An Investigation of the Sensitivity of Simulated Precipitation to Model Resolution and Its Implications for Climate Studies

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ABSTRACT

This paper examines the sensitivity of a regional atmospheric model to horizontal resolution and topographic forcing. The model is run for January and July month-long simulations over the European region at gridpoint spacings ranging from 200 to 50 km and with various topography configurations. Different precipitation parameterizations of complexity and structure similar to those used in present-day climate models are tested. When averaged over the whole continent, the precipitation amounts are more sensitive to gridpoint spacing than to topographic forcing. Topography mostly contributes to spatially redistributing precipitation, and its effect is dominant only over subregions characterized by complex topographical features (e.g., the Alps). Other variables, such as cloudiness, surface energy fluxes, and precipitation intensity distributions are also sensitive to resolution. Finally, simulated precipitation amounts vary with the parameterization scheme used at all resolutions. These results have important implications for climate modeling. They suggest that, when running a model on a wide range of horizontal resolutions, such as in a variable gridpoint spacing configuration, in "time slice" mode, or within a nested modeling system, the effects of physical forcings (e.g., topography) can be strongly modulated by the direct sensitivity of the model physics formulations to resolution.

1. Introduction

The issue of horizontal model resolution is central to the development of accurate regional climate change scenarios for impact assessments. Present-day coupled general circulation models (GCMs) are generally run at resolutions of a few hundred kilometers. This resolution can be increased globally or regionally (over areas of complex topography, coastlines and surface characteristics) using different techniques (e.g., Giorgi and Mearns 1991): the one-way nested modeling technique (e.g., Giorgi 1990), the use of variable resolution models (Deque and Piedlevre 1995), or the use of increased resolution GCM simulations in "time slice" mode (Cubasch et al. 1995).

Regardless of the method employed, there are various components to the debate concerning the effects of model horizontal resolution on regional climate simulation. For example, increased resolution in a GCM can lead to a better representation of the general circulation of the atmosphere, which in turn would affect the simulation of large-scale flow features determining the climate characteristics of a given region. This topic has been addressed in various works (e.g., Boville 1991; Boer and Lazare 1988) and is outside the purpose of the present paper. A second aspect of the problem is concerned with the local effects of improved representation of surface forcings, such as due to topography, large lakes, or coastlines. The nested modeling technique was specifically developed to simulate such effects and showed good performance in capturing the precipitation-producing mechanisms associated with topographic and lake forcing (e.g., Giorgi et al. 1994; Bates et al. 1995).

A third but not less important issue related to model resolution is that of the sensitivity of physics parameterizations, primarily for the precipitation process, to the horizontal model resolution itself, regardless of external or local forcings. It is well known that parameterizations of cumulus convection and resolvable-scale precipitation include parameters and assumptions (e.g., parameterization closures) that may be sensitive to the model horizontal gridpoint spacing and are often tested and tuned for specific resolutions. Therefore, when running a model at different spatial resolutions, the possibility exists that the direct sensitivity of a scheme to resolution may modify or even mask the effect of physical and dynamical forcings (e.g., topography). For this reason, without specific tuning of the parameterization, an increase in model resolution does not necessarily result in an improvement in the simulation of some aspects of precipitation. While the model response to high-resolution forcings reflects the impact

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of physical processes, the direct model sensitivity to resolution may be an artifact of possibly erroneous or inconsistent assumptions in the parameterization.

In this paper we investigate the sensitivity of different precipitation parameterizations to model resolution and the resolution of topographic forcing using a regional climate model (RegCM) driven by initial and lateral boundary conditions provided by analyses of observations. The strategy of this study is to compare results from a set of simulations employing the same boundary condition forcings but different resolutions and topography representations. In this way, since the driving large-scale fields are the same for each run, the effects of the internal model processes can be isolated.

