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FUTURE CLIMATE OF THEMEDITERRANEAN BASIN WITH PARTICULAR EMPHASIS ON CHANGES IN PRECIPITATION

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CONTENT

	Pages
INTRODUCTION	1
GLOBAL-MEAN CHANGES	2
Observed climate changes over the past 100 years	2
Future global-mean changes	4
REGIONAL CHANGES	8
Introduction	8
Climate scenarios	8
GCM limitations	9
GCM PROJECTIONS FOR THE MEDITERRANEAN BASIN	12

INTRODUCTION

Changes in climate between now and the middle of the 21st century are likely to be dominated by the influence of the greenhouse effect caused by increasing concentrations of carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), ozone (O_3) and chlorofluorocarbons (CFCs). These greenhouse gases individually and collectively change the radiative balance of the atmosphere, trapping more heat near the Earth's surface and causing a rise in global-mean surface air temperature. Substantial global warming is virtually certain; however, the attendant changes in climate at the regional level are highly uncertain. Except for a few regions, about all we can be sure of is that large changes <u>will</u> occur, but we cannot yet quantify these changes. Indeed, for precipitation we are as yet unable to specify even the sign of regional changes with any reliability. Nevertheless, some useful statements can be made with regard to the Mediterranean Basin, and it is of some value to review our present state of knowledge.

We begin by considering the global scale, highlighting the important distinction between equilibrium and transient changes in global-mean temperature. We will then consider published projections of the future climate in the Mediterranean Basin based on general circulation climate models. After critically reviewing these projections, we will give a concensus view of future climate.

GLOBAL-MEAN CHANGES

Observed climate changes over the past 100 years

Future changes in climate can be set in context by a brief description of changes that have occurred over the past 100 years. Both regionally and in terms of global-mean values, climatic conditions have fluctuated noticeably from year-to-year, decade-to-decade and on longer timescales. Consider the largest spatial scales first.

The near-surface air temperature averaged over the globe has increased by about 0.5°C since the late 19th century (Jones et al., 1986a,b,c). Fig. 1 illustrates this by showing hemispheric and global mean changes. Parallel changes in the temperature of the lower troposphere have also occurred; see Wigley et al. (1985) for a recent review.

As is evident in Fig. 1, this warming has not been a continuous upward trend. Nor has the warming been spatially homogeneous and trends have varied substantially from region to region. These spatial differences are illustrated in Fig. 2, which shows annual-mean temperature trends over the Northern Hemisphere for 1967-86. Note that, over the past 20 years, the Mediterranean Basin has warmed in the west and cooled in the east, so one cannot simply assume that changes in this region (or any other region) will follow the global-mean trend.

Noticeable changes in precipitation have also occurred this century on spatial scales from hemispheric downwards. Large-scale area-average precipitation changes are more difficult to quantify than temperature changes because of the higher spatial variability of precipitation and because of data homogeneity problems which are more difficult to overcome. For land-based data in the Northern Hemisphere, Bradley et al. (1987) show an upward trend from 1920 to the present in mid-to-high latitudes (35-70°N) and a marked downward trend in tropical-to-subtropical latitudes (5-35°N) (Fig. 3). On smaller spatial scales, some regions of the world have experienced marked changes on decadal time scales. The Sahel region of

Africa is a particularly striking example (Fig. 4).

Almost all other climate variables show evidence of a continually changing climate. For example, there have been noticeable changes in the atmospheric circulation of the North Atlantic as measured by the pressure gradient between the Azores High and Iceland Low - a trend towards a decreasing gradient. The extent of mountain glaciers has decreased over the 20th century (Meier, 1984), albeit with marked regional differences and inter-decadal variations. This has contributed noticeably to a general rise in sea level (Gornitz et al., 1982; Barnett, 1985). Of the 10-15cm global-mean rise, roughly one third can be attributed to each of mountain glacier melting, oceanic thermal expansion and melting from large ice sheets (Greenland and Antarctica) (Wigley and Raper, 1987).

In parallel with these changes in mean conditions, most measures of climate show changes in variability (as measured by variance and/or the frequencies of extreme events). Very few of these changes in variability (in contrast to changes in means) have been statistically significant. A noteable exception is England and Wales area-average precipitation, where there has been a significant trend towards more wet extremes in spring and more dry extremes in summer (Wigley and Jones, 1987).

A leading question is: What is/are the cause/causes of these changes? In the absence of changes in external forcing factors, the climate system would be expected to show considerable natural variability simply because of the complexity of interactions between the oceans, cryosphere, atmosphere and land surface and the range of timescales associated with these components. In addition, forcing undoubtedly occurs due to changes in planetary albedo (volcanic effects, cloud and surface changes), changes in solar irradiance, and changes in the concentrations of greenhouse gases (CO_2, CH_4, N_2O, O_3 and CFCs).

For the latter, the global-mean radiative forcing over the past 100 years has been large (around 2 Wm^{-2}) and this would be expected to cause a global-mean warming. Although a warming has occurred, and its magnitude is compatible with that expected to have resulted from the greenhouse effect, we are not yet able to positively relate cause and effect. Nevertheless, the qualitative agreement and the realistic physical basis of current climate models demands that the possibility of future global-mean warming, and the multitude of regional changes that would accompany such a warming, must be taken seriously.

Future global-mean changes

The magnitude of future global-mean warming will depend on future atmospheric trace gas concentrations, on the sensitivity of the climate system to external forcing and on delays due to the thermal inertia of the oceans. The greenhouse effect will cause changes in global-mean temperature on timescales of decades to centuries, with associated trends in all climate variables at smaller spatial scales. These changes will also have, superimposed on them, inter-annual to inter-decadal variability which is a natural characteristic of the unperturbed climate system.

In considering future climatic change, it is important to distinguish between the equilibrium and transient response of the climate system. We consider the equilibrium response first. For any given future greenhouse-gas forcing (ΔQ) there will be a corresponding equilibrium global-mean temperature change (ΔT_e) which is the temperature change that would finally be achieved when the climate system reached a new steady state. The relationship between ΔQ and ΔT_e is determined by the climate sensitivity.

The climate sensitivity can be expressed in terms of the equilibrium global-mean temperature change that would occur if the CO_2 level were doubled - the higher its value the higher the climate sensitivity. If the equilibrium temperature change due to a CO_2-doubling is ΔT_{2x}, then the equilibrium change for an arbitrary concentration change from C_o to C is

$$\Delta T_e = \Delta T_{2x}(\ell n(C/C_o)/\ell n2) \qquad \ldots(1)$$

Equ. (1) arises because ΔQ for CO_2 is logarithmic in concentration,

$$\Delta Q = 6.333 \, \ln(C/C_o) \qquad ...(2)$$

The value of ΔT_{2x} is uncertain, but it is thought to lie in the range 1.5-4.5°C (MacCracken and Luther, 1985; Bolin et al., 1986).

Since CO_2 is not the only greenhouse gas that must be considered in the forcing, it is often convenient to express the total forcing (ΔQ_T) in terms of an equivalent CO_2 concentration (C^*). This is defined by inverting equ.(2) to give

$$C^* = C_o \exp(\Delta Q_T/6.333)$$

where, by convention, C_o is taken to be the pre-industrial level of CO_2 alone (i.e. the other greenhouse gases are not considered in defining this reference level).

Relative to conditions in the mid to late 18th century, ΔQ_T is about 2.2 Wm^{-2} today (1985), with 1.4 Wm^{-2} of this being due to CO_2 changes and the rest due to changes in the other trace gases (mainly methane and the CFCs). Thus, the 1985 value of the equivalent CO_2 concentration is just over 390 ppmv (relative to a base level of 279 ppmv) compared with a concentration for CO_2 alone of 346 ppmv. A further forcing change of about 2.1 Wm^{-2} is expected between 1985 and 2025. This corresponds to an equivalent CO_2 concentration of about 550 ppmv in 2025, approximately double the pre-industrial CO_2 level.

Using ΔT_{2x} = 1.5-4.5°C, these values imply an <u>equilibrium</u> warming of 0.7-2.2°C over the period 1765 to 1985. Over the shorter period 1880-1985, for which ΔQ_T = 1.8 Wm^{-2}, ΔT_e lies in the range 0.6-1.9°C. Over 1985-2025, the implied equilibrium warming is 0.7-2.2°C. This is a "best guess" range, corresponding to the 2.1 Wm^{-2} projected forcing value. Allowing for uncertainties in future forcing, ΔT_e over 1985-2025 is expected to lie in the range 0.5-3.3°C. Although this is a very large range of uncertainty, even values at the lower end of the range correspond to very significant changes in climate.

