Simulation of the atmospheric response to soil moisture anomalies over Europe*

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SUMMARY

Positive and negative anomalies in the initial soil moisture over central Europe were introduced in turn into integrations of a global model of the July atmosphere. The results show that by modifying the partitioning of the turbulent fluxes such anomalies can have major effects on the modelled rainfall, humidity and temperature during the following 50 days over the anomaly area and that the anomalies can propagate into adjacent land areas. Similar results were obtained with a lower-resolution version of the model.

The effects on rainfall were already important on day 3 of the integration over the anomaly area and on day 6 over parts of Scandinavia outside the anomaly area.

The experiment was repeated with a version of the model in which the prevailing flow over Europe was less weak with increased moist westerly flow from the Atlantic. In this case the dry anomaly persisted for only about 20 days before it became too weak to affect the evaporation significantly.

1. INTRODUCTION

The moisture budget of the atmosphere over a region depends on three components:
(a) the net flux of water vapour and liquid water across the vertical boundaries of the region;
(b) the net upward flux of water vapour from the underlying surface (evaporation less condensation);
(c) the precipitation of rain and snow from the atmosphere to the surface.

In this paper we consider the importance of the second component (mainly evaporation) in determining the third (precipitation).

Mintz (1981) has noted that studies of the world water balance (Korzun 1974; Baumgartner and Reichel 1975) show that about 60% of the rainfall over Eurasia is re-evaporated, and he argued from an analysis of the water budget of the eastern and central USA that much of the continental rainfall is due to evaporation from the surface of the continent - partly because this increases the water vapour available for precipitation and partly because it provides the energy needed to initiate convection.

These studies imply that understanding and simulation of seasonal and longer term variations in rainfall over land require a realistic representation of land surface hydrological processes. For example changes in the availability of surface moisture may be partly responsible both for interannual variations in the poleward spread of rainfall over the African land mass in summer and for the long period variations in Saharan rainfall over the last 10000 years evident in the palaeoclimatological record. They also imply that it may be feasible for rainfall to be modified either naturally or artificially through changes in irrigation, land drainage or vegetative cover - a change from shallow to deep rooted vegetation may allow a more continuous supply of water to the atmosphere.

General circulation experiments made in recent years have generally confirmed the importance of evaporation over land for the precipitation process. Walker and Rowntree (1977) carried out two experiments using a tropical model with idealized geography in which the land surface over an 18° wide latitude belt was initially prescribed as dry and wet respectively. They found that the absence of evaporation in the dry case almost eliminated rainfall over the anomaly area; however, in the wet case the initially dry atmosphere over the anomaly area soon became moist allowing rainfall which, after about 10 days, balanced evaporation and so maintained the soil in a wet state. In a subsequent

* Some of the results of this paper were presented at the symposium on variations in the global water budget held at Oxford in August 1981, and a short version appears in the proceedings of that conference.
series of experiments, W. M. Cunningham (personal communication) used a global version of this model with realistic topography. The results confirmed the sensitivity of rainfall over the Saharan region to anomalies in the initial availability of moisture, both in the soil and in the atmosphere. Initial anomalies persisted for at least 50 days, partly due to a positive feedback in the moisture convergence.

Mintz (1981) reported recent experiments on the effects of global-scale soil moisture anomalies. He described three rather similar experiments which had been made by Suarez and Arakawa, Shukla and Mintz and Carson and Sangster in which global models were run for northern summer conditions with land either all dry or all wet. Over much of the land, precipitation in the dry cases was found to be considerably reduced by absence of evaporation. Even the southern (winter) continents equatorward of 40°S were affected. Shukla and Mintz obtained increased rainfall over parts of southern Asia but this was not evident in the other experiments. Only in Carson and Sangster’s experiment was the soil moisture allowed to vary with time. The results are instructive although it must be remembered that perpetual July conditions (radiation, sea surface temperature, etc.) were applied. By days 21–50 from the start of the dry integration there was substantial rainfall over most of the rainy regions of the tropics, though generally less than in the wet case. However, in middle latitudes there was still much less rainfall except locally near rainy eastern coasts. Even after 250 days there were still substantial land areas in the middle latitudes and the sub-tropics which were receiving more rain in the initially wet case.

The experiments discussed above were designed to investigate the effects of either rather large-scale anomalies in the summer tropics and sub-tropics or of global-scale anomalies. In this paper we report the results of experiments made to assess the response to an anomaly of relatively small horizontal scale in middle latitudes in order to approach the question of how large an anomaly needs to be to have a significant effect. This is an important question because most observed anomalies on the interannual time scale are likely to be on rather a small spatial scale.

