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Climate change prediction
Dry spells of 1988–90

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Climate change prediction

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Summary

A brief description of climate processes is given. Numerical models are described and an assessment of their ability to simulate climate and climate change is given. Results from simulations with increased atmospheric CO_2 concentrations are given, including the temporal evolution of the warming, and estimates of the geographical distribution of changes. The main sources of uncertainty are listed, and possible shortcomings in current research programmes are identified.

1. Introduction

In order to predict changes in climate, one must first identify and understand the various components of the climate system. These include the atmosphere, the oceans, the cryosphere (land–ice and sea–ice), the biosphere and, over geological time-scales, the geosphere. The fundamental process driving climate is radiation. The climate system is heated by solar radiation, which is generally strongest in low latitudes, and cooled by long (thermal or infra-red) radiation to space, which is more uniformly distributed with latitude. The resulting temperature contrast between low and high latitudes drives atmospheric and oceanic motions which transport heat polewards and thus tend to reduce the equator-to-pole temperature gradient. The resulting circulations and their interactions with orography and the biosphere determine the earth's climate.

Changes in the radiative heating of the earth will produce changes in its climate. Changes in solar heating can arise from external factors such as variations in incident solar radiation due to changes in the earth's orbit, or in the earth's reflectivity, for example, due to addition of volcanic aerosols to the atmosphere, and to changes in the reflectivity of the surface such as that caused by deforestation. The long-wave cooling can be modified by changes in atmospheric composition, notably by changes in the concentration of greenhouse gases. Other non-radiative factors such as changes in orography can also affect climate, but these occur on geological time-scales, and so will not be considered further here. Note that changes in climate will themselves produce changes in the earth's radiation budget: these are known as radiative feedbacks and are discussed later in this section.

In this paper, we are primarily interested in predicting changes in climates over the next century or so, hence it may be possible to neglect changes in some of the more slowly varying components of the system. The atmosphere and land surface, including seasonal snow cover, adjust to changes in heating on time-scales of a year or less. The seasonal mixed layer of the ocean (typically up to 100 m deep) and sea ice respond over periods of up to a decade or so. The oceanic warm water sphere (down to the permanent thermocline, typically at half a kilometre) has a thermal response time of several decades, whereas the deep ocean, extending to several kilometres, has a time-scale of centuries to millenia. The major ice-sheets vary over millennia, so apart from the substantial contributions from changes in accumulation and melting on sea level, they can be regarded as fixed in the present context.

In predicting climate change, it is generally assumed that for each specification of the factors affecting climate (solar heating, atmospheric conditions, orography), the earth's climate will move to a long-term steady state, known as the equilibrium climate, in which there is no net heating of the system. Note that this does not preclude considerable variability on interannual or even interdecadal time-scales, but there are no longer-term trends. If the factors affecting climate are changed (for example, by increasing the concentration of greenhouse gases), then the system will adjust slowly towards a new equilibrium state. The transition period is sometimes referred to as a 'transient' or 'time-dependent' climate change: the difference between the initial and final states is the 'equilibrium climate change'.

One possible method of predicting climate changes is to look for periods in the past when the factors affecting climate (solar parameters, greenhouse gas concentrations) were similar to those expected in the near future. One could then use reconstructions of the relevant past climates as a forecast for the future climate. Unfortunately, in the recent geological past, there are no close analogies to an increase in greenhouse gases, and so this method cannot provide reliable forecasts of climate over the next century or so.

The only alternative to the climate analogy approach is the use of physical-mathematical models of climate. Among such models, the most highly developed are general circulation models (GCMs). These are numerical models in which the equations of classical physics (including the laws of motion, and requirements for conservation of heat and mass) are solved in a 3-dimensional grid, with a horizontal spacing of 250–800 km, depending on the model concerned.

Many processes (for example, those associated with cloud or vertical mixing in the oceans) occur on much smaller scales than can be resolved by the model grid, and so are represented in a simplified or idealized way (parametrizations). These may be based on approximations of the underlying equations, data from observational studies or laboratory experiments, results from numerical experiments using a finer grid or, as a last resort, sensitivity experiments using the GCM itself. In a climate model, an atmospheric component (essentially the same as a weather prediction model) is coupled to a model of the oceans and sea ice, which may be equally complex.

Climate forecasts are derived in a different way from weather forecasts. A weather prediction model gives a description of the atmosphere's state up to 10 days or so ahead, starting from a detailed description of an initial state of the atmosphere at a given time. Such forecasts describe the movement and development of large weather systems, though they cannot represent very-small-scale phenomena, for example, individual shower clouds.

To make a climate forecast, the climate model is first run for a few (simulated) decades. The statistics of the model's output will be a description of the model's simulated climate which, if the model is a good one, will bear a close resemblance to the climate of the real atmosphere and ocean. The above exercise is then repeated with, for example, increased concentrations of the greenhouse gases in the model. The differences between the statistics of the two simulations (for example in mean temperature and interannual variability) provide an estimate of the accompanying climate change.

In this paper, reference will be made to atmospheric models run with prescribed sea-surface-temperature atmospheric models (referred to henceforth as AGCMs) coupled to an oceanic mixed layer, with or without allowance for ocean heat transport (A/MLMs) and atmospheric models coupled to full dynamical ocean models (CGCMs).

As alluded to above, changes in climate can lead to changes in the components of the earth's radiation budget which may be amplified (positive feedback) or reduced (negative feedback). The simulated strength of feedbacks (such as those due to cloud) vary from model to model. The relative strengths of the radiative feedbacks in different models have been analysed by substituting the globally averaged changes in the simple energy balance equation

$$\Delta T_s = \Delta Q/\lambda$$

where ΔQ is the applied radiative perturbation (W m^{-2}), ΔT_s is the equilibrium change in globally averaged surface temperature, and λ is the climate sensitivity parameter (W m^{-2} K^{-1}). For example, doubling CO_2 produces an increase of 4.4 W m^{-2} in the radiative heating of the troposphere and surface. If the system responded as a radiative black body, this would produce an equilibrium warming of 1.2 °C. Warming the atmosphere leads to an increase in atmospheric water content — water vapour is also a greenhouse gas — and estimates based on both observations (Raval and Ramanathan 1989) and models indicate that produces a positive feedback, increasing the warming to 1.7 °C.

Snow and sea ice reflect solar radiation back to space. The global warming associated with increases in greenhouse gases would reduce the aerial extent of snow and ice, increasing solar absorption, leading to a further increase in temperature. The magnitude of this feedback is generally much smaller than that due to water vapour, though estimates of the strength vary from model to model.

Clouds cool climate through reflecting solar radiation back to space, and warm climate through their greenhouse effect. In our present climate, it appears that the solar effect dominates. Any change in cloud amount, height or cloud radiative properties will alter the net effect of clouds on the earth's radiation budget.

Numerical studies suggest that doubling CO_2 amounts would lead to an equilibrium warming of 2–5 °C, much of the uncertainty being associated with uncertainties in the strength of cloud radiation feedbacks and associated processes (for example see Mitchell *et al.* (1989), Cess *et al.* (1989)).

Some scientists have suggested that the climate sensitivity is much smaller than indicated by models. A few, including Idso, and Newell and Dopplick considered the energy balance at the surface only or neglected the water vapour feedback (see Luther and Cess (1985) for references and a detailed explanation why such approaches are misleading). More recently, Lindzen (1990) and Ellsaesser (1989) have questioned the strength of the water vapour feedback in climate models. Lindzen argues that cumulus convection 'short circuits' much of the potential water vapour feedback by transporting heat from the boundary layer to the upper troposphere where it is more readily radiated to space. This process is undoubtedly represented in current general circulation models and provides, as expected, a strong negative feedback through changes in lapse rate (see, for example, Schlesinger and Mitchell (1987)). Lindzen and Ellsaesser also argue that as convection penetrates to higher levels in a warmer climate, the compensating subsidence will start from higher and therefore colder and drier levels. This would lead to a reduction in the absolute humidity in the upper troposphere in the descending branch of the Hadley circulation, and hence a 'local' negative water vapour feedback. Since Lindzen has yet not published the details of his argument, it is not possible to assess it quantitatively. Nevertheless, this mechanism is also included in current models (see, for example, Gregory and Rowntree (1990)). Indeed, on doubling CO_2 concentrations, the Meteorological Office high-resolution model produces decreases in absolute humidity in this region. It appears that other processes including the detrainment of warmer, moister air and lateral mixing of moistened air from the inner tropics and mid latitudes reduce the extent of the drying through deeper subsidence. The increased drying through subsidence appears to be responsible for the decreases in relative humidity in mid latitudes and the tropics which leads to the reduction in cloud in the upper troposphere and a strong positive cloud feedback found in many models (Mitchell and Ingram 1991). Thus, if models have exaggerated the positive water vapour feedback, it is also likely that they have *underestimated* the strength of the positive feedback associated with reductions in cloud amount. In summary, although tropical convection remains one of the major sources of uncertainty, it is unlikely that errors in the parametrization of convection lead to a gross overestimate of climate sensitivity (even removing the water vapour feedback *completely and globally* in a model with a sensitivity of 2.5 °C due to doubling CO_2 — the IPCC 'best guess' — would only reduce the equilibrium warming to about 1.4 °C).

2. Validation of numerical climate models

In this section, an attempt is made to answer the question 'To what extent should one believe results from climate models?' Firstly, the models are based on a firm physical basis with a minimum of adjustable parameters, as discussed in the previous section. Secondly, they show considerable skill in reproducing the large-scale features of current climate. Thirdly, they have been shown to be capable of reproducing many features of contemporary climate change, and some of the main features in more recent palaeoclimate. These last two points are expanded on below.

2.1 Simulation of present climate

The simplest way of validating a climate model is to prescribe present-day 'boundary conditions' and compare the long-term statistics of the resulting simulation with observational climatologies. This shows that AGCMs and A/MLMs have considerable skill in the portrayal of the large-scale distribution of pressure, temperature, wind and precipitation in both summer and winter, although this success is due in part to the constraints on sea surface temperature and sea ice. There has been a general reduction in the errors in more recent AGCMs as a result of increased horizontal resolution, improvements to the parametrization of convection, cloudiness and surface processes and the introduction of parametrizations of gravity-wave drag. For example, the three most recent simulations considered in the IPCC Working Group I Report (IPCC 1990) show a marked improvement in the simulation of the depth and position of the Antarctic circumpolar surface pressure trough (Fig. 1).

Changes in variability of climate are as important as changes in mean climate in the assessment of climate impacts. Hence, the ability of models to simulate the variability of current climate should also be assessed. The daily and interannual variability of temperature and precipitation have been examined but only to a limited extent. There is evidence that variability is overestimated in some models, especially in winter. The daily variability of sea level pressure can be well simulated, but the eddy kinetic energy in the upper troposphere (indication of the variability of the flow) tends to be underestimated. The level of interannual variability of global mean surface temperature in CGCMs (for example, Fig. 2) is comparable to that observed over similar time-scales if allowances is made for the estimated trend in the observed.

