Validation of a regional atmospheric model over Europe: Sensitivity of wintertime and summertime simulations to selected physics parametrizations and lower boundary conditions

By FILIPPO GIORGI and MARIA ROSARIA MARINUCCI

1National Center for Atmospheric Research, P.O. Box 3000, Boulder, Colorado 80307, U.S.A.
2Dipartimento di Fisica, Università degli Studi-L’Aquila, L’Aquila, Italy

(Received 12 January 1991, revised 10 June 1991)

Summary

This paper analyses wintertime and summertime simulations over Europe using the National Center for Atmospheric Research (NCAR) regional atmospheric model (MM4). The sensitivity of the model to selected physics parametrizations (explicit moisture scheme; slower release of condensation heat; horizontal diffusion on q-surfaces), and lower boundary conditions (sea surface temperature; initial soil moisture; snow cover) is examined. The simulation periods are January 1979 and June 1979: initial and lateral meteorological boundary conditions are provided by analyses of observations. The main focus of the analysis is on prediction by the model of surface air temperature and precipitation, but also sea-level pressure and upper-air variables are considered. In general, the model reproduces the main features of the synoptic events which prevailed during the two simulation periods. Mid-tropospheric and upper-tropospheric biases in the model are generally small. Surface maximum and minimum air temperature biases do not exceed a few degrees K over various regions of Europe. The largest model bias (2–3 K) is found for summertime minimum temperature. Wintertime surface temperatures are sensitive to the precipitation parametrizations tested, as these affect cloud formation and, in turn, the surface radiative fluxes. Also, temperature biases in the driving large-scale fields are partially transmitted to the surface air temperature calculations. Summertime temperatures are sensitive to the soil moisture content. Sea surface temperature variations influence land temperatures more strongly for wintertime than for summertime. Precipitation is sensitive to the parametrizations used. The explicit moisture scheme generally induces underprediction of precipitation. When this is not used, the biases are of the order of –10 to –20% of the observations. Summertime precipitation over the highest Alpine regions is overpredicted, except when diffusion of moisture and temperature along terrain-following g-surfaces is strongly reduced. Sea surface temperature variations substantially affect precipitation, especially over coastal areas, both in January and June. Summertime precipitation is significantly modified by the initialization of soil water content.

1. INTRODUCTION

In a previous paper, Giorgi et al. (1990) (hereafter referred to as GMV) described the first stages of the development of a Limited Area Model (LAM) nested in a General Circulation Model (GCM) for regional climate simulation over Europe. The models used were the Community Climate Model (version CCM1) of the National Center for Atmospheric Research (NCAR) and the NCAR/Pennsylvania State University meso-scale model (version MM4). The main focus was on present-day wintertime conditions, and the results clearly showed that the embedded model yielded a large improvement in the simulation of regional climatic detail compared to the GCM model alone. The nested MM4/CCM model produced regional distributions of climatic variables such as surface air temperature, precipitation, cloud cover, and snow cover, which compared well with available high-resolution climatic observations. It was concluded by GMV that, at present, the nested LAM/GCM modelling approach can provide a powerful tool for improving the simulation of possible regional effects of global climatic change over Europe.

The work of GMV was exploratory and illustrative in nature, and more testing and validation is needed to evaluate the potential and limitations of the model for the production of regional climate change scenarios. In fact, the first step in the development

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.
of a nested LAM/GCM regional climate model is an accurate validation of the LAM climatology over the area of interest, so that systematic model biases can be identified and, eventually, corrected. This can be achieved by simulating ensembles of observed weather events with the LAM, and comparing the model results with the observations of climatic variables for those events. The purpose of this paper is to undertake such an analysis of the MM4 climatology for the region encompassing western Europe and the Mediterranean basin.

Indeed, although the MM4 has been widely used in the last decade or so, it has not been applied extensively to the European region. Anthes and Kaiser (1979) first tested it by successfully simulating a case of Alpine cyclogenesis. The role of Alpine topography on cyclone development during a case of tropopause folding was then analysed by Jacobs and Kuo (1989), and, more recently, Vukicevic and Errico (1990) performed an analysis of the influence of topographic forcing on predictability over the area. In addition, GMV showed that, even with coarse-scale initial and lateral meteorological boundary conditions provided by the CCM1, the MM4 captured cyclogenetic processes in the lee of the Alps and over the Gulf of Genoa. All these studies employed versions of the MM4 differing in some of their physics parametrizations.

In this work we present a much more extensive and systematic study of the climatology of the MM4 over Europe. We analyse two sets of month-long MM4 simulations, for wintertime and summertime conditions, with driving meteorological initial and lateral boundary conditions provided by analyses of observations. This allows direct comparison with observations of surface variables for the simulated time periods. In our study we calculate objective measures of simulation skill, such as biases and threat scores, and focus on surface air temperature and precipitation—variables that are most often considered in climate change and climate impact studies. However, we also examine sea-level pressure and upper-air variables. We discuss a number of sensitivity experiments aimed at evaluating and testing the effects of different physics parametrizations and surface conditions on precipitation and surface air temperature. Some of these experiments were specifically designed for the purpose of understanding model biases and uncertainties identified by GMV, while others were designed to test the effect of precipitation parametrizations which are computationally inexpensive enough to allow their use in long-term simulations, and which have become available since the work of GMV.

In section 2 we first present a description of the model and of the objective measures of simulation skill considered in this work. In sections 3 and 4 we then analyse wintertime and summertime experiments, respectively. Finally, section 5 summarizes our main conclusions.

2. DESCRIPTION OF THE MODEL AND OF THE OBJECTIVE MEASURES OF SIMULATION SKILL

The version of the MM4 used here is the same as that described by GMV. The basic structure of the MM4 is described by Anthes and Warner (1978) and Anthes et al. (1987). It is a hydrostatic, compressible, primitive-equation model written in terrain-following \( \sigma \)-coordinates, where \( \sigma = (p - p_i)/(p_s - p_i) \), \( p \) is pressure, \( p_s \) is surface pressure, and \( p_i \) is the pressure at the model top. For our runs \( p_i \) is set at 1 mb, with 16 vertical \( \sigma \)-levels, 5 in the stratosphere, 6 equally spaced levels (\( \Delta \sigma = 0.1 \)) between \( \sigma = 0.25 \) and \( \sigma = 0.75 \), and 5 levels below \( \sigma = 0.75 \).

The following physics parametrizations are used here:
(a) Horizontal and vertical turbulent eddy diffusion

The MM4 includes a fourth-order horizontal eddy diffusion term, applied on \( \sigma \)-surfaces, for wind, temperature, and water vapour. The eddy diffusion coefficient, \( K_H \), is given by (Anthes et al. 1987)

\[
K_H = K_{H0} + \frac{1}{2} k^2 \Delta s^2 D
\]

where \( k \) is the Von Karman constant, \( \Delta s \) is the gridpoint spacing,

\[
D = \left\{ \left( \frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right)^2 + \left( \frac{\partial v}{\partial x} + \frac{\partial u}{\partial y} \right)^2 \right\}^{1/2}
\]

is the horizontal deformation (Smagorinsky et al. 1965) and \( K_{H0} \) is a background value equal to \( 1.5 \times 10^{-3} \Delta s^2 / \Delta t \) (\( \Delta t \) is the model time-step).

Tibaldi 1982 and Simmons (1986) have shown that horizontal diffusion on \( \sigma \)-surfaces may cause excessive warm-season precipitation in the presence of steep terrain by diffusing temperature and moisture along the mountain slopes and thus spuriously warming and moistening the mountain tops. To assess the importance of this effect we also tested a slightly modified formulation of the horizontal diffusion scheme in which the background value, \( K_{H0} \), which is usually dominant over the deformation term, is multiplied by the correction factor, \( H_g \), given by

\[
H_g = \frac{1}{1 + (\nabla h/0.001)^2}
\]

where \( \nabla h \) is the local topographical gradient, i.e. the larger of \( \Delta h_x / \Delta s \) and \( \Delta h_y / \Delta s \), where \( \Delta h_x, \Delta h_y \) are the differences in topographical heights between adjacent gridpoints in the \( x \) and \( y \) direction, respectively. The effect of the correction term \( H_g \) is to reduce the background horizontal diffusivity in areas of large topographical gradients. For our domain it is typically of the order of 0.1–0.02 corresponding to the main mountain systems, and thus considerably reduces diffusion along the mountain slopes. Note that various formulations of \( H_g \) were tested before choosing Eq. (2). Finally, vertical eddy diffusion is performed using a second-order term with a stability-dependent diffusion coefficient (Anthes et al. 1987).

(b) Radiative transfer

For radiative transfer calculations GMV coupled to the MM4 the package developed for the CCM1 (Kiehl et al. 1987). We use here the same arrangement. It includes computations of surface fluxes and atmospheric heating rates for solar and longwave radiation, under clear and cloudy sky conditions, and explicitly accounts for absorption/ emission by \( O_3 \), \( H_2O \), \( CO_2 \) and \( O_2 \), Rayleigh scattering, reflection between clouds and surface, and multiple reflection between cloud layers. Clouds are treated as grey bodies, with emissivity depending on the vertically integrated cloud water content. The fraction of a gridpoint which is covered by clouds is equal to \((RH - 80)^2/400\) (Slingo 1980) for relative humidity \((RH)\) greater than 80%, and to zero for lower relative humidity.

(c) Precipitation

For convectively unstable precipitation GMV used the cumulus parametrization of Anthes (1977) in the simplified form described by Anthes et al. (1987). This is a Kuo-
type parametrization, in which precipitation is initiated when the moisture convergence in a column exceeds a given threshold and the vertical sounding is convectively unstable. A fraction of the total moisture convergence precipitates, depending on the mean columnar relative humidity, while the remaining fraction is redistributed throughout the column in proportion to the dryness of the vertical gridpoint. Latent heat of condensation is redistributed between cloud top and bottom following a specified parabolic vertical heating profile which yields maximum heating in the upper half of the cloud layer.

While GMV, Giorgi and Bates (1989), and Anthes et al. (1989) found that this parametrization yielded good precipitation prediction for wintertime conditions, Giorgi (1991) demonstrated that it produced a large number of excessively strong gridpoint rain events, which he called 'numerical-point storm events'; especially over mountainous terrain. Giorgi (1991) attributed the occurrence of such events to a local dynamical feedback process whereby large releases of condensation heat during convective activity intensify local upward motions and, thus, moisture convergence. This in turn keeps the cumulus convection scheme locally active and leads to excessively large precipitation rates. Modifications to the standard Kuo scheme were thus introduced by Giorgi (1991), by which subgridscale latent heat of condensation and cumulus-cloud moistening produced by the Kuo scheme are not released instantaneously on the resolvable scale. Rather, they are accumulated into a three-dimensional buffer field and released with a timescale of 6 hours. In addition, they are subject to a horizontal diffusion operator. These modifications, which are hereafter referred to as 'slower condensation heat release', were shown to substantially reduce the occurrence of numerical-point storm events. In this paper we test the standard Kuo scheme as well as that including slower condensation heat release.

