high values of potential vorticity. The polar vortex also weakens in response to seasonal changes in radiative heating. Eventually the polar vortex is completely destroyed, and the anticyclonic circulation dominates the flow. During some events, the polar vortex may break down first in the upper stratosphere, and progressively later at lower levels (i.e. top down). During others, however, the vortex may break down first in the lower stratosphere and then later in the upper stratosphere (i.e. bottom up). A third kind of final warming is also found: events that involve the gradual weakening of an almost zonally symmetric westerly circulation. Such events occur when an earlier warming has markedly weakened the polar vortex. Because radiative timescales are much longer in the lower stratosphere than in the upper stratosphere, the polar vortex may recover in the upper stratosphere, while the westerly circulation (and associated potential vorticity gradients) remains weak in the lower stratosphere. Tropospheric disturbances are thereby unable to affect the upper stratosphere, which evolves quiescently to summer conditions. Along with these interannual differences in behaviour, final warmings in the NH show interannual variations in timing of up to two months, much greater than that found in the SH.

The flow regime during the build-up to the final warming in the SH shows much less interannual variability in the evolution and timing of events. Two flow regimes may be seen in spring. From about August through September, repeating episodes of “anticyclogenesis” occur, followed by eastward translation around the polar vortex and subsequent decay of the anticyclone. While the anticyclone is in the waning phase of its life-cycle, a second anticyclone may form over the site of its predecessor. This too undergoes eastward translation, so that, at a given time, a pair of diametrically opposite anticyclones may be found translating around an elongated polar vortex. The major axis of the polar vortex thereby rotates cyclonically, this rotation being signalled by the regular eastward progression of planetary wave number 2. During early October, there is a change of flow regime. Anticyclones begin to slow down over a preferred geographical region near the dateline. On some occasions, anticyclones generated during later episodes of anticyclongenesis catch up and merge with quasi-stationary anticyclones, leading to the development of a single, intense anticyclone. Eventually, with the polar vortex weakening systematically, the anticyclone replaces the polar vortex over the pole. Final warmings in the SH seem invariably to occur top down, and much later in the season (by up to two months) than their counterparts in the NH. More details are given in Lahoz et al. (1996).

5. Concluding Remarks

Interhemispheric differences between the stratosphere must, for the most part, be related to differences in the underlying troposphere. Despite several decades of research dynamical links between the troposphere and stratosphere in the extratropics are still imperfectly understood, undoubtedly because they involve highly nonlinear dynamics. The nature of the interhemispheric differences provide one of the most important “handles” on this problem that we have.

II. У параграфі 4.2 визначте видо-часові форми дієслів-присудків.

III. У параграфах 4.2 та 4.3 знайдіть та визначте ing- форми дієслів (Participle I, Participle II, Gerund).

IV. Складіть 5 запитань по змісту даного тексту.

КОНТРОЛЬНА РОБОТА №8
ВАРІАНТ №2

І. Зробіть письмовий переклад тексту за фахом:

Text TRANSPORT AND MIXING IN THE LOWERMOST STRATOSPHERE

1. Introduction

The transport and mixing of mass and chemical species between the stratosphere and troposphere is referred to as Stratosphere-Troposphere Exchange (STE). The process of STE is central to many aspects of atmospheric science, particularly as they relate to climate. Important examples of this include the impact of aircraft emissions on the ozone layer, the vertical structure of greenhouse gas distributions in the upper troposphere/lower stratosphere region, and midlatitude ozone depletion.

Until very recently, STE was regarded as a problem of tropopause dynamics, with attention focused almost exclusively on the mesoscale phenomenology of strong mixing events such as midlatitude tropopause folds and deep tropical convection (e.g. WMO, 1986). Most observational and modelling efforts on STE had therefore reflected this emphasis. However, over the last five years or so it has been explicitly recognized that STE is just one aspect of a global picture of transport and mixing of mass and chemical species, which is constrained by the dynamics of the whole stratosphere (Holton et al., 1995). The present paper summarizes recent progress in the area of STE, and identifies several outstanding issues.

2. The Global View

It has long been recognized that tropospheric air enters the stratosphere mainly in the tropics, and that this air moves polewards in the stratosphere. By mass conservation, this Brewer-Dobson circulation must close by stratospheric air returning to the troposphere in midlatitudes. It is quite true that a considerable part of the actual mass exchange across the tropopause appears to be accomplished by the mesoscale mixing phenomena mentioned above. However, in considering the global budget of long-lived chemical species, it has to be recognized that the STE occurring in the tropical and midlatitude tropopause regions is constrained by the transport achieved in the Brewer-Dobson circulation. Put simply, in order to achieve any mixing of chemical species in STE, a contrast between tropospheric and stratospheric air must exist, and the nature of this contrast involves the slower dynamics of the stratosphere itself.

Two new concepts lie at the heart of this current, more global view of STE. The first is the distinction between the stratospheric “overworld” and the “lowermost stratosphere”. To a first approximation, transport timescales in the overworld are determined largely by the slow diabatic circulation, with something like a two-year overturning timescale, while transport timescales in the lowermost stratosphere are determined by more rapid quasi-horizontal isentropic mixing by large-scale eddies. The lowermost stratosphere itself should be regarded as a set of quasi-horizontal outcropping layers, ventilated by the troposphere.

The second new concept is the distinction between tropical and midlatitude air in the overworld, clearly seen in aerosol concentrations and in chemical correlations. The sharpness of the transition suggests the existence of a subtropical mixing barrier, and allows for the relatively undiluted tropical ascent seen in water vapour measurements. This has led to a new conceptual model of straatospheric transport known as the "tropical pipe" (Plumb, 1996).

3. Tropical Ascent and Extratropical Descent

The rising tropical branch of the Brewer-Dobson circulation is clearly seen in chemical tracers such as N_2O and water vapour. The quantification of this upwelling is an important issue. The Lagrangian ascent is closely linked to the residual circulation, which is itself closely linked to the (transient) diabatic circulation. Since the atmospheric residual circulation cannot be reliably measured, there have been a number of attempts to estimate it from diabatic heating. However, it is well known that this calculation has certain difficulties, and the estimates therefore remain uncertain. (See Beagley et al. (1996) for an example of the ill-conditioned nature of the usual iterative method.)