(a) Effect of soil moisture on atmospheric relative humidity

Before considering the numerical experiments and their results it is useful at this point to consider the direct effect to be expected from a soil moisture anomaly on air flowing over it. This should allow an estimate of the horizontal scale of anomaly likely to be needed to affect precipitation. If the relative humidity of the air approaches 100%, precipitation is to be expected so we are interested in the effect on a measure of relative humidity, which we define as the ratio of the vertically integrated atmospheric water vapour Q (g cm⁻²) to the corresponding saturation vapour content Qs. This is increased by an increase in Q due to an excess of evaporation or evapotranspiration, E, over precipitation, P, and decreased by a rise in temperature, T, which increases Qs. Rowntree (to be published) discusses how the time-change of this mean relative humidity can be estimated in the absence of mean ascent or descent. For present purposes we take the column of the atmosphere, within which turbulent and convective fluxes and latent heat release are confined, to be the lower 500 mb and additionally assume that the net radiative cooling of this column of air balances the net radiative heating of the surface. Then taking middle latitude summer values of Q, 2.3 g cm⁻², and Qs, 3 g cm⁻², which are appropriate for a rather moist air mass liable to give rain if the relative humidity is increased, Rowntree shows that the change in mean relative humidity is about 0.06(E - P)d⁻¹ with (E - P) in mm d⁻¹. Slightly more than half of this (0.035(E - P)) is due to the change in Q, the rest to the change in temperature altering Qs. For moist ground in western Europe in summer, evapotranspiration, E, is about 3 mm d⁻¹ whereas for dry ground quite low evaporation rates have been observed, for example less than 1 mm d⁻¹ in eastern England in the summer of 1976 (Richards 1979). Such a modification of the partitioning of net surface radiation between sensible and latent heat fluxes can clearly affect quite rapid transformations of an air mass. Using the above estimate, with no rainfall and no mean vertical motion, mean relative humidity would increase by
0.18 d\(^{-1}\) in the wet case, 0.06 d\(^{-1}\) in the dry case. Clearly the likelihood of precipitation is much increased by the wet surface. As most of the air mass transformation due to surface heating occurs during a day-time period of about 12 hours, an air mass need only remain over a soil moisture anomaly for this period for its relative humidity to be affected by about 0.1. With winds of 5 m s\(^{-1}\) the horizontal scale of the anomaly need only be about 200 km.

A dry anomaly can persist only if the rainfall remains light enough for subsequent evaporation quickly to dry the ground again before the water penetrates to the root zone of the vegetation responsible for most of the evapotranspiration. The precise conditions depend on the detailed hydrology of the particular surface but it is clear that where rainfall due to advection moisture flux convergence from outside the anomaly area is as large as the potential evapotranspiration, reduction of evaporation due to a soil moisture anomaly is unlikely to persist for long. The smaller the horizontal scale of the anomaly, the greater will be the role of advection relative to evaporation in maintaining the rainfall, so that anomalies of small horizontal scale are limited both in the magnitude of their effect and their likely duration. A soil moisture anomaly may modify the moisture convergence by altering the pressure and flow fields, as noted in the experiments by Cunnington referred to earlier. By reducing evaporation more than precipitation a deficiency of surface moisture may be expected to affect not only the anomaly area but also the region downwind. It is thus possible for an anomaly to spread into adjacent regions. We shall obtain from the experiments discussed in this paper some guidance concerning the persistence and the spread of an anomaly field of limited horizontal scale in middle latitudes.

The main series of experiments reported here were made using as control an integration in which the surface flow over the anomaly area (Europe between the Mediterranean and the Baltic) was rather slack in a weak ridge from an Azores high, with easterlies over Scandinavia south of another anticyclone. Such a circulation is sufficiently like that of some summer and, more especially, spring months, for the results of these experiments to be instructive; however, most summer months have a more westerly flow pattern over Europe. An integration of the annual cycle with another version of this model (Slingo 1982) did generate a more westerly flow, so a further experiment was subsequently run using this as control. We will therefore be able to consider the evolution of an anomaly with two different circulation types.

2. THE MODEL

The models used for the experiments are developments of that described by Corby et al. (1977); changes are the inclusion of interactive soil moisture and snow, as described by Slingo (1982), and the moist instead of the dry gas constant is used. For the main series of experiments the climatological radiation scheme described by Corby et al., in which longwave radiative cooling is a function only of temperature, pressure and latitude, is retained; the later experiment with the annual cycle version of the model uses the radiation scheme described by Slingo, in which atmospheric absorption of solar and longwave radiation depends on moisture as well as temperature.

The models have five layers of equal mass and use a \(\sigma\)-coordinate system (\(\sigma = \text{pressure/surface pressure}\)). The horizontal grid gives a quasi-uniform resolution over the sphere. In the medium-resolution version used for the main series of experiments the grid length is approximately 330 km (3\(^\circ\) latitude); some results are given for a series of experiments with a low-resolution version of the model with a grid length of about 500 km, other features of the model being unchanged.

The treatment of surface fluxes of latent and sensible heat is discussed in detail by Corby et al. but it is appropriate in this context to give a brief description, as well as indicating the changes made to incorporate an interactive soil moisture variable. Essentially sensible heat flux, \(H\), evaporation, \(E\), are written in the form \(H \propto CV \Delta \theta\) and
\[ E \propto CV \Delta q \] each being proportional to the product of a surface transfer coefficient, \( C \) (which takes one of four values depending on stability and whether the surface is land or sea), a boundary layer wind speed, \( V \) (including an enhancement in unstable conditions), and \( \Delta \theta \) or \( \Delta q \). \( \Delta \theta \) and \( \Delta q \) represent the differences for potential temperature, \( \theta \), and specific humidity, \( q \), respectively, between air at the top of the boundary layer and air, originally with surface characteristics, taken to the top of the boundary layer. In going from the surface to the boundary layer top it is assumed that \( \theta \) may be enhanced by condensation and that \( q \) will be reduced to about 80% of the surface value. The formulation is identical to that described by Corby et al. except that in \( \Delta q = q_s - q_T \), where \( q_s \) is \( q \) at the top of the boundary layer, \( q_s \) is taken to be \( 0.8 \) \( a \) \( q_s(T_s) + (1 - a) q_T \), where \( q_s(T_s) \) is the saturation specific humidity at surface temperature \( T_s \) and \( a \) is a fraction \( = \) (soil moisture in cm)/10, but \( a \geq 1 \).