These numbers represent the extreme range of possible warming to which we would be committed in the year 2025. They are not the changes that would actually be observed by this time (viz. the transient-response changes). This is because, due to the large thermal inertia of the oceans, observed changes lag behind the equilibrium changes. The amount of lag is determined by the rate of ocean mixing, by the climate sensitivity, and by the rate of change of forcing. Because of this thermal inertia effect, only some 50-90% of the equilibrium warming may be observed at any given time.

Let us consider the consequences of this lag effect. Over 1880-1985, the expected equilibrium warming lies in the range 0.6-1.9°C. However, the expected <u>transient</u> warming (i.e. the value which should be compared with observations) is much smaller, lying in the range 0.4-1.1°C. This latter range can be compared with the observed warming of about 0.5°C. Since 0.5°C lies near its lower end, one might conclude that the climate sensitivity must lie near the lower end of the range 1.5-4.5°C. There are, however, a number of other ways to explain this discrepancy, so we cannot yet dismiss the higher sensitivity values.

An important consequence of the equilibrium/transient difference is that, no matter what happens to future greenhouse-gas concentations, we will always be committed to an additional warming as the climate system tends towards equilibrium. Thus, the difference between the 1880-1985 range of 0.4-1.1°C and the ΔT_e range of 0.6-1.9°C represents a "hidden warming", or current warming commitment, in the sense that it would result even if the concentrations of all greenhouse gases could be held constant at their 1985 values.

Over 1985-2025, the expected transient-response changes lie in the range 0.4-1.8°C. The additional warming commitment in 2025, which would occur even if concentrations remained constant at their 2025 levels, is 0.3-2.3°C. This commitment includes the present commitment in addition to the effects of future greenhouse-gas concentration changes. (The above results are based on the model of Wigley and Raper, 1987.)

REGIONAL CHANGES

Introduction

In the above, we have concentrated on global-mean temperature as a key measure of the state of the climate system. Changes in all climate variables, precipitation, evaporation, wind patterns and strengths, cloudiness, etc., will necessarily accompany changes in global-mean temperature. In the future, just as they have in the past, these changes will differ noticeably from region to region. Thus, for assessing impacts, global mean quantities are mainly of academic interest, although they do show that future climatic changes will probably be large and will probably occur more rapidly than any previous changes. For direct assessments of impacts, regional-scale (or smaller) details of future changes are needed - not only for temperature, but for a variety of climate variables. At present, our capability for predicting these details is limited and we must resort to the use of scenarios.

Climate scenarios

The most important method for obtaining information on possible future climates is to use an atmospheric General Circulation Model (GCM). However, because of deficiencies in current GCMs (described in more detail below), their outputs should only be considered as possible scenarios for future climatic change rather than predictions. A climate scenario, as defined in the literature, is intended to be an internally consistent picture of future climatic conditions. In addition to the use of GCMs, scenarios may be constructed by a number of other means (see, e.g., Pittock and Salinger, 1982; Palutikof et al., 1984). However, since these methods have not been applied to the Mediterranean Basin, we will concentrate here on GCM results. But first, it is important to keep in mind the limitations of GCMs. Note that, within the modelling community these are well-recognized problems. They are often, however, not realised by analysts who make use of GCM output.

GCM limitations

(1) Published GCM studies of the greenhouse effect consider only the equilibrium response to a doubled or quadrupled concentration of carbon dioxide. It is not yet known whether the regional patterns of climatic change for the transient response will correspond to those given by equilibrium studies (Webb and Wigley, 1985).

(2) Atmospheric GCMs must be coupled to ocean models in order to account for some of the ocean's thermal inertia effects. The ocean models currently in use for this purpose are extremely simple.

(3) In addition to their simplified treatment of the oceans, GCMs still have a number of recognized deficiencies in the way they model clouds, sea-ice effects, and land surface processes.

(4) Different GCM studies which attempt to describe the climate of a high-CO_2 world show regions of agreement and disagreement. Since regions of agreement can occur by chance, they cannot necessarily be accepted as regions where the predictions are more reliable. For example, on the global scale, most recent models give a similar value for ΔT_{2x} (of around $4^{o}C$), but different models give these similar answers for quite different reasons.

(5) The atmospheric models currently used in greenhouse-gas studies are unable to simulate the current ($1xCO_2$) atmospheric circulation at the regional level with any realism, although they do perform reasonably well in describing some of the large-scale features of the general circulation. A reliable "control-run" simulation is thought to be a necessary condition for reliability in any perturbation experiment.

(6) While GCMs can produce output with very short time steps (~10 minutes), the realism of these short-time-step results is in doubt. For current models, the limit of temporal resolution on which results

might have some meaning is of order one day. However, some models currently in use do not have a diurnal insolation cycle, which necessarily precludes realism even at the daily time scale, and the strange results produced by some models on the seasonal timescale, indicates serious deficiencies on timescales of months and longer.

(7) The current spatial resolution of GCMs is too coarse for most impact studies. Coarse resolution means that the orography in the models is highly smoothed and small-scale weather systems are non-existent. This means that one would not expect a GCM to produce reliable results even at the grid-point scale. At best, one can only hope for reliability on the scale of the area covered by tens of grid points - but even this is an optimistic hope at present.

Despite the deficiencies, there are some broad-scale GCM results which can be accepted with some confidence as predictions of future climate. These are summarized in Table 1.

In spite of the problems that plague current GCMs, they are the best tool we have for projecting future changes in climate at the regional level. A number of such projections have been used in impact studies, sometimes with implied faith in the reliability of the projections and other times more skeptically, with a primary aim being to examine climate sensitivities and to develop tools and methods for impact assessment (e.g. EPA, 1984; Parry et al., 1988). Surprisingly, such studies have invariably relied on a single GCM, and no detailed (i.e. regional-scale) comparisons have been made of the projections of different GCMs. Just such a comparison is made in the next section of this paper.

Table 1: A selection of model results from equilibrium GCM experiments for a doubling or quadrupling of atmospheric CO_2 concentration, together with an estimate of the confidence that can be placed in these results. "Unknown" indicates that knowledge of possible future changes is zero. This applies to all items not mentioned.

MODEL RESULTS	CONFIDENCE
<u>Global scale</u> (i.e. global mean values)	
Warming of lower troposphere	High
Increased precipitation	High
Cooling of stratosphere	High
Warming of upper troposphere (especially the tropics)	Moderate
<u>Zonal-mean to regional scale</u>	
Reduced sea ice	High
Enhanced polar warming in Northern Hemisphere (especially winter half year)	High
Increased P-E in high latitudes	High
More absolute high temperature extremes	High
Increased continental summer dryness	Moderate
Stronger monsoon	Moderate
More tropical storms	Unknown
More/less blocking	Unknown
Greater/less interannual variability	Unknown
Spatial detail in general	Unknown
Rainfall extremes	Unknown

GCM PROJECTIONS FOR THE MEDITERRANEAN BASIN

Seasonal-mean values of the equilibrium changes in climate (temperature and precipitation) due to a doubling of the atmospheric CO_2 level are shown in Figs. 5-12. (Note that, for "CO_2" we can read "equivalent CO_2" in order to account for the other greenhouse gases.) These Figures show the results from four different GCMs, each having a mixed layer ocean: the widely used GISS model (Goddard Institute for Space Studies; Hansen et al., 1984), the GDFL model (Geophysical Fluid Dynamics Laboratory; Manabe and Stouffer, 1982), the Community Climate Model of the National Center for Atmospheric Research (CCM; Washington and Meehl, 1985) and the OSU model (Oregon State Univeristy, Schlesinger and Gates, 1985). Further details of these models are given by Schlesinger and Mitchell (1985, 1987).

These four models are similar in their basic physics, but they have major differences in the way they handle sea ice, clouds and surface processes. They also differ in the way they solve the various partial differential equations (GISS and OSU solve these in grid-point form, while GFDL and CCM use a spectral method), and in their resolution (OSU has four times the horizontal resolution of GISS, with GFDL and CCM in between; but OSU has much coarser vertical resolution than any of the other models). The GISS and OSU models are known to perform poorly in simulating present-day regional-scale climatic conditions (Santer, 1987), with GISS producing somewhat more unrealistic results than OSU. The "control run" performances of the GFDL and CCM models have not been assessed in detail.