The mean tropical upwelling is balanced by diabatic heating: the tropical lower stratosphere is cooled below radiative equilibrium. The associated poleward mass flux in the extratropical stratosphere is balanced by wave drag. Yulaeva et al. (1994) have thus argued that the annual cycle in lower stratosphere tropical temperatures is controlled by the annual cycle in extratropical wave drag, and some diagnostic support for this connection has been provided by Rosenlof (1995). However, a detailed quantitative theoretical between extratropical wave drag and tropical temperatures remains lacking.

The extratropical branch of the Brewer-Dobson circulation descends through the lowermost stratosphere. The long radiative timescales in this region of the atmosphere mean that the steady-state, "downward controlled" limit is not achieved on seasonal timescales, and there is therefore likely to be a lag between the extratropical wave drag and the downwelling through the extratropical tropopause. In addition, the mass of the stratosphere itself undergoes a seasonal "breathing" as the tropopause moves up and down. These processes has been quantified by Appenzeller et al. (1999). In the Northern Hemisphere, they find a lag between the maximum mass flux into and out of the lowermost stratosphere, which appears to explain the known seasonal cycle of STE based on radioactive tracers. Note that cycle is in marked contrast with the seasonal cycle of the rate of tropopause folds, confirming the global rather than local view of STE.

4. Mixing and Filamentation

The mean residual circulation described above is only part of what determines chemical transport; mixing is also important. It is believed that quasi-horizontal, nearly-isentropic motions are the dominant contributor to mixing in the stratosphere, and recent attention has focused on two regions. The first is the edge of the tropics. A number of recent studies have shown that the tropical pipe is very leaky below about 22 km or so, yet a quasi-barrier is still evident in chemical tracer correlations. This means that the tracers can be used to infer mixing rates. For example, there is a clear imprint of the tropospheric annual cycle in CO_2 in the lowermost stratosphere (Boering et al., 1994), implying significant detrainment of the tropical upwelling, and a "short-circuiting" of the Brewer-Dobson circulation; a similar conclusion has been reached on the basis of water vapour measurements by Dessler et al. (1995). Entrainment into the tropics has also been inferred from chemical profiles, and has been quantified by using a variety of chemical species with different lifetimes (Volk et al., 1996); these studies conclude that up to 50% of the ascending tropical air at 21-22 km is of extratropical origin. Much effort is now being exerted to better quantify the leakiness of the tropical pipe, as it has significant implications for the impact of aircraft No_x emissions on the ozone layer.

The other region of great interest is the lowermost stratosphere. Laminae seen in vertical profiles of O_3 and other tracers, previously attributed to inertia-gravity waves, are now clearly linked to filamentation processes. These filaments/laminae are respectively the horizontal/vertical manifestations of sloping sheets of tracer, arising from the combined effects of horizontal strain and vertical shear (a generic result of layerwise quasi-two-dimensional turbulence). The process has been quantified theoretically by Haynes and Anglade (1996), who argue that for conditions characteristic of the lower stratosphere, the ratio of horizontal to vertical length scales should be expected to be around 250. (It is a coincidence that ratio is close to Prandtl's ratio N/f.) The limiting process is small-scale vertical diffusion. Balluch and Haynes (1996) use the observed horizontal scale of tracer filaments to infer a vertical diffusivity of no greater than 10^{-2} m^2s^{-1}, which is much smaller than estimates from radars. A global climatology of lamination rates has been estimated from horizontal gradients of temperature and tracer, and shows that the process occurs year-round in the lowermost stratosphere.

This view of mixing is very different from classical diffusion, and exhibits several characteristic features of chaotic advection. The underlying process is that of large eddies acting on a sharp tracer gradient, i.e. "stirring". Homogenization occurs first at large scales; the process of mixing is analogous to shuffling a deck of cards. (Or to paraphrase Garrett (1983), if one releases red dye, one sees red streaks rather than a pink cloud.) Thus advective development of small scales is a precondition for mixing. (It follows that simulation models with insufficient horizontal resolution of tracers will require unrealistically large horizontal diffusivities to achieve a given flux.)

The filamented nature of tracer fields in the lower stratosphere is clearly seen in aircraft measurements, and has implications for (quasi-horizontal) tracer spectra. Classical two-dimensional turbulence arguments predict a k^{-1} power-law scaling, while stratospheric observations show something much steeper, close to k^{-2}. This discrepancy can be resolved by noting that if a sharp tracer edge is filamented, then a tracer cross-section will produce a set of jumps; this gives a k^{-2} "singularity spectrum" at small scales (Saffman, 1971). Since stretching rates in the lower stratosphere are highly intermittent and spatially inhomogeneous, and since filament lifetimes are estimated to be only about two weeks (Haynes and Anglade, 1996), one expects a range of scales characterized by occasional rapid injection of filaments that are then mixed away, giving a k^{-2} spectrum. (More realistically, one expects a multi-fractal spectrum since the tracer edge is not just a Heaviside function.)

A major question is how to quantify the mixing and filamentation process in the lowermost stratosphere. Note the difference between PV and chemical tracers in this regard; radiative processes should erase PV filaments, leaving chemical "fossils" of exchange (P.H. Haynes, personal communication, 1995); this is analogous to C. Gibson's "fossil turbulence" in the oceanic mixed layer.

It is perhaps worth remarking that in this view of mixing, a sharp edge in the PV is more a symptom than a cause of a mixing barrier. The extent to which this is true depends on the extent to which the PV can be regarded as locally passive; the Rossby-wave critical layer serves as a useful paradigm, since the PV mixing is determined by the kinematics of the large-scale velocity field.

5. Discussion

The above discussion highlights the fact that the scientific framework for the study of STE is changing extremely rapidly at present; we are in a time of enormous transition. New conceptual frameworks have emerged over the last five years, and are still evolving. The innovative use of chemical measurements, both from aircraft and from satellites, is providing new possibilities for quantification of transport and mixing – especially in the tropics, where analyzed winds are of dubious quality. We appear to be moving towards a Lagrangian view of the chemical circulation of the stratosphere, which is an exciting prospect. (It is somewhat reminiscent of the kind of insight long familiar to oceanographers in their ability to characterize water masses by their T-S relations.)