Soil moisture is incremented by rainfall and snowmelt and decreased by evaporation and runoff. Runoff occurs only to restrict soil moisture to a maximum value of 20 cm.

This formulation is based on that of Manabe (1969) and its form is supported by the observational data analysed by Priestley and Taylor (1972). An interpretation of it is that a fraction, \( a \), of the ground is assumed to be moist with surface mixing ratio equal to the surface saturation value, while the remaining fraction, \( 1 - a \), is assumed to have the same \( q \) as that of the air.

The surface snow-free albedo, \( z \), is defined as a function of latitude, as in Corby et al., with values of about 0.17 at latitudes around 50°N where the soil moisture anomalies are imposed. With seasonal snow-cover, which is represented explicitly in these versions of the model, the albedo is \( z = 0.12 \beta^{1/2} \), where \( \beta \) is the snow water equivalent in mm, with an upper limit of 0.6. This feature is unlikely to have any effect in these experiments with anomalies in Europe, in which conditions appropriate to July are assumed with initial snow-depths zero except over Greenland and Antarctica, where initial values are set at 60 m to maintain permanent snow-cover.

There are a number of relevant features of the models which could be made more elaborate. The radiative transfer does not depend on modelled cloud, nor, in the main series of experiments, on modelled humidity. The albedo is kept independent of soil moisture – Idso et al. (1975) and Norton et al. (1979) have reported increases in albedo by 0.1 to 0.15 due to a change from wet to dry soil or wet to dry vegetation.

3. The experiments

For the main series of experiments the low- and medium-resolution models were each integrated for 50 days from an initial global data set compiled from real data for most of the northern hemisphere (for 27 May 1977) and day 60 of a previous July integration for the southern hemisphere. The constants used in the radiation scheme were those appropriate for northern summer derived from calculations of Rodgers as described by Corby et al.; sea surface temperatures and ice extent were prescribed from climatology for mid-June. For the control integration initial soil moisture was prescribed as in Fig. 1(a) for the medium-resolution model, with 5 cm except in some arid regions with no soil moisture; specifications were generally similar for the low-resolution model, with minor changes due to different locations of grid points.

In the anomaly experiments the models were again integrated for 50 days starting from initial conditions which were identical to those of the control experiments except for the soil moisture content for the area of Europe indicated for the medium-resolution models in Fig. 1(a); this area extends from the Atlantic nearly to the Black Sea between the Mediterranean and the Baltic. The area was similar in the low-resolution model, the model rows affected being 42°F, 47°F, and 51°F observed with 43°F, 46°F, 49°F, and 52°F N in the medium-resolution model. Dry and wet experiments were run with initial soil moisture over the anomaly area set to 0 and 15 cm respectively. The nomenclature of all the experiments is defined in Table 1. Generally the senses of the significant differences
Figure 1. Soil moisture (mm) for experiment MC (a) day 0, (b) days 41–50 mean. The grid points at which anomalous values of initial soil moisture were applied are marked by filled circles in (a). Symbols +, ■, × define the areas (south Scandinavia, Norway, south and central Spain) referred to in Fig. 9.

<table>
<thead>
<tr>
<th>Model</th>
<th>Experiment series</th>
<th>Grid length (km)</th>
<th>Initial soil moisture over the anomaly area (cm)</th>
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<tbody>
<tr>
<td></td>
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<td>Dry (0)</td>
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<tr>
<td>Low resolution</td>
<td>L</td>
<td>500</td>
<td>LD</td>
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<tr>
<td>Medium resolution</td>
<td>M</td>
<td>333</td>
<td>MD</td>
</tr>
<tr>
<td>Annual cycle</td>
<td>A</td>
<td>333</td>
<td>AD</td>
</tr>
</tbody>
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* Initial soil moistures over Europe generally exceeded 15 cm in experiment AW.

(wet − control) and (control − dry) are similar. Therefore for brevity, and to maximize the signal, results will mostly be shown only for the wet and dry anomaly experiments.

The initial data for the anomaly experiments with the annual cycle version of the model were taken from day 376 (June 7 of the second year) of the run described by Slingo.
The soil moisture was set to zero at the same points over Europe as for experiment MD. It should be noted that soil moisture in middle latitudes at day 376 of the annual cycle run was generally greater than the 5 cm specified for the main series of experiments. In particular, over the anomaly area and to the north over Scandinavia and east over Russia soil moisture generally exceeded 15 cm. Since evaporation is little affected by soil moisture till soil moisture is well below 10 cm, this prevents the anomaly over Europe affecting the evaporation in these adjacent regions in experiment AD.

The area adopted for the anomaly is similar to that affected by the European drought of 1976 (though parts of Scandinavia were also dry then). Rainfall for February to June 1976 was less than 80% of normal over most of the anomaly area with below 50% over most of the north-western quadrant.

4. The medium-resolution control experiment

The model’s surface tends to dry out over most middle latitude land in summer, as is shown by a comparison of the initial and days 41–50 mean soil moisteres (Fig. 1) for Europe, western Asia and North Africa. The only parts of this area to show an increase are tropical Africa and India and small regions of Scandinavia and Russia near 60°N. Regions initially having zero soil moisture still have mostly less than 10 mm at days 41–50, exceptions being the central Sahara and south-eastern Arabia. Experiments with the low-resolution version of this model (Carson and Sangster 1981) indicate that rainfall in these regions is inversely related to the albedo, which is rather low (about 0.22) in this model.