Two parametrizations are tested for non-convective rain. In the first, that used by GMV, all supersaturated water vapour at a given gridpoint is instantaneously precipitated. The second, which treats resolvable-scale cloud processes in a more realistic fashion, is the explicit moisture scheme of Hsie et al. (1984). This consists of prognostic equations for cloud water and rainwater mixing ratio, and includes advection by resolvable-scale winds, diffusion by subgridscale motions, condensation/evaporation of cloud water and rainwater, autoconversion of cloud water into rainwater, aggregation of cloud water by rainwater, and gravitational settling of rainwater. The water production terms are based on the bulk formulation of Kessler (1969). Note that the explicit moisture scheme only treats resolvable-scale cloud and rain processes, therefore it has been used only in conjunction with the cumulus-cloud scheme representing convective processes. Cumulus-cloud and explicit moisture schemes are coupled as described by Zhang et al. (1988). Note that, when using both the simple precipitation of supersaturated water and the explicit moisture scheme, the fractional cloud cover for radiative transfer calculations is obtained from Slingo's (1980) formula.

(d) Boundary-layer physics

As in GMV, we use an explicit, medium-resolution boundary-layer formulation with five levels at approximately 40, 110, 310, 730, 1400 m above the surface. For stable or near-neutral conditions, the vertical turbulent transport of momentum, heat, and moisture in the boundary layer is represented by an eddy diffusion term. The eddy diffusivity follows Blackadar (1979) and depends on the vertical wind shear, the parcel mixing length, and the local Richardson number. For unstable conditions, the turbulent vertical transport of heat and moisture is mostly brought about through a dry convective adjustment
scheme. In addition to the standard temperature adjustment, the scheme redistributes atmospheric moisture vertically so that the relative humidity is equal in the adjusted layers, and the vertical integral of water vapour mass is conserved. The performance of this scheme was compared with that of others by Giorgi and Bates (1989).

(e) Surface physics

Calculations of the physics at the surface are carried out using the package of the physics/soil hydrology at the surface developed at NCAR by R. Dickinson and collaborators (the Biosphere-Atmosphere Transfer Scheme, or BATS). BATS is a model of land-surface processes designed to describe the role of vegetation in modifying the surface fluxes of momentum, heat and moisture. The version of BATS used here is described by Dickinson et al. (1986). The scheme comprises a vegetation layer, a snow layer, a surface soil layer 10 cm thick, and a deep soil layer, or root zone, 1–2 m thick. At a given gridpoint (or grid-box), a seasonally-dependent fraction of surface covered by vegetation is specified and the remaining fraction is assumed to be covered by bare soil. Prognostic equations are solved for the temperature of the surface soil layer and of the root zone using a generalization of the force-restore method of Deardorff (1978). At gridpoints where vegetation is present, the temperature of canopy air and canopy foliage is calculated diagnostically by means of an energy-balance formulation taking account of sensible heat, latent heat, and radiative fluxes.

The soil hydrology calculations include predictive equations for the water content of the surface soil layer and the root zone. These equations account for precipitation, snowmelt, canopy foliage drip, evapotranspiration, surface runoff, infiltration below the root zone, and diffusive exchange of water between soil layers. Snow depth is prognostically calculated from snowfall, snowmelt, and sublimation. Precipitation is assumed to fall in the form of snow if the temperature at the atmospheric model level closest to the surface is less than or equal to 271 K (Auer 1974).

Fluxes of sensible heat, water vapour, and momentum at the surface are calculated using a standard surface-drag coefficient formulation based on surface-layer similarity theory. The drag coefficient depends on the surface roughness length and on the atmospheric stability in the surface layer. The surface evapotranspiration rates depend on the availability of soil water, which is a prognostic variable and varies with time. BATS can currently accommodate 15 vegetation types, soil textures ranging from coarse (sand), to intermediate (loam), to fine (clay), and different soil colours (light to dark) for the soil-albedo calculations. These are described by Dickinson et al. (1986).

BATS and the CCM1 radiative transfer scheme were coupled to the MM4 in order to facilitate its use for climate applications. Their use requires the assignment of a land-surface and vegetation type for each gridpoint, as well as the concentrations of CO\textsubscript{2} and O\textsubscript{3}. The former were obtained from the standard MM4 land-use data archive (Giorgi and Bates 1989). For CO\textsubscript{2} and O\textsubscript{3} we use a constant mixing ratio of 330 ppmv, and the mid-latitude summer and winter vertical profiles of McClatchey et al. (1971), respectively. All these quantities are the same as in GMV.

The model domain and topographical field are shown in Fig. 1. The model uses a Lambert conformal projection, with a domain of 4000 × 3700 km\textsuperscript{2} size centred over the western Mediterranean. The horizontal gridpoint spacing is 70 km. At this resolution the main features of the Alps, the Iberian plateau, the Pyrenees, the Balkans, the Apennines and the Grampian mountains in Scotland are captured.

The periods of simulation are January and June 1979. Sea surface temperatures and meteorological initial and lateral boundary conditions (wind components, tempera-
ture, water vapour mixing ratio, and surface pressure) necessary to drive the model runs are interpolated from the European Centre for Medium Range Weather Forecast (ECMWF) original IIb global analysis of FGGE (First GARP Global Experiment) data (Bengtsson et al. 1982; Mayer 1988; Trenberth and Olson 1988). These time periods were chosen for the good quality, relatively high resolution of the global analysis, and easy access of this analysis at NCAR.

The ECMWF data have a resolution of $1.875 \times 1.875$ degrees in the horizontal, are distributed on 15 pressure levels extended to 10 mb and are spaced at intervals of 12 hours. Vertical interpolation from the analysis fields to the model grid is linear in pressure for wind and relative humidity, and linear in the logarithm of pressure for temperature. The lateral boundary conditions are provided using the relaxation technique described by Davies and Turner (1977) and Anthes et al. (1987). This includes a Newtonian and a diffusion term gradually applied to the outermost four gridpoint rows of the domain which drive the model solution toward the specified boundary-condition values. At these gridpoints it is the lateral-boundary-conditions' terms rather than the model's internal physics that dominate the solution of the model equations. Only at the outermost gridpoint row is the model solution forced to be exactly the same as the boundary value. For this study, only analysis fields at 12-hourly intervals were available. At this interval it is possible that fast-moving systems may cross the boundaries of the domain without being 'detected' by the lateral boundary conditions. Future work should investigate the effect of the frequency of update of large-scale lateral boundary conditions.

A few comments on the ECMWF analysis would be useful (Bengtsson et al. 1982; Uppala 1986; Arpe 1986). First, the IIb analysis, and thus our model, included not the observed but the climatological sea surface temperatures for the FGGE period. The
ECMWF model used for the analysis included smoothed topography, interactive radiative transfer and cloud calculation, turbulent vertical eddy diffusion, a Kuo cumulus convection scheme, full hydrologic cycle and computed ground temperatures, although the diurnal cycles were not incorporated in the calculations. Overall, the IIIb analysis was considered to be of good quality (Bengtsson et al. 1982). The main inaccuracies, which may affect our study, were found in the fields of humidity analysis. The humidity analysis scheme was in the early stages of development (in fact the humidity fields were to a large extent provided by the model's first-guess forecasts; Uppala 1986) and the analysis tended somewhat to underestimate the amounts of lower tropospheric moisture (Arpe 1986). Subsequent analyses have improved the quality of the humidity fields; we plan to use these analyses for future tests of the model.

Concerning the initialization of BATS variables, surface soil and canopy temperatures are set equal to that of the lowest level of the model and the root-zone temperature is initialized with an average of the lowest level of temperature of the model interpolated from the ECMWF data-set for the whole period of simulation. Initial snow depths are set to zero for simplicity. Finally, the soil water content of the surface soil layer and of the root zone (expressed in cm of water), $w_s$, is initialized using the formula suggested by Giorgi and Bates (1989) and used by GMV:

$$w_s = w_{\text{wilt}} + M_A (w_{\text{sat}} - w_{\text{wilt}})$$

(3)

where $w_{\text{wilt}}$ is the soil water content at which transpiration ceases; $w_{\text{sat}}$ is the water content at saturation (see Dickinson et al. 1986); and $M_A$ is the moisture-availability parameter used in the standard MM4. This is specified for each surface type and varies for most of Europe from 30% (grassland) to 60% (forest) for wintertime conditions, and from 15% (grassland) to 30% (forest) for summertime conditions (see Anthes et al. 1987). Unfortunately, somewhat of a mismatch is evident between the complexity of the physics package at the surface in the model and the crudeness of the initialization of the surface variables. This is due to the lack of data necessary for such initialization. Hopefully, with the development of remote-sensing techniques, it will be possible to obtain surface information of sufficient quality to be used in relatively sophisticated models of the physics at the surface.

Because we are mostly interested in the model's climatological accuracy rather than the model's day-to-day forecast skill, instead of carrying out simulations of individual 2-day or 3-day weather events we performed relatively long (month-long) simulations encompassing several events and selected objective measures of simulation skill based on the (30-day) aggregated characteristics of this ensemble of events. The first measure we called model large-scale bias, i.e. the deviation in the model prediction of midtropospheric and upper-tropospheric variables from the driving large-scale fields. Although the model's large-scale climatology is ultimately driven by the analyses which are used to provide the lateral meteorological boundary conditions, the internal model physics, in general, produces a shift between the model solution and the driving large-scale fields. This shift, which is measured by the large-scale bias, gives an indication of the differences between model-produced and driving large-scale climatologies.

To calculate the model's large-scale bias for a given region, at a given model $\sigma$-level (or pressure level), $k$, we first define the instantaneous, mass-weighted, spatial-mean error at the level $k$ and time $t$, $E_{m,k}(t)$, as

$$E_{m,k}(t) = \frac{\sum_i \sum_j \left[ p_{i,j,k}^M(t) F_{i,j,k}^M - p_{i,j,k}^A(t) F_{i,j,k}^A \right]}{\sum_i \sum_j p_{i,j,k}^A(t)}$$

(4)

where $i$ and $j$ are the horizontal gridpoint indexes in the eastward and northward
directions, respectively; $F$ is the meteorological variable of interest (e.g. temperature, moisture, wind); and the superscripts $M$ and $A$ refer to the model and the analysis, respectively. The instantaneous errors are calculated every 12 hours, i.e. at the times when analysed data are available for verification. The model's large-scale bias at the level $k$ is then computed by time-averaging Eq. (4) over the whole period of simulation.

Note that the calculation of the large-scale bias is made on the model grid rather than the analysis grid. Since the analysis data has a resolution lower than that of the model data, it cannot capture some mesoscale features predicted by the model. While this could substantially affect the calculation of spatial-mean-square errors, it is not important for the biases.