The single largest challenge in STE over the next few years is likely to be developing a quantitative diagnostic framework in which to interpret measurements and compare them with models. In particular, we need to get beyond the classical K-theory that represented the state of knowledge about a decade ago (WMO, 1986); "leaky pipes" are a first step in this regard. A second point is that most diagnostic approaches to transport and mixing remain based on a zonally averaged framework, which is clearly inadequate in the lowest part of the stratosphere.

II. У параграфі 2 визначте видо-часові форми дієслів-присудків.

III. У параграфі 4 знайдіть та визначте ing- форми дієслів (Participle I, II, Gerund).

IV. Складіть 5 запитань по змісту даного тексту.

КОНТРОЛЬНА РОБОТА №8
ВАРІАНТ №3

I. Зробіть письмовий переклад тексту за фахом:

<u>Text</u> OZONE TRENDS

<u>Abstract</u>

The current state of knowledge of trends in ozone is reviewed. Recent analyses of ozone data have confirmed the general features of stratospheric ozone depletion reported previously. These results are discussed in the light of the most recent international assessment (WMO, 1995). Trends in total ozone calculated from the recently revised TOMS data are slightly less than those found using the combined SBUV-SBUV/2 record and the ground-based network. The three data sets are in broad agreement to within about 1% per decade or so, i.e. at the edge of the claimed instrument stability. The trends in total ozone are primarily the result of ozone destruction in the lower stratosphere between altitudes of 15 and 25 km. There is uncertainty in the magnitude of the trends below 20 km as the trends found from the SAGE I/II record are much larger than those found from the ozonesonde network at northern mid-latitudes (the only region where a real comparison can be made). Recent work on the altitude registration of the SAGE I data has brought the SAGE I/II trends into good agreement in the upper stratosphere. A joint SPARC-IOC assessment of the trends in the vertical distribution of ozone underway. Possible causes of the observed ozone trends are discussed.

<u>Introduction</u>

Accurate determination of the magnitude and nature of trends in ozone is needed if the relative contributions of the various possible causes of decadal changes in ozone are to be distinguished. The Antarctic ozone hole is a good example where ozone measurements provided tight constraints on the possible causes of the rapid springtime depletion. High quality measurements of total ozone since the 1950s provided an excellent climatology which showed that the decline first observable in the late 1970s was a new phenomenon and not part of the natural variability, at least on a 20-30 year timescale (Farman et al., 1985). These same measurements showed that total ozone dropped rapidly from the end of August and that extensive loss did not take place in the polar night, but only upon the return of sunlight. Ozonesonde measurements of the vertical distribution soon confirmed the latter point and, more importantly, showed that the loss occurred in the lower stratosphere, principally at altitudes between 15 and 20 km. These observations tightly constrained possible explanations, especially when allied with the chemical and aerosol measurements made in the second half of the 1980s.

There is no other region where available ozone measurements provide such tight constraints. (Certainly not the Arctic where there are no records of comparable length and quality, but where a number of recent studies have empirically determined the chemical ozone loss in particular winters. While this is partly due to the large size of the ozone loss over Antarctica which dwarfs the uncertainties in the ozone measurements and partly due to the relative homogeneity inside the Antarctic vortex, it is also testament to the perseverance of the people who made the measurements. Our understanding of global trends on ozone has been reviewed periodically in the WMO assessments, most recently in 1994.

Our picture of the multi-decadal variations on ozone come from the ground-based network, principally Dobson spectrophotometers, which was mainly in the northern mid-latitudes in the early years. The quality of these records is very patchy and has been greatly improved over the last ten years. The magnitude of the observed trends is comparable with the uncertainties in the measurement, so that quality of measurement is still an issue of prime concern.

It is now 18 years since the Nimbus 7 satellite was launched included the TOMS and SBUV instruments in its layload. The two instruments worked for unexpectedly long times, until 1993 and 1990, respectively. Subsequent versions of both types have been launched and it is possible to combine records from the different instruments providing great care is taken with their absolute calibrations. In practice the only independent means of assessing this is by comparison with the ground-based stations, a process which inevitably reduces their true independence.

Finally it is important to remember that the limits on statistical significance imposed by natural variability and particularly by uncertainty in instrumental calibration do not necessarily equate with scientific importance. Two particular instances stand out – ozone over the tropics where a small change in ozone could have large radiative impact, and ozone in the lower stratosphere which is just plain hard to measure with any accuracy at a global level.

Total Ozone

Recent analyses of total ozone data have confirmed the general features of stratospheric ozone depletion reported in the most recent international assessment (WMO, 1995), where the global trends were based on the combined Nimbus 7 SBUV/2 records for the period from 1979 to 1994 (see also Hollandsworth et al., 1995). Since 1979 annually average ozone decreased by 4-5% globally and by about 7% at mid-latitudes in each hemisphere. At northern mid-latitudes in winter and spring the decrease since 1979 is about 11%. The seasonal variation in the changes is less pronounced in the southern hemisphere. These changes were in good agreement with those found from the ground-based network. Bojkov and Fioletov (1995) have analysed the same ground-based records for Europe and North America since 1957.

The magnitude and significance of the ozone depletion in the 1970s are unclear. The trend analyses in the report of the International Ozone Trends Panel (WMO, 1990) indicated statistically significant ozone loss in the 1970s, as do the analyses in WMO (1995). However this is not obvious in the time series shown in Figure 2 and 6 of Bojkov and Fioletov (1995). It would be useful to resolve this issue to increase our confidence in the behaviour of the stratosphere when the chlorine loading eventually drops below the 1980 value (around 2 ppbv Cl).

