J8.11 SUMMER DRYNESS AND GREENHOUSE WARMING: A PROCESS STUDY FOR THE MIDWESTERN UNITED STATES.

Sonia I. Seneviratne*, Jeremy S. Pal, Elfatih A.B. Eltahir, and Christoph Schär

1. Introduction

Earlier GCM studies have suggested that an enhancement of greenhouse warming might increase the occurrence of summer droughts in mid-latitudes, especially in southern Europe and central North America (e.g. Manabe et al. 1981, Wetherald and Manabe 1995, Kattenberg et al. 1996, Wetherald and Manabe 1999). This could represent a severe threat for the agriculture of the regions concerned, where summer is the main growing season.

However, most studies featuring enhanced summer dryness in mid-latitudes use very simple representations of the land surface (''bucket'' models; see Manabe 1969), a fact which could potentially exaggerate some of the mechanisms leading to summer dryness. Indeed, the bucket-type models are known to overestimate daytime evaporation when energy and soil moisture are available (Dickinson and Henderson-Sellers 1988, Henderson-Sellers et al. 1996). This behaviour is likely to emphasize simulated decreases in soil moisture (Gates et al. 1996). It is thus desirable to investigate this issue with models including realistic representations of the land surface processes.

2. Experimental Design

The present experiments are performed with a modified version of the National Center for Atmospheric Research's Regional Climate Model (RegCM); a detailed description of the NCAR RegCM can be found in Giorgi and Mearns (1999) and references therein. This regional climate model includes a land surface scheme of intermediate complexity (the Biosphere-Atmosphere Transfer Scheme BATS, see Dickinson et al. 1993).

A surrogate climate-change scenario following the methodology proposed by Schär et al. (1996) is used for the simulation of a warmer climate. The control runs (CTL) are initialised with and driven by observations and reanalysis data, while the sensitivity experiments (WARM) are forced by a modified set of initial and lateral boundary conditions. The modifications consist of a uniform 3K temperature increase and an attendant increase of specific humidity (unchanged relative humidity). The atmospheric CO_2 concentration of the sensitivity experiments is set to four times its pre-industrial value. The simulations are conducted for the springs and summers of 4 years, corresponding to drought (1988), normal (1986, 1990) and flood (1993) conditions.

* *Corresponding author address*: Sonia I. Seneviratne, Institute for Atmospheric and Climate Science ETH, Winterthurerstrasse 190, 8057 Zurich, Switzerland; e-mail: sonia@geo.umnw.ethz.ch

The model domain covers all of the contiguous United States and parts of Canada and Mexico (see Figure 1). It is centered at 40N/95W and comprises 129EWx80NS grid points, with a horizontal grid spacing of 55.6 km.

The region of focus for the analysis is the American Midwest (outlined in Fig. 1). This subdomain extends from about 36N to 48N, and 99W to 87W. In the result section, we make comparisons with the results of the process study by Wetherald and Manabe (1995; hereafter referred to as WM95); it should be noted that our focus region is located at somewhat lower latitudes than those investigated by WM95 (45N to 60N).

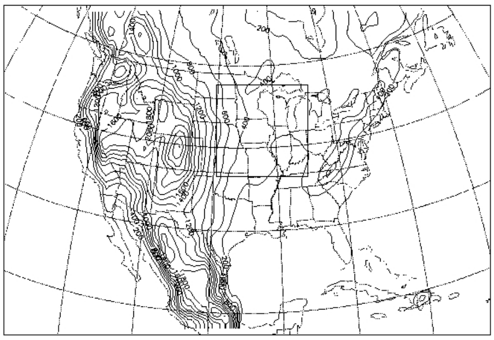

Fig 1. Computational domain and topography [m] used for the numerical simulations. The Midwest analysis region (outlined box) is also indicated (approximately 36N to 48N, 99W to 87W).

3. Validation of the Control Integrations

The control integrations of four spring and summer seasons show good agreement with observations. Figure 2 displays the temporal evolution of simulated precipitation over the Midwest for 1988 and 1993 and the average of the four years. The model is able to capture the interannual variability of precipitation and simulates well the different evolutions observed in the normal and extreme years.

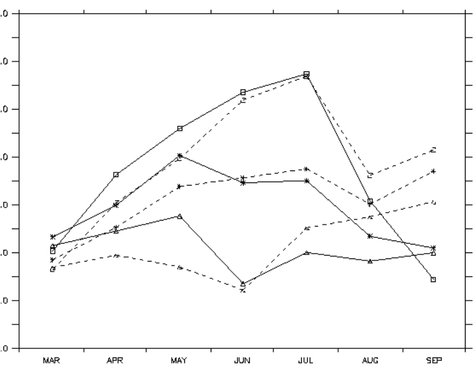

Fig 2. Observed (dashed) and simulated (solid) monthly precipitation [mm/d] over the American Midwest for 1988 (triangles), 1993 (squares) and the average of the four years (asterisks): 1986, 1988, 1990 and 1993. Observations are from the USHCN dataset (Karl et al. 1990). The values are spatial averages over the box outlined in Figure 1.