The model domain covers the European region, which is characterized by substantial topographic variability. The focus of the analysis is on precipitation, although other model variables that affect the surface hydrologic and energy budgets are analyzed, such as cloud cover and the surface radiative sensible heat and latent heat fluxes. Analysis of the model performance in simulating the structure of synoptic events is also presented. Sets of month-long simulations are performed for summer and winter conditions in order to include synoptic events of varying characteristics. Note that although the parameterizations tested in this work are specific to the model used, they include assumptions and complexity of formulations comparable to those of other regional and global climate models, so that our conclusions are more general in nature.

The paper is organized as follows. Sections 2 and 3 describe the model characteristics and the experiment design, respectively. Results are discussed in section 4, and the main conclusions and implications of our findings are discussed in section 5.

2. Model description

The dynamical component of our RegCM is essentially the same as that of the National Center for Atmospheric Research (NCAR)–The Pennsylvania State University Mesoscale Model MM4 (Anthes and Warner 1978; Anthes et al. 1987). This is a hydrostatic, compressible, primitive equation, terrain-following $\sigma$ vertical coordinate model, where $\sigma = (p - p_{top})/(p_{top} - p_{bottom})$, $p$ is pressure, $p_{top}$ is the pressure specified to be the model top, and $p_{bottom}$ is the prognostic surface pressure. The model vertical structure includes 14 $\sigma$ levels with the model top at 80 mb. Five levels are below $\sigma = 0.8$, six equally spaced levels ($\Delta \sigma = 0.1$) are located between $\sigma = 0.8$ and $\sigma = 0.2$, and three levels are above $\sigma = 0.2$. The version of RegCM used for this study is that described by Giorgi et al. (1993a,b). This is a second-generation RegCM that includes detailed representations of radiative transfer, planetary boundary layer (PBL), surface physics, and cloud and precipitation processes.

The RegCM radiative transfer package is the same as that of the newest version of the NCAR Community Climate Model (CCM2) (Briegleb 1992; Hack et al. 1994). Radiative transfer calculations include the contributions of $CO_2$, $O_3$, $H_2O$, $O_2$, and clouds. In the presence of clouds the scattering and absorption parameterization of Slingo (1989) is used, whereby the optical properties of the cloud droplets (extinction optical depth, single scattering albedo, and asymmetry parameter) are expressed in terms of the cloud liquid water content and a specified effective droplet radius equal to 10 $\mu$m. The scheme requires as input two key cloud variables, the cloud fractional cover and the cloud water content, which are interactively calculated based on the model precipitation parameterization as described in section 2a.

The surface physics calculations are performed using the Biosphere–Atmosphere Transfer Scheme (or BATS) version BATS-1E (Dickinson et al. 1993). BATS is a state-of-the-art surface package designed to describe the role of vegetation and interactive soil moisture in modifying the surface–atmosphere exchanges of momentum, energy, and water vapor. It comprises a vegetation layer (BATS currently includes 15 vegetation types), a snow layer, and a three-layer soil water model (a 10-cm-thick surface soil layer, 1–2-m-thick root zone, and a 5-m-thick deep soil layer). Equations are solved for the soil temperature, soil water content, and, in the presence of vegetation, the temperature of canopy air and canopy foliage. The soil hydrology calculations account for precipitation, snowmelt, canopy foliage drip, evapotranspiration, surface runoff, infiltration below the root zone, and diffusive water movement. Sensible heat, water vapor, and momentum fluxes at the surface are calculated using a standard surface drag coefficient formulation based on surface layer similarity theory.

The RegCM employs a medium-resolution PBL scheme, developed by Holtslag et al. (1990), with five levels in the lowest 1.5 km of the atmosphere, at approximately 40, 110, 310, 730, and 1400 m above the surface. The vertical eddy flux within the PBL is given by an eddy diffusion term plus a “countergradient” term describing nonlocal transport due to deep convective plumes in the PBL. The eddy diffusivity follows a nonlocal parabolic profile between the surface and the PBL top. Details on various aspects of the scheme can be found in Holtslag et al. (1990) and Holtslag and Boville (1992).