On regional scales, there are significant errors in all models. A validation of A/MLMs for five selected regions (typically 4×10^6 km^2) showed errors in area average surface temperature of 2–3 °C (IPCC 1990). This is small compared with the average seasonal range of temperature of 15 °C. Errors in mean precipitation for the same five regions ranged from 20–50% of the observed average. All the recent models reproduce the northern summer monsoon rainfall maximum over

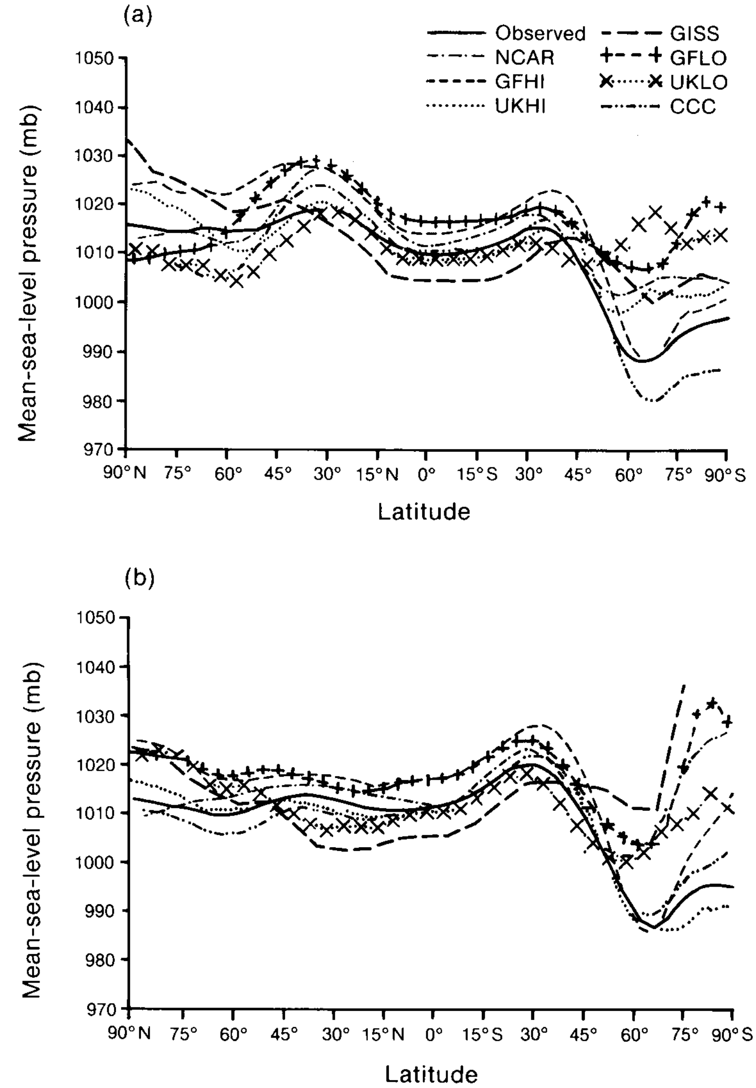

Figure 1. Zonally averaged mean-sea-level pressure (mb) observed (Schutz and Gates 1971, 1972) and modelled for (a) December, January and February, and (b) June, July and August using the following models: NCAR (National Center for Atmospheric Research) (15 spectral waves), GISS (Goddard Institute for Space Studies) (8° latitude × 10° longitude), GFLO (Geophysical Fluid Dynamics Laboratory) (15 spectral waves), UKLO (Meteorological Office) (5° × 7.5°), GFHI (Geophysical Fluid Dynamics Laboratory) (30 spectral waves), UKHI (Meteorological Office) (2.5° × 3.75°) and CCC (Canadian Climate Center) (32 spectral waves). The more recent high-resolution runs referred to in the text are GFHI, UKHI and CCC.

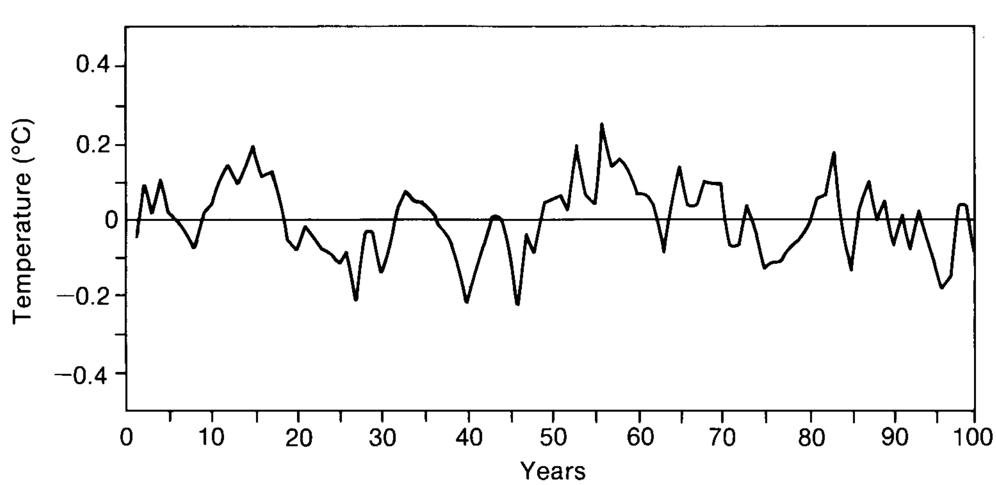

Figure 2. Temporal variation of the deviation of global mean surface air temperature (°C) of the Geophysical Fluid Dynamics Laboratory coupled ocean–atmosphere model from its long-term average (Manabe *et al.* 1991).

south-east Asia (Fig. 3), but in the example shown, the mean rainfall over south-east Asia is substantially less than observed as the simulated rain-band does not extend far enough to the north and east. Given the large errors in the simulation of present-day regional climate, confidence in the simulated changes in regional climate must be low.

Models of the oceanic circulation simulate many of the observed large-scale features of ocean climate, especially in low latitudes, although their solutions are sensitive to resolution and to the parametrization of sub-grid-scale processes such as mixing and convective overturning. It is particularly important that vertical mixing in the ocean is reproduced correctly as this determines the rate at which the ocean responds to increases in greenhouse gases. This can be validated to some extent by simulating the spread of passive tracers, such as that of tritium following the atomic bomb tests in the 1950s and 1960s. Current models are capable of reproducing the main features of the spread of passive tracers (for example, Fig. 4), though there may be errors in the detailed changes (for example, in Fig. 4(b), underestimating the strength of penetration near 30–50° N). In CGCMs, it has proved necessary to add empirical adjustments to the ocean surface fluxes in order to reduce errors in the simulated climate.

2.2 Simulation of contemporary climate change and recent palaeoclimate.

AGCMs have been used to simulate the atmospheric response to prescribed anomalies in sea surface temperature (SST), and have been notably successful in simulating the response to tropical SST anomalies. Large-scale positive SST anomalies occur in eastern tropical Pacific during the occurrence of El Niño, and models have simulated successfully the associated observed anomalies in precipitation and circulation (for example, Fennessy and Shukla (1988)). Other AGCMs have reproduced the relationship between large-scale SST distribution and summer rainfall in the Sahel (for example, Folland *et al.* (1989)).

A/MLMs have reproduced some of the large-scale features of the mid-Holocene climate, only the changes in the earth's orbital parameters being specified, and also the last glacial maximum, the distribution of land ice and the prevailing CO_2 concentration being the only prescribed changes (for example, COHMAP members (1988)).

2.3 Validation summary

The most recent models are able to simulate the large-scale features but not the regional-scale (400–2000 km) details of present climate. AGCMs reproduce the

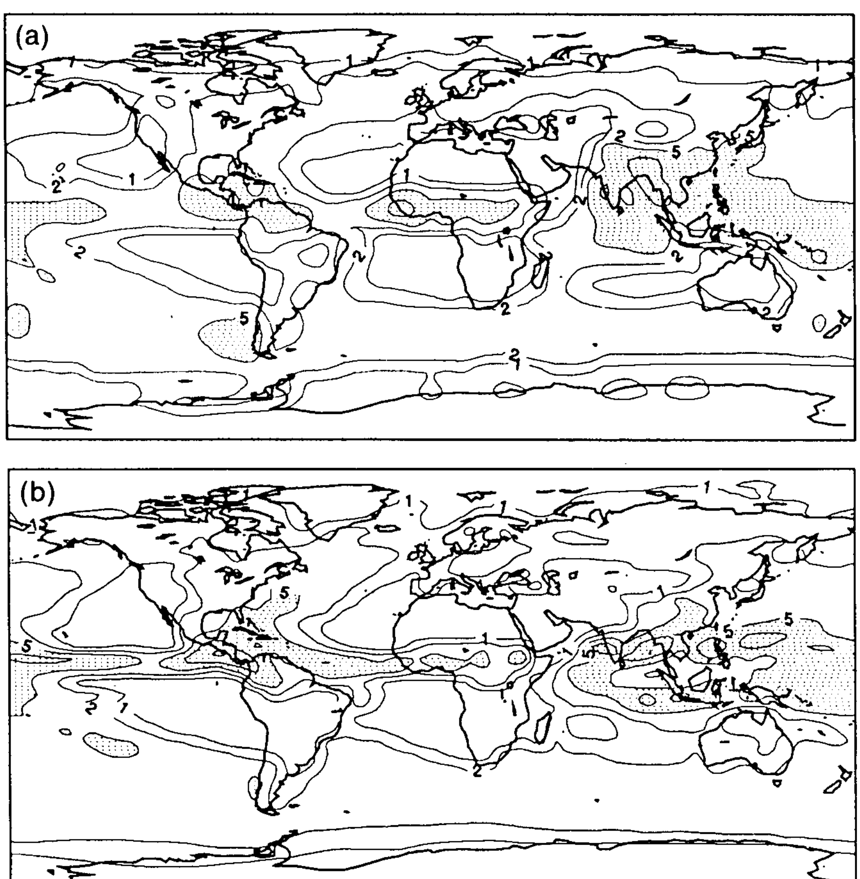

Figure 3. Preciptation (mm day^{-1}) for June, July and August, shown stippled where greater than 5 mm day^{-1}. (a) Observed (Jaeger 1976), and (b) simulated (Meteorological Office high-resolution model (2.5° × 3.75°)).