The aridity of the northern continents at days 41–50 is probably exaggerated by the initial specification of the soil moisture content at 5 cm. Although this value produces only a slight reduction in evaporation from the potential value, further drying steadily reduces evaporation so that a reduction in rainfall is to be expected from the results of the global anomaly experiments discussed earlier. This drying out is perhaps accelerated, at least in the western parts of the middle latitude land masses, by the weakness of the surface westerlies carrying moisture from the oceans. This weakness is evident from the sea level pressure map shown by Corby et al., which is similar in pattern to that of the control experiment, with weak flow over central Europe and easterlies north of about 60°N over Europe on the southern flank of a polar anticyclone.

Because the middle latitude summer continents tend to dry out in the control experiment the rainfall distribution does not reach an equilibrium state. This is demonstrated in Fig. 2 by the 10-day means of evaporation and rainfall for the anomaly area shown in Fig. 1(a). In the control experiment soil moisture, evaporation and rainfall all decrease monotonically through the period. The marked drop in precipitation between the first and second 10-day periods is probably due to the model’s adjustment from the observed initial state towards its own equilibrium. That the days 11–20 rainfall is near the model’s equilibrium is confirmed by the rainfall of the wet case (MW) which stays near 2–2.5 mm d⁻¹ throughout the rest of the 50 days. Because this amount is less than the potential evapotranspiration of 3 to 4 mm d⁻¹, both the wet case and the control dry out. Since changes in atmospheric moisture content are small relative to soil moisture changes it follows from the moisture budget that there is a net moisture flux divergence over the area which decreases as the ground dries out. A larger decrease in flux divergence might have occurred if adjacent land areas, especially those upwind, were not drying out at the same time, as shown in Fig. 1.

One might expect from the dependence on soil moisture that evaporation would decrease linearly as soil moisture decreases from 10 to 0 cm. This does not occur because as \( a \) (soil moisture in cm/10) decreases, the surface temperature and saturation mixing ratio increase so as to maintain the balance between net radiation and the upward turbulent fluxes — ground heat storage being small. The relation between \( E \) and \( a \) can be estimated from the data used in Fig. 2 to be approximately \( E \propto a^{1/2} \).
5. THE DEVELOPMENT OF DIFFERENCES DUE TO SOIL MOISTURE ANOMALIES

(a) Initial developments

In this section we illustrate the development of the predictions for the first week of the medium-resolution experiments with dry and wet anomalies in the soil moisture over the area indicated in Fig. 1(a). As discussed in section 1, the direct effect of a dry anomaly is to modify the partitioning of the surface turbulent heat flux between the sensible and latent parts so as to increase the temperature and decrease the specific humidity of the atmosphere, particularly the boundary layer.

Figure 3 shows the differences between MW and MD in the temperature and humidity of the bottom (~200 mb deep) model layer after 48 hours. Temperatures in the dry case are 4 K higher over most of the anomaly area with local increases of over 8 K. Specific humidities are lower by 1 to 3 g kg\(^{-1}\). The average differences are 3.8 K and 1.7 g kg\(^{-1}\), giving relative humidities of about 90% and 50%; the increase in sensible and decrease in latent heat content of the bottom layer are 752 and 816 J cm\(^{-2}\) d\(^{-1}\) respectively. As the differences in sensible and latent heat flux are 753 and 854 J cm\(^{-2}\) d\(^{-1}\) respectively over this period, about half the differences have remained in the bottom 200 mb over the anomaly area. Of the remainder, part has been advected out of the anomaly area mostly to the west due to the generally easterly low-level flow prevailing at the initial time, while part has been lost through greater rainfall in the wet case. The differences in rainfall up to this time (not shown) are small, though on the second day rainfall rates exceeding 1 mm d\(^{-1}\) spread north from a rain area over southern Europe to about 48°N near 5°E in the wet case but only 45°N in the dry case. More marked differences in rainfall are evident on the third day (Fig. 4). In the dry case, the rainfall area which was over southern France on day 2 has disappeared whereas in the wet case much of France has over 2 mm d\(^{-1}\). The region of heavy rain over south-east Europe has shrunk in the dry case and areas with over 1 mm d\(^{-1}\) outside the anomaly area over southern Spain and over and to the west of the British Isles are absent or much smaller. There is at this stage little change in the rainfall to the north of the anomaly because of the mainly northerly flow in these regions.

With the increase in lower tropospheric temperatures in the dry case there are decreases in surface pressure of 2 to 3 mb over most of the anomaly area by day 3 with
little change in adjacent regions. This association of decreased pressure and decreased soil moisture and rainfall should not be surprising since the pressure change is consistent with the temperature increase. Similar changes were evident in the soil moisture experiments of Walker and Rowntree (1977) for an idealized North Africa; it can also be related to observed conditions over the Sahara. The fall of pressure leads to a more marked trough over France and southern Britain but has little evident effect on the rainfall distribution, the rain area associated with the trough, though developing in the dry case, yielding generally below 2 mm d\(^{-1}\) on days 4 and 5 compared with peak values exceeding 5 mm d\(^{-1}\) in the wet case (not shown).