Since the focus of this paper is on precipitation, the resolvable-scale and cumulus convection schemes are described in some detail in the next sections.

a. Resolvable-scale precipitation

Two schemes are tested in this work for the simulation of resolvable-scale precipitation. The first, which we call “implicit,” consists of instantaneous precipitation of supersaturated water when relative humidity of 100% is exceeded at a given grid point. When this
scheme is used, cloud fractional cover $f_d$ and liquid water content $w_d$ are calculated as follows. At a given grid point, $f_d$ is 0 if the local relative humidity RH is lower than a given threshold $R_{th}$, and it is given by

$$ f_d = \frac{(RH - R_{th})^2}{(1 - R_{th})^2} $$

for $RH \geq R_{th}$. This scheme is based on the parameterization proposed by Slingo (1980). The relative humidity threshold is a linear function of the model grid-point spacing $\Delta x$ and varies from 0.9 for $\Delta x \leq 10$ km to 0.75 for $\Delta x \geq 100$ km. This formulation was designed to roughly account for resolution effects and yields values of $R_{th}$ close to those proposed by Slingo (1980). The cloud water content $w_d$ varies linearly as a function of temperature from 0.015 g m$^{-3}$ at 265 K to 0.15 g m$^{-3}$ at 295 K, so as to roughly account for altitude and latitude effects and to produce values in the range used for the CCM2.

The second scheme for resolvable-scale precipitation is a simplified version of the explicit moisture scheme of Hsieh et al. (1984). The Hsieh et al. scheme consists of prognostic equations for cloud water and rainwater mixing ratio including advection by resolvable-scale winds, diffusion by subgrid-scale motions and bulk formulations of condensation/evaporation of cloud water and rainwater, autoconversion of cloud water into rainwater, aggregation of cloud water by rainwater, and gravitational settling of rainwater (Kessler 1969). This full explicit scheme is computationally too expensive to be used in climate mode, since it adds about 30%--50% of computation time to the RegCM. Therefore, a simplified version of it, including only an equation for cloud water, was developed for coupling with the RegCM. Cloud water is formed when supersaturation is attained, it is then advected, it can reevaporate, and it is converted into rainwater via the Kessler-type autoconversion term

$$ P_{conv} = k_d (w_d - w_h). $$

Rainwater is then immediately precipitated out. In Eq. (2), $k_d$ is an autoconversion rate equal to $10^{-3}$ s$^{-1}$ (Hsieh et al. 1984) and $w_h$ is a cloud water threshold that is given the same temperature dependence as in the values of $w_d$ used for the implicit scheme (see above). This is done in order to maintain consistency between the implicit and explicit schemes in the definition of cloud water content for radiative transfer, since the cloud water content calculated by the explicit scheme is directly used in the radiation calculations. The fractional cloud cover for the explicit scheme is assumed to be equal to 1 when cloud water is present.

Both the implicit and explicit schemes treat only resolvable-scale cloud and rain processes and are used only in conjunction with a cumulus cloud scheme representing convective processes. The cumulus cloud and explicit moisture scheme are coupled as described by Zhang et al. (1988), whereby the resolvable-scale precipitation schemes are called in the code after the convection schemes.

b. Convective precipitation

Two parameterizations of convective precipitation are tested in this work. The first is a Kuo-type formulation described by Anthes et al. (1987) and Anthes (1977). In this parameterization, 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 (the $b$ factor) is redistributed throughout the column proportionally to the dryness of the vertical grid point. Latent heat of condensation is redistributed between cloud top and cloud bottom following a specified parabolic vertical heating profile that yields maximum heating in the upper half of the cloud layer. Also included in the scheme are the modifications of Giorgi (1991), by which latent heat of condensation and moisture produced by cumulus convection are released at the gridpoint scale with a characteristic time of two hours.