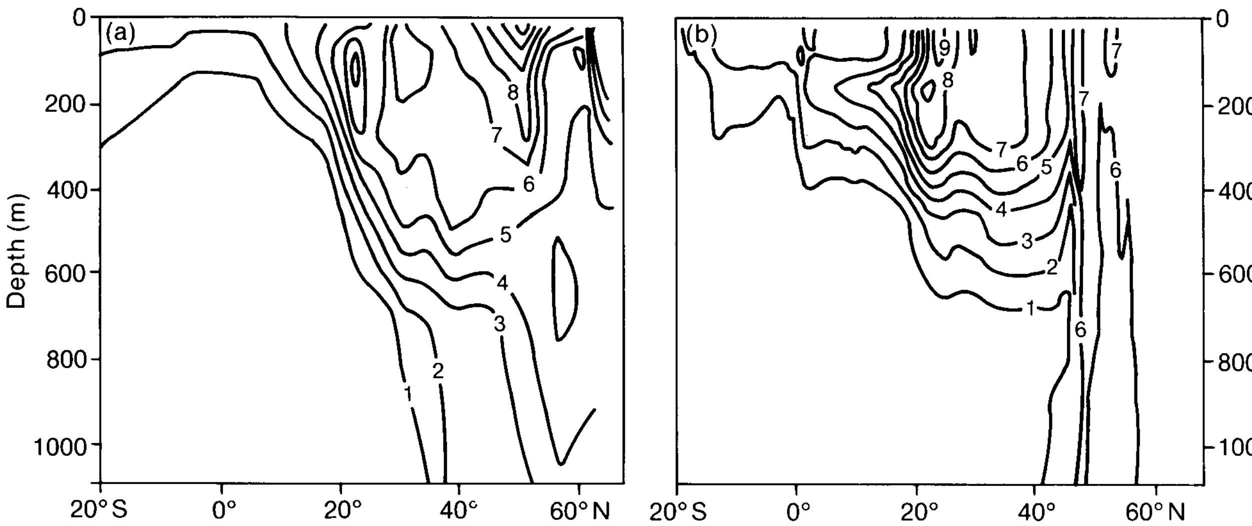

Figure 4. Tritium in the GEOSECS section in the western North Atlantic approximately 1 decade after major bomb tests. (a) GEOSECS observations, and (b) as predicted by a 12-level model in tritium units (Sarmiento 1983).

atmospheric response to contemporary SST anomalies, and A/MLMs are capable of reproducing the major changes in climate over the last 18 000 years provided certain boundary conditions (CO_2 concentrations, orbital parameters, major ice sheets) are prescribed. This, combined with the physical basis on which the models are developed, gives us some confidence in their ability to predict future changes in climate, particularly on larger scales. The least reliable aspects of models are their treatment of sub-grid-scale processes, and their ability to simulate variations in climate on a regional scale. Aspects of the simulation of present-day climate by CGCMs may be inadequate unless corrective measures, such as adjustments to the surface fluxes, are made.

3. Simulation of climate change

The simulation of climate change due to increases in greenhouse gases is now considered. Most of the understanding of greenhouse-gas-induced climate change is based on equilibrium experiments based on mixed-layer models. In the last year or so, several studies have been made of the time-dependent response to increases in greenhouse gases using CGCMs. However, results presented here will be based largely on the equilibrium experiments which have been analysed in much greater detail. Furthermore, the rates of increase in the time-dependent coupled experiments do not correspond to the IPCC scenarios of increases in greenhouse gases. Hence, in order to calculate the evolution of temperature resulting from the IPCC scenarios, we have used results from one-dimensional upwelling diffusion models calibrated using the more complex A/MLMs and CGCMs.

3.1 Equilibrium climate sensitivity

The equilibrium sensitivity of mixed-layer models to doubling CO_2 ranges from 2 to 5 °C. The range is due to uncertainties associated with sub-grid-scale parametrification, particularly those associated with cloud. Some of the more recent studies attempt to allow for changes in the microphysical properties of cloud — these more detailed, but not necessarily more accurate, models give an equilibrium sensitivity of 2–4 °C.

The observed change in temperature (1860–1990) seems to be consistent with an equilibrium warming of 1–3 °C, allowing as far as possible for natural variability and assuming other factors such as the effect of sulphate emissions on cloud albedo have not affected the warming due to increases in greenhouse gases. On the basis of modelling studies, the sensitivity of climate due to doubling CO_2 is most likely to lie between 1.5 and 4.5 °C, and, in view of the observational evidence, a 'best guess' of 2.5 °C was chosen to illustrate the IPCC scenarios.

3.2 Equilibrium and transient climate change

When the radiative heating of the earth–atmosphere system is changed by increases in greenhouse gases, the atmosphere will respond (by warming) immediately. The atmosphere is closely coupled to the oceans, so the oceans also have to be warmed; because of their thermal capacity this takes decades or centuries. This exchange of heat between the atmosphere and ocean will act to slow down the temperature rise forced by the greenhouse effect.

Consider the concentration of greenhouse gases in the atmosphere rising to a new level and remaining constant thereafter. The radiative forcing would also rise rapidly to a new level. This increased radiative forcing would cause the atmosphere and oceans to warm, and tend towards a new equilibrium temperature. Commitment to this equilibrium temperature rise would occur as soon as the greenhouse gas concentration had changed. But at any time before equilibrium is reached, the actual temperature will only have risen by part of the equilibrium temperature change — known as the realized temperature change.

One CGCM (Manabe *et al.* 1991) predicts that, for the case of a steady increase in radiative forcing similar to that currently occurring, the realized temperature at any time is about 60–80% of the committed temperature. If the forcing was to be held constant, temperatures would continue to rise slowly, but it is not certain whether it would take decades or centuries for most of the remaining rise to equilibrium to occur.

3.3 Changes in global mean temperature based on the IPCC scenarios

The following results are derived from an upwelling-diffusing model, assuming that the temperature of surface water sinking in high latitudes does not change. The upwelling velocity is taken as $4\ m\ yr^{-1}$ and the diffusivity as $0.63\ cm^2\ s^{-1}$. The IPCC 'business-as-usual' scenario, in which effective CO_2 concentrations double over pre-industrial levels by 2020, gives a warming of 1.3–2.6 °C above pre-industrial levels at 2030, corresponding to a prescribed climate sensitivity of 1.5–4.5 °C (Fig. 5). The 'best guess' sensitivity gives a warming of 1.8 °C at the time of doubling, which is reduced to about 1.5 °C in the other scenarios (Fig. 6). Scenario B assumes an effective doubling of CO_2 by 2040, and scenario C by about 2050. Note that the changes in emission scenarios take some time to produce an effect — this is a result of the slowness with which both the gas concentrations and the oceans respond — much of the warming immediately after 1990 can be attributed to the system 'catching up' the effect of previous emissions.

3.4 Patterns of climatic change due to an effective doubling of CO_2

Only one AGCM has been used in both an equilibrium experiment coupled to a mixed-layer ocean, and in a long time-dependent experiment coupled to a deep ocean model (A CGCM, Stouffer *et al.* (1989)) (Fig. 7). The equilibrium sensitivity of this model to doubling CO_2 is 4 °C. In most regions heat is mixed down to the main thermocline, at about 500 m; and the surface temperature change at the time of doubling CO_2 in the time-dependent experiment (Fig. 7(b)) is similar to the equilibrium response (Fig. 7(a)), but reduced by 20–40% (Fig. 7(c)). The exceptions are round Antarctica, where there is little or no warming in the time-dependent experiment, and over the northern North Atlantic and north-western Europe, where the reduction is 40–60%. In these regions, a deep wind-driven circulation (circumpolar ocean) or deep convection (North Atlantic) mix heating down to several kilometres. The surface warming is greatest in high northern latitudes, and north of 30° S is generally greater over land than over the ocean at the same latitudes, as found in equilibrium experiments. The warming in northern high latitudes is generally greatest around the sea-ice margins in late autumn and early winter, and around the snow-line over the continents in spring. Over the low-latitude oceans and moist areas of the low-latitude continents, the warming is small relative to the global mean.

The changes in the hydrological cycle are qualitatively similar in these equilibrium and transient experiments. Hence, the remainder of this section is based on results from equilibrium experiments. Precipitation generally increases in high latitudes throughout the year and in mid latitudes in summer (Fig. 8). Precipitation generally increases in the tropics, these changes are accompanied by shifts in the main tropical rain-bands which vary from model to model, so there is little consistency in results for a particular region.

The warming produces an increase in global mean precipitation and evaporation. Enhanced precipitation increases soil moisture levels in high northern latitudes in winter. Most models produce a drying of the land surface over the northern mid-latitude continents in summer (for example, see Fig. 9) as a result of earlier snow melt, enhanced evaporation and, in some regions, reduced precipitation.

Predicted changes in the day-to-day variability of weather are uncertain. However, simulated episodes of high temperature become more frequent simply due to

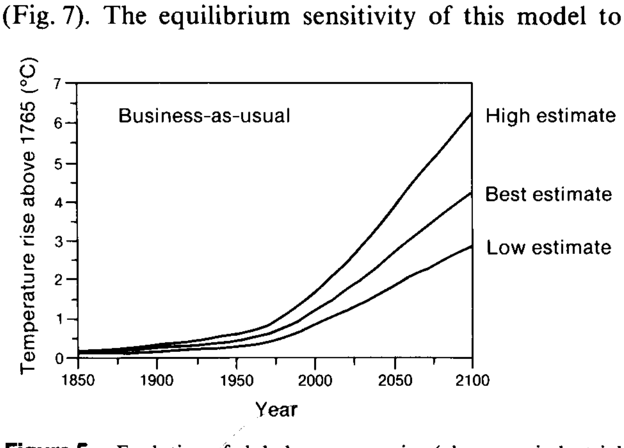

Figure 5. Evolution of global mean warming (above pre-industrial temperatures) assuming IPCC 'business-as-usual' scenario. The curves correspond to an equilibrium sensitivity to doubling CO_2 of 4.5, 2.5 and 1.5 °C, respectively (IPCC 1990).

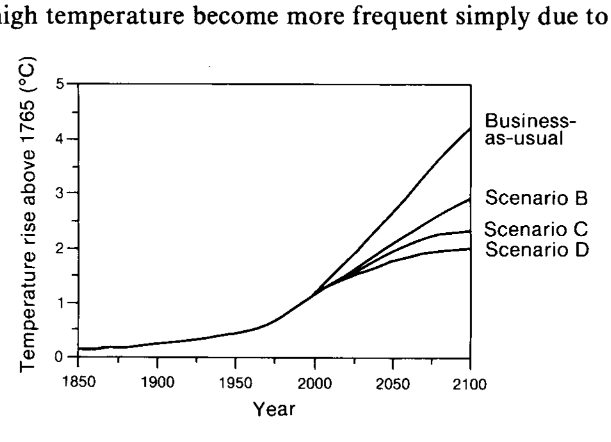

Figure 6. Evolution of global mean warming (above pre-industrial temperatures) assuming an equilibrium sensitivity to doubling CO_2 of 2.5 °C for the IPCC scenarios (IPCC 1990).