Before discussing the results produced by these models, we need to assign some time in the future to which they might apply. The key point here is that they are <u>equilibrium</u> results for $2xCO_2$. That is, they do not simulate the transient response to continually changing CO_2 levels. As noted above, the spatial details of the transient respose need not look like those for the equilibrium response. If we ignore this complication (as we must, since no transient response experiments with suitable GCMs have been published), then the main equilibrium/transient difference will be one of timing. Thus, if an equivalent $2xCO_2$ concentration level were reached in, say, 2025, the climate would not achieve the equilibrium configuration corresponding to this level for a number of years. The magnitude of the delay is uncertain, since it depends on properties of ocean heat transport which are presently rather poorly understood, but a delay of 15-35 years is a reasonable guess. This means that the climate changes shown in Figs. 5-12 should be representative of the change between pre-industrial times and some time around 2040 and 2060. Since the changes that have occurred between pre-industrial times and now are much less than many of those shown in Figs. 5-12, we can say that these Figures also represent the changes that might occur between now and 2040-2060.

But.... can we put any faith in these results? Certainly, we cannot believe any of the spatial differences that the models show occurring between different parts of the study areas. Some of these differences are so large and they occur over such small distances, that they are little short of ridiculous - they serve only to expose serious defects in the physics of the models concerned. In all cases, the spatial differences vary from model to model: some models show greater temperature changes in the south, some in the north; some models show marked east-west contrasts in precipitation changes, while others do not; and so on. These model-to-model differences hardly instill any confidence in the performance of any individual model.

The only common feature that we can have any faith in, is a large-scale warming in all seasons. There are only small differences between the seasons, especially if one averages the results of all models together: a warming of about 3.5°C spread uniformly over the seasons would be the best guess based on these model results. This is slightly less than the global mean changes produced by these particular models.

The common features in the precipitation changes are: little change in spring (exception, OSU); little change in summer (exception, GFDL); and an increase in precipitation in autumn (exception, GFDL). Winter changes vary markedly from model to model. Note that, even in the common features noted above, one must average over the whole Basin before any coherent picture emerges. To summarize these results, the best guess would be: no major precipitation changes, with a hint of a slight increase in autumn precipitation.

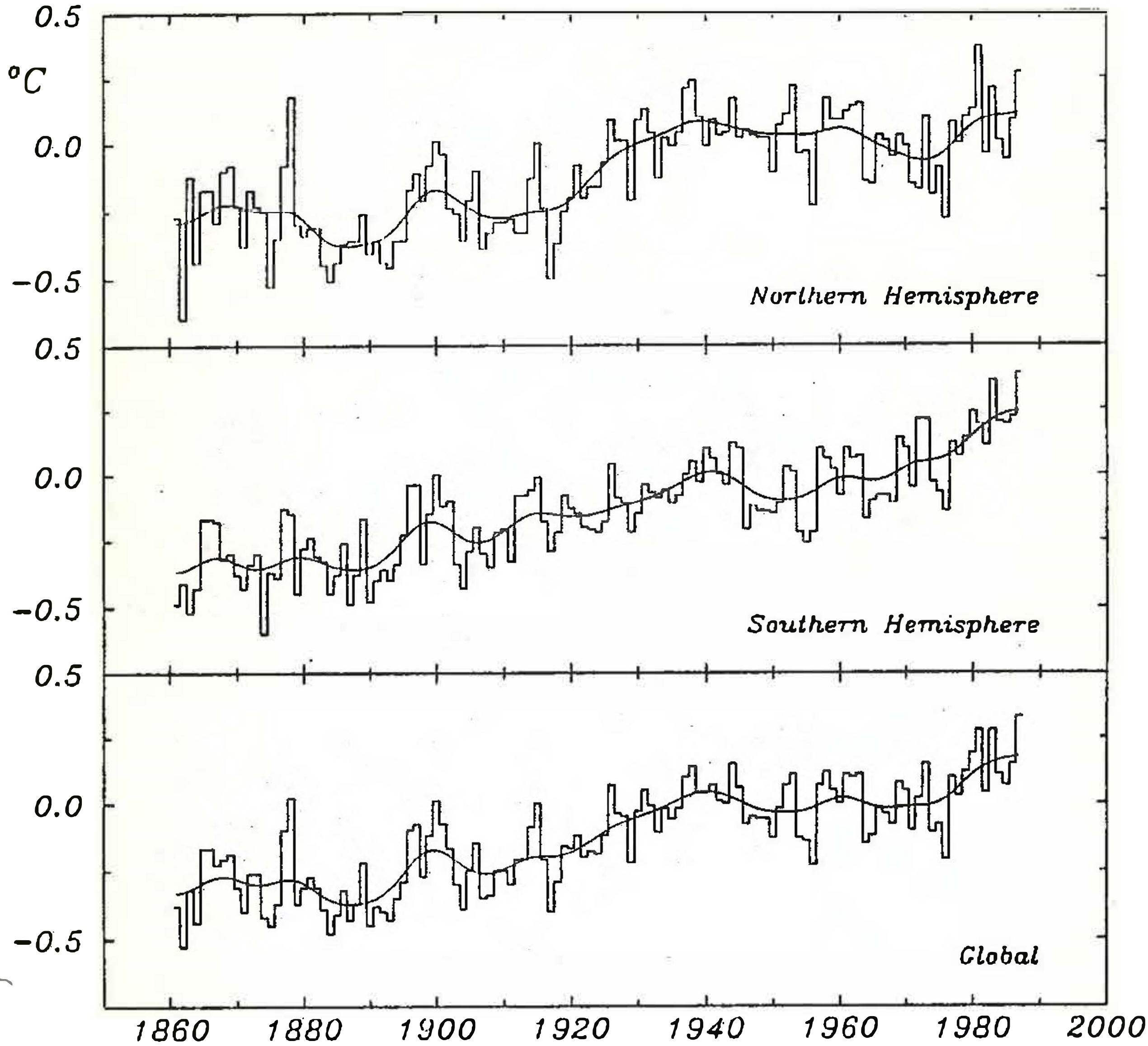


Fig. 1: Hemispheric and global mean temperature changes based on land and marine data.