Trends in total ozone found from the recently available TOMS v.7 data (combined records from the instruments on Nimbus 7 and Meteor 3) are slightly smaller than those found using the combined SBUV-SBIV/2 record and the ground-based network. The three data sets are in broad agreement to within about 1% per decade or so, i.e. at the edge of the claimed instrument stability. The most obvious impact of the revisions in the TOMS data is to change the area of statistically significant trends, in particular in the northern mid-latitude in autumn and in the southern mid-latitudes. This change should not be exaggerated as the trends have changed from being just above the 95% significance level to just below.

The reduction of the magnitude of the early winter loss over northern mid-latitudes is intriguing, as trend analyses of the TOMS v.7 data in an equivalent coordinate system (following Randel and Wu, 1995) indicate that there is still loss outside the Arctic vortex in early winter (I.Kilbane-Dawe, private communication), whereas trends calculated in geographic coordiates do not.

The other feature worthy of further investigation is the significance of the trends in the tropics. The tropical trends derived from the SBUV-SBUV/2 record in WMO (1995) were about 2% per decade and were not significant when allowance was made for the uncertainty in the instrumental stability. The trends from TOMS v.7 are also statistically insignificant in the tropics and are closer to zero. The two records thus agree within experimental uncertainty, but a trend of –1 or –2% per decade in the tropics could still have a large radiative impact.

Vertical Distribution of Ozone

WMO (1995) reported that the bulk of the decline in total ozone over northern mid-latitude has taken place at altitudes between 15 and 25 km. Between 20 and 25 km there is good quantitative agreements between the trends since 1979 found in the ozonesonde and SAGE I/II records. However while both records show statistically losses over northern mid-latitudes at altitudes between 15 and 20 km, there is disagreement on the magnitude of the reduction with SAGE giving trends up to -20±8% per decade and the ozonesondes giving an average of -7±3% per decade. Ozone losses of up to 10% per decade are observed at higher altitudes (35-45 km).

The altitude registration is recognised problem of the SAGE I measurements. In WMO (1995) it was assumed that the correction to the SAGE I data should be 300 m. Wang et al. (1996) have reassessed the issue and concluded that a latitude variation is more appropriate. Using empirically determined values for the correction ranging from 200-400 m, they obtain trends from the SAGE I/II record that are in much better agreement with those from SBUV/2 at high altitudes than those reported in WMO (1995). Measurements just above the tropopause were not considered in this analysis.

The reason for the apparent disagreement between 15 and 20 km is not known. One suggestion is an error in the SAGE measurement in the presence of aerosol. A new treatment of aerosol absorption is being introduced into the SAGE II algorithm. The effect on trends is not currently known, but will be covered in the joint SPARC-IOC assessment of our knowledge of trends in the vertical distribution with particular attention being paid to the lower stratosphere. A report is being prepared with chapters on: a) the instruments and their algorithms; b) a comparison of measurements (SAGE, ozonesonde, Umkehr, SBUV, lidar, UARS, etc.); and c) determination of trends. It is hoped that this will be printed in early 1998.

Causes of Trends

While it is beyond the scope of this paper to discuss the possible causes of the observed ozone loss in anything other than very general terms, certain aspects of trend analysis are worth considering. There are strong chemical and dynamical influences on stratospheric ozone and so it is reasonable to suspect that either or both could be contributing to the observed long-term, global changes. These is no doubt that rapid chemical ozone loss occurs in the

Antarctic and Arctic vortices. The signal is large and a whole host of chemical measurements allow unambiguous identification of perturbed halogen chemistry as the cause.

More attention is now being paid to examining the time series of ozone measurements to see if the observed interannual variations can provide information on possible mechanisms. For example, Solomon et al. (1996) used a 2D model to investigate the effect of observed aerosol amounts on the trends calculated for the chlorine and bromine changes. Their calculated ozone changes following major volcanic eruptions in the 1980s and 1990s correlated with variations in the observed total ozone record, although the calculated ozone loss was two-thirds that observed (see also Solomon et al. and Jackman et al. in this volume). This discrepancy may be reduced when the models are compared with TOMS v.7 measurements. However a similar discrepancy exists between ozone loss rates found in the Arctic vortex and photochemical calculations indicating that a problem may remain.

It is clear that chlorine and bromine compounds lead to ozone depletion, but can they account quantitatively for the observed trends or are other factors involved ? Analyses of total ozone trends routinely include terms to describe the influence of the solar cycle and quasibiennial oscillation on total ozone. With 17 years of global satellite data and 40 years of ground-based observations – i.e. longer than the periods of these or other natural phenomena which could affect ozone – the trends (assumed to be linear) are robust to possible errors in the description of these phenomena. It is still possible, however, that the attribution of the observed variations in total ozone to these phenomena is not correct. For example, Solomon et al. (1996) suggested that some of the apparent influence of the solar cycle on total ozone may actually be related to the timing of the recent volcanic eruptions.

Long term changes in the circulation of the atmosphere connected with climate change could cause trends in total ozone. Changes in temperature have been observed (Naujokat and Pawson, this volume; Ramaswamy, this volume). How much do the observed temperature changes result from the ozone changes? The observed regional variations in the ozone trends have been attributed to the observed changes in the geopotential heights in the low stratosphere and upper troposphere. An increase in the tropopause height at Hohenpeissenberg (420m over 30 years) has been suggested as contributing to the observed long term change in ozone there, in both the column and the vertical distribution. It is important to find out on what scales such changes are occurring. Is it a redistribution of stratospheric ozone leading to local changes with no net global change or is there a more fundamental change in the stratosphere leading to both regional and global changes in stratospheric ozone ? The most unlikely scenario would seem to be one in which there is no change in the climatology of the stratosphere. GCM model studies do indicate that increased CO^2 could modify the chemistry and dynamics of the stratosphere with consequent changes in the ozone distribution. Changes in the chemical composition of the stratosphere are thus likely to be occurring in the presence of a changing dynamic structure of the stratosphere. Identifying, attributing and predicting changes in ozone (both in total and in the vertical distribution) thus becomes very tricky.

The author thanks both the UK Department of Environment and the Environment and Climate programme of the DG-XII of the European Commission for their support.