The second scheme tested here was developed by Grell (1993) and is described in detail in Grell et al. (1994). In the Grell scheme, clouds are pictured as two steady-state circulations caused by an updraft and a downdraft with no direct mixing between cloudy air and environmental air except at the top and bottom of the circulations. A simplistic cloud model is used that assumes that the mass flux in the updraft and downdraft ($m_u$ and $m_d$, respectively) is constant with height, and the originating levels of updraft and downdraft are given by the levels of maximum and minimum ambient moist static energy, respectively. The scheme is activated when a parcel lifted from the updraft originating level eventually attains moist convection. Condensation in the updraft is calculated by lifting of a saturated parcel, and rainfall is given by

$$ R = I_c m_u (1 - \beta), $$

where $I_c$ is the amount of condensation integrated over the whole depth of the updraft normalized by the updraft mass flux and $\beta$ is the fraction of updraft condensate that reevaporates in the downdraft; that is, $1 - \beta$ is the precipitation efficiency. For our studies, the $\beta$ parameter varies from 0.3 to 0.5 depending on the vertical wind shear (Grell et al. 1994). The heating and moistening feedback to the large scale is determined by compensating mass fluxes and detrainment at cloud top and bottom. Also included in the scheme is the cooling effect of moist convective downdrafts (Grell et al. 1994).

Because of the simplicity of the cloud model and because of Eq. (3), any closure assumption can be
adopted to complete the scheme, which relates the mass flux at the bottom of the updraft to the large-scale forcings. For this study we use the closure

$$m_b = \frac{\text{ABE}^\prime - \text{ABE}}{\text{NA} \Delta t},$$

where $\text{ABE}$ is the buoyant energy available to a cloud, $\text{ABE}^\prime$ is the production of available buoyant energy by the large-scale motions during the time step $\Delta t$, and NA is the rate of change of available buoyant energy per unit of mass flux (Grell et al. 1994). This closure is similar to that adopted by Arakawa and Schubert (1974) and basically states that cumulus clouds stabilize the environment as fast as the large scale destabilizes it.

When cumulus clouds are formed by either the Kuo or the Grell scheme, the gridpoint fractional cloud cover is such that the total cover for the column extending from the model-computed cloud-base level to the cloud-top level (calculated assuming random overlap) is a quadratic function of horizontal gridpoint spacing, varying from 1 for $\Delta x = 10$ km to 0.3 for $\Delta x = 200$ km. This formulation roughly accounts for resolution effects, and for the larger scales yields values similar to those adopted in early versions of the CCM. The average water content within a cumulus system is assumed to be equal to 0.15 g m$^{-3}$, that is, the maximum value attainable by resolvable-scale clouds. Both for resolvable and cumulus clouds, the thickness of a radiative cloud layer is assumed to be equal to that of the model layer. Also, because excessive cloudiness tends to form in the bottom model levels (e.g., Giorgi et al. 1993a,b), clouds generated in the three bottom atmospheric layers (a depth of about 500 m from the surface) are not considered to be radiatively active.

3. Experiment design

a. Model setup and experiments

The model domain used for this work is shown in Fig. 1. It is of 6000 km $\times$ 4500 km size and it encompasses western Europe and the Mediterranean Basin. To investigate the effect of model resolution, experiments for three different values of $\Delta x$, 200, 100, and 50 km, were performed over this domain. The model topography for $\Delta x = 200$ and 50 km is shown in Figs. 1a,b. At the highest resolution, the main features of the Alps, the Iberian Plateau, the Pyrenees, the Balkans, the Apennines, and the Grampian Mountains in Scotland are captured. Surface vegetation types were obtained for all experiments by interpolating from the same dataset used in the standard MM4 (e.g., Anthes et al. 1987), and soil water content was initialized as a function of vegetation as described by Giorgi and Bates (1989) and Giorgi and Marinucci (1991).