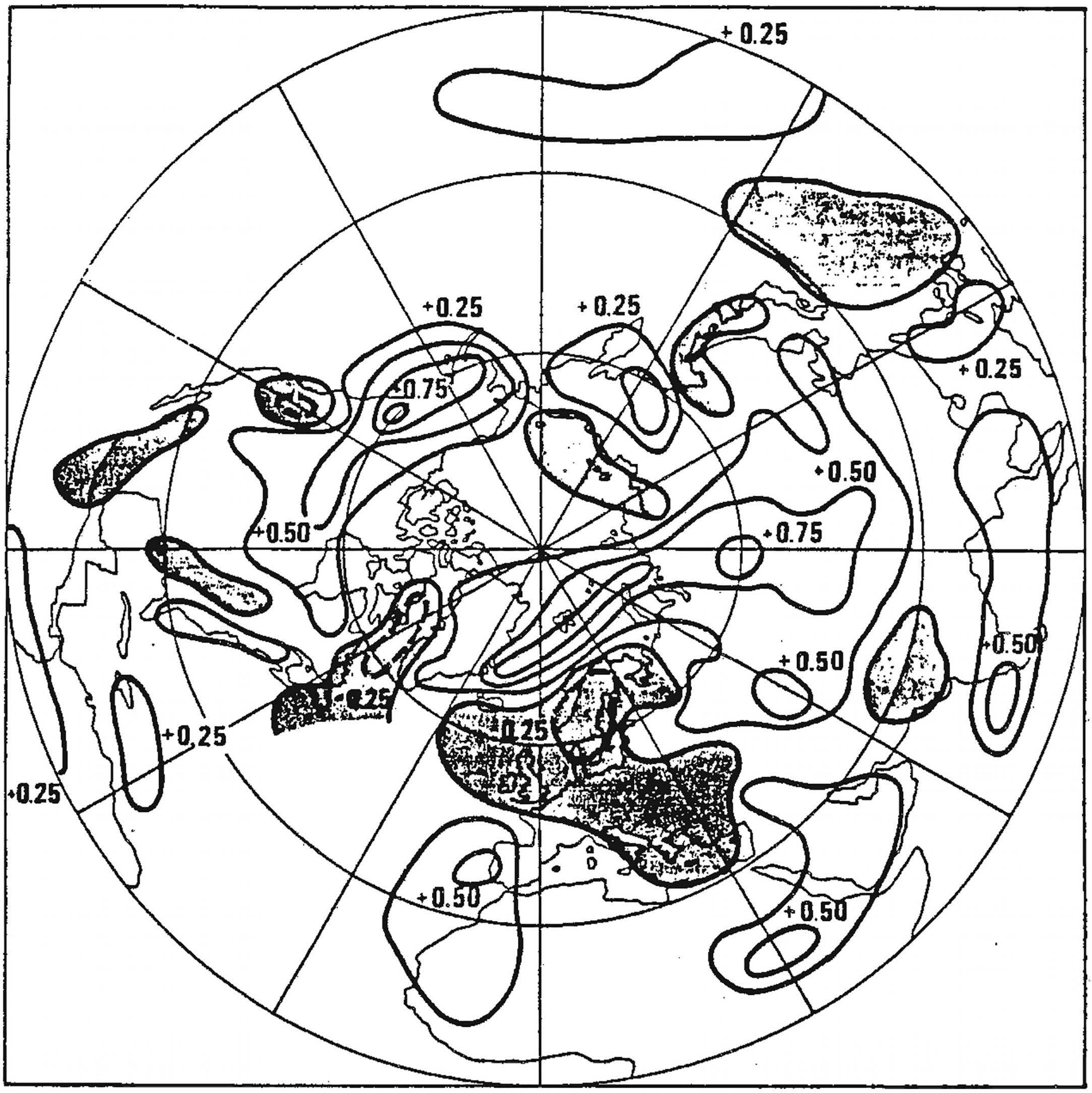


Figure 2: Linear trend in temperature over Northern Hemisphere land masses, 1967-86 (Units, ^{o}C per decade). Shaded areas show regions of cooling.

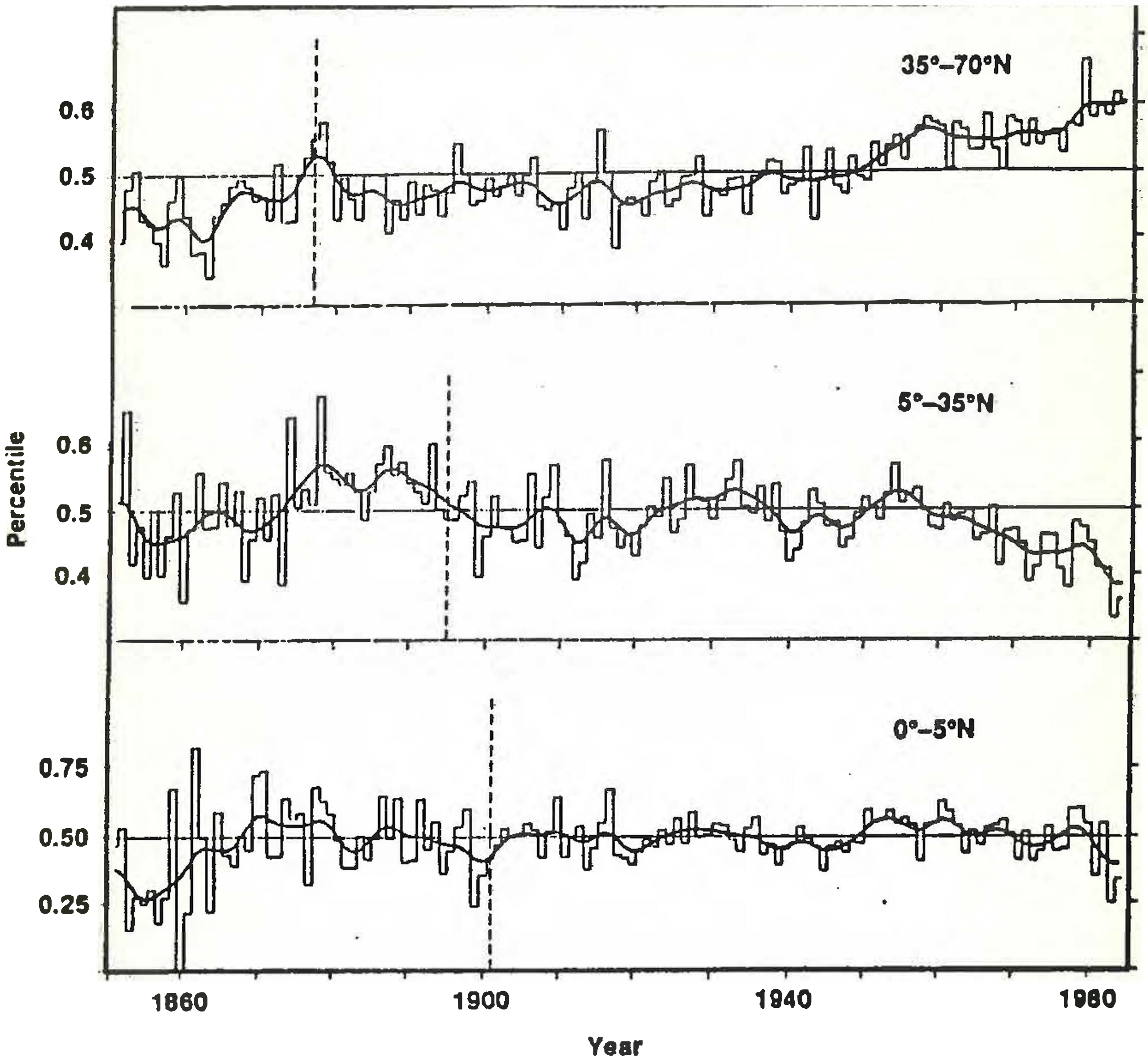


Fig. 3: Precipitation changes for different parts of the Northern Hemisphere (land data only).