II. У параграфі "Vertical Distribution of Ozone" визначте видо-часові форми дієслів-присудків.

III. У параграфі "Total Ozone" знайдіть та визначте ing- форми дієслів (Participle I, II, Gerund).

IV. Складіть 5 запитань по змісту даного тексту.

КОНТРОЛЬНА РОБОТА №9

ВАРІАНТ №1

I. Зробіть письмовий переклад тексту за фахом:

<u>Text</u> MONITORING AND UNDERSTANDING OF UPPER TROPOSPHERE - LOWER STRATOSPHERE WATER VAPOR

The distribution of water vapor in the upper troposphere and lower stratosphere (UT/LS), and its future evolution, is of major importance for understanding the radiative balance of the earth-atmosphere system, as well as tropopause region dynamics and chemistry. The present state of knowledge of this distribution is presented, with some general considerations for a measurement strategy. This is followed by a discussion of plans for determining the present water vapor distribution, through the use of satellite and in-situ data, and initial plans using mainly in situ measurements to understand the mechanisms that maintain that distribution. Some other desiderata and conclusions are presented in the closing sections.

The material in this paper has been derived from the preliminary draft of the Prospectus for a SPARC Plan for the Study of Lower Stratosphere/Upper Troposphere Water Vapor (Gille et al., 1996).

1. <u>The Importance of Lower Stratosphere – Upper Troposphere Water</u>

Water vapor has long been recognized as crucial in determining the radiative balance of the Earth. Recently it has been pointed out the far infra-red component of the water vapor emission may be a significant influence on the water vapor greenhouse effect. Recent work clearly demonstrates that most of the greenhouse trapping due to water vapor absorption occurs in the middle and upper troposphere. In the tropical troposphere a relatively small change of the water vapor abundance (17%) will produce the same greenhouse forcing as doubling the CO_2 concentrations. Since 17% is at best comparable to the uncertainty in radiosonde measurements of upper troposphere humidity, those data are not sufficient for studies of Earth's radiative balance. More accurate knowledge is required.

Water vapor is also an excellent tracer for stratosphere-troposphere exchange processes, as illustrated by the recent observations of the so called tape recorder effect, and in distinguishing between stratospheric and tropospheric air in the "middleworld". Finally, water vapor is the source of hydroxyl radicals, responsible for much of the oxidizing capacity of the atmosphere, and therefore critical for removing many anthropogenic compounds released into the atmosphere. It is also involved in hydrolysis reactions that are important for the removal of reactive chlorine and nitrogen species. A good knowledge of its distribution and variations is needed to improve our knowledge of UT/LS chemistry.

It appears that, to understanding these processes, water vapor distributions need to be known with an accuracy no worse than 10%, and a precision of 5-10%. The spatial and temporal resolution depends on the application of the data; studies of some mechanisms will require smaller spatial and temporal measurement scales than those necessary for establishing a baseline climatology.

2. Present Knowledge of the Water Vapor Distribution

It has long been recognized that only if nearly all air entering the stratosphere is "freeze dried" on entering the tropical stratosphere can the observed aridity of the stratoshere be explained. The precise processes by which this dehydration occurs, and the temporal and spatial variability of the dehydration process, remain very uncertain. A review of the status of this subject
is provided in Holton et al. (1995).

According to current understanding, vertical advection by a non-locally controlled global-scale meridional circulation, driven by wave breaking in the extratropical stratosphere, slowly pulls air upward and poleward across the isentropes (accompanied by diabatic heating), and thus circulates air dehydrated by passage through the tropical tropopause throughout the stratosphere. The annual variation of tropopause temperature, and hence of saturation mixing ratio at the tropopause, is clearly preserved in the vertical profile as air is pulled up from the vicinity of the tropical tropopause (Mote et al., 1996).

Recent analyses of satellite and in situ aircraft water vapor measurements by Rosenlof et al. (1996) indicate that the Southern Hemisphere lower stratosphere is drier than the Northern Hemisphere, as a result of the interplay of a number of factors. High southern latitudes are dehydrated by the low temperatures in the Antarctic polar vortex in late winter and spring. The tropical regions are controlled by the seasonal cycle in tropical water vapor, while mid-latitudes see an interplay between quasi-horizontal exchange from the tropics and large scale vertical descent. One result is that the base of the "overworld" is near 60 mb in the tropics, and thus good observations are required at least up to that level.

Our understanding of processes controlling the water distribution is still very uncertain in the layer of the atmosphere in the vicinity of the tropical tropopause in which water vapor mixing rations must change from 100's of ppmv to ~3 ppmv. In order to understand the crosstropopause flux of water vapor in the tropics it appears necessary to determine the relative roles of overshooting convection and large-scale cirrus. Careful in-situ radiative and microphysical measurements as well as modeling studies are required.

3. An Overall Measurement Strategy

Given the importance of UT/LS water vapor, a number of issues immediately arise. The first is that we do not really know the present distribution of water vapor and its temporal variations in the tropopause region well enough; this must be the first question to address. This will be of limited use unless we understand the processes that maintain this distribution, including the range of scales which are involved, and the critical phase changes. Finally, this understanding must be incorporated in a hierarchy of models, to allow future water vapor distributions to be predicted.

4. Improved Determination of the Present Distribution of UT/LS Water

Determination of the global distribution of water vapor requires global observations. Useful data in this part of the atmosphere have been provided by 3 types of observation with complementary characteristics. Research satellite observations have obtained global long term coverage. These have all involved looking at the earth's limb, either in emission or in occultation, to provide vertical resolutions as fine as about 1 km, and horizontal resolution from a few hundred to a few thousand km. The great advantage of the satellite techniques is that a single instrument is used everywhere, and takes a large number of observations, so that random errors can be averaged out, and spatial and temporal variations seen. However, because of the indirect nature of the observations, the accuracy is more difficult to determine, and usually must be verified through comparison with in-situ observations, which have their own uncertainties. Numerous satellite data sets are available for use now, and more are expected in the next few years.

In situ observations from aircraft and balloons generally have a well-determined absolute accuracy. More recently, ground-based active remote sensing techniques have been developed which can provide high vertical resolution, and extensive time series at a single location.