The periods of simulation are January 1991 and July 1991, two months in which several storm systems traveled across the European region producing substantial precipitation over the continent (see section 4a). Sea surface temperatures and meteorological initial and lateral boundary conditions (wind components, temperature, water vapor mixing ratio, and surface pressure) necessary to drive the model runs are interpolated from the European Centre for Medium-Range Weather Forecasts (ECMWF) analyses of observations, which are available on a spectral T42 grid (approximately 3° resolution) (Trenberth and Olson 1988). The ECMWF driving data are distributed on 15 pressure levels and are available at intervals of 12 hours. Linear interpolation in time is performed between ECMWF 12-h fields.

Boundary conditions from the ECMWF data are provided using the relaxation procedure of Davies and Turner (1977). This consists of Newtonian and diffusion terms that gradually drive the model solution of
wind components, temperature, water vapor mixing ratio, and surface pressure toward specified large-scale values inside a lateral “buffer” area (Anthes et al. 1987). We use here the modified form of this method suggested by Giorgi et al. (1993b), in which an exponential function is multiplied by the boundary relaxation term, which varies from 1 at the boundary toward 0 at the edges of the buffer zone.

The list of experiments performed in this work is given in Table 1. Basically, two sets of experiments were performed, one using the Kuo cumulus scheme and the implicit resolvable precipitation scheme (KI configuration) and one using the Grell cumulus scheme along with the simplified explicit moisture scheme (GE configuration). This allows us to test different available precipitation parameterization schemes in only two sets of experiments. For each of the two model configurations, five experiments were conducted, with different resolution and topography. In experiments KI(GE)-200, KI(GE)-100, KI(GE)-50, Δx = 200 km, 100 km and 50 km, respectively, and the topography is that of the corresponding resolution. In KI(GE)-100/200 and KI(GE)-50/200, Δx is equal to 100 km and 50 km, respectively, but the topography is bilinearly interpolated from that of the Δx = 200 km grid. Therefore, comparison of the results from various experiments allows us to isolate the effects of topography and resolution. The buffer zones for the various resolution experiments were adjusted to be of the same size (approximately 500–750 km in the middle to lower troposphere) by changing the value of the exponential correction function of Giorgi et al. (1993b). Note that the range of model resolutions studied here is representative of current global and regional climate models.

b. Criteria for model analysis and observation datasets

The main purpose of the present study is to examine the sensitivity of the model precipitation parameterizations to model resolution as compared to the sensitivity to topographic forcing. For this reason, most of the analysis is based on the intercom-

comparison between runs with various resolutions and topographies. In addition, since the precipitation process is closely related to different atmospheric variables and forcings, the analysis is not limited to precipitation but is extended to other physical and dynamical variables. In particular, in addition to precipitation, we analyze cloudiness, surface air temperature, sea level pressure (SLP), and the surface energy and water budgets.

In the evaluation of the model performance, use is made of three standard measures of deviation, that is; the bias, the root-mean-square error (rmse), and the correlation coefficient. For a variable $\phi$, the bias is defined as the time average (over 12-h snapshots) of the error, defined as

$$ E = \frac{\sum_i [\phi_i^M - \phi_i^O]}{N_t}, $$  

(5)

where the summation can be carried out either over a set of grid points or over a set of station locations $N_t$, and the superscripts $M$ and $O$ in Eq. (5) refer to model and observations (or analysis), respectively. The rmse and correlation coefficient (CORR) are defined as the time averages of

$$ \text{rmse} = \left( \frac{\sum_i [\phi_i^M - \phi_i^O]^2}{N_t} \right)^{1/2} $$

(6)

$$ \text{CORR} = \frac{\sum_i (\phi_i^M - \bar{\phi}^M)(\phi_i^O - \bar{\phi}^O)}{\left[ \sum_i (\phi_i^M - \bar{\phi}^M)^2 \sum_i (\phi_i^O - \bar{\phi}^O)^2 \right]^{1/2}}, $$

(7)

where the overbar in Eq. (7) indicates a spatial average of the quantity $\phi$ over the area under consideration (e.g., the interior of the domain).