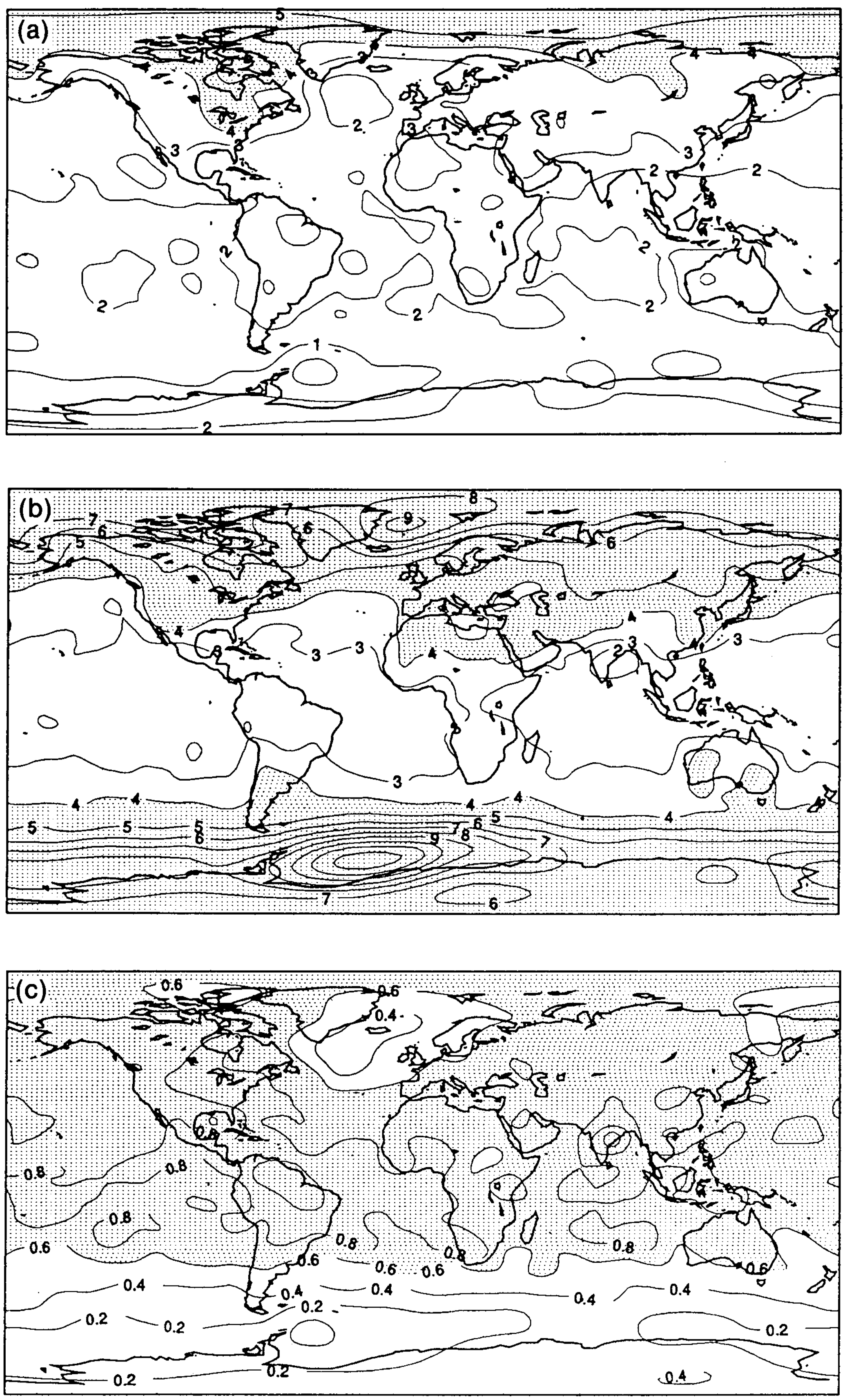

Figure 7. Geographical distribution of (a) time-dependent response of surface air temperature (°C) in the Geophysical Fluid Dynamics Laboratory coupled ocean–atmosphere model to a 1% per year increase of atmospheric CO_2. Shown is the difference between the 1% per year perturbation and the control run for the 60–80 year period when the CO_2 approximately doubles (Stouffer *et al.* 1989), (b) equilibrium response of surface air temperature (°C) in the atmosphere–mixed layer ocean model to a doubling of atmospheric CO_2 (Stouffer *et al.* 1989), and (c) ratio of time-dependent to equilibrium responses shown above (Manabe *et al.* 1991).

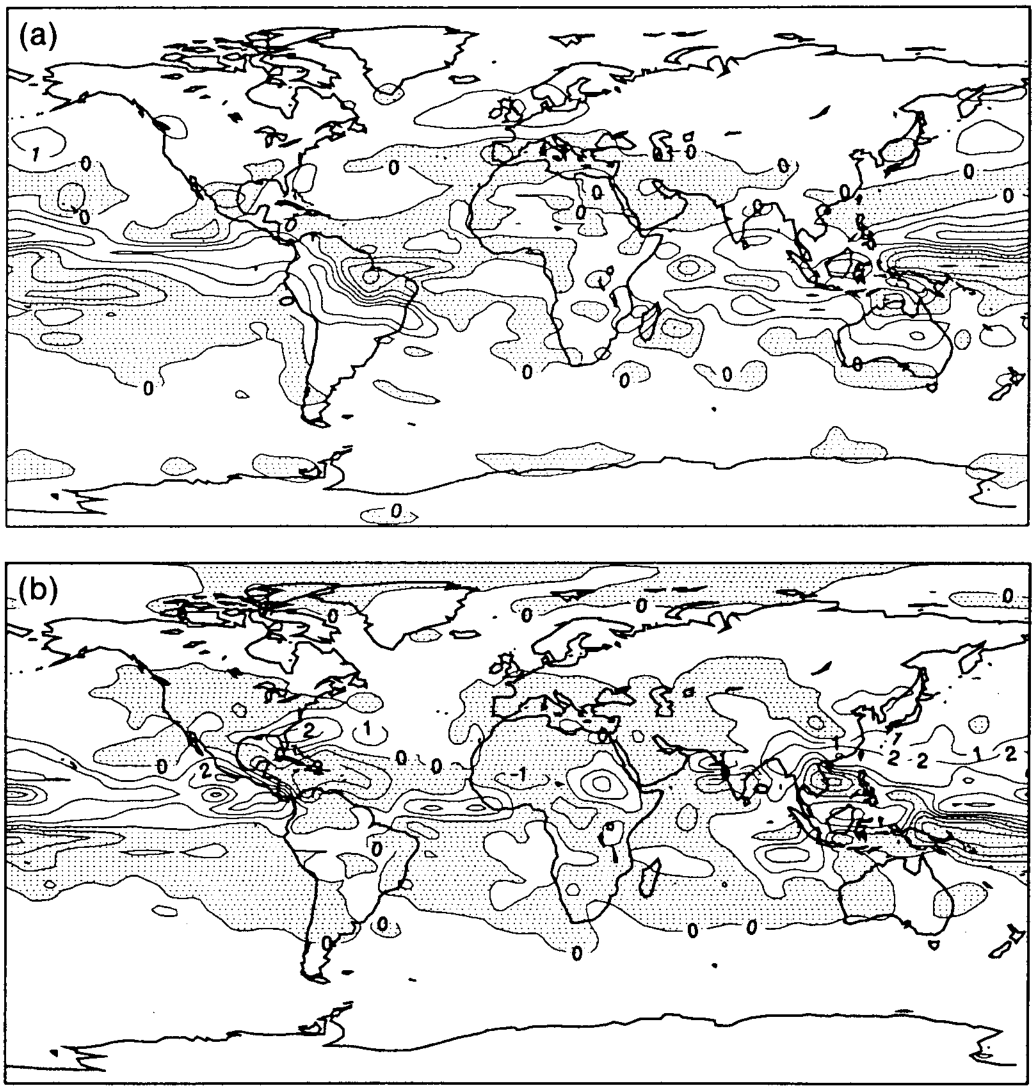

Figure 8. Simulated changes in precipitation-rate equilibrium following a doubling of atmospheric CO_2 concentrations in the Meteorological Office 2.5° × 3.75° resolution model (mm day^{-1}, contours at 0, ±1, +2 and +5 mm day^{-1}, with areas of decrease stippled) for (a) December, January and February, and (b) June, July and August.

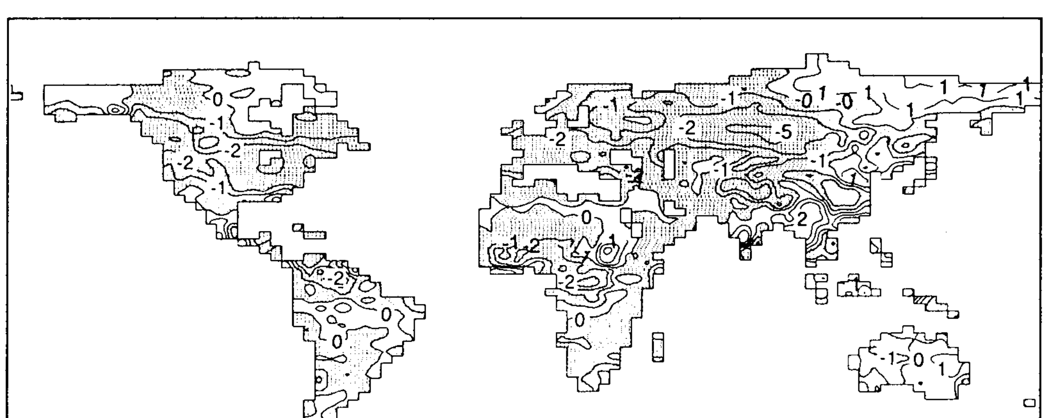

Figure 9. Simulated changes in soil moisture at equilibrium due to doubling atmospheric CO_2 in June, July and August. Contours at 1 cm intervals with areas of decrease stippled (Canadian Climate Center model with a horizontal resolution of 32 spectral waves (Boer (1990) personal communication).

the substantial increases in mean temperature. There is some evidence of an increase in convective precipitation. Numerical experiments show a reduction in mid-latitude storms, but as current models only resolve the larger disturbances, this does not rule out the possibility of smaller but more intense storms. An increase in the maximum intensity of tropical cyclones might be expected on theoretical grounds due to the increased availability of latent heat, but changes in the intensity of tropical disturbances simulated in different A/MLMs are inconsistent.