The approach to develop a global water vapor and water climatology is then to make use of present and future satellite observations to obtain global, long term observations of the spatial and temporal variations of water vapor, to intercompare various satellite observations to reduce systematic effects, and to anchor them as firmly as possible to coincident in-situ and groundbased observations. However, obtaining and interpreting observations with in situ local instruments and satellite measurements that may average over considerable distances, and keeping small time differences, is not easy.

5. Understanding the Mechanisms that Maintain the Distribution of Water Vapor

5.1. Tropical Measurements

The first measurement program is proposed to be a longitudinal, seasonal survey and mechanistic study in the tropics. Accurate and precise measurements of the relative humidity (water vapor) and condensate (liquid and ice) mass over a wide geographical area and by season are important for assessing the validity of present inputs into mesoscale and general circulation models. Recently the presence of subvisible cirrus clouds, particularly in the tropical western Pacific, have been modeled, and their role in heating the upper troposphere and possible influence on vertical motions near the tropopause have been suggested. The mechanism by which the tropical lower stratosphere is dried is still the subject of significant debate.

The three regimes to be considered initially would be the convectively active western Pacific, the subsidence region in the Pacific, and the monsoon controlled Indian Ocean. Also of importance are the continental regions in tropical South and Central America, and Africa.

Such a set of measurements would detail the differences in the water distribution in these very different regimes as well as tracking how they change with season. These observations should also be useful in understanding the importance of these regimes in transporting water (and other constituents) from the troposphere into the stratosphere.

5.2. Extra-Tropical Measurements

Outside of the tropics the existence, cause, and impact of long-term changes in the water distribution also need to be addressed. Currently there is only one station (Boulder, CO) with a long-term lower stratospheric water vapor measurement program. There is evidence from this record that water vapor in the lower stratosphere is increasing (Oltmans and Hofmann, 1995) and that near the tropopause this increase is larger than that expected from the increase of methane in the atmosphere. To corroborate and extend these findings a Southern Hemisphere mid-latitude site such as Lauder, New Zealand should be established. Long-term satellite measurements will be required to make this determination on a global basis. Linking these with a S.H. in situ measurement site is important.

6. Water Vapor, Cloud, Climate Modeling and Prediction

An active modeling activity will need to be part of this activity. An range of mechanistic and first-principle models will be used to interpret the observations, predict future effects, and incorporate the result and test treatments before they are incorporated in large GCM's.

7. Additional Considerations

There is a critical need for the development of new instruments to meet the needs of present and future investigations. These might include low cost "operational" balloon instruments; an instrument for a long duration balloon; and instrumentation for unpiloted airborn vehicles (UAV).

The measurement campaigns described here will take extended efforts as well as logistical and financial support from numerous agencies and countries. Every effort should be made to develop these plans in conjunction with other international programs, especially of the WCRP. The GEWEX is a particularly good example. Cooperation on programs with IGAC should also be explored.

8. Concluding Remarks

This paper has summarized the reasons why a better understanding of the present and future distribution of water vapor in the upper troposphere and lower stratosphere is of critical importance for an understanding of the earth's radiative balance, and thus the evolution of its climate. It has further laid out a very preliminary plan for improving the understanding of the present distribution of water vapor, making use of existing observations and measurement capabilities, and for improving our understanding of the basic physical mechanisms that maintain that distribution. Finally, it has pointed out that a vigorous modeling program to exploit this evolving understanding is needed, and identified needs for developments in instrumentation.

II. У параграфі 1 визначте видо-часові форми дієслів-присудків.

III. У параграфах 5.1, 5.2, 6.7 знайдіть та визначте інфінитиви, об′єктний та суб′єктний інфінітивні звороти.

КОНТРОЛЬНА РОБОТА №9
ВАРІАНТ №2

I. Зробіть письмовий переклад тексту за фахом:

Text AN UPDATE OF THE EQUATORIAL QBO AND ITS VARIABILY

Introduction

The Quasi-Biennial Oscillation (QBO) of zonal winds and temperature in the lower and middle tropical stratosphere was discovered independently by Reed et al. (1961) and Veryard and Ebdon (1961). Since then, the QBO has often been described in literature; the last complete update was given by Naujokat (1986). Ten years after this publication, we want to give another update of the observed equatorial QBO. In addition to the description of the monthly mean zonal winds we will also focus on vertical wind shears as they are closely related to both temperature anomalies and the occurrence of meridional advection cells associated with the QBO, the latter being of importance for the transport of trace gases in the lower equatorial stratosphere.

It is generally known that the reversal of winds is not a purely periodic phenomenon. Instead, period lengths vary between less than two and nearly three years, and the transition of westerly to easterly wind regimes seems to be phase locked to the northern hemisphere annual cycle. A convincing physical mechanism explaining both the variability of period lengths and its phase locking to the annual cycle is still missing. After discussing the characteristics of the variability of QBO period lengths, however, we will finally propose such a mechanism and give some qualitative and quantitative arguments supporting our thesis.

Zonal Winds

The figure shows a time-height series of monthly mean zonal winds; westerly winds are shaded. Since 1976, the data set is based on daily radiosonde ascents at Singapore; prior to that, radiosonde data from Canton Island (Jan 1953 – Aug 1967) and Gan/Maledives (Sep 1967 – Dec 1975) have been used. Monthly mean profiles were calculated from the daily radiosonde soundings.

The main features of the QBO can easily be identified in the figure; Alternating easterly and westerly wind regimes are propagating downwards from mid into lower stratosphere. Although the average period length is 28 months, the QBO is by no means a periodic oscillation of the zonal wind. Short cycles as the one between 1959 and 1961 with a cycle length of less than two years exist as well as cycles with lengths of nearly three years as the one between 1963 and 1966.

The frequency distribution of zonal winds above 60 hPa is bimodal: either easterly or westerly winds prevail. Both easterly and westerly winds reach maximum values between 15 and 20 hPa. Easterlies are significantly stronger (typically up to –30 m/s) than their westerly counterparts (usually not more than 18 m/s). The strongest wind values ever observed were –41 m/s (at 10 hPa) and 28 m/s (at 15 hPa), respectively. While propagating downward, zonal wind speeds weaken; below 60 hPa, the two branches of the distribution are merging into one. Easterly and westerly wind regimes can still be separated in the time-height section of zonal winds, but occasionally easterlies from the upper troposphere can reach up to the 70 hPa level, increasing the number of wind reversals in this level. Usually, westerlies do not propagate further down than 90 hPa.