For verification of precipitation and surface air temperature we use an observational station data-set developed by the Climate Analysis Center of the United States National Meteorological Center (NMC) available at NCAR, which reports daily precipitation and daily maximum and minimum surface air temperature at stations located throughout the World (documentation for this data-set is available from the NCAR Data Support Section). We used data from 308 European stations for January 1979 and 364 for June 1979. Only stations were selected for which at least 25 of what are defined as 'valid extreme temperature (or precipitation) reports' were available during the selected month. As an example, the locations of the stations used for verification of the January 1979 simulations are shown in Fig. 2. Unfortunately, the station distribution is not uniform, with maximum density over central Europe and minimum density over the Mediterranean countries. We performed a few tests in which we added about an extra 300 stations including a lower number of valid reports and found that the results differed little (a few tenths of a degree for temperature and less than 20% for precipitation) from those

![Figure 2](image_url)  
Figure 2. Location of the stations used for validation of model surface air temperatures and precipitation for January 1979. Also shown are four climatic subregions considered in this work: M—Mediterranean region, WE—Western Europe, CE—Central Europe, A—Alpine region.
obtained with the smaller but more complete data-set. Therefore, we decided to use for our calculations the set of stations in Fig. 2. (The station distribution for June 1979 is similar to that of Fig. 2). Also shown in Fig. 2 are four European climatic subregions considered in this work: the Mediterranean region (including southern France and the Iberian, Italian, and Balkan peninsulas), the Alpine region, western Europe (the British Isles, most of France, the Benelux countries and western Germany), and Central Europe (most of the eastern European countries along with western Soviet Union).

For precipitation and maximum/minimum daily temperature we calculate what we call the model surface bias $B_a$, i.e. the average deviation of the simulated variable, $a$, (precipitation and temperature) from observations. This is defined as

$$B_a = \frac{1}{N_S} \sum_{n=1}^{N_S} (a^M_n - a^O_n)$$

(5)

where $N_S$ is the total number of daily reports for all stations in a given region for each simulated month, $a^O_n$ is the observed daily value of the variable $a$ at a given station, and $a^M_n$ is the corresponding model value linearly interpolated from the MM4 grid to the station location. As explained later (section 3(b), Eq. 7), in the calculation of the temperature biases a correction is added to the station data to account for differences in the station and model elevations.

A second measure of precipitation simulation skill is the 30-day accumulated precipitation threat score, $T_{P_T}$, for the precipitation threshold $P_T$, which is defined as

$$T_{P_T} = \frac{C_{P_T}}{O_{P_T} + F_{P_T} - C_{P_T}}$$

(6)

In (6), $O_{P_T}$ is the number of stations with observed precipitation accumulated over the whole simulation which is in excess of $P_T$, $F_{P_T}$ is the corresponding number of model forecasts at the station locations, and $C_{P_T}$ the number of stations where both observed and forecast precipitation exceeds $P_T$. The precipitation threat score is a measure of the model spatial and temporal simulation skill. It measures the model accuracy in predicting the area that, in a given period, receives an amount of precipitation above a given threshold, varying from 0 to 1. A threat score of 1 indicates a perfect forecast and the accuracy of the forecast decreases as the threat score decreases. Only days for which observed values are available are included in Eqs. (5) and (6).

It is evident from the choice of our objective measures of simulation skill that the present analysis is limited in scope and is specifically directed at evaluating the model climatology as measured by the model performance in reproducing the average characteristics of ensembles of events. In general it is desirable that the model reproduces not only average climate variables but also higher-order climate statistics, such as distribution of daily precipitation. Evaluation of the model's higher-order statistics, however, probably requires larger samples and will be performed when longer model runs become available.

In the following sections we describe our verification analysis for wintertime and summertime simulations.

3. ANALYSIS OF WINTERTIME SIMULATIONS

As a test of the model performance for wintertime conditions, we carried out simulations of January 1979, and more specifically of the 30-day period beginning 1 January 1979, 0000 GMT, and ending 31 January 1979, 0000 GMT. We performed six
experiments. The first, JANCON, includes the most comprehensive physics packages, i.e. standard MM4 horizontal diffusion, the CCM1 radiation code, BATS, the medium resolution planetary boundary-layer scheme, the explicit moisture scheme and the modifications to the standard Kuo scheme to allow slower release of condensation heat. Since JANCON contains the most comprehensive model physics, for discussion purposes we considered it as our control simulation. To test the effect of an explicit moisture scheme, slower condensation heat release, and a correction factor to the horizontal diffusion coefficient (Eq. (2)), we carried out experiments JANNOEX, JANGMV and JANDIF, respectively. In the case of JANNOEX we removed the explicit moisture scheme and replaced it with precipitation of supersaturated water, in JANGMV we removed the explicit moisture scheme as well as the slower release of condensation heat (i.e. the model is run with the same configuration as by GMV), while in JANDIF we added the horizontal diffusion correction factor to the model configuration of JANNOEX.

In the next experiment, JANTM3, the model physics is as in JANNOEX, but the temperatures in the large-scale driving fields are decreased by 3 K (in addition, the water vapour mixing ratio is modified in the initial and lateral-boundary-condition fields so as to keep the relative humidity in this experiment equal to that of JANNOEX). Experiment JANTM3 was designed to evaluate the extent to which temperature biases in the driving large-scale fields are transmitted to the surface-temperature calculations. The main motivation for this experiment was to assess if, in the simulations of GMV, the 2–4 K cold tropospheric bias in the driving CCM1 contributed to the 2–3 K cold surface temperature bias of the nested MM4.

Another source of uncertainty identified by GMV was the snow field initialization. Although BATS calculates snow depth in a physically consistent way from snowfall, snowmelt, and sublimation, the initial snow field of GMV was interpolated from the driving CCM1, which simply assumed a uniform snow depth of 10 equivalent mm of water for land areas north of about 42°N. Not having any quantitative information on snow depth for January 1979, we initialized it with zero in most of our runs. To test the importance of snow initialization we carried out a run, JANSN, in which, with all conditions as in JANNOEX, the snow depth was initialized as by GMV, i.e. with a value of 10 equivalent mm of water for land areas north of 42°N.

Finally, in a previous paper, Giorgi (1990) found that the simulation of precipitation over the coastal states of North America is sensitive to the sea surface temperature of upwind coastal oceanic waters. Often, sea surface temperature values are known only approximately, especially for climate conditions different from the present. To evaluate the importance of such uncertainty, as well as the role that variations in sea surface temperature can have on surface air temperature and precipitation, we performed an experiment, JANSSTP3, in which all conditions are as in JANNOEX, but the sea surface temperatures within our domain are increased uniformly by 3 K. Note that most of our runs do not include the explicit moisture scheme since this increases by about 20% the computational time necessary to run the model.

In the next subsection we first present a description of the main synoptic events which occurred during January 1979 and the ability of the MM4 in reproducing them; then we discuss results from the various sensitivity experiments described above.

(a) Description of synoptic events occurring during January 1979

To illustrate the synoptic patterns which prevailed during January 1979, and the skill of the MM4 in reproducing such patterns, we compare here sea-level pressure fields, produced by the model at different times during the JANCON simulation, with those of
the ECMWF analysis of observations. Although model results are presented only for the JANCON run, the basic conclusions of this subsection do not differ significantly for the other runs.

Wintertime weather over western Europe is generally characterized by the passage of North Atlantic cyclonic disturbances associated with the Icelandic low, and Mediterranean storms originating south-west of the Iberian peninsula. Several such disturbances occurred during January 1979. The first major system entered our domain on 4 January, when a deep low-pressure cell moved towards the north-western coasts of France from the North Atlantic. As shown by Fig. 3(a), the low reached the French coasts on about 5 January and travelled to south-eastern Europe, where it eventually dissipated on 6 January. A cold front associated with this system swept northern Spain, France, northern Italy and the Balkan peninsula, inducing widespread precipitation there (Fig. 3(b)). The model reproduces well the location of the main cell as it enters the continent and dissipates over south-eastern Europe, although the values of sea-level pressure in the model are lower than in the observations.

![Figure 3](image)

By 7 January an anticyclonic system reached the Mediterranean basin and central Europe where it resided until 9 January. On 9 January a low entered the western Mediterranean from the Moroccan region and started moving in a north-easterly direction. This disturbance reached the Gulf of Genoa on 10 January (Fig. 4(a)), deepening owing to the ocean heat fluxes, and then continued in its north-eastward direction until it dissipated over the Balkans. The path of this storm is well simulated by the model, which yields substantial precipitation amounts along the storm’s track (Fig. 4(b)). Systems similar to that of Fig. 4, which are defined as weather type A in the Air Ministry’s (1962) booklet, occur quite frequently during wintertime and were observed also in the simulations of GMV.

As shown in Fig. 5(a), the Atlantic system which is located north of the British Isles on 10 January (Fig. 4(a)) reached central Europe on 12 January. As the low crossed Germany and Poland, the effect of the Alps topographical barrier produced a deformation of the cell, by which the original low elongated in a southward direction (Fig. 5(a)). On
13 January the disturbance split, with the primary system moving towards north-eastern Europe and a secondary one following a south-eastern path over the Hellenic peninsula (Fig. 5(b)). Precipitation is produced over central and, especially, south-eastern Europe (Fig. 5(c)), where low pressure persisted until about 17 January. Alpine cyclogenetic processes of the type described here are rather common synoptic events over Europe, and are strictly connected with the presence and shape of the Alps (Buzzi and Tibaldi 1978; Buzzi and Speranza 1983; Mesinger and Pierrehumbert 1986). They are an important cyclogenetic mechanism over the Mediterranean basin, especially during the winter. The motion and deformation of the primary cell are captured reasonably well by the model, which also shows the formation of the secondary cell over south-eastern Europe. Sea-level pressures of the primary cell are, however, 4–6 mb higher in the model than in the observations. In GMV it was illustrated how Alpine cyclogenetic processes similar to that shown in Fig. 5 often occur, also when the MM4 is driven by large-scale CCM meteorological fields.

On 14 January the Azores high, which during winter is usually positioned over the North Atlantic, moved over western and central Europe where it persisted for about 7 days, inducing cold meridional flow and dry conditions over most of the continent. Pushed northward by the high-pressure region, a disturbance (not shown) grazed the western European coasts between 19 and 21 January, and induced precipitation mostly over the Iberian peninsula.

The synoptic situation changed abruptly when two fast-moving low-pressure systems moved across Europe on 22–25 January (not shown). One had an evolution similar to that of Fig. 4. It entered the Straits of Gibraltar, travelled through the western Mediterranean over the Gulf of Genoa, where it attained its maximum strength, and then dissipated over the Balkans. The other followed a northern track along the Baltic Sea coasts. These storms induced heavy precipitation over southern France, the Pyrenees, the Gulf of Genoa, and northern Italy, and lighter precipitation over the central European coasts.