For verification of the model performance in reproducing spatial precipitation patterns we use the precipitation threat score $T_p$, defined as

$$ T_p = \frac{C_{pr}}{O_{pr} + F_{pr} - C_{pr}}, $$

(8)

### Table 1. List of experiments performed in this work. (All experiments were completed for January 1991 and July 1991.)

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Resolution (km)</th>
<th>Topography resolution (km)</th>
<th>Cumulus scheme</th>
<th>Resolvable precipitation scheme</th>
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<td>Grell</td>
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<td>Grell</td>
<td>Explicit</td>
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where $O_P$ is the number of stations with observed monthly precipitation in excess of the precipitation threshold $P_T$, $F_P$ is the corresponding number for model forecasts, and $C_P$ is the number of stations where both observed and forecast monthly precipitation is in excess of $P_T$.

The threat score as defined by Eq. (8) is a measure of the accuracy of the model in forecasting the area that, for the simulated month, receives an amount of precipitation above a given threshold. It varies 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. The threat score as applied to the 30-day average precipitation is thus basically a measure of spatial forecast skill. In addition, for precipitation, we also analyze not only the monthly means but also the frequencies and intensity distributions of daily precipitation, since these are key elements for the application of climate model output to impact assessments.

For verification of precipitation and surface air temperature, a station observation dataset was assembled for the two simulated months. This dataset was obtained from the Climate Analysis Center of the United States National Meteorological Center (documentation for this dataset is available from the NCAR Data Support Section) and includes monthly averaged data from 1252 European stations in January 1991 and 1407 stations in July 1991. Verification of atmospheric variables is carried out by comparison with the driving ECMWF analyses.

4. Results and analysis

a. Brief description of prevailing synoptic events during the simulated months

The prevailing synoptic situations that occurred during the two simulated months are here described by presenting SLP maps for the ECMWF analyses at specific dates and comparing them with corresponding maps of model-produced SLPs. Note that the model simulations were started five days prior to the beginning of the month to allow for model spinup (e.g., Anthes et al. 1989).

January 1991 was dominated by two distinct flow regimes. The first half of the month, until about 13 January, was characterized by a prevailing zonal flow with a series of rapidly moving Icelandic storm systems traveling across northern Europe. An example of these systems is given in Figs. 2a,b. A few cases of secondary Alpine cyclogenetic lows also occurred, mostly affecting the eastern Mediterranean Basin. Conversely, the second part of the month was dominated by the development of a blocking high over central Europe (Figs. 2c–f), which formed on about 13 January and persisted until about 29 January. Associated with this blocking high was a strongly meridional flow carrying a flux of cold and dry Siberian air over the eastern Mediterranean. The path of North Atlantic cyclonic systems was deflected around the blocking high over Iceland and northern Russia. Meanwhile, two cases of pronounced cycloonic events over the central Mediterranean occurred around 15 and 27 January, as illustrated in Figs. 2c–f.

The first 10 days of July 1991 were mostly characterized by a high pressure region over central and northern Europe, with occurrence of weak cycloonic systems. The first intense Atlantic storm moved over the regions northwest of the British Isles around 11 July (Figs. 3a,b), where it persisted for a few days. This was followed by a rapid succession of three low pressure systems that moved across the northeastern Atlantic, the British Isles, the North Sea, and the Scandinavian Peninsula between 16 and 25 July (examples are given in Figs. 3c–f). The last few days of the month were then dominated by high pressure over the Scandinavian region, which deflected the storm path over the western and central European regions.