Wind Shears

When considering transport in the lower equatorial stratosphere, vertical shear zones associated with transitions of the QBO are more important than the zonal winds themselves, as the former are accompanied not only by temperature anomalies, but also by meridional circulation cells. Both westerly and easterly shear zones regularly exceed values of 10 m/s/km. Delays in the downward propagating wind regimes are found in the downward progression of shears as well. Strongest wind shears of both westerly and easterly shears occur between 25 and 50 hPa which is just that altitude region in which most of the observed delays of downward propagating zero wind lines occur.

Although easterly and westerly shear zones are initiated well separated in time at the 10 hPa level, westerly shear zones approach the easterlies from above as they propagate downward with time. This results in a convergence of westerly and easterly shear zones around the 50 hPa level occurring prior to each transition from easterly to westerly wind regimes.

Relative to zero wind lines, the position of strong (i.e. stronger than 5 m/s/km) vertical wind shears is not symmetric. Instead, strong wind shears tend to occur within weak easterlies, i.e. after the onset of easterlies or prior to the onset of westerlies.

Finally, we would like to point out that shear zones are rather thin vertical structures in the equatorial lower stratosphere. Strong shear zones – regardless of being easterly or westerly – usually do not cover more then 3 to 4 km in altitude.

Period Lengths

It has already been pointed out that the average period length of the QBO is 28 months at all levels. The shortest period ever observed at 30 hPa was 20 months (1959-1961), the longest 35 months (1966-1969). Above 30 hPa, easterly winds last longer than their westerly counterparts; below the 30 hPa level, easterlies are shorter than westerlies.

An inspection of the lengths of individual cycles suggests an increase of period lengths. A least square estimate indeed results in a positive trend of typically 1.8 months/decade in all levels; the estimated 2σ confidence intervals of more than 2.6 months/decade, however, indicate that this trend is statistically not significant.

The variability of period lengths is connected with the delay of the downward propagating wind regimes, especially of the easterlies. However, westerlies can be delayed as well, as the progressions of westerlies during the northern hemisphere winters 1960/61, 1968/69 and 1979/80 demonstrate. This can be clearly seen in frequency distributions of the downward propagation speeds of the zero wind lines: Westerlies propagate downward with a typical speed of 1 km/month, sometimes being slightly delayed around the 30 hPa level. Easterlies propagate downwards significantly slower (typically 0.5 km/month) and often show significant delays in the altitude range between 25 and 50 hPa.

The coupling of these delays to the annual cycle (Dunkerton and Delisi, 1985; Dunkerton, 1990) can be demonstrated by plotting the number of easterly onsets for each month at the 50 hPa level which shows a strong seasonal dependence; during northern hemisphere winter, only very few onsets took place. Alternatively, a pronounced annual cycle is found in the downward propagation speeds of easterlies between 25 and 50 hPa, with minimum values during northern hemisphere winter.

As a possible cause for the varying period lengths the variation of the wave forcing which drives the QBO has been suggested by Dunkerton (1990) and Geller (1996, personal communication); the coupling to the annual cycle nevertheless remains unexplained. However, the following argument excludes changes in the wave driving as the primary cause of the variability of the QBO: In the absence of wave driving, the temperature anomalies associated with vertical shear zones would be radiatively damped on a timescale of, say, 20 days, resulting in the decay of strong shears. Instead, observations show that strong vertical shear zones are maintained over several months even if the downward propagation of a wind regime stalls. Thus, the wave forcing maintaining the shear zones and driving the QBO must continue even if the downward propagation of shears is delayed. Whatever causes delays in the QBO therefore does not reduce the wave forcing significantly (although variations in the driving waves will certainly affect the QBO period length to a certain degree). The signature of wave forcing in driving the QBO therefore should be a strong forcing when strong vertical shears are present. This is the pattern shown by scatter plots of wind acceleration vs. shear at the 15 hPa level; at those levels where the development of the QBO is subject to delays, however, strong easterly shears coincide with weak net accelerations.

During the last few years, the annual cycle in the lower tropical stratosphere has been reinterpreted, and is now understood as a consequence of the annual cycle in the diabatic meridional circulation driven by extratropical planetary waves. One of the prominent features of this concept is the tropical tape recorder (Mote et al., 1996; also see references therein) in water vapour.

However, the annual cycle in the tropical uprising branch of the global meridional diabatic circulation also explains an important part of the observed variability in QBO period length by causing a seasonally varying nonlinear vertical advection of QBO shear zones. Obviously, this concept works in a qualitative sense: Strongest uprising (and therefore the most significant delay of QBO shears) occurs during northern hemisphere winter. The amplitude of the annual cycle in the residual circulation is in the order of 0.1 mm/s; coupling into QBO shear zones of more than 5 m/s/km results in additional forcings in the order of 0.4 mm/s/d, which happens to be the magnitude of the net forcing driving the entire QBO.

In order to obtain amore quantitative argument, we reconstructed a climatological wave forcing of the QBO by subtracting the vertical advection term from the observed wind accelerations. An estimation of the involved vertical motions due to the annual cycle in the meridional circulation is based on a climatology of tropical temperatures throughout the stratosphere presented by Reid (1994). Scatter plots of the reconstructed wave forcing then show the expected behaviour of strong wave forcing occuring in common with strong vertical shears, supporting our thesis that the missing part in the variability of QBO period lengths is indeed given by variations in the diabatic circulation.

The complete results of our analysis will be submitted for publication in the near future.

II. У параграфі “Zonal Winds” визначте видо-часові форми дієслів-присудків.

III. У параграфах “Introduction” та “Wind Shears” знайдіть та визначте інфінитиви, об′єктний та суб′єктний інфінітивні звороти.