By day 5 (Fig. 5) the low which had formed over southern Britain by day 4 of both experiments has developed a trough eastwards over northern Germany and there is advection of air from central and eastern Europe into southern Scandinavia between the trough and an anticyclone over central Scandinavia. The pressure pattern in MD (not shown) is similar with, as at day 3, pressures generally lower by up to 4 mb, over the anomaly area. The effects of the soil moisture anomaly are thus advected northward into Scandinavia as is evident in Fig. 6 which shows that differences in the bottom layer's
specific humidity exceed 1 g kg\(^{-1}\) well north of the 52\(^\circ\)N limit of the anomaly. During the following 24 hours, substantial rainfall (Fig. 7) occurs in the wet case over southern Scandinavia ahead of the surface trough as it moves slowly northeast towards the Baltic. Over 2 mm of rain falls in a belt linking regions of heavier rain over north Germany and Norway. Yet in the dry case there is only a somewhat weakened rain area over Norway, with less than 0.5 mm over Denmark. Thus as early as the sixth day the effects of the anomaly are already evident over land outside the anomaly area. These land areas will dry out so that the area with reduced evaporation expands.

(b) Subsequent developments

Figure 8 shows the rainfall for the wet and dry cases averaged over the first 10 days. This confirms the impression gained from the daily maps with the rainfall maxima over western and south-eastern Europe both considerably weaker in MD; there are particularly large reductions over eastern Europe north of 48\(^\circ\)N - from 5 to 0.5 mm d\(^{-1}\) at one point. Outside the anomaly area there are decreases over Denmark and southern Sweden and over Spain and Italy near 40\(^\circ\)N.
In the following 10 days (not shown) the differences continue to develop and expand, though eastern Europe is drier than in days 1–10 in both experiments. Almost all Scandinavia is drier in MD and slight decreases are evident over north-east Africa and Arabia.

The development of the mean soil moisture for the anomaly area and three nearby regions which also develop substantial differences is shown in Fig. 9 for the dry case, wet case and control. The general drying noted in Fig. 1 is evident except for the Norway area, though even this area dries in MD – this is not so in the low-resolution experiment (section 6). Over the anomaly area the control case is almost as dry as MD by day 50, and this drying of the control case is probably responsible for the small differences between it and the dry case over Spain by this time. The most striking differences outside the anomaly area are over south Scandinavia, which is drier than the anomaly area at days 40 and 50 in experiment MD.

6. THE TIME-AVERAGED EFFECTS AT DAYS 21–50

Although the model does not reach an equilibrium state during the 50 days of the experiments, time-averaging helps to remove some transient features of the simulations. Averages have therefore been calculated for days 21–50 and are discussed in this section.
Figure 7. Rainfall (mm) during the sixth day of experiment (a) MW, (b) MD.

Statistical assessments of the differences between the dry and wet experiments are discussed in section 7. Some results of the low-resolution experiments are shown. These experiments' results are generally very similar to those of the medium-resolution experiments and so increase one's confidence in the conclusions.

(a) Rainfall

There is a striking contrast between the mean rainfalls for cases MD and MW (Fig. 10) with reductions by a factor of three or more in MD over most of the anomaly area and much of Sweden. These results provide a clear answer to the questions posed in the introduction concerning the impact and the persistence of an anomaly of this horizontal scale in middle latitudes, and the possibility of its extending to affect other areas. It is particularly interesting that precipitation in the western central part of the anomaly area in the wet case is still sufficient to balance the potential evaporation of about $4 \text{mm d}^{-1}$. This is confirmed by the soil moisteres (not shown), which exceed the initial value of 15 cm at four of the rainiest points in the days 21–50 mean; in the dry case these points have at most $1.55 \text{mm d}^{-1}$ rainfall over this period. In the eastern part of the anomaly area, rainfall is well below potential evaporation in both MW and MD and the lifetime of the wet anomaly is clearly limited to the time taken for soil moisture to decrease to values at which it restricts evaporation – it is thus a function of: (a) the excess of the initial value (15 cm here) above the critical value (10 cm); (b) the excess of evaporation over precipitation.
Figure 8. Rainfall (mm d$^{-1}$) of experiments (a) MW, (b) MD for days 1-10.
Figure 9. Soil moisture at ten-day intervals for experiments MW, MC, MD averaged over areas defined in Fig. 1: (a) anomaly area, (b) south Scandinavia, (c) south and central Spain, (d) Norway.
Figure 10. Rainfall (mm d⁻¹) averaged for days 21–50 of experiments (a) MW, (b) MD.
The rainfall difference pattern (Fig. 11) delineates more clearly the regions outside the anomaly area where rainfall has been affected. As well as the large areas of Scandinavia and the Iberian peninsula evident at days 1–10 there are decreases over Turkey and much of North Africa in the dry case which extend in a rather broken fashion to the Persian Gulf. The differences over North Africa are partly in a desert region which the model commonly fails to keep dry. Although there is still too much rainfall in the dry case, these results indicate that a moisture source to the north over Europe can contribute to the model's rainfall in this area.

The rainfall distributions for the low-resolution model are generally similar to those of Fig. 10 with an equally striking contrast between the dry and wet cases over the anomaly area (Fig. 12). There can be no doubt that in both models, rainfall over Europe is greatly affected by surface moisture availability. There are some important differences outside the anomaly area, notably over northern Scandinavia where rainfall is increased in the dry case. This increase is associated with lower surface pressure (Fig. 17) suggesting increased cyclonicity which is probably a chance occurrence. The maximum differences in rainfall are over 5 mm d$^{-1}$ over central Europe, south-east of the maximum in (MW – MD). The northern regions of the Sahara have slightly more rainfall in the dry case but most of North Africa is drier as in (MW – MD). An interesting feature not shown by the map is that the decreases over Africa extend well south of the equator. There is also a tendency to this in (MW – MD).