Between 25 January and the end of the month, weather over Europe was dominated by a system of two Atlantic storms, which are shown in Fig. 6(a-b). They are both deep.
The southernmost of these crossed the coast of Portugal on 28 January, and after about two days had crossed the whole European continent in a north-eastward direction reaching the north-western regions of the Soviet Union. Precipitation associated with this system, up to several cm, fell over the Iberian peninsula, southern France, the Alpine region, and most of Czechoslovakia, Poland and the north-western Soviet Union (Fig. 6(b)). The northernmost cell, with sea-level pressures as low as 980 mb, moved more slowly. It originated over the north-eastern Atlantic and affected mostly the British Isles and northern France. These rapid sequences of frontal passages are known to be associated with the Icelandic low, whose effect is to steer alternately cold and warm air over northern Europe, sometimes producing a final outbreak of very cold air from Greenland and Iceland over central Europe. Figure 6(a-b) indicates how model results and observations show general agreement in the description of the evolution of these low-pressure cells, although the model-produced and observed paths of the southern European cell do not coincide and the sea-level pressures in the northern European cell differ by up to 9 mb.
Finally, Fig. 7 shows the 30-day accumulated precipitation predicted in JANCON for January 1979. The model produces at least 2.5 cm of rain/snowfall over most of central and western Europe. The wettest regions are the Iberian peninsula, with a maximum exceeding 25 cm over the Pyrenees, southern France, southern Italy and areas of the Alps and Balkans, which are affected by five storms during the month. Substantial precipitation is also simulated over Ireland, Scotland and southern Norway.
The discussion presented in this section illustrates how several storm systems characterized by different storm paths, structure and intensity, developed over Europe during January 1979. The comparison with observations, although limited to a selected number of sea-level pressure 'snapshots', shows that the model captured these storm events and reproduced their basic structure, intensity and evolution. However, modelled and observed sea-level pressures differed in some instances by several mb. Especially over mountainous terrain, this could be due to the procedures used to extrapolate pressure at sea level (temperatures below the ground are estimated from the temperatures of the lowest atmospheric level assuming a standard 6.5 K km\(^{-1}\) lapse rate).

Note that no blocking events occurred during January 1979, therefore the present run cannot provide information on the model's ability to reproduce low-frequency-variability characteristics over the region. Also note that analysis and model-produced sea-level pressures are not equal at the boundaries of Figs. 3 to 6 because the outermost gridpoint rows of the domain—i.e. the only gridpoints where the model and the analysis fields are forced to be exactly the same (Anthes et al. 1987)—are not included in these figures, as well as not in all MM4 maps presented in this paper. In the next section we proceed to analyse more quantitative measures of model simulation skill.

\[(b)\] Simulation skill analysis

We first consider the tropospheric large-scale bias as defined in section 2. Again, we use for this discussion results from the JANCON case, remembering that our conclusions are essentially valid also for the other sensitivity tests. Figure 8(a–c) shows the large-scale-bias profiles for temperature, water vapour and zonal wind for land and ocean areas in the interior of the model domain (interior of the domain is hereafter defined as not including the 5 gridpoint rows closest to the lateral boundaries, since these are strongly affected by the relaxation boundary-condition terms).

In general, the temperature and water vapour biases are small in the mid to upper troposphere and increase near the surface and the top of the model. This is an indication that, while in the mid regions of the vertical domain advection from the lateral boundaries exerts a strong forcing on the model solution, near the surface and top of the model the model's processes for the physics at the surface, and the rigid top boundary conditions adopted in the model (Anthes et al. 1987) become increasingly important. Note that, because of the coarse horizontal and lower tropospheric vertical resolution of the analysis (analysis fields are available only at 1000 mb and 850 mb), the values of the near-surface analysis fields interpolated onto the model grid, which are employed to calculate the large-scale biases, may not be accurate. In particular, they may not account in detail for surface processes and boundary-layer structure. Conversely, the model includes a full description of surface processes and an explicit description of the boundary layer.

The temperature and water vapour biases vary in sign and magnitude throughout the troposphere. Temperature biases are mostly negative, being largest in the region \(\sigma = 0.70-0.90\), and decreasing to \(\leq 1\) K in the mid to high troposphere. Near the surface they are of opposite sign over land and ocean. Water-vapour biases are less than 0.25 g kg\(^{-1}\) in magnitude above \(\sigma = 0.9\) and are maximum over ocean surfaces where the model evaporation is largest. The magnitudes of the bias are generally similar to those found by GMV when the model was driven by GCM-produced large-scale lateral boundary conditions.

The zonal-wind bias is positive throughout most of the troposphere, except near the surface and the top boundary, again an indication of the increasing importance of the model's internal physics there. The positive zonal-wind bias in the troposphere is a feature which was found also by GMV for the same region, and by Giorgi and Bates
Figure 8. MM4 large-scale bias for case JANCON as a function of $\sigma$ over land and ocean areas: (a) temperature; (b) water vapour; (c) zonal wind.
ATMOSPHERIC MODEL VALIDATION

(1989) and Giorgi (1990) for the western United States. This overall intensification of the jet by the model seems thus to be a fairly consistent product of the scale transition from the coarse-resolution driving fields to the higher-resolution model grid. One of its consequences is to induce convergence near the downwind lateral boundary of the domain, which in turn can trigger convective precipitation. As shown in Fig. 7, such an effect is not important in this simulation. However, it was observed in the simulations of GMV.

Figures 9(a, b) show the 300 mb zonal wind (i.e. at jet-stream level) in the ECMWF analysis, averaged over the whole simulation, and the difference between the average zonal wind in the model and that in the analysis. A maximum in the jet stream, exceeding 30 ms$^{-1}$, is located over the central Mediterranean, while the Atlantic jet west of the British Isles is weaker, about 14 ms$^{-1}$. As shown by Fig. 9(b), the modification of the driving upper-tropospheric wind by the model is spatially quite irregular. Areas of maximum and minimum reaching up to several ms$^{-1}$ are scattered in various regions throughout the domain. Similar patterns were found at different altitudes. This result, and the fact that the wind differences of Fig. 9(b) are much smaller than the values of the wind shown in Fig. 9(a), would indicate that, over Europe, at least for this simulation, the model did not substantially modify the average large-scale circulations of the driving analysis. As another example, large-scale biases for the 500 mb geopotential heights did not exceed 10 m throughout the domain (not shown). A similar result was found by GMV, although in that work it was observed that the nested MM4 did alter to some extent the average storm track of the driving CCM1.

Turning our attention to the surface biases, Table 1 presents the maximum and minimum daily surface air temperature bias for all January experiments, for the whole set of stations in Fig. 2, and for stations lying within the four European climatic subregions depicted there. In the calculation of the bias we equated the model surface air temperature approximately to the temperature at the lowest model $\sigma$-level ($\sigma = 0.995$), which is generally located at an altitude of 30–40 m from the ground, and we excluded the first day of simulation in order to reduce errors due to the initialization of surface temperatures. This is the same procedure as was adopted by GMV. It probably introduces a positive bias for the minimum temperature and a negative bias for the maximum
TABLE 1. Maximum (TMAX) and minimum (TMIN) temperature bias (K) for different January 1979 simulations. Biases are presented as obtained using all stations and using stations in the four European climatic subregions of Fig. 2.

<table>
<thead>
<tr>
<th></th>
<th>JANCON</th>
<th>JANNOEX</th>
<th>JANGMV</th>
<th>JANDIF</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TMAX</td>
<td>TMIN</td>
<td>TMAX</td>
<td>TMIN</td>
</tr>
<tr>
<td>All stations</td>
<td>0.377</td>
<td>2.167</td>
<td>-0.500</td>
<td>0.567</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>-1.533</td>
<td>0.206</td>
<td>-2.068</td>
<td>-0.908</td>
</tr>
<tr>
<td>Western Europe</td>
<td>1.194</td>
<td>2.716</td>
<td>0.342</td>
<td>1.321</td>
</tr>
<tr>
<td>Central Europe</td>
<td>1.421</td>
<td>3.260</td>
<td>0.672</td>
<td>1.047</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>0.149</td>
<td>2.421</td>
<td>-0.470</td>
<td>0.740</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>JANTM3</th>
<th>JANSN</th>
<th>JANSSTP3</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TMAX</td>
<td>TMIN</td>
<td>TMAX</td>
</tr>
<tr>
<td>All stations</td>
<td>-1.976</td>
<td>-1.138</td>
<td>-0.715</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>-3.448</td>
<td>-2.322</td>
<td>-2.182</td>
</tr>
<tr>
<td>Western Europe</td>
<td>-0.900</td>
<td>-0.179</td>
<td>0.070</td>
</tr>
<tr>
<td>Central Europe</td>
<td>-1.701</td>
<td>-1.041</td>
<td>-0.190</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>-1.990</td>
<td>-1.121</td>
<td>-0.664</td>
</tr>
</tbody>
</table>

Temperature, since observed temperatures are typically taken a few metres from the ground. Also, we applied to the model temperatures a correction, ΔT, to account for differences between the elevation of the stations, \( H_s \), and that of the smoothed model topography at the station location, \( H_m \). This correction is given by

\[
ΔT = (H_m - H_s) \cdot Γ
\]

(7)

where \( Γ \) is the average standard tropospheric lapse rate of 6.5 K km\(^{-1}\). On average, the values of ΔT were of the order of 0.2 to 0.7 K for different regions of Europe. Daily maximum and minimum temperatures in the model were inferred from hourly model output.

In the JANCON simulation, the overall daily maximum temperature bias is small, 0.4 K, while the minimum temperature bias is higher, 2.2 K. When looking at the regional distribution of the biases we find that for maximum daily temperature the model is too cold by 1.5 K over the Mediterranean region, while it is too warm by up to 1.2 and 1.4 K over western and central Europe, respectively. A similar trend is observed for minimum daily temperatures, with largest positive bias over central Europe of about 3.3 K. Overall, however, the mean daily temperature bias does not exceed a few degrees K in any region.

Interestingly, while the land model biases near the ground of Fig. 8(a) (which were calculated with respect to the driving ECMWF analysis) are negative, the JANCON temperature biases in Table 1 are mostly positive. This would imply that the temperatures in the ECMWF analysis near the ground are higher than those of the station data-set. A number of factors may contribute to this result:

1. The biases of Fig. 8(a) are calculated over all land areas, whereas those of Table 1 are calculated over a sample of sometimes relatively sparse point values.

2. The biases of Fig. 8(a) are calculated using 12-hourly ‘snapshots’ at 0000 GMT and 1200 GMT, and not from the maximum and minimum daily temperatures.
3. The ECMWF analysis does not include the diurnal cycle.

4. The ECMWF temperature fields are interpolated to the MM4 lowest $\sigma$-level from relatively coarse horizontal (1.875° × 1.875°) and vertical (levels at 1000 mb and 850 mb near the ground) data-sets. As explained in section 2, the vertical interpolation of temperature is linear in the logarithm of pressure, and errors can be introduced by such interpolation owing to differences in ECMWF analysis and MM4 topographies, and to the coarse vertical resolution of the analysis near the ground.