The spatial representation of precipitation over the whole domain is satisfactory as well; in particular, the extreme dry and wet months, June 1988 and July 1993 are both very well captured by the model (not shown).

4. Results of the Sensitivity Experiments

Figure 3 presents the mean temporal evolution of precipitation, evapotranspiration and runoff (sum of surface runoff and baseflow) over the Midwest subdomain in the CTL and WARM integrations. The magnitude of the observed changes is small compared to the results of WM95. The highest differences in precipitation and evapotranspiration are of the order of 0.3 mm/day, while they are about two to three times larger in the simulations of WM95. Moreover, runoff is almost unaffected in the simulated months.

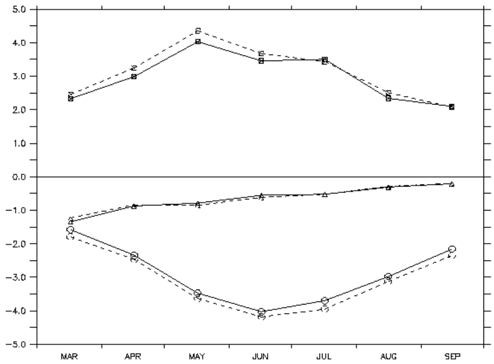

Fig 3. Temporal evolution of precipitation (squares), evapotranspiration (circles) and runoff (triangles) over the Midwest subdomain (outlined in Fig. 1) in the CTL (solid) and WARM (dashed) integrations (average over all years). The values are given in mm/day.

As far as the sign of the changes is concerned, evapotranspiration is higher in the WARM experiments than in the CTL integrations in all the months simulated. Precipitation is higher in the WARM experiments from March to June due to an enhancement of convective activity during these months (not shown). Later in the year (from July to September), precipitation differences between the CTL and WARM simulations are negligible.

The temporal evolution of the net input of water in the soil (P-E-R) over the Midwest subdomain is shown in the bottom half of Figure 4. From March to June, the increase in precipitation is higher than the increase in evapotranspiration, and this extra input of water can thus be stored in the soil. This is the case because the soil is not at saturation (see top half of Figure 4). In July, the increase in evapotranspiration remains substantial, while precipitation is the same as in the CTL integrations. For this reason, there is an enhanced depletion of soil moisture during this month, but due to the higher storage of water during spring, soil moisture in the WARM simulations reaches lower values than in the CTL simulations only by the end of August.

This sequence of events is well apparent on the top half of Fig. 4 which displays the temporal evolution of the soil moisture saturation in the root zone. It should be noted that

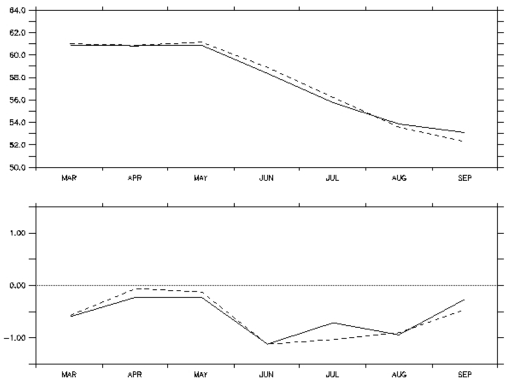

Fig 4. Temporal evolution of the relative soil moisture content in the root zone in [% of saturation] (top) and net input of water in the soil (P-E-R) in mm/day (bottom); the values are averaged over the Midwest subdomain (outlined in Fig. 1); CTL (solid lines) and WARM (dashed lines) integrations (average over all years).

the differences between the CTL and WARM experiments are again very small. The highest (positive or negative) changes are of the order of 0.5-0.8% of soil saturation or 2.5-5.5 mm in the root zone, the soil layer of relevance for plant growth. In the total column of soil, changes range from +7.1 mm (July) to -3.3 mm (September). For comparison, WM95 report mean soil moisture decreases of the order of 10 to 30 mm in their simulations.

Figure 5 displays the mean summertime (June to August) differences in moisture saturation of the root zone (1 to 2 m depth) between the WARM and CTL simulations over the whole model domain. It shows that there is no important drying in the focus region. Indeed, the highest drying peaks observed in the Midwest are of the order of 2% of the saturation water content, i.e. about 10-20 mm. Furthermore, many parts of this region display no changes in soil moisture at all or even some signs of soil wetting (for instance in the states of Missouri, Kansas, and South and North Dakota). A possibly interesting feature is the large drying observed in the western Gulf Coast region around 30N which is caused by a large decrease of summer precipitation (not shown).

Fig 5. Average summer (June-August) changes in relative soil moisture content [% of saturation] in the root zone (1-2 m depth) for the 4 years simulated (WARM-CTL). Shading occurs above and below differences of +2% and -2% of saturation. A weak spatial smoothing has been applied for display purposes.