(b) Evaporation

The differences in evaporation (Fig. 13) over the anomaly area in the days 21–50 mean are still very much as forced by the initial anomalies, even over eastern Europe where the rainfall is much too small to maintain the wet anomaly for long. The large evaporation differences obtained here arise both because the soil has remained very dry in
Figure 12. The difference in rainfall (mm d$^{-1}$) between experiments (LW - LD) for days 21–50.

Figure 13. The difference in evaporation (mm d$^{-1}$) between experiments (MW - MD) for days 21–50.
the dry case and because the eastern edge of the wet anomaly region is subject to intense drying by evaporation into air being advected from the arid regions to the east.

Outside the anomaly area the evaporation differences reflect the differences in rainfall. Over desert regions such as the Sahara where the soil is originally dry (Fig. 1), any precipitation which falls is quite quickly evaporated, and evaporation is obviously limited by rainfall. Over Scandinavia the decreased rainfall in the dry case leads to a reduction of soil moisture and evaporation. Once evaporation is limited here, it is not possible to assess whether subsequent decreases in rainfall are due to advection of drier air from the anomaly area or to the local soil moisture deficit.

(c) Atmospheric moisture

As discussed in section 4, the moisture content of the model's bottom layer shows a large response within two days; Fig. 14 shows a response of similar magnitude in the days 21–50.

![Figure 14. The difference in specific humidity (g kg⁻¹) at σ = 0.9 between experiments (MW - MD) for days 21–50.](image)

21–50 mean with differences of up to 5 g kg⁻¹ over eastern Europe. This is clearly a direct effect of the reduction in the local surface moisture source in the dry case. The changes extend well outside the anomaly area to the south over Africa and Arabia. It is most probable that this is due to advection of drier air in case MD, though, as discussed above, a local soil moisture feedback can help to maintain the anomalies. It is likely that this has happened where the differences in ω at σ = 0.9 have a maximum in the central Sahara. The large horizontal scale of the differences in atmospheric moisture content is strong evidence for regarding the differences in rainfall over North Africa and Arabia as being due to the European soil moisture anomaly. The differences in ω at higher levels (σ = 0.7, 0.5) (not shown) are of similar sign and horizontal scale, except that coastal regions of north-west Europe have increased moisture in the dry case. Flow is westerly at upper
levels over this area so that moist oceanic air is advected from the Atlantic and the anomaly has little effect on $q$. The difference patterns for the low-resolution model (not shown) are of generally similar character at $\sigma = 0.9$ though the largest differences over Europe are only $3 \text{ g kg}^{-1}$. At $\sigma = 0.7$ and 0.5 the area of increased moisture over Europe in the dry case is more extensive.

(d) Temperature

The differences in temperature at $\sigma = 0.9$ (Fig. 15) are quite similar in pattern to the differences in specific humidity at that level, just as they were after two days (Fig. 3). Again the differences over the anomaly area are clearly due to the local soil moisture while the general warming over North Africa and Arabia in the dry case must be due originally to advection though with some local soil moisture feedback effects. Relative humidity is of course considerably reduced in MD in these areas of increased temperature and decreased specific humidity, so that both are likely to contribute to the decreases in rainfall. At $\sigma = 0.7$ (not shown) there is only weak ($\sim 1 \text{ K}$) warming in the dry case and at $\sigma = 0.5$ the wet case is warmer over all the anomaly area except the British Isles and a narrow coastal strip of the nearby continent. This weakening of the changes with height is much less marked in the low-resolution model (not shown).

(e) Sea level pressure

General circulation models, like observed data, show marked variability in sea level pressure from one year to another. Mitchell (1982) shows this for winter for the seasonally varying version of this model; in summer standard deviations are substantially lower (Mitchell, personal communication) with typically values of 2 mb or less in middle latitudes. The differences between cases MW and MD (Fig. 16) are thus of a magnitude such
that it is difficult to be confident whether they are systematic effects of the anomalies or are due to the model’s natural variability. Some guidance on this comes from a comparison of \((MW - MC)\) and \((MC - MD)\) (not shown) and from the statistical tests in section 7. There is agreement in the signs of these differences over the anomaly area but little more elsewhere than would be expected by chance. Comparison with the low-resolution model results (Fig. 17) leads to the same conclusion – there is no more agreement outside the area of direct forcing than would be expected by chance. In particular the pattern over the Atlantic is largely reversed between Figs. 16 and 17, with opposite changes in the maxima near Iceland and Newfoundland. The falls over the anomaly area in the dry cases were already evident in the first few days and were discussed in section 5. Maximum
differences here exceed 3 mb with both the low- and medium-resolution models, not only over the anomaly but some distance to the east.

7. Statistical aspects

It is desirable to assess the statistical significance of the differences. Since there are not a large number of similar experiments available to allow calculation of the model's inherent variability it has been necessary instead to assess the consistency with time of the differences between experiments, by using Student's t-test to determine the statistical significance of the departure from zero of the mean difference for the period days 1–50. The standard deviation necessary for this has been derived from the differences \( d_1, d_2, \ldots, d_5 \) for the five constituent 10-day periods, days 1–10, 11–20, \ldots, 41–50, so that \( t = \sqrt{5} \frac{d}{\sigma} \), where \( \sigma = \left( \frac{1}{4} \sum d_i^2 \right)^{1/2} \) and \( d = \frac{1}{4} \sum d_i \), the summations extending from \( i = 1 \) to 5. This procedure involves the assumption that the five 10-day periods are effectively independent, so that, when used in statistical tests, a mean formed from them has four degrees of freedom.