Also, the results of the JANCON simulation in Table 1 contrast with those of GMV, who found a cold January mean surface air temperature bias of 2–3 K over most of Europe when the MM4 was nested in the CCM1. To understand this difference better, we need first to look at the processes which affect the surface temperature during wintertime conditions. Figure 10 shows the average diurnal cycles of absorbed solar, net infrared, sensible-heat and latent-heat fluxes at the surface for our simulation. The averages are taken over all land gridpoints in the interior of the domain and over the entire period of simulation (not including the first day).

The absorbed solar flux reaches a maximum of about 175 W m$^{-2}$ during daytime and decreases to zero during night-time. Conversely, the diurnal variation of net infrared loss from the surface has a much less pronounced diurnal cycle, with a maximum of about 45 W m$^{-2}$ and a minimum of about 33 W m$^{-2}$. The daytime loss of sensible heat reaches a maximum of about 37 W m$^{-2}$, while during night-time the sensible-heat flux changes sign (about $-20$ W m$^{-2}$). Finally, the maximum upward latent-heat flux is about 36 W m$^{-2}$ during daytime, with negligible latent-heat flux during night-time.

The only component of the surface energy budget which is missing from Fig. 10 is that due to heat exchanges between surface and deep soil, i.e. the root-zone layer. Supposing that after the first few days of simulation the temperature of the surface reaches thermal equilibrium with the energy fluxes that determine it, the energy flux from the root zone to the surface can be approximately estimated as being that necessary to balance the fluxes shown in Fig. 10. This yields a maximum daytime flux from the surface to the root-zone layer of about 57 W m$^{-2}$, and a night-time flux from the root-zone layer to the surface of about 13 W m$^{-2}$. For this run, the root-zone temperature was, on average, lower than the daytime maximum surface soil temperature by about 3 K, and it was higher than the night-time surface soil temperature by about 2 K (the root-zone temperature does not respond to the diurnal cycle). In summary, during daytime, the absorption of solar radiation gives the largest contribution to the surface energy budget, while the other fluxes have a magnitude which is 3 to 5 times smaller. During night-time, surface infrared cooling and sensible-heat exchanges with the atmosphere dominate, but diffusive heat exchange with the deep soil is also important.

Given this picture, the main differences between the model configurations of case JANCON and the model of GMV are:

(i) Precipitation parametrization. Unlike the model of GMV, JANCON includes the explicit moisture scheme and slower condensation heat release during convective activity. This can affect precipitation and atmospheric water-vapour amounts, which in turn can influence cloud cover and surface infrared and solar radiative fluxes.

(ii) Driving large-scale fields. While the MM4 is here driven by an analysis of observations, hence by realistic large-scale temperatures, in the model of GMV it was driven by the CCM output of a simulation which had a cold January temperature bias of 2–4 K. The cold bias in the driving fields is felt by the MM4 in two ways. The first is in the
Figure 10. Average diurnal cycle of absorbed solar, net upward infrared, sensible-heat (positive upward), and latent-heat (positive upward) fluxes at the surface for all land gridpoints in the interior of the domain in the JANCON simulation. Units are W m$^{-2}$.

initialization of the root-zone temperature. This is done by averaging over the entire period of simulation the bottom-level air temperatures obtained from the driving large-scale fields. If these are affected by a cold bias, as in the case of the CCM1 simulation of GMV, then also the initial root-zone temperature will have a cold bias, which in turn can affect the heat exchange with the surface. Secondly, the cold bias in the driving large-scale temperatures can be transmitted to the interior of the domain through advection from the lateral boundaries. This effect can be especially significant at the surface during wintertime when the energy fluxes are at their minimum.

(iii) Snow initialization. While in JANCON we initialized the snow-depth field with a value of zero, in the model of GMV the initial snow depth was interpolated from the standard CCM1, in which this is specified to be 1 cm of equivalent liquid water at all land points north of about 42°N. This factor can be significant, since the snow has a higher albedo than, and different heat capacity from, the underlying soil. To test the importance of these three factors, experiments JANNOEX, JANGMV, JANTM3, and JANSN were carried out.

The effect of removing the explicit moisture scheme and slower condensation heat release can be evaluated from the results of cases JANNOEX and JANGMV. From Table 1 it can be seen that the average maximum temperature bias and the minimum temperature bias become more negative in JANNOEX than in JANCON by about 0.9 K and 1.6 K, respectively. The cooling in JANNOEX compared to JANCON occurs in all regions of the domain, but is more pronounced in the continental interiors, i.e. over central Europe and the Alpine region. This is mostly due to a change in the model cloud cover. The average cloud cover over land decreases from 75% in JANCON to 58% in JANNOEX. This is caused by the absence in JANNOEX of processes like cloud water and rain water evaporation, which cause an increase in precipitation (see later), and thus a drying of the atmosphere. Decrease in cloud cover induces increases in infrared radiation loss and solar radiation absorption at the surface. In the model, the
increase in net infrared loss occurs throughout the whole day, and, on average, amounts to $15 \text{ W m}^{-2}$. Increase in absorption of solar radiation occurs only during the daytime, when it attains a maximum value of $25 \text{ W m}^{-2}$.

Because the increase in absorbed solar radiation is significant for only a few hours a day, the contribution of the infrared flux dominates, and the surface undergoes a net radiant-energy loss in JANNOEX compared with JANCON. This in turn causes a decrease in the root-zone temperature. The temperature of the root zone does not respond to a diurnal cycle of insolation. However, it responds with a characteristic time of a few weeks to changes in average energy inputs. For example, although the root-zone temperatures have the same initial values in the two runs, after about 20 days they have asymptotically reached values lower by about 2 K in JANNOEX compared with JANCON. Lower root-zone temperatures imply lower upward heat flux from the root zone to the surface soil layer. The net result of this effect, when added to the infrared and solar radiation effects, is to reduce average surface soil and vegetation temperatures in JANNOEX by about 2.5 K during night-time, and by about 1.5 K during peak daytime hours. This in turn causes the changes in surface air temperature bias shown in Table 1.

Note that the value of 75% average cloud cover over land obtained in the JANCON run appears high. Although there are areas of central and north-western Europe for which the average January fractional cloud cover exceeds 70%, the average January cloudiness over the Mediterranean countries is in the range of 50–70% (Warren et al. 1986). The excessive simulated cloud cover in JANCON is probably due to the use of the relative-humidity-dependent formula of Slingo (1980) in conjunction with the explicit moisture scheme. An alternative way of specifying fractional cloudiness at a gridpoint when using the explicit moisture scheme would be to assume 100% cloud cover only if the cloud water mixing ratio is different from zero or exceeds a given threshold. This would probably lead to lower total column cloud amounts.

Removal of slower release of condensation heat from the Kuo scheme does not produce a substantial effect. Surface air temperatures are about 0.3–0.5 K higher in JANGMV than in JANNOEX. These are induced by small changes in the surface radiative fluxes associated with variations in the vertical distribution of clouds. Compared to JANNOEX, fewer mid-tropospheric clouds, but slightly more lower-tropospheric clouds were produced in JANGMV.

To evaluate the effect of decreasing the driving large-scale temperatures we can compare results from experiments JANNOEX and JANTM3. The maximum and minimum daily temperature biases are more negative in JANTM3 than in JANNOEX by 1.5 K and 1.7 K, respectively. The surface cooling in JANTM3 is slightly more pronounced over the continental interior regions than over the coastal regions owing to the influence of the oceanic waters. In the JANTM3 run the maximum and minimum temperature bias for all regions is negative, with values ranging from $-0.9$ K to $-3.4$ K.

The surface temperature cooling in JANTM3 is mostly due to the combined effect of advection of colder air from the boundaries and decreased heat transfer from the root zone to the surface, associated with colder root-zone temperatures. On average, these were lower in JANTM3 than in JANNOEX by about 1.5 K. Average surface sensible heat, latent heat, infrared and solar fluxes over land did not differ in the two runs by more than $2–4 \text{ W m}^{-2}$. It thus appears that if a temperature bias is present in the driving large-scale fields this is, at least partially, transmitted to the surface air temperatures in the interior of the domain.

Conversely, the results from case JANSN show that the model surface temperatures are not very sensitive to the initialization of snow depth. Decreased absorption of solar radiation associated with higher albedos in the presence of the larger snow covers in
JANSN (about 5 W m\(^{-2}\)) contributes to a surface cooling in JANSN compared with JANNOEX of only a few tenths K.

In summary, comparison of cases JANCON, JANNOEX, JANGMV, JANTM3 and JANS suggests that significant contributions to the differences in biases observed in JANCON and in the model of GMV are given by: (i) the cold bias in the driving large-scale CCM tropospheric temperatures of GMV, which can be transmitted to the interior of the domain through advection from the lateral boundaries, and reduced heat transfer from the deeper soil to the surface; and (ii) the use in JANCON of the explicit moisture scheme, which decreases precipitation, increases cloud cover, and thus decreases infrared radiation loss from the ground. This last effect shows how, as is the case for GCM simulations, correct prediction of cloud climatology and/or cloud radiative properties is critical for an accurate prediction of surface temperatures.

Comparison of the biases for cases JANNOEX and JANDIF shows that horizontal diffusion on terrain-following \(\sigma\)-surfaces significantly affects temperature prediction over mountainous areas. Average maximum and minimum temperatures over the Alpine region are, in fact, some 1.4 K and 1.8 K lower in JANDIF than in JANNOEX, respectively. This is due to the decreased diffusion of relatively warm low-elevation air along the topographic slopes towards the mountain tops. This cooling effect is also significant over the Mediterranean region, while over flatter western and central Europe the reduced diffusion of relatively cool high-elevation air from the adjacent mountain areas causes a slight warming. Overall, as can be expected, temperature over land is lower by a few tenths of a degree in JANDIF than in JANNOEX owing to decreased temperature diffusion along coastal slopes.

Finally, the importance of sea surface temperature variations on the surface air temperatures is evident from the comparison of the biases in the JANNOEX and JANSSTP3 cases. Averaged over the whole set of stations the maximum and minimum temperature biases vary by 1.2 K and 1.1 K, both becoming positive in JANSSTP3. This is mostly due to the advection of warmer air from the oceans onto the continents. The bias variations are more pronounced for the coastal regions, such as the Mediterranean region and western Europe, where they exceed 1.4 K and 1.2 K for maximum and minimum temperature, than for the continental interior regions.

Precipitation biases and threat scores for precipitation thresholds, varying from 0.25 cm to 7.5 cm, are reported for the seven January experiments in Tables 2 and 3. (Because there were only a small number of stations, we calculated the threat scores for the whole set of stations). As expected, the model is mostly sensitive to the use of different precipitation schemes, i.e. for cases JANCON, JANNOEX, and JANGMV. Looking at the biases for these three cases, we see that for the whole set of stations they decrease in magnitude from \(-40\%\) in JANCON, to \(-22\%\) in JANNOEX, to \(-12\%\) in JANGMV. Inclusion of the explicit moisture scheme in place of simple precipitation of supersaturated water vapour thus reduces precipitation by about 20\%. This is mostly due to the processes included in the explicit moisture scheme of cloud water evaporation, before this is converted into rainwater, and evaporation of falling rainwater before it reaches the ground. Slower release of condensation heat further reduces precipitation by about 10\%. As discussed by Giorgi (1991), the effect of slower release of condensation heat is much more important during summertime than during wintertime.