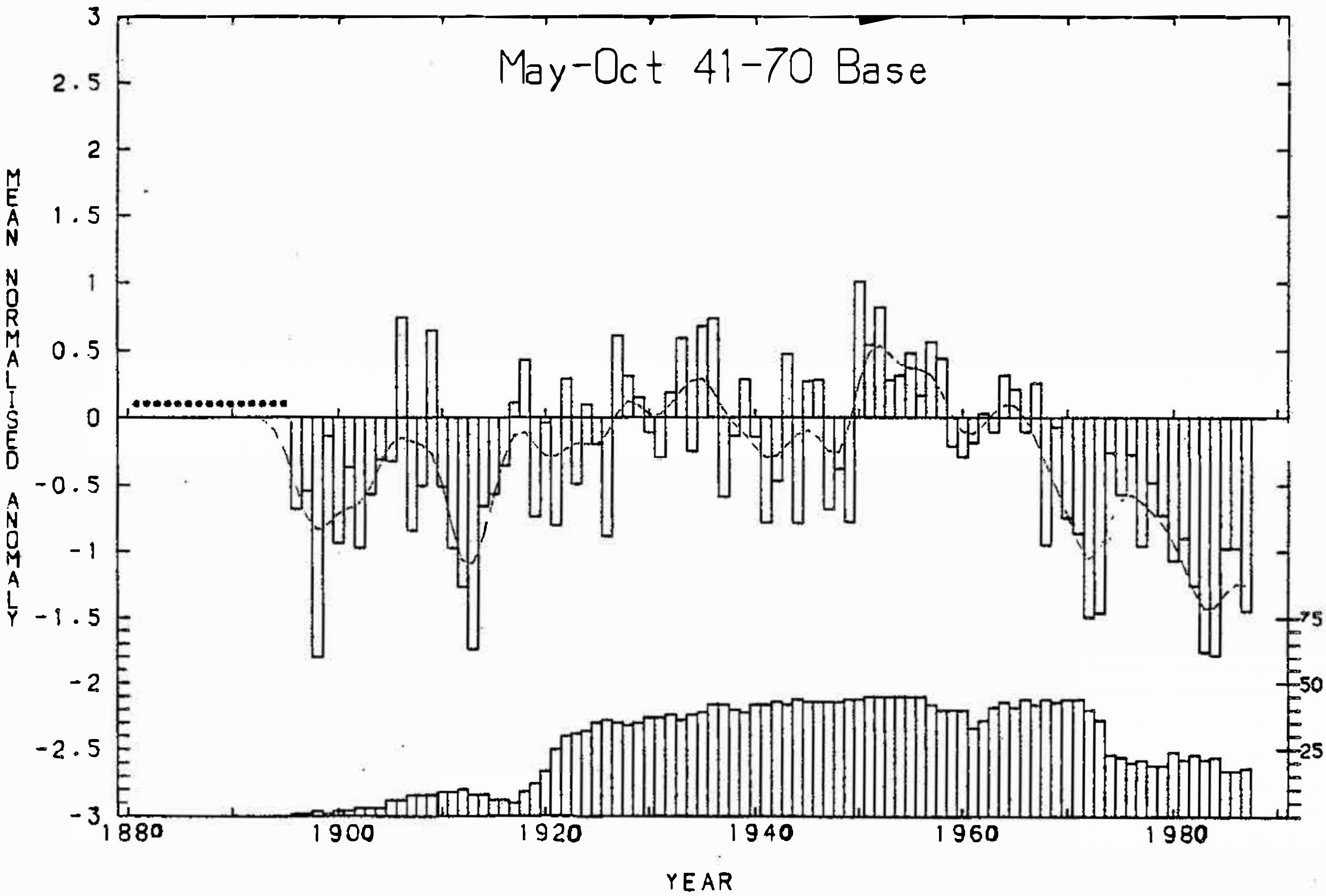
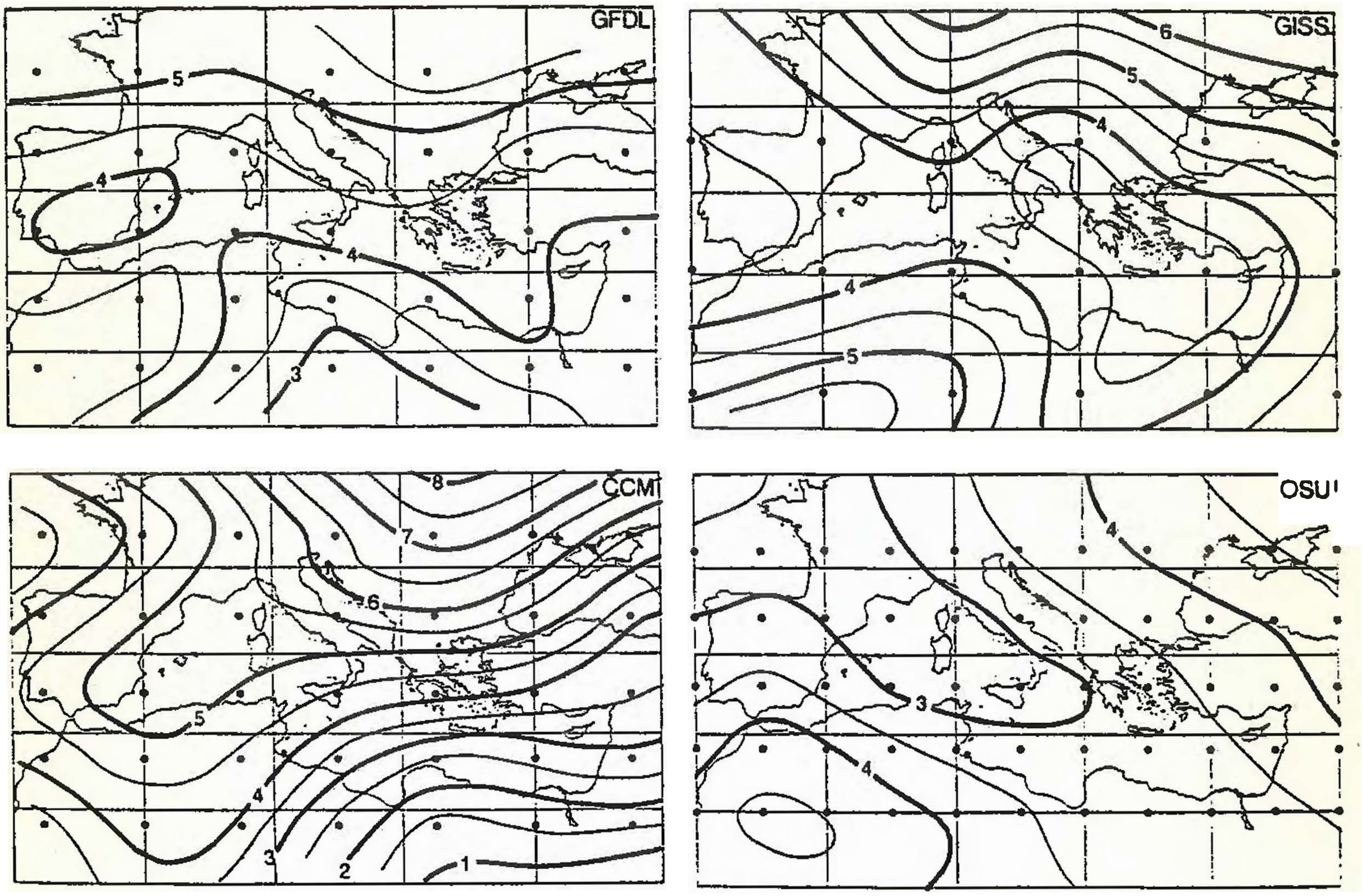


Fig. 4: Precipitation changes in the Sahel zone of west Africa.



WINTER TEMPERATURE

Fig. 5: Winter (DJF) temperature changes (in ^{o}C) due to a doubling of the CO_2 concentration; results from four independent GCMs.

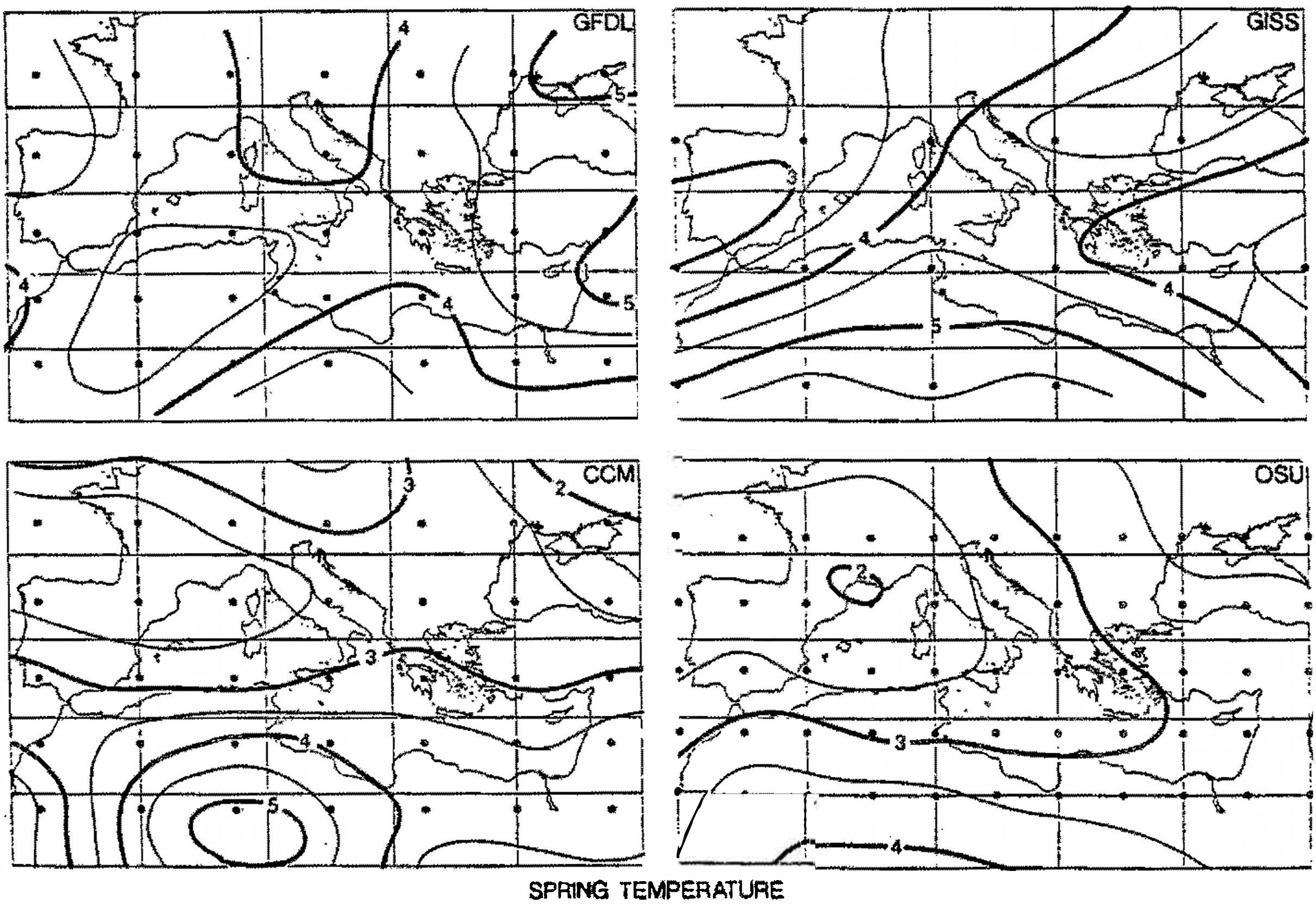


Fig. 6: Spring (MAM) temperature changes (in ^{o}C) due to a doubling of the CO_2 concentration; results from four independent GCMs.