3.5 Estimating regional climate change: pre-industrial times to 2030 (IPCC 'business-as-usual' scenario)

The reader is reminded of the limited ability of current climate models to simulate regional climate change. In deriving the results below, the following assumptions have been made:

(a) The 'best guess' of the *magnitude* of the global mean equilibrium increase in surface temperature at 2030 is 1.8 °C (this is consistent with a climate sensitivity of 2.5 °C).

(b) The patterns of equilibrium and transient climate change are similar (this may be approximately true for the regions considered, but not for the high-latitude Southern Ocean, see Fig. 7).

(c) The regional changes in temperature, precipitation and soil moisture are proportional to the global mean changes in surface temperature (this will be approximately valid except in regions where the changes are associated with a shift in the position of steep gradients, for example where the snowline retreats, or on the edge of a rain-belt which is displaced).

Although it is hard to justify some of these assumptions on rigorous scientific grounds, the errors involved are substantially smaller than the uncertainties arising from the threefold range in climate sensitivity. The results below (IPCC 1990) are derived using equilibrium results from three high-resolution models (see caption to Fig. 1) and scaling them to give the appropriate change in global mean temperature. For a climate sensitivity of 1.5 °C the changes should be reduced by 30%, and for a climate sensitivity of 4.5 °C they should be increased by 50%. All the changes below are averages over the region — the range arises because of the use of different models.

3.5.1 Central North America (35–50° N, 85–105° W)

The warming ranges from 2 to 4 °C in winter and 2 to 3 °C in summer. Precipitation increases range from 0 to 15% in winter whereas there are decreases of 5–10% in summer. Soil moisture decreases in summer by 15–20%.

3.5.2 South-east Asia (5–30° N, 70–105° E)

The warming ranges from 1 to 2 °C throughout the year. Precipitation changes little in winter and generally increases throughout the region by 5–15% in summer. Summer soil moisture increases by 5–10%.

3.5.3 Sahel (10–20° N, 20° W–40° E)

The warming ranges from 1 to 2 °C. Area mean precipitation increases and area mean soil moisture decreases marginally in summer. However, there are areas of both increase and decrease in both parameters throughout the region which differ from model to model.

3.5.4 Southern Europe (35–50° N, 10° W–45° E)

The warming is about 2 °C in winter and ranges from 2 to 3 °C in summer. There is some indication of increased precipitation in winter, but summer precipitation decreases by 5–15%, and summer soil moisture by 15–25%.

3.5.5 Australia (10–45° S, 110–155° E)

The warming ranges from 1 to 2 °C in summer and is about 2 °C in winter. Summer precipitation increases by around 10%, but the models do not produce consistent estimates of the changes in soil moisture. The area averages hide large variations at the sub-continental level.

3.6 Simulation summary

Evidence from both modelling and observational studies suggest that the equilibrium global mean warming due to doubling atmospheric CO_2 is most likely to lie in the range 1.5–4.5 °C. Assuming an effective doubling of CO_2 by 2020 and an equilibrium sensitivity of 2.5 °C, simple models calibrated using the more complex GCMs predict a warming of just under 2 °C above pre-industrial levels by 2030. The warming is expected to be greatest over the higher northern latitudes in winter and least over the southern ocean throughout the year. Simulated precipitation increases in middle and high latitudes in winter, but soil moisture generally decreases in northern mid latitudes in summer. The magnitude of changes in precipitation is generally 0–15%. Little confidence can be placed in the variations in the changes on the regional scale (less than 2000 km).

4. Reducing uncertainties: current research programmes and their possible shortcomings

The assessment of recent results by IPCC gives an opportunity to review existing research programmes and to identify areas where further initiatives are required. The major sources of uncertainty in the simulation of climate change arise from:

(a) Lack of knowledge concerning the processes leading to the formation and dissipation of clouds, and determining their radiative properties. Furthermore, an improved understanding of the relevant microphysical processes needs to be matched with improved parametrization of these processes in large-scale models of climate if the current uncertainties in climate sensitivity are to be reduced. This issue is addressed to some extent by the World Climate Research Programme (WCRP) through its Global Energy and Water Cycle Experiment (GEWEX) and the ongoing International Satellite Cloud Climatology Project, though it is not obvious, for example, that clouds and related processes are the top priority in GEWEX. Thus, it may be necessary to review widely spread programmes such as GEWEX to ensure that the major sources of uncertainty are given due priority.

(b) Lack of knowledge concerning the processes leading to the vertical mixing of heat into the deep ocean, particularly in high latitudes. The World Ocean Circulation Experiment (WOCE) of the WCRP was set up to describe the ocean circulation at all depths during a 5-year period (1990–1995). The area under consideration was belatedly extended to cover the high-latitude southern oceans and does not include the Norwegian Sea. In view of the apparent importance of high-latitude oceanic mixing in recent CGCM experiments, there may be a need for additional field experiments in high latitudes to supplement WOCE. The activities of the Joint Global Flux Study (part of the International Geosphere–Biosphere Programme of the International Council of Scientific Unions) is also relevant to the problem of ocean mixing.

(c) Lack of knowledge concerning the parametrization of tropical convection, which appears to be associated with an uncertainty of a factor of two in the magnitude of the sensitivity in the tropics. This is presumably addressed as part of GEWEX, but may need to be made more explicit.

(d) Lack of knowledge of land-surface processes and the interaction between climate and ecosystems. This gives rise to uncertainties in the simulations of important hydrological characteristics of the land surface such as soil wetness and evaporation. This is addressed explicitly in GEWEX and in the International Satellite Land Surface Climatology Programme.

Other factors which limit progress are insufficient computing power and lack of scientists with experience in the relevant topics. Note that in this paper we have assumed that changes in greenhouse gases are prescribed, and so additional uncertainties due to the effect of changes in climate on greenhouse gas concentrations are ignored.

Acknowledgements

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Were the dry spells of 1988–90 worse than those in 1975–76?

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Summary

Evidence is put forward to show that over England and Wales rainfall has been decreasing, with the greatest deficit in Northumbria. Compared with 1975–76, the recent dry spells are not so severe in terms of rainfall amount but are worse in terms of effective precipitation as shown by an examination of MORECS (Meteorological Office Rainfall and Evaporation Calculation System) data for an area on the border of Kent and East Sussex. In conjunction with rainfall levels generally there is some reason to believe that this may be typical of much of eastern and southern England.

1. Introduction

'After three of the driest years ever recorded, parts of Britain are facing a serious water shortage'. One might be forgiven for thinking those words applied to the last three years, but that was the introductory sentence of the June 1965 *Geographical Magazine.* Dr R.G. Allen (1965), then Director of the Water Research Association, wrote about the problems of water demand, irrigation and rainfall deficiencies, and the anxieties being expressed about the future. Since then many parts of the United Kingdom have experienced two major dry spells, one in the mid 1970s, and the other within the last 3 years. This paper compares, in general terms, the recent dry weather with that of the previous decade, together with a closer examination of the effect of the rainfall deficit for one small area of southern England.

2. Rainfall over England and Wales since 1970

The periods of dry weather which have affected virtually all of England and Wales in the last two to three years, have revived memories of the exceptional drought of 1975–76, catalogued in the *Atlas of Drought in Britain* (Institute of British Geographers 1979), and raised further anxieties about receding water levels. In the south and east of England especially, views were expressed in newspapers, and on radio and television, that any prolongation of the dry weather beyond the summer of 1990 may lead to severe water problems in 1991. Water authorities are very concerned about declining water levels in wells and aquifers, and in some cases groundwater levels are approaching, or falling below, the minimum historical levels (Institute of Hydrology 1990). The very warm and dry weather which prevailed for much of 1989, preceded by a drier than normal winter (which was also the warmest this century), led to hose-pipe bans during the summer. Fears of worse to come were somewhat ameliorated by the very wet winter of 1989/90 (the wettest since 1914), but the return of the warm and dry weather in 1990 wiped out most of the benefits of recharged groundwater levels, and there were reports of a continued hosepipe ban in early 1991 in a few areas. Provisional totals for the winter of 1990/91 show the season's rainfall to have been just above normal, at 104% of the 1941–70 average.

Running means, calculated over a period of years, will give a smoothed guide to changes in rainfall amounts, but a more direct and immediate response to excess or shortfall in rainfall totals can be gauged by using cumulative differences from the average monthly values. Fig. 1 shows the accumulated differences, month by month, from the 1941–70 average for England and Wales. This has been used as the best available starting point since almost all the monthly values after 1970 have been based on the CARP (Comprehensive Areal Rainfall Program) technique devised by Shearman and Salter (1975). The graph quite clearly demonstrates the marked deficit which reached its lowest point in August 1976, together with a secondary minimum in the autumn of 1978. Through much of the next 10 years monthly rainfall totals were often above average but a further decline began in 1988.

The extent of the decrease in rainfall has varied across England and Wales, and the differences may be gauged by examining the areal rainfall for each of the water company regions in England and Wales. Fig. 2 shows the total rainfall from January 1971 to September 1990, expressed as a percentage of the 1941–70 average. Each region shows a deficit, and is most marked in Northumbria. The decline in rainfall amounts, if it is sustained or becomes more marked, could have serious consequences should the general demand for water continue to increase.

3. Comparison of recent rainfall totals with the 1970s

The drought of the 1970s reached its nadir in August 1976, for the following months of September and

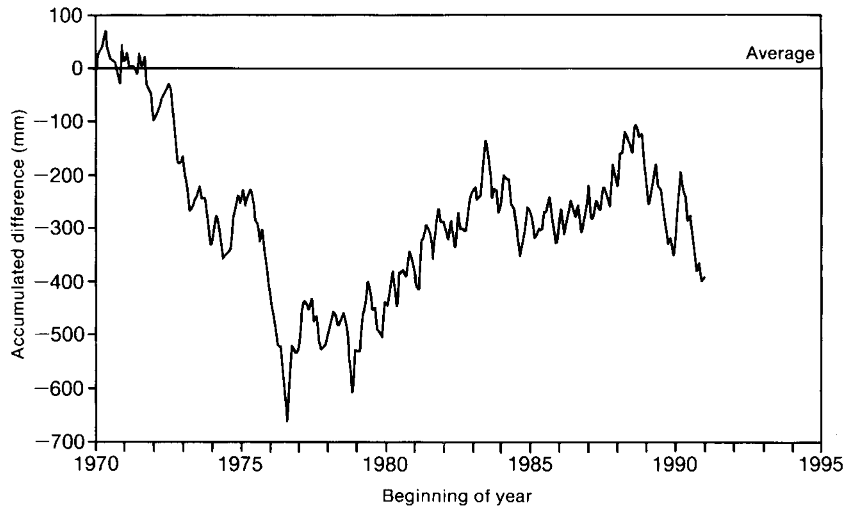

Figure 1. Cumulative differences from the 1941–70 average of monthly rainfall since 1970 over England and Wales. Values since September 1990 are still provisional.