Both in January and July, the model generally captured the occurrence and evolution of the main synoptic events and flow patterns, although the intensity and location of the cycloonic centers in the model did not always closely follow the ECMWF analysis. During both months, a variety of storm systems and synoptic flow configurations affected the European region. This is important for our study, since the behavior of the parameterizations can be tested under a variety of conditions. Because our interest is on the implications of the parameterization behavior for climate studies, we do not specifically analyze any of the individual events of Figs. 2 and 3 but focus on the aggregated characteristics of the model response for the entire simulated periods.

b. Simulation of synoptic patterns

As measures of model skill in simulating synoptic weather patterns we here employ the biases, rmse, and correlation coefficients as defined in section 3b and calculated by comparison with the ECMWF analysis fields. They were first calculated for each 12-h snapshot using all points in the interior of the domain (i.e., by excluding areas within the buffer zone) and then were averaged throughout the entire simulated months. Table 2 shows SLP biases, rmse's, and correlation coefficients for all experiments.

SLP biases, rmse's and correlation coefficients are generally in the range of 2–2.5 mb, 4.1–4.7 mb, and 0.93–0.95 for January and 0.4–2.3 mb, 2.7–4 mb and 0.86–0.92 in July, respectively. For the 500-mb height (not shown) the biases, rmse's, and correlation coefficients have values in the range of 2 to $-7$ m, 27 to 35 m, and 0.98 in January and $-3$ to 4 m, 19 to 30 m, and 0.95 to 0.98 in July, respectively. As a reference, for SLP the average rmse's based on persistence, that is, calculated with respect to the initial state, were in the range of 14–15 mb in January and 7–8 mb in July,
Fig. 2. Observed (ECMWF analysis) and simulated sea level pressure at different times for the KI-50 km January 1991 experiment. (a) 1200 UTC 6 January, ECMWF analysis; (b) 1200 UTC 6 January, RegCM KI-50 experiment; (c) 0000 UTC 15 January, ECMWF analysis; (d) 0000 UTC 15 January, RegCM KI-50 experiment; (e) 0000 UTC 27 January, ECMWF analysis; (f) 0000 UTC 27 January, RegCM KI-50 experiment. Units are millibars, and the contour interval is 4 mb.
Fig. 3. Observed (ECMWF analysis) and simulated sea level pressure at different times for the KI-50 km July 1991 experiment. (a) 1200 UTC 11 July, ECMWF analysis; (b) 1200 UTC 11 July, RegCM KI-50 experiment; (c) 1200 UTC 18 July, ECMWF analysis; (d) 1200 UTC 18 July, RegCM KI-50 experiment; (e) 0000 UTC 25 July, ECMWF analysis; (f) 0000 UTC 25 July, RegCM KI-50 experiment. Units are millibars, and the contour interval is 4 mb.
Table 2. Sea level pressure bias (mb), rmse (mb²), and correlation coefficients (CORR) for different experiments.

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<td>1.3</td>
<td>1.3</td>
<td>0.8</td>
<td>2.0</td>
<td>1.6</td>
<td>1.4</td>
<td>1.1</td>
<td>1.1</td>
</tr>
<tr>
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<td>2.9</td>
<td>3.1</td>
<td>2.7</td>
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<tr>
<td>CORR</td>
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<td>0.88</td>
<td>0.90</td>
<td>0.88</td>
<td>0.91</td>
<td>0.92</td>
<td>0.89</td>
<td>0.91</td>
</tr>
</tbody>
</table>

while the persistence average correlation coefficients were in the range of 0.4–0.5 both in January and July. Note also that, based on persistence, the rmse’s rapidly increased and the correlation coefficients rapidly decreased after the first several days of spinup. Therefore, in the present experiments, the model showed a generally good performance in reproducing the aggregated characteristics of the dominant events during the simulated months.

Note that a positive SLP bias of 1–2 mb is observed, similar to that found in previous experiments (Giorgi et al. 1993a,b). This seems to be a characteristic feature of the RegCM (and the MM4), and it may be partially due to the interpolation procedure from the model surface to the sea level, which assumes an average temperature lapse rate of 6.5°C. One of the reviewers also suggested that this bias may be caused by the lack of inclusion of water vapor sources and sinks in the continuity equation for total air (e.g., Savijarvi 1995).