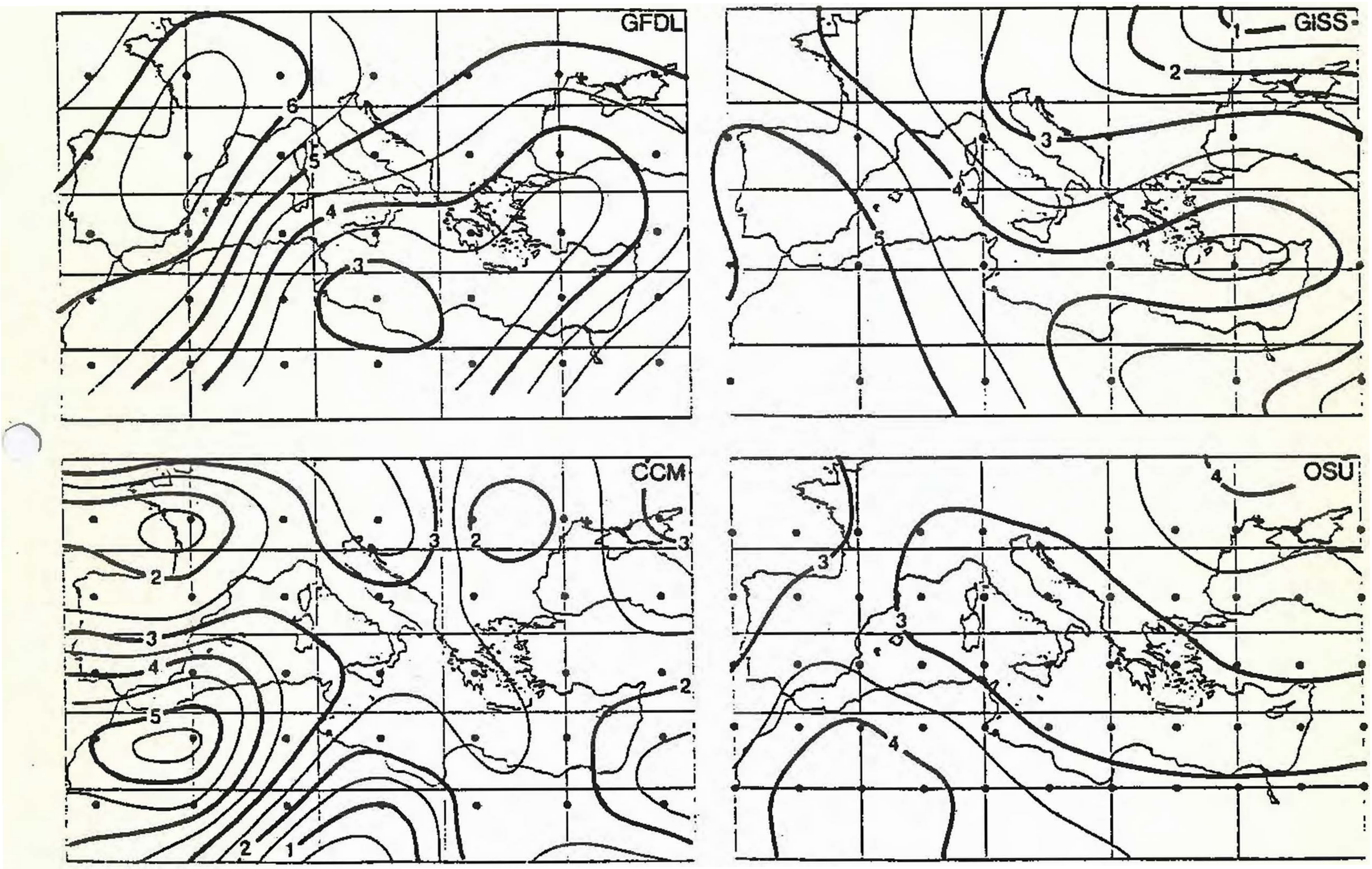


Fig. 7: Summer (JJA) temperature changes (in ^{o}C) due to a doubling of the CO_2 concentration; results from four independent GCMs.

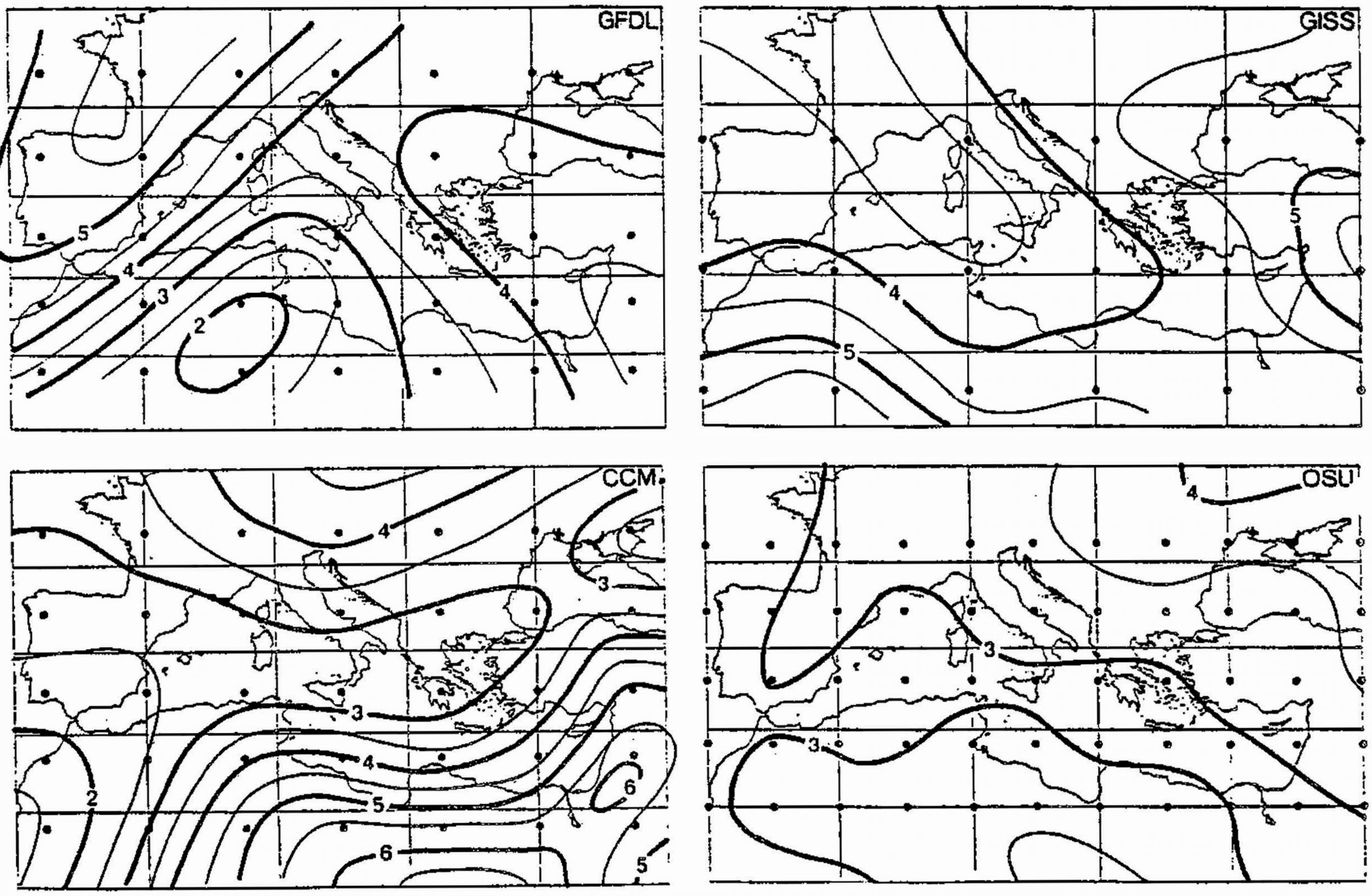


Fig. 8: Autumn (SON) temperature changes (in ^{o}C) due to a doubling of the CO_2 concentration; results from four independent GCMs.