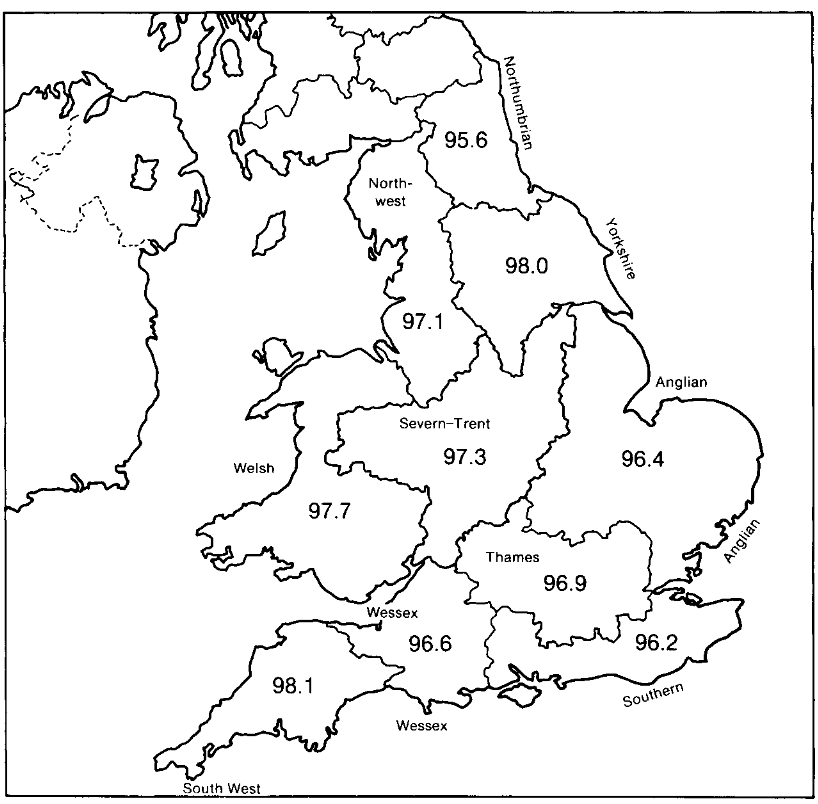

Figure 2. Percentage of the 1941–70 average rainfall over the water company regions from January 1971 to September 1990.

October were very wet, both receiving nearly twice the normal amount. Fig. 3 shows the percentage of 1941–70 rainfall over the period May 1975 to August 1976 (inclusive) together with the calculated return period, for a rainfall event specifically starting in May. It should be noted, however, that wet or dry spells can start in any month: thus there are 12 times as many opportunities and the return periods given in Figs 3–5 can be divided by 12 to give an approximate assessment of events starting at any time. In the more recent dry spell the equivalent period is from August 1988 to November 1989, inclusive: Fig. 4 shows the values for that 16-month span, and Fig. 5 values for the longer period of 26 months from August 1988 to September 1990. This last period includes the exceptionally wet winter of 1989/90.

4. The effect on ground moisture levels

The effect of a dry spell on both the available water content (AWC) for plants and crops, and general water supplies may be serious enough in the short term, but potentially disastrous if the drought is prolonged. The

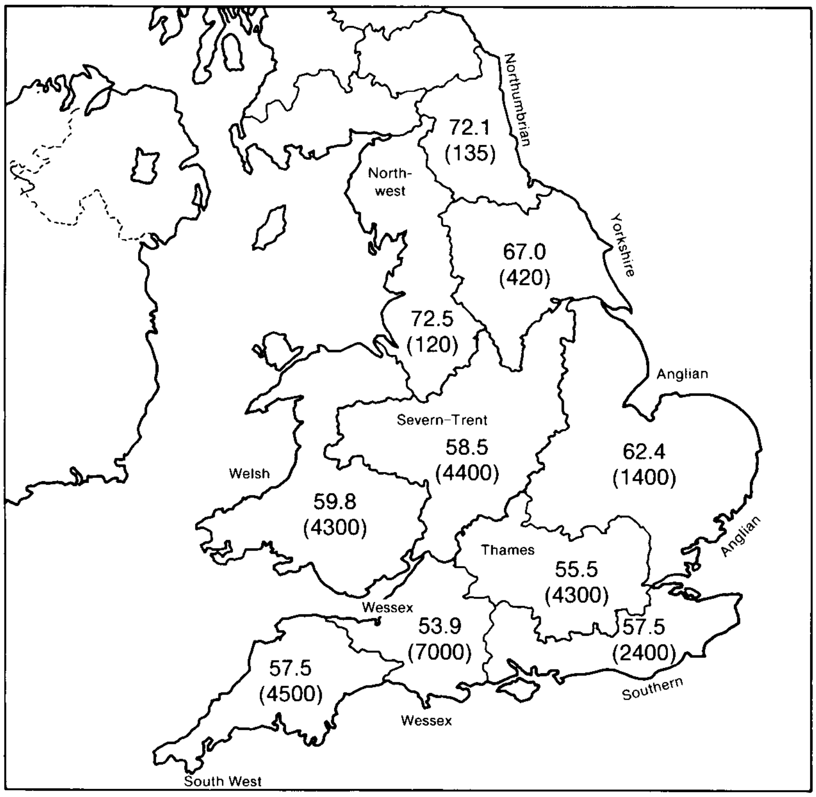

Figure 3. As Fig. 2 but for the period from May 1975 to August 1976. The associated return period in years is stated within the brackets.

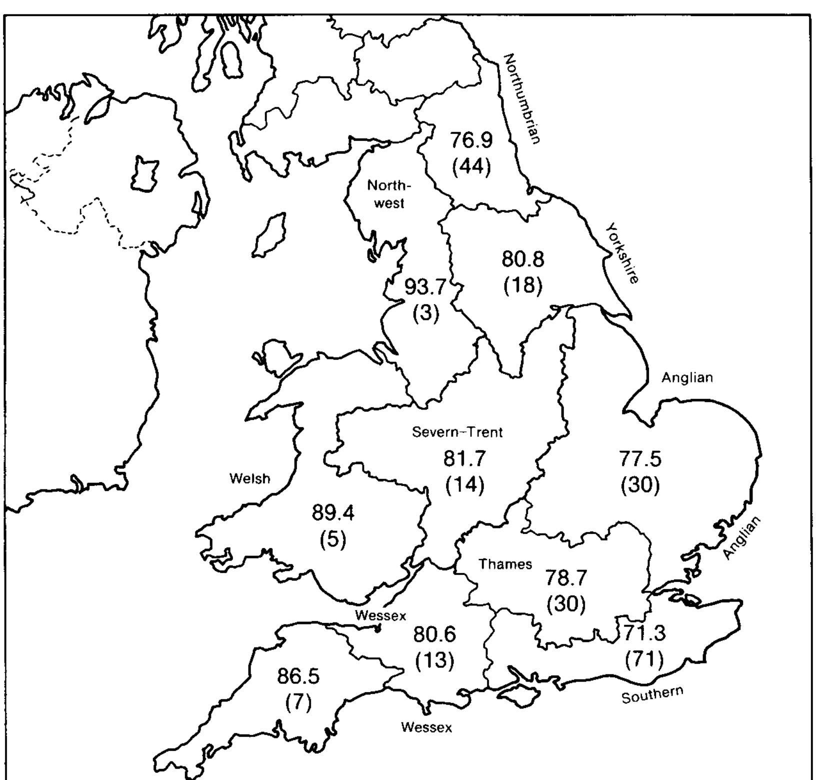

Figure 4. As Fig. 3 but for the period from August 1988 to November 1989.

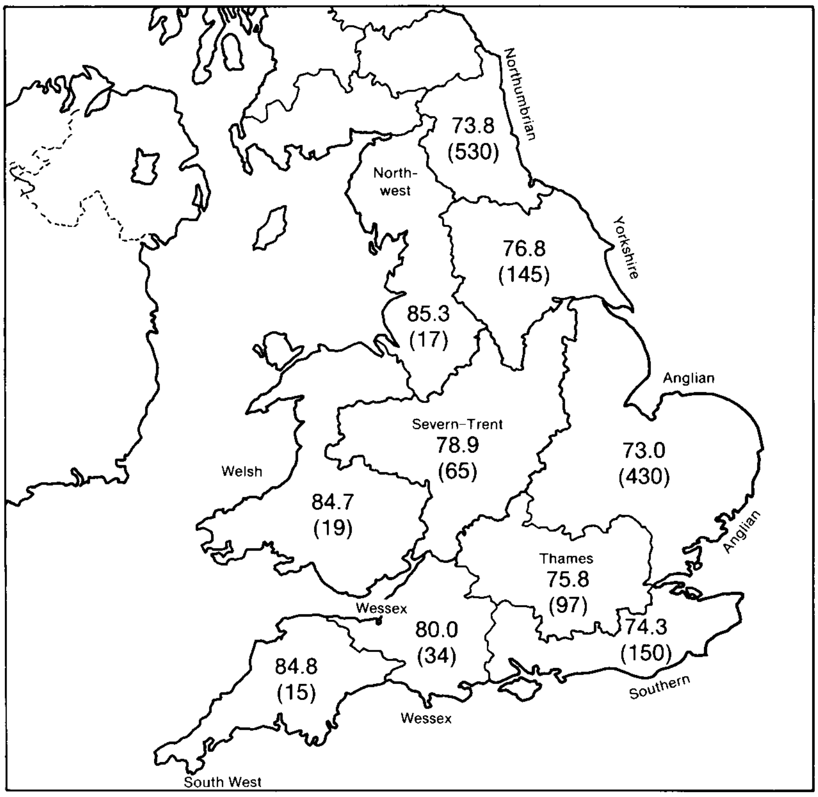

Figure 5. As Fig. 3 but for the period from August 1988 to September 1990.

dry periods during the 1970s, coupled at times with the exceptional warmth of 1975 and 1976 caused widespread concern. Since both 1989 and 1990 have shown similar characteristics, it is instructive to compare these years with those of 1975–76 in terms of effective precipitation and soil moisture deficits. A study covering the whole of England and Wales is outside the scope of this paper, but a comparison over a small area has been undertaken.

The MORECS (Meteorological Office Rainfall and Evaporation Calculation System) Bulletin has been designed to provide estimates of weekly and monthly values in the form of data within a grid of 40-kilometre squares. An account of the physical processes can be found in Hydrological Memorandum No. 45 (Thompson *et al.* 1981). The core of the scheme is a system which determines the evapotranspiration from various crop and land surfaces. It is broadly defined as a combining of the processes of evaporation from the surface of the earth and transpiration from vegetation, and can be calculated by taking into account a number of factors including sunshine, wind speed, temperature and vapour pressure as well as the state and height above ground of the crops. Once the values of these two processes have been assessed the soil moisture deficit (SMD) can be evaluated by the differences between the rainfall received and the evaporation; and the effective precipitation (EP) is the excess left after the ground has absorbed the precipitation, i.e. run-off from the ground into rivers, aquifers, etc.