КОНТРОЛЬНА РОБОТА №9
ВАРІАНТ №3

I. Зробіть письмовий переклад тексту за фахом:

Text THE MONSOON CLIMATE OF EAST AFRICA

The monsoon winds blow in response to the differential heating between continent and the Ocean. The continent heating is strongest when the sun is overhead at a given location. In the equatorial region the sun is overhead at any given location twice in each year. This causes two distinct monsoons to flow to this zone of maximum heating and creating a region of low level airmass convergence called the Intertropical Convergence Zone (ITCZ). The two monsoons over eastern and central Africa are the northeast monsoon air current occurring in northern winter season and the southeast/southwest monsoon air current which occurs during the northern summer season. During the intermediate seasons, equinoctial periods (March - May, September - November) both monsoons are present with one withdrawing while the other is advancing. These monsoons, during the transitional seasons, bring in “equatorial air” from the Indian Ocean. This equatorial air has a conditionally unstable lapse rate and responds rapidly to low level convergence with widespread cloudiness, showers and thunderstorms. The following subsections give details of the monsoons of the region.

1.1 Northeast Monsoon

The northeast monsoon air current has its roots in the high pressure centre over Arabia and India. Generally the northeast monsoon air is mainly of continental origin with a moderate sea trajectory and hence is dry. It is known (Findlater, 1971; Anyamba, 1983) to be not only shallow but also divergent. This southern summer monsoon air enters East Africa in a northeasterly direction but bifurcates immediately over northern Kenya into two airstreams. One stream flows southwards along the East African Coast and over mainland Tanzania while the other flows westwards into Uganda, southern Sudan and eastern Zaire. The vertical extent of the monsoon has been shown (Anyamba, 1983) to be below 500 hPa level and is capped aloft by an easterly air current. The strength of the monsoon air is variable but the mean peak resultant wind speed ranges from 7 to 10 metres per second. Anyamba (1983) has presented details of the mean streamline and isotach patterns of the monsoon air current at several levels of the East African troposphere. A similar observational analysis has also been presented by Kiangi et al. (1981).

Sadler et.al. (1987) have shown that the northeast monsoons at the gradient level recurve and become westerlies near their southern limit. The curvature over the southwest Indian Ocean depicts two troughs to be associated with these equatorial / near-equatorial westerlies. The one on the equatorward side of the westerlies has its origin in the northern hemisphere while that to the south has its origin in the southern hemisphere. These oceanic westerlies are shown to be truly equatorial during April.

The occurrence of episodes of active weather conditions in the region during the dry season sometimes do assume large scale dimensions. In such situations the normally dry northeast monsoon becomes very active weatherwise in eastern Kenya and northeastern Tanzania. These rainy episodes in January and February occur when upper level troughs (typically at 500 hPa level) in the weaterlies over eastern Mediterranean Sea undergo massive equatorwards extension (deeping). The subsequent upper level evacuation of air from the east African region towards the trough tends to be associated with marked veering of the monsoon air at low levels to become mainly an easterly air current which is conductive to increased funnelling of moisture into the area. This easterly air current is often confluent and on undergoing orographic lifting over the eastern high grounds in Kenya and northeastern Tanzania lead to rainy episodes over large areas.

1.2 Southeast/Southwest Monsoon

The cool and moist southeast/southwest monsoon air flows from the Mascarene high in the south. These monsoon currents are shallow, being confined to the lowest three kilometres above mean seal level (Findlater, 1969a; 1971).

An inversion layer exists with base near 700 hPa and top near 600 hPa level (Kiangi et al. 1981). This inversion hinders further cloud development rendering them shallow and therefore unable to precipitate. This season (June to August) is therefore cold and dry. There are some areas, however, which receive considerable amount of rainfall during the southeasterly monsoon season. These are western Kenya, eastern Uganda, northwestern Tanzania and central Africa (Tomsett, 1969). These rains are caused by the interaction between the southeast monsoon current, the Congo Airmass and the Lake Victoria thermally induced mesoscale circulation (Kiangi et al. 1981). It has been proposed (Semazzi, 1980) that the quasi-permanent orographically induced trough over central Africa intensifies and subsequently merges with the Lake Victoria Trough during this season. On occasion during this season, high pressure forms over the southeastern Atlantic/Zaire basin throwing westerlies into the eastern Africa region. These westerlies are relatively moister and unstable, and give widespread rains over eastern Africa.

1.3 The Rainy Seasons

Most parts of East Africa receive rains twice in a year. These rainy seasons occur during transitional months, that is when the circulation is changing from one monsoon regime to the other. The season occurring during mid-October to mid-December is locally called the "Short Rains" and the one from mid-March to early June is the "Long Rains". It is also observed from Table 1 that some areas receive rain only once in a year. The areas with only one rainy season are found in the extreme south of the region and also in the northwest starting from western Kenya.

1.3.1 The "Short Rains" seasons

From October the Near-Equatorial Trough (NET) south of the equator takes the role of ITCZ and moves first southwards and then northwards and the rains also move with it. In its southward journey the ITCZ causes rains over Kenya, Uganda, and northern Tanzania (north of about latitude 5^oS) from mid-October to mid-December, by the end of December and gives rains in southern Tanzania from November to March.

During the period of the "Short Rains", northerlies are replacing southeasterlies in the lower troposphere. Note that during October the prevailing flow over the west Indian Ocean is westerly (Sadler et al., 1987) similar to that depicted in Fig. 2 for April.

1.3.2 <u>The "Long Rains" seasons</u>

From the end of January the ITCZ starts to move northwards following the overhead sun. It again causes rains over northern Tanzania, Kenya and Uganda from mid-March to early June. This period is known as the "Long Rains" season. By April the ITCZ is back to about 2^oS and is about to give up its role as ITCZ and function as NET south of the equator. At this stage NET north of the equator assumes the role of ITCZ, and moves northwards taking the monsoons with it to India.

During the season of the Long Rains, the southerlies are pushing forward from the south in the lower troposphere along the coast of East Africa till the Low Level jet gets established by the end of June.

II. У параграфі "Northeast Monsoon" визначте видо-часові форми дієслів-присудків.

III. У параграфах "Introduction" та "Northeast Monsoon" знайдіть та визначте інфінитиви, об′єктний та суб′єктний інфінітивні звороти.

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