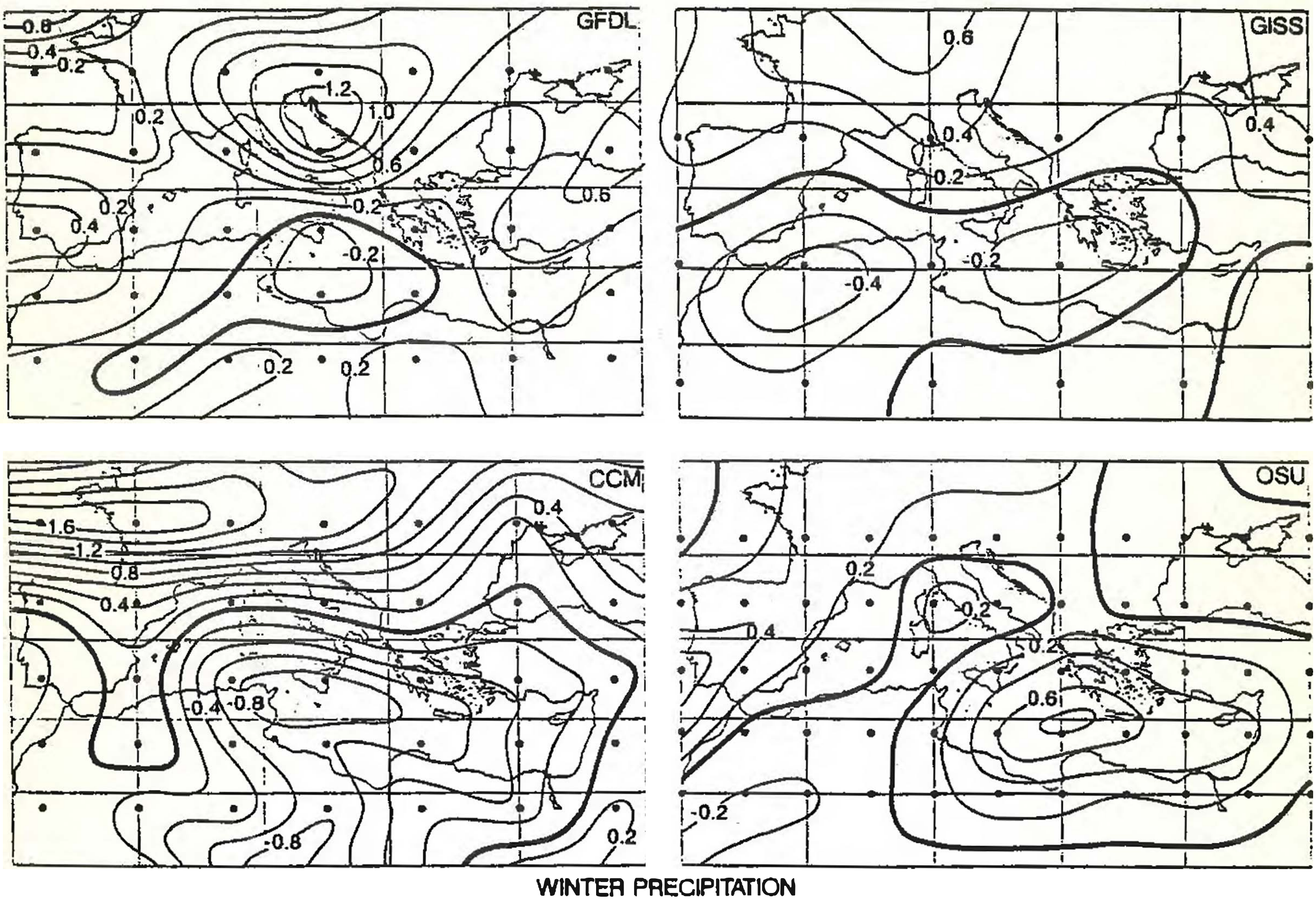


Fig. 9: Winter (DJF) precipitation changes (in mm/day) due to a doubling of the CO_2 concentration; results from four independent GCMs.

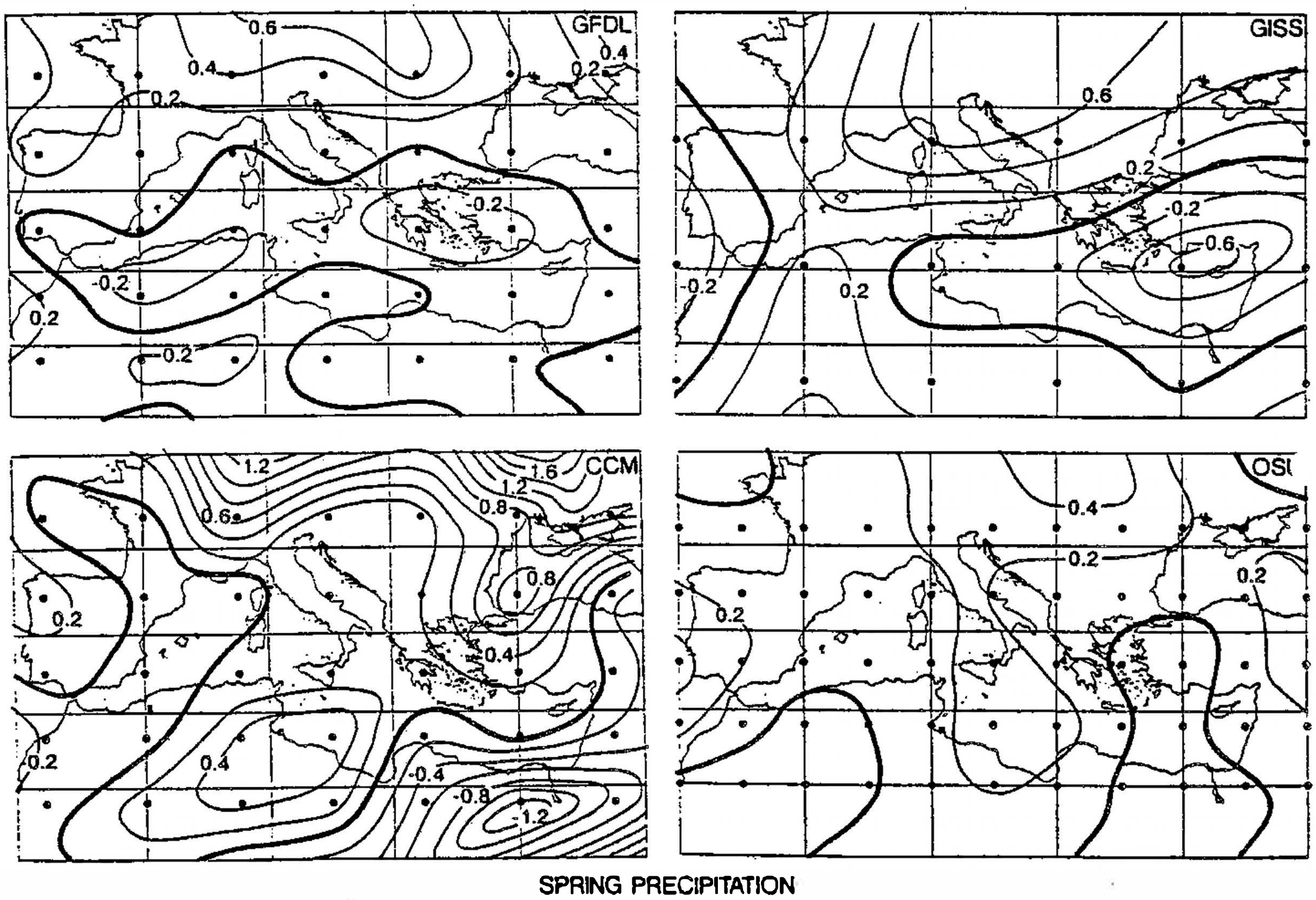


Fig. 10: Spring (MAM) precipitation changes (in mm/day) due to a doubling of the CO_2 concentration; results from four independent GCMs.