The variables SMD and EP, together with the rainfall amount are important indicators of conditions for farmers, growers, hydrologists, and many others. They show when irrigation may be necessary (as SMD values become relatively large), and if regulatory measures by water companies are necessary when water levels in boreholes and aquifers fall below acceptable levels during long periods with negligible or no EP. In the summer months the amount of evapotranspiration from surfaces exceeds the average rainfall over much of Great Britain, so that the community is dependent on the land reaching its maximum water capacity in winter months, and reservoirs and underground water resources are replenished.

In the Southern water region the rainfall for the 6 months from August to January provides, on average, about 60% of the annual average (based on the 1941–70 average), and around one-third of the annual average rainfall occurs in the three months from November to January. This pattern appears to be repeated in the other water regions of England and Wales. The latter period is of great importance, for evapotranspiration and soil moisture deficit are at or close to the minimum, and the contribution by effective precipitation at its highest. A dry winter can lead to a serious water shortage if the following summer is dry, which indeed proved to be the case in 1988–89. Tables I–III, for grassland use, show the end of month values of rainfall, SMD and EP, together with the 1961–85 long-period average values, for MORECS grid square 173, which

Table I. Comparison of MORECS values (mm) for grid-square 173 — rainfall

Month	1961–85 average	SD	Rainfall 1975	1976	1988	1989	1990	1991
Jan.	73	34	142	19	186	27	107	90
Feb.	50	30	24	31	47	53	116	37
Mar.	61	31	108	18	87	69	4	38
Apr.	51	28	45	11	36	94	51	
May	58	24	79	23	48	2	8	
Jun.	57	32	15	9	14	57	58	
Jul.	47	21	23	42	79	33	11	
Aug.	57	29	28	10	42	29	36	
Sep.	73	51	144	127	47	29	34	
Oct.	71	49	45	141	73	75	126	
Nov.	83	43	82	145	34	39	74	
Dec.	82	34	39	84	17	126	61	

Notes:
SD = Standard deviation.
In the current MORECS model SMD for grass cannot exceed 125 mm.

Table II. Comparison of MORECS values (mm) for grid-square 173 — soil moisture deficit (for grass)

Month	1961–85 average	SD	Soil moisture deficit 1975	1976	1988	1989	1990	1991
Jan.	7	1	0	7	0	23	0	2
Feb.	2	2	5	1	9	1	4	0
Mar.	7	8	1	17	1	13	4	13
Apr.	20	18	23	60	18	4	50	
May	42	27	36	100	55	92	108	
Jun.	60	30	98	124	91	107	108	
Jul.	80	26	118	125	95	122	124	
Aug.	79	36	125	124	100	125	109	
Sep.	56	39	18	37	103	125	109	
Oct.	36	37	6	0	63	77	21	
Nov.	11	23	0	0	41	54	1	
Dec.	1	1	0	0	38	2	0	

Notes: as Table I

Table III. Comparison of MORECS values (mm) for grid-square 173 — effective precipitation (for grass)

Month	1961–85 average	SD	Soil moisture deficit 1975	1976	1988	1989	1990	1991
Jan.	62	32	124	9	170	0	89	76
Feb.	35	29	17	13	33	11	88	23
Mar.	35	27	79	0	39	41	0	16
Apr.	14	18	19	0	0	34	0	
May	7	11	20	0	0	0	0	
Jun.	5	14	0	0	0	0	0	
Jul.	0	0	0	0	0	0	0	
Aug.	0	>0	0	0	0	0	0	
Sep.	10	25	0	0	0	0	0	
Oct.	25	37	6	76	0	0	0	
Nov.	42	45	63	134	0	0	36	
Dec.	63	30	32	79	0	58	45	

Notes: as Table I

covers most of the Kent and East Sussex border area, for the years 1975–76 and 1988–91. In particular Table II shows that while the SMD levels were at or near the theoretical possible maximum of 125 millimetres in 1975, 1976, 1989 and 1990, the last of these years showed the longest period of deficit in excess of 100 millimetres. Table III reveals that the number of months with nil effective precipitation in 1988, 1989 and 1990 was greater than in either 1975 or 1976.

5. Conclusion.

Figs 1 and 2 show that rainfall has declined over England and Wales as a whole over the last 20 years. Comparison of the areal rainfall and the return periods (Figs 3–5) show that the dry spell of 1975–76 was quite exceptional, and neither the 16-month period in 1988–89 (Fig. 4) nor the 26-month period 1988–90 (Fig. 5) are anything like so severe. It is noticeable, however, that over the longer of the two time-spans, the return periods have become much greater, and notably so in Northumbria. From the point of view of Table III the effect of the 1988–90 dry spells, and particularly the long, dry period in 1990, appears to have had a greater effect on underground water reserves than in 1975–76. Two of the three winter months 1990/91 had below normal effective precipitation, and March 1991 was also less than average*. The water companies may have good reason to be concerned about future supplies in the face of these facts, especially as the 1990/91 winter rainfall for England and Wales as a whole is below normal.

* This article was written in the early spring of 1991. Since then there has been one very dry month (May) and one very wet month (June) over almost all of England and Wales. Both months were cooler than normal.

Thus it must be concluded that for at least one small area of England the precipitation contribution to water storage systems and aquifers is markedly less than normal and apparently less than in 1975–76. If this pattern has been repeated across much of southern and eastern England (as supported by Fig. 5), then the anxieties expressed by water and river authorities appear fully justified, and even Wales and western districts of England may have cause for concern. Though the provisional winter rainfall figure of 228 mm, for England and Wales, is 4% below the 1941–70 average, it is close enough to normal to warrant the need for above-average rainfall (and below-average evapotranspiration) in the following months if the situation is to be ameliorated.

Acknowledgements

Thanks are due to D. Hollis for computing the rainfall return periods, J. Fullwood for advice on aspects of the MORECS system, and Dr J. Brownscombe for helpful discussions about precipitation statistics as well as the calculations for Fig. 1.

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Correspondence

Discussion on the review of *Pilot's weatherpack*

I am very grateful to have B.K. Lloyd's fair review of *Pilot's weatherpack* (*Meteorol Mag*, **120**, 149–150). Perhaps it should be made clear that the Case 1 analysis was principally the work of S.G. Cornford (now retired), whereas all subsequent accidents were investigated later by W.S. Pike, who appreciates the valid point made regarding inclusion of a case featuring carburettor icing. R. Reynolds kindly organized the pack's printing.

Aircraft piston-engine induction-system icing (commonly known as carburettor icing) is caused by internal cooling within the fuel system where a sudden drop in temperature (of up to 30 °C) occurs at the carburettor venturi, due to fuel vaporization and air expansion there. Although this is a gradual process which causes a progressive decline in engine power, it is often described to pilots as being 'stealthy', because sometimes the first sign of this mechanical problem developing is when the engine(s) stutter or fail completely in flight. Commonly, the pilot has overlooked sufficient remedial application of carburettor heat, available at the flick of a switch in most modern aircraft, but at the simultaneous cost of increased fuel consumption.

Accidents result if the ice cannot be melted before a forced landing has to be made. Damaging occurrences are also reported due to the inability of the engine(s) to develop sufficient power for a successful take-off to be achieved. Of the 10–15 incidents where carburettor icing is suspected each year, only half of these result in accident reports to the Air Accident Investigations Branch (AAIB) of the Department of Transport, and fatalities are, also, quite rare — occurring approximately once every three years on average.

Frequently, other factors are cited as contributing to an accident (e.g. airframe icing, fuel contamination, mechanical failure, etc.) and the AAIB Bulletin or Report will mention that 'conditions conducive to severe carburettor icing' were in existence. Often this is because the hard evidence rapidly disappears following an accident! Water found in a fuel system could be either pre-existing contamination or melted ice.

A more comprehensive guide to the various forms of carburettor icing can be found in the CAA's *Aeronautical Information Circular* (AIC No. 59/1990 (Pink 8)). Carburettor icing is more likely to occur with MOGAS fuel (because of its greater volatility and higher possible water content) than when AVGAS fuel is used. Also, some types of aircraft appear to be more prone than others.

Above all it should be mentioned that carburettor icing is not just a winter phenomenon, but it is even more likely to appear on warm days in conditions of high humidity. An interesting diagram in the abovementioned AICs (Fig. 1) relates the risk of carburettor icing to (1) engine power setting, (2) temperature, and (3) humidity/dew-point temperature. It is noticeable that at 'low descent-power' a serious risk of carburettor icing exists for ambient temperatures as high as 32 °C when combined with dew-point temperatures of between 16 and 20 °C. Probabilities of carburettor icing also increase whenever aircraft are operating in or near cloud, fog and precipitation.

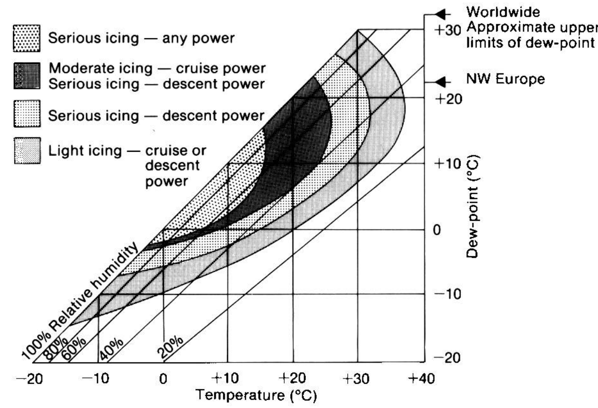

Figure 1. Carburettor icing in air free of cloud, fog or precipitation. The risk and rate of carburettor icing will be greatest when operating in cloud, fog and precipitation.

If we have the opportunity to revise the *Pilot's weatherpack* (now titled *Weather as a hazard*) to include an accident featuring carburettor icing in interesting or unusual circumstances, we will do so.

W.S. Pike

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Reply by B.K. Lloyd

Thank you for giving me the opportunity to reply to Mr W.S. Pike's letter.

I agree with all that Mr Pike has said, and to identify an incident totally attributable to carburettor icing would be difficult. However, I was particularly pleased with his final paragraph. An article featuring carburettor icing should be included if he were to have the opportunity to revise the *Pilot's weatherpack* (now titled *Weather as a hazard*).

May I point out that the AICs that he has quoted have now been reissued under AIC 59/1990 (Pink 8) dated 2 August. Another source of information is the *General aviation safety sense* series, particularly leaflet Nos 3a (winter flying) and 14 (piston engine icing) which may be of some use. These leaflets are prepared by the Safety Promotions Section and the Public Relations Department of the Civil Aviation Authority.