In summary, the WARM experiments show little sensitivity to the applied forcing. The main changes observed are a wetter spring with enhanced convective activity (from March to June/July), followed by a period with drier climatic conditions (July-September). Although the summertime depletion of soil moisture in the WARM experiments is higher than in the CTL integrations, it is generally balanced by the higher volume of water stored in spring, when convective precipitation is enhanced. Due to this compensating mechanism, the WARM integrations start presenting signs of soil drying by late August only. All these changes are of very small magnitude compared to the results of earlier studies on this issue.

5. Conclusions

The relatively mild changes observed in our simulations can mainly be explained by two factors. First, the soil is not fully saturated in spring and can thus absorb the extra precipitation occurring during this season. In the simulations of WM95 however, there are no compensating effects for the increases in evapotranspiration, because most of the enhanced spring precipitation is lost to runoff.

Second and perhaps more importantly, increases in evapotranspiration are relatively moderate; this contributes towards restricting soil moisture depletion occurring in summer. As mentioned in the introduction, the bucket model is known to overestimate latent heat flux compared to sensible heat flux when energy and soil water are available, a fact which might exaggerate the simulated summer drying in climate change simulations conducted with GCMs including this representation of land surface processes.

However, our simulations entail various simplifications, which might question some of the results. First, our methodology does not allow for global changes in the synoptic-scale circulation patterns. Possible shifts in the storm tracks could be important features of climate change and are at present still difficult to predict (e.g. Kattenberg et al. 1996, IPCC 2001). Second, it is possible that the slight moisture deficit displayed by the integrations towards the end of the summer might gain importance in multi-year simulations. Third, some factors which were not accounted for in the present simulations (e.g. changes in aerosol concentrations, vegetation feedbacks) might also play an important part in this issue.

Despite the aforementioned limitations, our results pinpoint the importance of land-surface processes in climate integrations and suggest that the risk of enhanced summer dryness in the studied region might be less acute than previously assumed.

6. References

Dickinson, R.E., and A. Henderson-Sellers, 1988: Modelling tropical deforestation: A study of GCM land-surface parametrizations. *Q. J. R. Meteor. Soc.*, **114**, 439-462.

Dickinson, R.E., A. Henderson-Sellers, and P.J. Kennedy, 1993: Biosphere-atmosphere transfer scheme (BATS) version 1e as coupled to the NCAR community climate model. NCAR technical note, 72 pp.

Gates, W.L., A. Henderson-Sellers, G.J. Boer, C.K. Folland, A. Kitoh, B.J. McAvaney, F. Semazzi, N. Smith, A.J. Weaver, and Q.C. Zeng, 1996: Climate models - evaluation. in: Climate Change 1995, The science of climate change, Contribution of working group I to the second assessment report of the Intergovernmental Panel on Climate Change. Houghton JT and others (eds.), 285-357. Cambridge University Press.

Giorgi, F., and L.O. Mearns (1999) Introduction to special section: Regional climate modeling revisited. *J. Geophys. Res.*, **104**, D6, 6335-6352.

Henderson-Sellers, A., K. McGuffie, and A.K. Pitman, 1996: The project for intercomparison of land-surface parametrization schemes (PILPS): 1992 to 1995. *Clim. Dyn.*, **12**, 849-859.

IPCC 2001, Third Assessment Report WG I. Cambridge University Press, in press.

Karl, T.R., T.N. Williams Jr., T.N. Quinlan, and T.A. Boden, 1990: United States Historical Climate Network (HCN) serial temperature and precipitation data. *Technical Report Environmental Science Division, Publication No. 3404*, Oak Ridge National Laboratory, Oak Ridge, TN. 389 pp.

Kattenberg, A., F. Giorgi, H. Grassl, G.A. Meehl, J.F.B. Mitchell, R.J. Stouffer, T. Tokioka, A.J. Weaver, and T.M.L. Wigley, 1996: Climate models - projections of future climate. In: Climate Change 1995, The science of climate change, Contribution of working group I to the second assessment report of the Intergovernmental Panel on Climate Change. Houghton JT and others (eds.), pp. 285-357. Cambridge University Press.

Manabe, S.,1969: Climate and ocean circulation. Part I: The atmospheric circulation and the hydrology of the earth’s surface. *Mon. Wea. Rev.*, **97** , 739-774.

Manabe, S., R.T. Wetherald, and R.J. Stouffer, 1981: Summer dryness due to an increase of atmospheric CO2 concentrations. *Clim. Change*, **3**, 347-386.

Schär, C., C. Frei, D. Lüthi, and H.C. Davies (1996) Surrogate climate-change scenarios for regional climate models. *Geophysical Research Letters*, **23**, 6, 669-672.

Wetherald, R.T., and S. Manabe, 1995: The mechanisms of summer dryness induced by greenhouse warming. *J. Climate*, **8**, 3096-3108.

Wetherald, R.T., and S. Manabe, 1999: Detectability of summer dryness caused by greenhouse warming. *Clim. Change*, **43**, 495-511.