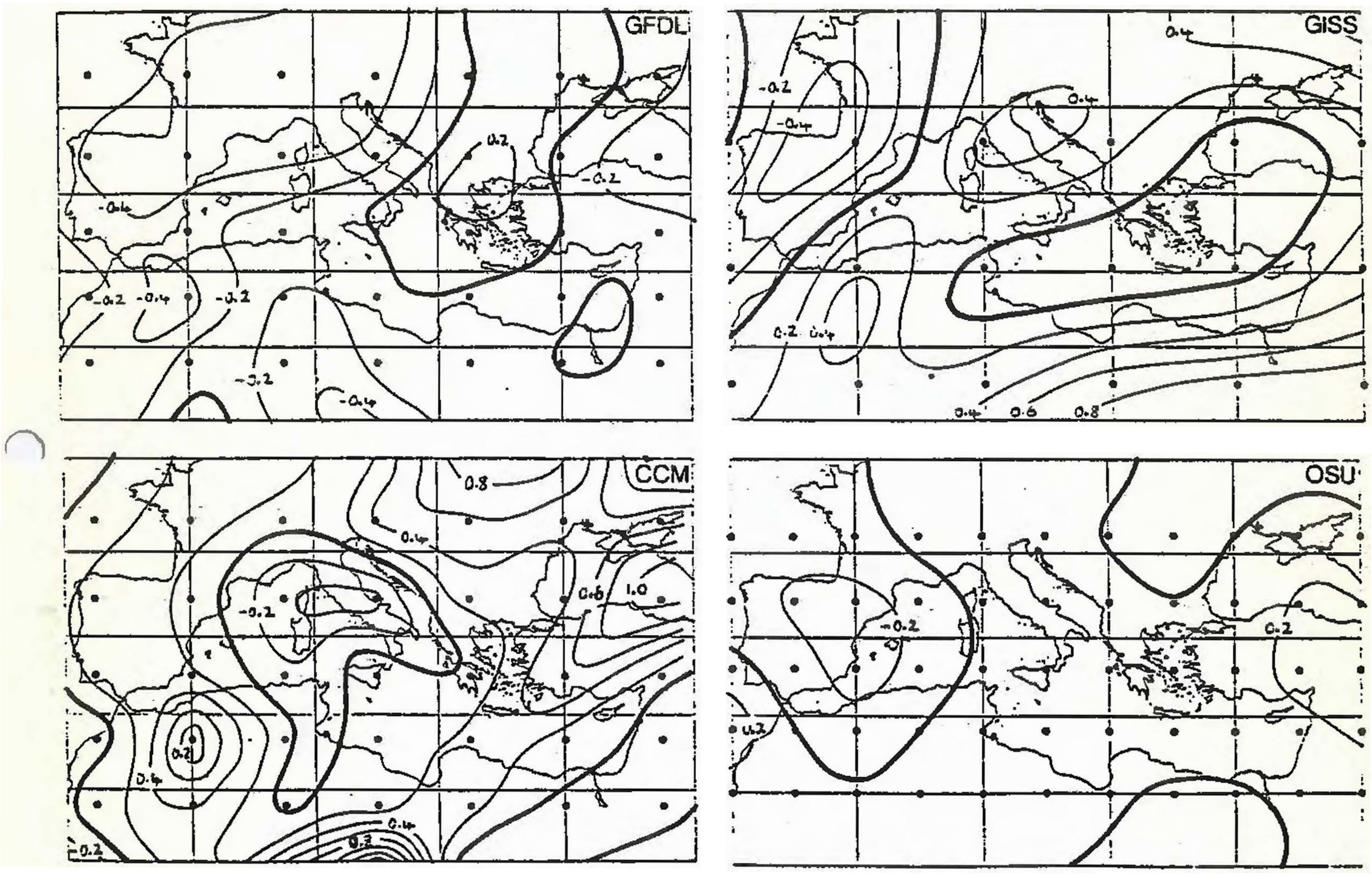


Fig. 11: Summer (JJA) precipitation changes (in mm/day) due to a doubling of the CO_2 concentration; results from four independent GCMs.

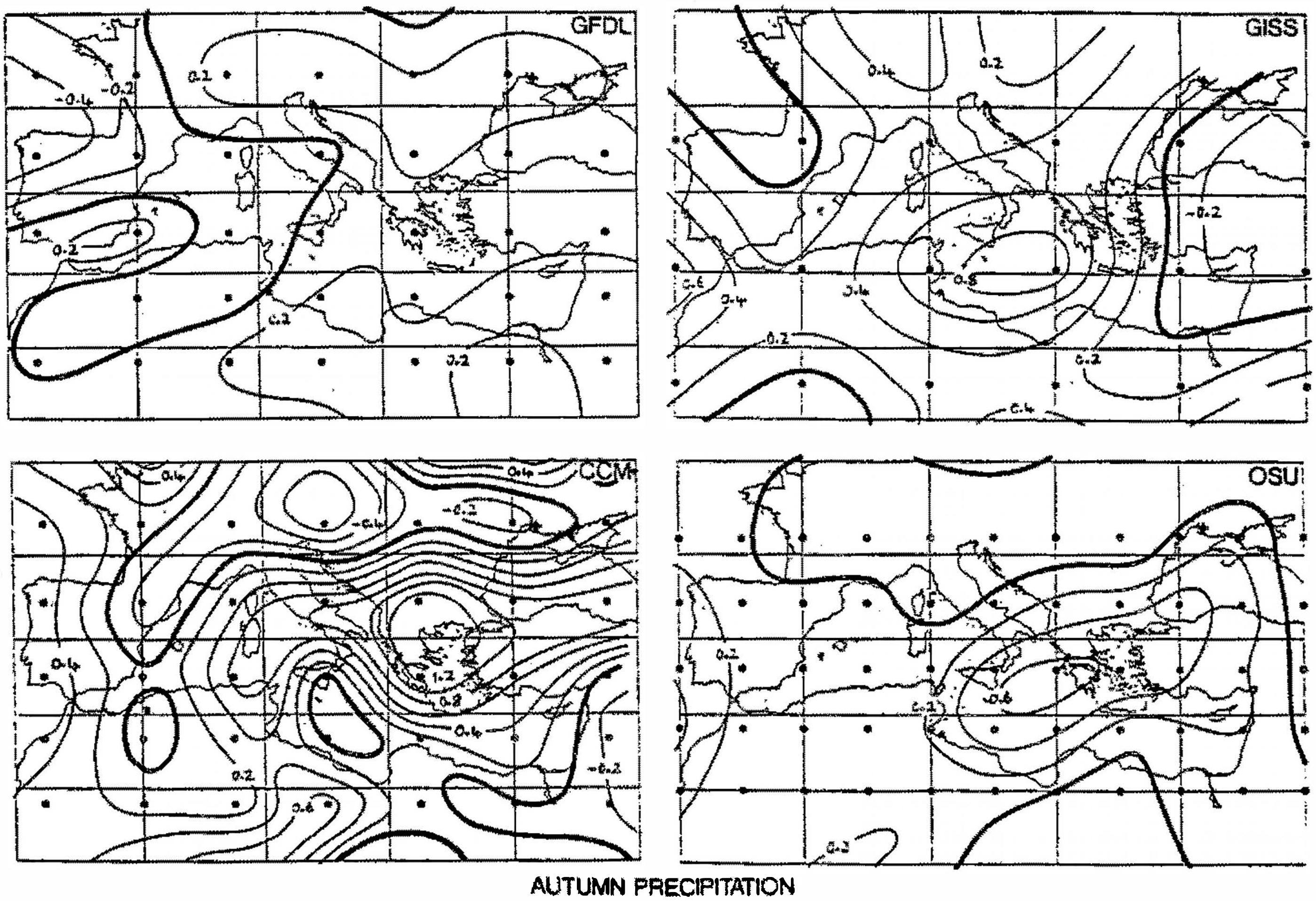


Fig. 12: Autumn (SON) precipitation changes (in mm/day) due to a doubling of the CO_2 concentration; results from four independent GCMs.