Overall, it appears that in these simulations the model underestimates precipitation, especially over the Mediterranean region and central Europe. However, for cases JANGMV and JANNOEX the biases are less than 10\% over western Europe and the Alpine region. Although the model physics parametrizations are certainly a big factor in producing the negative bias, uncertainties in the driving large-scale water-vapour fields
TABLE 2. AVERAGE DAILY OBSERVED PRECIPITATION (CM) AND DAILY PRECIPITATION BIAS (CM) FOR DIFFERENT JANUARY 1979 SIMULATIONS. RESULTS ARE PRESENTED AS OBTAINED USING ALL STATIONS AND USING STATIONS IN THE FOUR EUROPEAN CLIMATIC SUBREGIONS OF FIG. 2 IN PARENTHESES THE BIASES ARE REPORTED AS PERCENTAGES OF OBSERVED PRECIPITATION.

<table>
<thead>
<tr>
<th></th>
<th>Observed Precipitation</th>
<th>JANCON Bias</th>
<th>JANNOEX Bias</th>
<th>JANGMV Bias</th>
<th>JANDIF ° Bias</th>
</tr>
</thead>
<tbody>
<tr>
<td>All stations</td>
<td>0.258</td>
<td>-0.103 (-40%)</td>
<td>-0.056 (-22%)</td>
<td>-0.031 (-12%)</td>
<td>-0.079 (-31%)</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>0.383</td>
<td>-0.174 (-45%)</td>
<td>-0.101 (-26%)</td>
<td>-0.079 (-21%)</td>
<td>-0.119 (-31%)</td>
</tr>
<tr>
<td>Western Europe</td>
<td>0.211</td>
<td>-0.056 (-26%)</td>
<td>-0.018 (-8%)</td>
<td>0.014 (7%)</td>
<td>-0.053 (-25%)</td>
</tr>
<tr>
<td>Central Europe</td>
<td>0.178</td>
<td>-0.082 (-46%)</td>
<td>-0.050 (-28%)</td>
<td>-0.042 (-24%)</td>
<td>-0.060 (-34%)</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>0.291</td>
<td>-0.120 (-41%)</td>
<td>-0.066 (-23%)</td>
<td>-0.021 (-7%)</td>
<td>-0.091 (-31%)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>JANTM3 Bias</th>
<th>JANSN Bias</th>
<th>JANSSTP3</th>
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<tbody>
<tr>
<td>All stations</td>
<td>-0.057 (-22%)</td>
<td>-0.057 (-22%)</td>
<td>-0.028 (-11%)</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>-0.111 (-29%)</td>
<td>-0.102 (-27%)</td>
<td>-0.073 (-20%)</td>
</tr>
<tr>
<td>Western Europe</td>
<td>-0.002 (-1%)</td>
<td>-0.019 (-9%)</td>
<td>0.022 (10%)</td>
</tr>
<tr>
<td>Central Europe</td>
<td>-0.057 (-32%)</td>
<td>-0.051 (-28%)</td>
<td>-0.034 (-19%)</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>-0.071 (-24%)</td>
<td>-0.067 (-23%)</td>
<td>-0.040 (-14%)</td>
</tr>
</tbody>
</table>

TABLE 3. THIRTY-DAY ACCUMULATED PRECIPITATION THREAT SCORES FOR DIFFERENT JANUARY 1979 SIMULATIONS AS CALCULATED USING ALL STATIONS.

<table>
<thead>
<tr>
<th>Precipitation threshold (cm)</th>
<th>JANCON</th>
<th>JANNOEX</th>
<th>JANGMV</th>
<th>JANDIF</th>
<th>JANTM3</th>
<th>JANSN</th>
<th>JANSSTP3</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.25</td>
<td>1.00</td>
<td>1.00</td>
<td>1.00</td>
<td>1.00</td>
<td>1.00</td>
<td>1.00</td>
<td>1.00</td>
</tr>
<tr>
<td>1.00</td>
<td>0.97</td>
<td>0.99</td>
<td>0.99</td>
<td>0.99</td>
<td>0.99</td>
<td>0.99</td>
<td>0.99</td>
</tr>
<tr>
<td>2.50</td>
<td>0.67</td>
<td>0.78</td>
<td>0.79</td>
<td>0.74</td>
<td>0.76</td>
<td>0.78</td>
<td>0.80</td>
</tr>
<tr>
<td>5.00</td>
<td>0.36</td>
<td>0.49</td>
<td>0.53</td>
<td>0.46</td>
<td>0.50</td>
<td>0.49</td>
<td>0.50</td>
</tr>
<tr>
<td>7.50</td>
<td>0.15</td>
<td>0.33</td>
<td>0.33</td>
<td>0.18</td>
<td>0.30</td>
<td>0.34</td>
<td>0.40</td>
</tr>
</tbody>
</table>

probably also contribute. We have already mentioned how the main inaccuracies in the ECMWF IIIb analyses were found in the moisture fields, and in particular how these analyses tended to underestimate lower-tropospheric moisture. This underestimation of moisture in the driving analyses could be a contributing factor to the negative precipitation bias in the model.

Comparison of cases JANNOEX, JANSN, and JANTM3 shows that precipitation simulation was not sensitive to snow-cover initialization or to the reduction of the large-scale driving temperatures. The latter result is interesting, since in JANTM3 the relative humidities were adjusted to be the same as in JANNOEX, and therefore the driving large-scale water-vapour loadings in JANTM3 were lower than in JANNOEX. Note, however, that in the case of JANTM3 the sea surface temperatures were not modified. This caused more pronounced water-vapour deficits between the ocean surface and the atmosphere, which in turn resulted in increased evaporation and lower atmospheric instability. Evidently, this process was sufficient to balance the effect of reduced atmospheric water vapour in the initial and lateral boundary conditions.

Simulated precipitation is, however, quite sensitive to sea surface temperature variations and decreased horizontal diffusion. Increased sea surface temperatures contribute to enhance precipitation over land in two ways. The first is through increased
evaporative moisture supply to the travelling Atlantic and Mediterranean storms before they move over the continent. The second is through enhanced atmospheric instability associated with the sensible-heat fluxes from the warmer ocean waters. The average model precipitation over land stations increases by about 12% from JANNOEX to JANSSTP3, leading to an overall bias in JANSSTP3 of about –11%. As expected, the effect is most pronounced over western Europe, where the model precipitation changes by 18%, and is least pronounced in percentage terms over the Mediterranean and in absolute precipitation over central Europe. A significant reduction of precipitation in the case of JANDIF compared to JANNOEX can be observed in all regions. Over the Mediterranean and Alpine regions this can probably be attributed to the surface cooling effect discussed above (see Table 1), which tends to inhibit convective activity.

The monthly precipitation threat scores indicate a perfect prediction skill of areas that received small amounts of precipitation during the month (up to 1 cm). As the threshold increases, the scores decrease. For a threshold of 2.5 cm, the scores are still quite high, 0.67–0.80, while for thresholds of 5 and 7.5 cm they are in the range of 0.36–0.53 and 0.15–0.40, respectively. The case which includes the explicit moisture scheme shows the worst scores, because it has the lowest amounts of simulated precipitation.

The scores of Table 3 are similar in magnitudes and trends to those found by Giorgi and Bates (1989) for the western United States. They basically indicate that the model reproduces areas of light to moderate precipitation well. This is in agreement with the correct model simulation of storm paths and evolution illustrated by Figs. 3–6. The scores indicate larger model uncertainties in the reproduction of areas of heavy precipitation. For our simulation, this is basically due to the general model underprediction of precipitation discussed earlier. In interpreting this result, however, one should keep in mind that the threat scores for the larger precipitation thresholds are obtained over a relatively small set of stations, and thus errors at any single one may contribute heavily to the total score. Figure 7 and the simulations of GMV show that most model precipitation maxima correspond to the main mountain systems, i.e. that the model captures the topographic forcing of these systems on precipitation.

4. Analysis of summertime simulations

The selected 30-day summertime simulation period begins on 1 June 1979, 0000 GMT, and ends on 1 July 1979, 0000 GMT. For this period we performed five sensitivity experiments. In four of them, JUNCON, JANNOEX, JUNDIF and JUNSSTP3, the model is run with the same configurations as in JANCON, JANNOEX, JANDIF and JANSSTP3, respectively. Given the computational cost of a 30-day simulation with the MM4, experiments analogous to JANGMV, JANSN and JANTM3 were not carried out because the study of Giorgi (1991) showed that, for summertime conditions, the MM4 without modifications for slower release of condensation heat produces too many numerical point storm events over mountainous terrain. Little or no snow is generally present over Europe during the summertime, and, during summertime, local surface energy fluxes would dominate relatively small changes of large-scale temperature advection in determining the surface temperatures.

Because, however, the surface sensible-heat and latent-heat fluxes become dominant components of the surface energy budget, we added a simulation in which, with model configuration as in JANNOEX, the initial soil moisture content is equal to twice the amount predicted by Eq. (3). Thus, while in the case of JANNOEX the initial soil moisture content relative to saturation is in the range of 15–30% for the European region,
in JUNSW2 it is in the range of 30–60%. Given the uncertainties in estimating and modelling soil water content, such an experiment, which we refer to as JUNSW2, can provide a useful assessment of the sensitivity to this variable of simulated climate. A similar simulation was not carried out for January 1979 because Giorgi (1990) showed that wintertime temperatures and precipitations are little affected by soil moisture content.

The analysis presented in this section is similar to that of section 3. We first describe the prevailing synoptic events that occurred during June 1979 and their model representation, and then analyse the measures of simulation skill for the various sensitivity tests.

(a) Description of synoptic events occurring during June 1979

As done for January 1979, we describe the main synoptic events occurring during June 1979, using daily sea-level pressure maps obtained from the driving ECMWF data analysis, and compare them with sea-level pressures obtained from the JUNCON model simulation.

We present only illustrative examples of how the model reproduced observed conditions. In general, because of reduced meridional temperature gradients, the mid-latitude jet stream is weaker during the summertime than during the wintertime, Atlantic and Mediterranean cyclonic activity affecting Europe is reduced and summer weather over European region is mostly affected by local convective systems. Convective activity is usually most intense in mountainous areas owing to enhanced destabilization associated with high-elevation surface heating and low-level flow convergence over and around mountainous terrain.