Some assessments of significance have also been made by comparing the mean of the five 10-day means from experiment MD with the mean of the five from MW using Student's t-test. This procedure has the advantage that there are more degrees of freedom (8 instead of 4). However, it is unsatisfactory for most variables because of the trends during the 50 days; these occur not only in rainfall and soil moisture but also in temperature, the northern continents warming steadily from the initial (late May) data as shown for the anomaly area in Table 2. Despite this a t-test between the five 10-day means of MW and those of MD in Table 2 shows that the probability of them belonging to the same population is less than 1% (\( t = 3.3, 8 \) degrees of freedom); however, this technique fails to show the control case to be different from either of the anomaly cases at the 10% level. The use of the t-test on the differences, on the other hand, yields high significance levels despite the fewer degrees of freedom and the tendency for the differences to increase (MC – MW) or decrease (MD – MC) with time as the control dries out. The latter is an unavoidable problem with non-stationary difference patterns. Because of these difficulties, the results of significance tests reported in the rest of this section all refer to the application of t-tests to the five 10-day mean differences.

The field most directly related to soil moisture, evaporation (Fig. 18), shows highly significant (\( < 1\% \)) differences over the anomaly area and along 55°N immediately to the north and differences significant at the 10% level over southern Sweden. Differences of opposite sign are significant over adjacent sea areas to the west and north over which drier, warmer air is advected in the dry case and over western Russia near 40°E where the soil became moist early in the integration. The t-test for sensible heat flux gives very similar results except that signs are reversed.

For rainfall, Fig. 19 shows significant differences for most of the anomaly area though differences over parts of the Balkans and western Russia, where differences
Figure 18. Student's $t$ for the difference in evaporation between experiments (MW - MD) (see text for details). Isolines drawn for 10%, 5%, and 1% significance levels (4 degrees of freedom).

Figure 19. Student's $t$ for the difference in rainfall between experiments (MW - MD).

become small as the soil dries out in the wet case, are not significant. Outside the anomaly area, much of southern Scandinavia and central Spain are significantly different. The differences over North Africa are significant at only a few grid points probably because they do not develop till after the first ten days.

Surface pressure (Fig. 20) is significantly lower in the dry case over much of the anomaly area and also to the east as far as 40°E. The significant areas over the Atlantic have, as noted earlier, changes of opposite sign in the low-resolution experiment and probably arise by chance.

Figure 20. Student's $t$ for the difference in surface pressure between experiments (MW - MD).
8. Experiments with the annual cycle model

The sea level pressure pattern for experiment AW averaged for the first 10 days of the experiment (Fig. 21) shows a much more westerly flow over northern Europe than that shown in Corby et al. (1977) which is typical of experiments MD, MC and MW. As would be expected in the real atmosphere the flow varies somewhat from one 10-day period to another but the ridge near 45°-50°N and troughs at 60°-65°N are characteristic of most of the subsequent 10-day periods of AD and AW over Europe. Troughs and ridges move

Figure 22. Rainfall (mm d⁻¹) averaged for days 1–10 of experiments (a) AW, (b) AD.
eastward from the Atlantic bringing alternating wet and dry spells with occasional secondary depressions near 50°N producing considerable rainfall, as shown for AW at days 1–10 in Fig. 22(a).

In the initial absence of soil moisture over western and central Europe in AD evaporation must initially be zero and from the arguments presented earlier one would expect rainfall to be reduced over the anomaly area. This expectation is borne out by Fig. 22(b) which shows rainfall to be typically 30 to 70% of that in Fig. 22(a). There are also reductions over southern Scandinavia and central Spain as shown for MW and MD in Fig. 8. The mean pressure pattern for AD (not shown) is very similar to that in Fig. 21 with falls of about 2 mb over the anomaly area as in Fig. 16. Although in AD rainfall exceeds evaporation, which averages around 1 mm d$^{-1}$ over the period, so that soil moisture and evaporation are increasing, evaporation is still below 2 mm d$^{-1}$ over much of the anomaly area in the days 11–20 mean and rainfall (not shown) continues to be generally less than in AW.

However, around day 20 a depression moves into central Europe and persists for several days. As might be expected there is heavy rainfall over much of the anomaly area in association with this cyclonic flow. The rain in AD increases evaporation to around 3 mm d$^{-1}$ over the northern half of the anomaly area for the rest of the 50-day integration. This is too close to the 3 to 4 mm d$^{-1}$ of AW for any significant impact on rainfall to be expected.

Averaged for days 21–50 (Fig. 23) the rainfall is heavier in AD over most northern

![Figure 23. The differences in rainfall (mm d$^{-1}$) between experiments (AW - AD) for days 21–50.](image)

and eastern parts of the anomaly area. Evaporation (not shown) and rainfall do remain less over the south-western part of the anomaly area and over central Spain. The differences in this region are statistically significant at up to the 1% level (Fig. 24) and the decreases near 50°N 15°E are significant at the 5% level.

9. **Discussion and conclusions**

In all the experiments we have made with soil moisture anomalies, there has been a
substantial impact through evaporation on precipitation over the anomaly area. A dry anomaly in soil moisture reduces evaporation and increases the surface temperature and sensible heat flux in the manner discussed by Walker and Rowntree (1977) and vice versa for the wet anomaly. These changes lead directly, within a day or two, to increased atmospheric temperatures and decreased absolute and relative humidities, which tend to inhibit precipitation. This reduction of precipitation occurs despite a fall in surface pressure associated with the warming of the atmosphere.