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Review

Land surface evaporation: measurement and parameterization, edited by T.J. Schmugge, J.-C. André. 179 mm × 260 mm, pp. xv+424, *illus.* New York, Springer Verlag, 1991. Price Dm.168.00. ISBN 0 387 97359 1.

It is by now well known that the atmosphere responds to variations in its surface boundary conditions. Several model experiments in the last two decades have shown that this sensitivity occurs on a number of spatial scales, ranging from that of a mesoscale model grid-square of some hundred square kilometres to the general circulation grid-square size of several thousands of square kilometres. This book is the result of a workshop held in October 1988 in Banyuls and aims to present a review of land surface evaporation processes from both a modelling and measurement point of view. The combination of these two aspects in a single book make it stand out from other workshop proceedings. In effect, the title is slightly misleading as the various contributions deal not only with evaporation but cover many more aspects of the land surface energy and water balance. The book draws heavily on examples from the HAPEX-Mobilhy experiment which took place in an area of 100 km × 100 km in south-western France in 1986. The principal focus of this experiment was to measure the model hydrological and atmospheric fluxes on the scale of a general circulation grid-square.

The first chapters present the reader with a state-of-the-art review on the modelling of land surface processes in a variety of atmospheric models: global climate models, numerical weather prediction models and mesoscale models. They do not only describe the physical background of the parametrizations but also show how sensitive the models are to these parametrizations. In the case of climate modelling, examples are shown of the use of these models to predict the consequences of Amazonian deforestation. The two chapters on numerical weather prediction models and mesoscale models show how model results are critically dependent on the initialization of the land surface parametrizations. Unfortunately there is considerable overlap and repetition in the content of these chapters as all present in some detail the respective land surface parametrizations used in these models. A particularly relevant problem raised in these chapters is how the ultimate products of large-scale meteorological models, fluxes representative of areas of several hundreds of square **kilometres**, are to be tested. Current practice, and several examples of this are given in this book, is to test the one-dimensional representation against local flux data obtained over several hundreds of **metres**. Although this is an obvious necessary requirement for a parametrization scheme to be acceptable, it clearly is not a sufficient one. A dramatic illustration of this problem is given in the chapter on climate modelling where in one case the prediction of run-off agrees very well with observations and in the other case not at all. Why does the small-scale physics appear to work on the large scale in the first one, but not in the second?

The section on measurement describes in detail the various measurement techniques currently available to measure the various components of the energy balance. An important aspect of these chapters is that they cover a range of scales from micrometeorological measurements, using profile and eddy correlation methods, to flux measurements from aircraft, which integrate over several kilometres, to remote sensing measurements. The section provides insight into the effects of land surface variability on the structure of the surface layer and the use of aircraft to estimate fluxes over inhomogeneous or inaccessible terrain. Although aircraft have considerable potential in estimating area-averaged fluxes, it is pleasing to see the aircraft work described in a realistic way, so that future planning of large-scale field experiments can draw from the experiences of earlier experiments.

Overall the book presents a valuable overview of current knowledge of land surface processes and makes, through its reference to the HAPEX-Mobilhy experiment, a case for well planned large-scale international field experiments. Any researcher interested in hydrology and meteorology, from plot or catchment to continental or global scale, will not be disappointed by purchasing this well produced book.

A.J. Dolman

Books received

The listing of books under this heading does not preclude a review in the Meteorological Magazine at a later date.

The earth as transformed by human action, edited by B.L. Turner II, W.C. Clark, R.W. Kates, J.F. Richards, J.T. Mathews and W.B. Mayer (Cambridge University Press, 1991. £75.00) is the culmination of an undertaking involving the examination of the toll taken on the world by the human race while taking technical and social strides forward. The papers included are aimed at a broad audience. ISBN 0 521 36357 8.

Global environmental change, edited by R.W. Corell and P.A. Anderson (Berlin, Heidelberg, New York, London, Paris, Tokyo, Hong Kong, Springer-Verlag, 1991. DM 158.00) contains papers presented at a NATO workshop on the subject. The workshop's three central themes were the needs to: develop the global observation and monitoring systems, better understand basic Earth system processes and improve models of global change. ISBN 0 387 53128 9.

Satellite and radar photograph — 2 July 1991 at 1700 UTC

During the afternoon of 2 July 1991 a slow-moving front extended from the north-west tip of Denmark to eastern Spain (see Fig. 1). On the COST-73 (Co-operation in Science and Technology) image shown in Fig. 2 this front can be seen as a band of cloud and patchy rain; just east of the front there there is an area of heavy convective rain (marked A in Fig. 2) with intensities of over 30 mm h^{-1} in a few places, and a large area of medium and high cloud spreading north and east from it. This system formed shortly before 1500 UTC when several intense convective cells started to develop over southern Belgium in the warm air near a shallow wave on the front. The cells moved northwards and became organized into a band running inland from the coast; an area of high-level outflow cloud spread north-east from them over the next few hours. By 1700 UTC (Fig. 2) the cells had organized into a mesoscale convective system (MCS) with a band of convective rain over 100 km long, new developments spreading north from its north-western end, and with the anvil of medium- and high-level outflow cloud covering more than 25 000 km^2. To the north-east of the outflow cloud a small thunderstorm (marked B in Fig. 2) can be seen over The Netherlands with its own upper-level outflow cloud spreading away from it.

The MCS continued to expand and develop over the next few hours as it moved over the North Sea. The intensity of precipitation gradually decreased, but the area covered by it increased as lighter rain started to fall from the outflow cloud. By 2100 UTC the MCS had overtaken and engulfed the smaller thunderstorm (B in Fig. 2) and the combined area of medium- and high-level outflow cloud covered around 125 000 km^2. After this time the system started to shrink and become less organized, finally dissipating in the early hours of 3 July.

R.B.E. Lilley

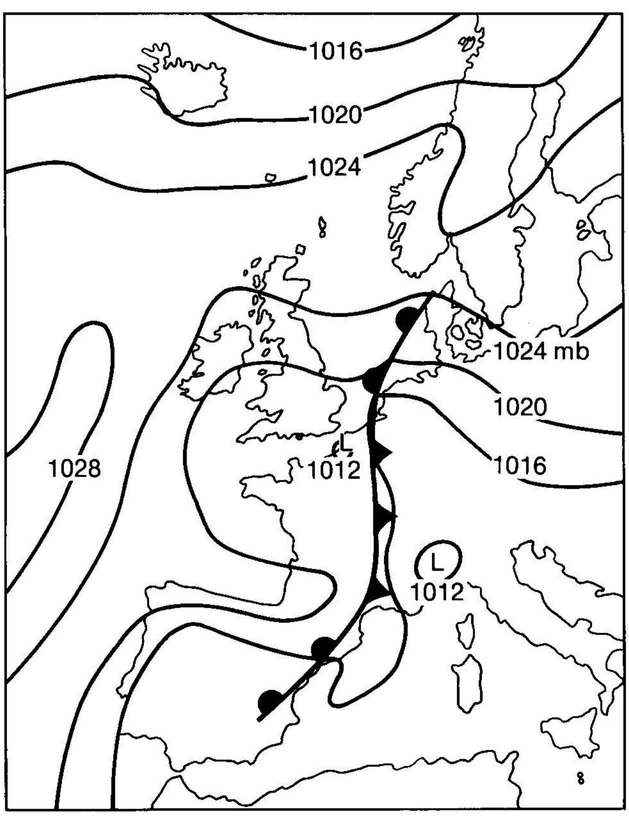

Figure 1. Surface analysis for 1800 UTC on 2 July 1991.

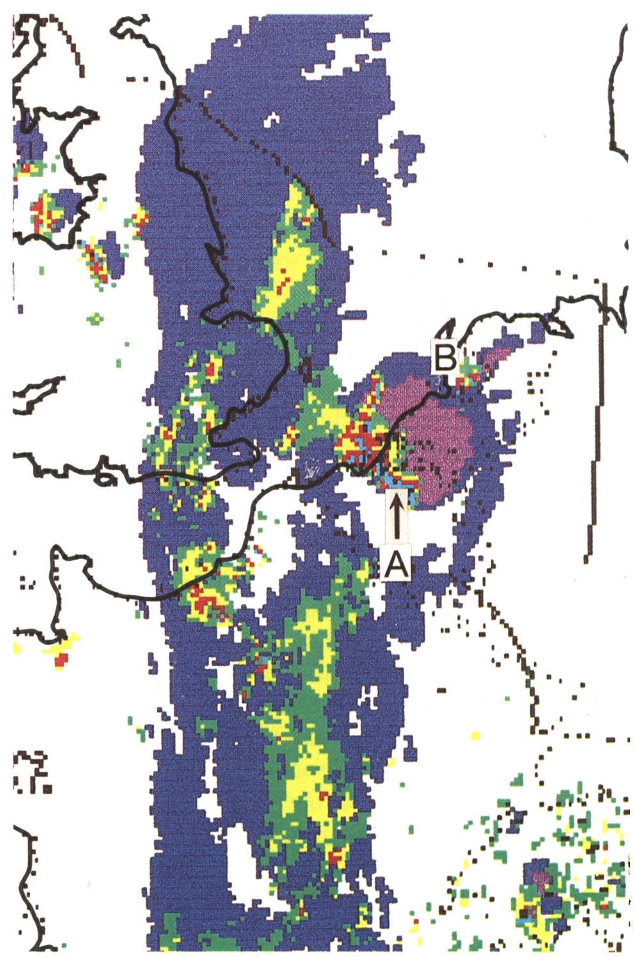

Figure 2. COST-73 European radar and satellite composite picture for 1700 UTC on 2 July 1991. Cloud is shown from the Meteosat infra-red data, with dark blue indicating cloud-top temperatures from −15 °C to −45 °C, and purple indicating tops colder than −45 °C. Rainfall rates (mm h^{-1}) shown are: green (0.3–1), yellow (1–3), red (3–10), light blue (10–30) and black (> 30). See text for explanation of lettering.

GUIDE TO AUTHORS

Content

Articles on all aspects of meteorology are welcomed, particularly those which describe results of research in applied meteorology or the development of practical forecasting techniques.

Preparation and submission of articles

Articles, which must be in English, should be typed, double-spaced with wide margins, on one side only of A4-size paper. Tables, references and figure captions should be typed separately. Spelling should conform to the preferred spelling in the *Concise Oxford Dictionary* (latest edition). Articles prepared on floppy disk (Compucorp or IBM-compatible) can be labour-saving, but only a print-out should be submitted in the first instance.

References should be made using the Harvard system (author/date) and full details should be given at the end of the text. If a document is unpublished, details must be given of the library where it may be seen. Documents which are not available to enquirers must not be referred to, except by 'personal communication'.

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