This trend is reflected in our simulation of June 1979. The first low-pressure cell entered the British Isles from the north Atlantic on 5 June, and moved across central Europe until it dissipated over Poland on 8 June. Substantial precipitation throughout western and central Europe, and especially over the Alps, is produced by the frontal system associated with this storm. Sea-level pressure on 6 June and precipitation during the period 5–8 June are shown in Fig. 11(a, b). As can be seen, the model reproduces the low-pressure region extending from the North Sea to the Italian peninsula, as well

Figure 11. (a) Model-simulated (solid curve) and observed (dashed curve) sea-level pressure on 6 June, 0000 GMT. Units are mb, the contour interval is 4 mb, and highs and lows for the observations are slanted. (b) Accumulated model precipitation for the period 0000 GMT, 5 June to 0000 GMT, 8 June. Units are cm and the contour lines are 0.5, 1.5, 2.5, 5, 10, 20 cm.
as the high-pressure area which starts building over the eastern Atlantic. This high-pressure cell moves over the continent in the following few days, where it persists until 15 June, inducing relatively dry conditions and a flow of cool Atlantic air over central Europe.

On 15 June a second northern European disturbance moves over the North Sea and eventually over central Europe. This is illustrated in Fig. 12(a). The model produces a cyclonic system which is deeper than in, and located somewhat to the north-east of, the observations, and which induces intense precipitation (up to a maximum of 19 cm) between 15 and 20 June over the Alps, central and northern Italy, southern Germany, Yugoslavia, Austria, Czechoslovakia, Poland and western Soviet Union (Fig. 12(b)). The frontal zone of Fig. 12(a), which separated a low-pressure region situated over eastern Europe from a dry, high-pressure cell over the eastern Atlantic and western Europe, moved slightly eastward in the following days and persisted until about 25 June, producing wet conditions over the Balkans and dry conditions over France and the Iberian peninsula. During the last five days of the month a cyclonic cell developed over the Scandinavian peninsula, and a wide but weak front swept over central Europe producing light precipitation over the higher mountains.

The examples of Figs. 11–12 again illustrate how, also for June 1979, the model captures the basic structure and evolution of the main synoptic patterns which characterized the weather of the month, although, as in the January simulation, observed and model-produced sea-level pressures over land show discrepancies of up to several mb (see the discussion at the end of section 3(a)). Figure 13 presents the model-produced total monthly precipitation. The wettest regions are the Alps, eastern Europe and the Balkans, northern Great Britain and the southern Norwegian mountains. These are the areas mostly affected by the synoptic disturbances described in this section. Conversely, high pressure and drier conditions prevailed over France, southern Great Britain and the western Mediterranean regions.

The model precipitation maxima clearly correspond to the major local topographical features. This is an indication of the strong topographical forcing of summertime precipitation in the model (Giorgi 1991). In this simulation the highest Alpine regions receive
model precipitation totals exceeding 30–40 cm, with a maximum of about 86 cm corresponding to the highest elevation gridpoint. This is probably overpredicted in the run JUNCON (see section 4(b)) although, in general, a distinct summer maximum of rainfall is observed over the Alps, which locally exceeds several cm per month (World Survey of Climatology 1970).

(b) Simulation skill analysis

As in section 3(b), we first analyse the large-scale biases for temperature, water vapour and zonal wind in experiment JUNCON. These are shown in Fig. 14(a–c). As occurred in the wintertime simulation, temperature and water-vapour biases are smaller in the mid to upper troposphere than near the surface and the top of the model. In particular, the temperature bias near the surface over land exceeds 3 K, and the water-vapour bias exceeds 1.5 g kg\(^{-1}\), i.e. they are larger than in wintertime, which is to be expected. More during summertime than in wintertime, do model surface processes dominate advection from the lateral boundaries and determine the values of surface temperature and moisture.

The effect of the model jet acceleration observed for the January simulation is present in the June experiment only in the upper troposphere and is less pronounced. The average 300 mb wind in the analysis and the 300 mb zonal-wind bias are shown in Fig. 15(a, b). The Atlantic jet for this simulation is located to the west of the British Isles and a secondary maximum is indicated by a ridge over the western Mediterranean. The June zonal-wind bias shows a more coherent structure than for the JANCON simulation, with the occurrence of jet acceleration corresponding with the North Atlantic jet, and with jet weakening over the western Mediterranean. This causes the disappearance in the model (not shown here) of the western Mediterranean ridge that can be observed in
Figure 14. MM4 large-scale bias for case JUNCON as a function of $\sigma$ over land and ocean areas: (a) temperature; (b) water vapour; (c) zonal wind.
Fig. 15(a). Although the flow acceleration is not large, a possible consequence of this flow deflection, which could be significant when applying the model to climate studies, is a modification of the model-simulated storm paths compared to those of the driving-condition fields. It can be expected that large-scale flow modifications by the model nesting will be more pronounced for summertime than for wintertime conditions, because, during the warm season, advection from the lateral boundaries is at its minimum and the model physics forcings at their maximum.

Table 4 presents the maximum and minimum temperature bias for all June 1979 experiments and for the various regions of our domain. In the JUNCON case the overall minimum temperature bias is about 3 K, with a maximum of about 3.9 K over the Alpine region, and a minimum of about 2.2 K over western Europe. The maximum temperature bias is smaller, varying from 0.6 K over the Mediterranean region to 2 K for western Europe, with an average value over all stations of 1.5 K. In the JUNNOEX case the maximum temperature bias increases, and the minimum temperature bias decreases, by a few tenths of a degree. A more pronounced sensitivity is observed in the case of JUNSW2. Compared with JUNNOEX, the average peak daytime temperatures decrease by 1.6 K, yielding an overall bias of 0.15 K. For this case the maximum temperature bias over the Alpine region and central Europe becomes small, that over the Mediterranean region is negative and that over western Europe is reduced to 1.1 K. The minimum

<table>
<thead>
<tr>
<th></th>
<th>JUNCON</th>
<th>JUNNOEX</th>
<th>JUNDIF</th>
<th>JUNSW2</th>
<th>JUNSSTP3</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>TMAX</td>
<td>TMIN</td>
<td>TMAX</td>
<td>TMIN</td>
<td>TMAX</td>
</tr>
<tr>
<td>All stations</td>
<td>1.556</td>
<td>2.937</td>
<td>1.806</td>
<td>2.774</td>
<td>1.028</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>0.611</td>
<td>2.642</td>
<td>0.783</td>
<td>2.347</td>
<td>-1.088</td>
</tr>
<tr>
<td>Western Europe</td>
<td>2.019</td>
<td>2.199</td>
<td>2.373</td>
<td>2.213</td>
<td>1.805</td>
</tr>
<tr>
<td>Central Europe</td>
<td>1.788</td>
<td>3.563</td>
<td>2.016</td>
<td>3.327</td>
<td>1.974</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>1.445</td>
<td>3.913</td>
<td>1.635</td>
<td>3.761</td>
<td>-0.099</td>
</tr>
</tbody>
</table>
temperature biases in JUNSW2 are also reduced, but only by less than 0.6 K. Finally, the maximum and minimum temperature biases in JUNSSTP3 are 0.2–0.9 K larger than in JUNNOEX for the various regions of the domain, and a general cooling, most pronounced over the Alpine and Mediterranean regions (1–2 K), occurs with the reduced horizontal diffusion coefficient of experiment JUNDIF. Note that the biases of Table 4 are more consistent with the lower tropospheric temperature biases calculated from the ECMWF data (Fig. 14(a)) than in the January simulation.

To interpret these results in terms of surface processes we can again look at the average diurnal cycles of absorbed solar, net infrared, sensible-heat and latent-heat fluxes at the land surface. For experiment JUNCON these are shown in Fig. 16. Compared to the wintertime fluxes shown in Fig. 10, the summertime sensible-heat and latent-heat fluxes become more important. During the daytime they reach peak values of about 332 and 185 W m\(^{-2}\), respectively. These are smaller than the peak solar flux (730 W m\(^{-2}\)), but larger than the net infrared flux (120 W m\(^{-2}\)). During night-time the infrared loss (75 W m\(^{-2}\)) dominates the sensible-heat (−35 W m\(^{-2}\)) and latent-heat (10 W m\(^{-2}\)) fluxes. By comparing these two fluxes, as we have done in section 3(b), we obtain a contribution from the energy exchange between surface soil and deep soil of 95 W m\(^{-2}\) during peak daytime hours, and 50 W m\(^{-2}\) during night-time. In summary, daytime temperatures are dominated by solar insolation and sensible-heat and latent-heat fluxes, while night-time temperatures are still mostly determined by the net infrared radiation budget.

As in the wintertime simulations, the main effect of replacing the explicit moisture scheme with instantaneous precipitation of supersaturated water vapour is to increase precipitation (see later) and reduce cloud cover. This has a twofold effect: first, the average cloud cover over land decreases from 43% in JUNCON to 24% in JUNNOEX—most of the change occurring in the mid and upper troposphere—which causes an increase in net infrared loss and solar absorption at the surface; second, enhanced precipitation results in higher soil moisture contents and increased evaporation. On average, the net infrared radiation loss, peak daytime absorbed solar radiation flux, and peak daytime latent-heat flux were higher for JUNNOEX than for JUNCON by 10 W m\(^{-2}\), 40 W m\(^{-2}\),

![Diagram](image)

**Figure 16.** Average diurnal cycle of absorbed solar, net upward infrared, sensible-heat (positive upward), and latent-heat (positive upward) fluxes at the surface for all land gridpoints in the interior of the domain in the JUNCON simulation. Units are W m\(^{-2}\).
and 15 W m$^{-2}$, respectively. These changes forced the lower nighttime and higher peak daytime temperatures in JUNNOEX, implied by the biases of Table 4.

Doubling the initial soil moisture values has a strong influence upon the sensible-heat and latent-heat fluxes. The effect is not only to increase the availability of initial soil water for evaporation, but also to increase precipitation (see later) and keep the soil water contents at higher levels in JUNSW2 than in JUNNOEX. In the case of JUNNOEX the average peak daytime latent-heat flux over land is 200 W m$^{-2}$, with corresponding average peak sensible-heat flux of 340 W m$^{-2}$. The situation is reversed in the case of JUNSW2, where the average peak daytime latent-heat flux increases to 305 W m$^{-2}$, and the average peak daytime sensible heat flux decreases to 268 W m$^{-2}$. This change in Bowen ratio causes an average decrease in peak surface air temperatures of 1.6 K in run JUNSW2, which is reflected in the maximum temperature biases of Table 4. During night-time the surface flux differences between JUNNOEX and JUNSW2 are small, of the order of 3–7 W m$^{-2}$, thus the surface temperatures in the two runs do not differ greatly.

The effect of sea surface temperature variations on surface temperature is much less pronounced than during wintertime. Only over the Mediterranean region and western Europe do temperature changes due to a 3 K increase in sea surface temperature reach 0.6–1 K. Over the other regions is only of the order of a few tenths of a degree, generally larger during the change night-time than during the change daytime. This, again, emphasizes that there is greater forcing exerted by local processes during summertime than during wintertime. Finally, as discussed in section 3(b), the cooling observed in case JUNDIF over the mountainous regions of our domain is due to decreased diffusion of relatively warm low-elevation air along the mountain slopes.