In the experiments reported here, the dry and wet anomaly cases tended to converge despite the reduction in rainfall with the dry anomaly. There is no evidence here of the intransitivity or lack of convergence to the same quasi-equilibrium state such as that over the Sahara in the experiments of W. M. Cunnington. This may be because of the different locations of the anomalies or because of the different models. Cunnington used a radiation scheme in which long-wave fluxes depended on humidity and temperature, whereas the humidity dependence was lacking in our experiments except for AD and AW and in those the dry anomaly was swamped by heavy rain after about 20 days. The humidity feedback will cause the anomaly region to emit more radiation to space and so cool more quickly in the dry case. This cooling tends to raise surface pressures and reduce the low-level inflow of moist air; similar effects due to surface albedo increases were discussed by Charney (1975). Neither our experiments nor Cunnington’s included model-dependent cloud, which could also modify the results; Charney et al. (1977) found that the effects of cloud on the solar radiation were largely balanced by the effects on the long-wave radiation.

In the experiments with a weak or easterly flow over Europe, the wet and control cases tended to converge towards the dry case although for the wet case convergence was slow with no clear sign of convergence after 50 days in the wettest part of the anomaly area. However, with the erosion of the wet anomaly, it appears unlikely that these regions would remain wet indefinitely. In the experiment with more westerly flow over northern Europe convergence occurred much more quickly when the ground became wet in the dry case during the cyclonic days 21–30 period, although soil moisture was still generally low enough for the region of substantial decreases in evaporation to expand again if a dry spell occurred. Clearly more rapid convergence is to be expected where the equilibrium state is wet than where it is dry; this is also evident in the global anomaly experiments of Carson and Sangster (1981). This is partly because of the form of the relation between evaporation and soil moisture, which is such that a small increase in soil moisture from a low value changes evaporation more than does a similar decrease from a large value. In our experiments it is also partly due to the rather large value (15 cm) of soil moisture used.
to initialize the wet anomaly region. Although the reasons for the different convergence rates are thus partly artificial, it is certainly true in the real world that a small amount of rainfall on a dry surface can change the evaporation to near the potential value, and so make a much greater difference than a little evaporation from wet ground because in the latter case water can be drawn from below the surface both by diffusion and root transfer.

The relation of evaporation to ground moisture and the treatment of moisture in the ground are matters currently being given much attention (WMO 1982). We make no claim that the quite simple treatment adopted for these experiments is a sufficiently accurate approach. Mintz (personal communication) has suggested that the potential evaporation calculated using the area-average surface temperature is too high in cases where soil moisture is so lacking that of the total grid box area only a small fraction, which would be wetter and so cooler than average, would be evaporating at all. He points out that in this context it may be inappropriate to apply the empirical approach of Priestley and Taylor (1972) in which the potential evaporation is for a freely transpiring cooler surface. The effect of this error would be to overestimate the evaporation which resulted from a small amount of rain on an arid surface. However, another criticism of the present approach is that where rain falls on a dry surface, it may in reality all be available for evaporation. The two suggested errors tend to cancel and it is not obvious that correction of either by itself will give a more realistic parametrization. There is no doubt that less simple soil moisture and evaporation treatments, representing the sub-grid-scale nature of much precipitation, soil moisture diffusion using multi-layer models, the vegetative root system, the vegetative canopy and evaporation of intercepted precipitation, and surface and sub-surface runoff, should be tested in climate models to assess the sensitivity to these parametrizations.

A particularly interesting aspect of the results is the propagation of the anomaly. In the experiment with the annual cycle version there is little scope for propagation because of the wetness of most of the surrounding land which, with the present soil model, prevents the evaporation being affected by any reduction in the precipitation caused by the dry anomaly. The other experiments show considerable propagation. In all the experiments the effects of the anomalies on rainfall extend north of the anomaly area into southern Scandinavia. This is due to the advection of dry (wet) air from the dry (wet) anomaly region in the southerly flow ahead of low pressure troughs where much middle latitude precipitation occurs. The southward spread of the anomalies to the convergence zone south of the Sahara was an unexpected result. However, in a dry region such as the Sahara, which lacks its own surface moisture source, the atmospheric humidity must depend on the moistness of air advected into the region; evidently the anomaly area provides a substantial fraction of this in the model.

The mechanism for the propagation of the anomaly generally appears to be the advection of modified air from the anomaly area. It appears that the changes in the mass and flow fields through the effects of the anomalies on temperature, convergence and vertical motion do not play a major role. They must have some impact over the anomaly area but the direct effects of soil moisture are dominant there. The eastward extension of the surface pressure falls in the dry cases (Figs. 16 and 17) is associated with falls in 500 mb height (not shown) near the Black Sea and increased rainfall near 50° N 40° E. It seems quite probable that this upper troughing is a downstream effect of an upper ridge forced over the dry anomaly. This aspect of the results requires further investigation.

In conclusion, we have found that a mid-latitude soil moisture anomaly of the limited horizontal scale on which such anomalies are observed can have important effects on the precipitation, humidity and temperature in a general circulation model not only over the anomaly but also over a considerable area of the adjacent land. The persistence of the anomaly is dependent on the subsequent atmospheric circulation, and on the nature (dry or wet) of the anomaly and of the model’s local climatology.
REFERENCES