The precipitation biases and threat scores for all the summertime cases are shown in Tables 5 and 6. The JUNCON simulation underpredicts precipitation by about 27%. The bias is negative over all regions with values of $-30\%$ to $-38\%$, except over the Alps, where it is slightly positive. This is consistent with the results of Giorgi (1991), who found that this model configuration underestimated precipitation over most of the western United States, except over the highest mountain ranges.

Much of the precipitation underprediction is due to the use of the explicit moisture scheme. When this is removed (JUNNOEX), the bias for all stations becomes $-6\%$, and varies from $-21\%$ over central Europe, to a minimum of $-7\%$ over the Mediterranean region, to a maximum of $+34\%$ over the Alps. Compared with the results of JUNNOEX, precipitation is considerably reduced over the Alpine region, to the point, in JUNDIF, of resulting in a negative bias ($-15\%$). This is consistent with the findings of Tibaldi (1982) and Simmons (1986). It confirms that, at least when using a Kuo-like cumulus scheme, excessive convective activity and precipitation over mountain tops is induced by spurious diffusion of relatively warm and moist low-elevation air up steep slope-following o-surfaces. This effect is more pronounced for summertime than for wintertime conditions.

Increasing the initial soil moisture content has a substantial effect on precipitation. The overall bias varies from $-6.5\%$ in JUNNOEX to $+15\%$ in JUNSW2, corresponding to an increase in model-produced precipitation of the order of 20%. This effect is felt especially over the drier and less mountainous regions, the Mediterranean region, western Europe and central Europe, where precipitation increases by about 20–30%. Over the Alps the increase is less than 5%. We emphasize that, since shallow convection processes are not included in the model, evaporated water vapour is not efficiently removed from the boundary layer, where it tends to accumulate and eventually to condense. As a consequence, the model may show an excessive sensitivity to soil moisture content. The
TABLE 5. AVERAGE DAILY OBSERVED PRECIPITATION (CM) AND DAILY PRECIPITATION BIAS (CM) FOR DIFFERENT JUNE 1979 SIMULATIONS. RESULTS ARE PRESENTED AS OBTAINED USING ALL STATIONS AND USING STATIONS IN THE FOUR EUROPEAN CLIMATIC SUBREGIONS OF FIG. 2. IN PARENTHESES THE BIASES ARE REPORTED AS PERCENTAGES OF OBSERVED PRECIPITATION

<table>
<thead>
<tr>
<th></th>
<th>Observed precipitation</th>
<th>JUNCON Bias</th>
<th>JUNNOEX Bias</th>
</tr>
</thead>
<tbody>
<tr>
<td>All stations</td>
<td>0.231</td>
<td>-0.063 (-27%)</td>
<td>-0.015 (-6%)</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>0.124</td>
<td>-0.043 (-35%)</td>
<td>-0.009 (-7%)</td>
</tr>
<tr>
<td>Western Europe</td>
<td>0.175</td>
<td>-0.053 (-30%)</td>
<td>-0.020 (-11%)</td>
</tr>
<tr>
<td>Central Europe</td>
<td>0.279</td>
<td>-0.106 (-38%)</td>
<td>-0.058 (-21%)</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>0.534</td>
<td>0.026 (3%)</td>
<td>0.183 (34%)</td>
</tr>
<tr>
<td></td>
<td>JUNDIF Bias</td>
<td>JUNSW2 Bias</td>
<td>JUNSSTP3 Bias</td>
</tr>
<tr>
<td>All stations</td>
<td>-0.042 (-18%)</td>
<td>0.035 (15%)</td>
<td>-0.015 (-6%)</td>
</tr>
<tr>
<td>Mediterranean region</td>
<td>-0.011 (-9%)</td>
<td>0.012 (10%)</td>
<td>0.014 (11%)</td>
</tr>
<tr>
<td>Western Europe</td>
<td>-0.035 (-20%)</td>
<td>0.021 (12%)</td>
<td>-0.006 (-3%)</td>
</tr>
<tr>
<td>Central Europe</td>
<td>-0.060 (-21%)</td>
<td>0.022 (8%)</td>
<td>-0.064 (-23%)</td>
</tr>
<tr>
<td>Alpine Region</td>
<td>-0.080 (-15%)</td>
<td>0.206 (39%)</td>
<td>0.092 (17%)</td>
</tr>
</tbody>
</table>

TABLE 6. THIRTY-DAY ACCUMULATED PRECIPITATION THREAT SCORES FOR DIFFERENT JUNE 1979 SIMULATIONS AS CALCULATED USING ALL STATIONS

<table>
<thead>
<tr>
<th>Precipitation threshold (cm)</th>
<th>JUNCON</th>
<th>JUNNOEX</th>
<th>JUNDIF</th>
<th>JUNSW2</th>
<th>JUNSSTP3</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.25</td>
<td>0.93</td>
<td>0.95</td>
<td>0.98</td>
<td>0.97</td>
<td>0.96</td>
</tr>
<tr>
<td>1.00</td>
<td>0.80</td>
<td>0.87</td>
<td>0.89</td>
<td>0.90</td>
<td>0.89</td>
</tr>
<tr>
<td>2.50</td>
<td>0.49</td>
<td>0.61</td>
<td>0.70</td>
<td>0.71</td>
<td>0.65</td>
</tr>
<tr>
<td>5.00</td>
<td>0.31</td>
<td>0.32</td>
<td>0.42</td>
<td>0.39</td>
<td>0.34</td>
</tr>
<tr>
<td>7.50</td>
<td>0.32</td>
<td>0.27</td>
<td>0.44</td>
<td>0.38</td>
<td>0.33</td>
</tr>
</tbody>
</table>

effect of increased sea surface temperature is felt mostly over the Mediterranean region and western Europe, where precipitation increases by 20% and 10%. Precipitation over central Europe and especially the Alpine region is somewhat lower in JUNSSTP3 than in JUNNOEX. We should also recall that the already discussed low moisture bias in the driving ECMWF IIIb analysis possibly contributes to the generally negative precipitation biases of Table 5.

The threat scores are generally lower than for the wintertime simulations, which is consistent with the results that Giorgi and Bates (1989) and Giorgi (1991) obtained for the western United States, and can be expected from the more localized nature of summertime precipitation. As in the case of the wintertime runs, the scores decrease as the precipitation threshold increases, and increase for the experiments which show larger negative precipitation biases. For a threshold of 0.25 cm the scores are higher than 0.9. They decrease to 0.8–0.9 for a threshold of 1 cm, 0.5–0.7 for a threshold of 2.5 cm, and 0.27–0.44 for thresholds of 5 and 7.5 cm.

5. SUMMARY AND CONCLUSIONS

As part of the development of a regional climate model for western Europe and the Mediterranean basin initiated by GMV, in this paper we have carried out a preliminary validation analysis of the climatology of a version of the NCAR regional meteorological model, MM4, for the European region. Our version of the MM4 model includes packages of detailed radiative transfer and of the physics and soil hydrology of the surface. The
purpose of the study is to identify relevant model biases and inaccuracies by comparing wintertime and summertime model simulations with the observations that are available for the simulated time period.

The periods of the simulations are January and June 1979, and the initial and lateral meteorological boundary conditions driving the model are taken from the ECMWF IIb analysis of FGGE observations. Model-produced maximum and minimum surface air temperatures and precipitation are compared with station observations. Compared to the model of GMV, we test here two additional precipitation parametrizations, i.e. an explicit moisture scheme for stable precipitation, and modifications introduced by Giorgi (1991) to the Kuo cumulus-cloud scheme of the standard MM4 model, allowing slower release of condensation heat. We analyse the sensitivity of the model to reduced horizontal diffusion of wind, temperature and moisture on $\sigma$-surfaces in areas of steep terrain, and to the specification of surface conditions such as sea surface temperature, soil-moisture, and initial snow-cover distribution.

Based on the analysis of sea-level pressure fields, the model reproduced the main characteristics of the structure and evolution of the synoptic events which occurred during the two periods of simulation. Compared to the large-scale driving analysis, the model did not show large biases in the prediction of variables of the mid and upper troposphere. For our experiments, wintertime maximum and minimum daily surface-air-temperature biases generally did not exceed a few degrees K for various regions of Europe. Removal of the explicit moisture scheme decreased surface temperatures by 0.5–2 K in the wintertime owing to decreasing cloudiness and the consequent increase in associated net surface infrared radiation loss. The surface temperatures were not sensitive to the inclusion of slower condensation heat release and to snow initialization. In the January simulations, cold biases in the large-scale driving conditions were transmitted to the model surface air temperatures by advection from the lateral boundaries, and reduced heat transfer to the surface from the deep soil. This could, at least partially, explain the cold bias in the surface air temperature observed in the model of GMV. For summertime conditions, maximum surface air temperature biases were generally less than 1–2 K and mostly sensitive to the soil moisture content, while minimum air temperature biases were of the order of 2–3 K. The response of surface air temperature over land to variations in oceanic sea surface temperatures was more pronounced during wintertime than during summertime. Both for January and June decreased diffusion of relatively warm low-elevation air along steep slopes induced significant cooling over mountainous areas.

Precipitation was sensitive to the parametrizations tested. For wintertime and summertime, the use of the explicit moisture scheme led to substantial precipitation under-prediction. When this scheme was removed, the precipitation biases for our simulations were of the order of −10% to 30%. Possible biases in the ECMWF driving moisture fields may have contributed to this result. Diffusion of moisture and temperature along mountain slopes leads to overprediction of precipitation over the Alps during summertime. When this diffusion was substantially reduced, June precipitation decreased there by a factor of 1.6. Precipitation amounts in the summertime run were rather sensitive to initialization of soil water contents. Variations in sea surface temperature significantly affected precipitation over land, especially over the coastal areas of western Europe. Thirty-day precipitation threat scores were close to unity for light precipitation thresholds (1 cm) and decreased to 0.3–0.5 for heavy precipitation thresholds (5 to 7.5 cm). They were slightly higher for wintertime than for summertime precipitation.

In conclusion, although limited in scope, our analysis shows that, given good driving large-scale conditions, model physics parametrizations can be selected which yield realistic descriptions of synoptic events and which give errors in the simulation of
important climatic variables such as surface air temperature and precipitation that are
within reasonable ranges. Care has to be taken, however, to correctly initialize or
specify quantities such as soil water content and sea surface temperature, as they affect
the calculations of precipitation and surface temperature; and to reproduce realistic
climatic distribution of cloud cover, as this strongly influences the surface energy budget.

ACKNOWLEDGEMENTS

We thank S. Tibaldi, G. Visconti and the anonymous reviewers for their comments
and suggestions, also D. Joseph and W. Spangler of the Data Support Section of NCAR
for providing the station observation data-set. This work was completed while one of the
authors (M.R.M.) was a visitor at the Climate and Global Dynamics Division of NCAR.
This research was partially supported by a fellowship to M.R.M. from the Fondazione
Scientifica San Paolo, Italy.

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