Overall, the aggregated SLP (and 500-mb height) model results are not very sensitive to model resolution, topography, and physics parameterizations. A general improvement in the measures of skill is found with increasing resolution, and the experiments employing low-resolution topography with high-resolution grid show values intermediate between the corresponding full low- and high-resolution experiments. The sensitivity of the SLP (and 500-mb height) climatology to the different precipitation schemes is small. In summary, all experiments show similar climatological skill in the simulation of synoptic system structure.

c. Precipitation analysis

1) Average precipitation

As illustrative examples of the model simulation of precipitation for the two months, Figs. 4a–d show the average January and July precipitation for experiments KI-200 and KI-50. In January 1991 precipitation maxima are found in the northwestern Atlantic and the central Mediterranean in correspondence of prevailing storm tracks. In July 1991 three regions of maximum precipitation occur over the British Isles, the Iberian Peninsula, and central eastern Europe. The different resolution runs generally show similar broad precipitation patterns; however, at the higher-resolution experiments, finer detail is present, especially in areas of complex topography and coastlines, such as the Iberian Plateau and the Pyrenees and the western Alps and the Balkans.

For a more quantitative analysis, Figs. 5a–f show the average precipitation over land and ocean areas in the interior of the domain and its breakdown in convective and resolvable-scale components. Also shown are the average precipitation amounts in an Alpine region extending from 45° to 48.5°N and from 5° to 15°E. Table 3 presents the precipitation biases for the whole European region (land only) and for the Alpine subregion, which were calculated by interpolating the model precipitation to the station locations and comparing the model results with the station dataset described in section 3b. Finally, Table 4 reports the 30-day precipitation threat scores for all experiments and different precipitation thresholds.

In January over land, all experiments produce similar amounts of precipitation, in the narrow range of 1.42–1.55 mm day⁻¹. The range of January results is somewhat wider over ocean, 2.30–3.20 mm day⁻¹. It is interesting, however, that the partitioning between convective and nonconvective precipitation is rather different when using the Grell and the Kuo convective schemes. With the Grell scheme, the fraction of convective precipitation is negligible over land and it is in the range of 5%–15% of total precipitation over ocean. Conversely, when using the Kuo scheme the percentage of January convective precipitation is of the order of 40%–50% over land and 45%–60% over ocean. This is in contrast to the July results, which show larger fractions of convective precipitation in the Grell than in the Kuo scheme over land.
These results illustrate the different dominant mechanisms that force convection in the two schemes. In the Kuo scheme, the primary triggering factor is the vertically integrated large-scale moisture convergence. Initiation of convection also requires unstable vertical profiles, but a perturbation to the temperature and moisture fields at the level of highest moist static energy is added in the calculation of the vertical convective stability. In the Grell scheme, the dominant triggering mechanism is large-scale buoyancy production. As a result, the Grell scheme mostly responds to lower-level heating and destabilization, which over land is maximum in the summer and negligible in the winter. Conversely, the Kuo scheme still produces some winter convection because of the use of the convergence criterion and of the perturbation temperature and moisture in the stability calculations.

Note that observational studies (e.g., Houze et al. 1976; Herzegh and Hobbs 1980; Matejka et al. 1980) have indicated that convective rainbands also occur at
Table 3. January and July precipitation bias for the European and Alpine regions calculated by comparison with station observations. Units are percentage of observed precipitation.

<table>
<thead>
<tr>
<th></th>
<th>KI-200</th>
<th>KI-100/200</th>
<th>KI-100</th>
<th>KI-50/200</th>
<th>KI-50</th>
<th>GE-200</th>
<th>GE-100/200</th>
<th>GE-100</th>
<th>GE-50/200</th>
<th>GE-50</th>
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<td>35.3</td>
<td>34.1</td>
<td>34.1</td>
<td>38.3</td>
<td>34.2</td>
<td>37.0</td>
<td>46.5</td>
<td>33.3</td>
<td>43.3</td>
<td>34.2</td>
</tr>
<tr>
<td>Alps</td>
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<td>-10.0</td>
<td>30.1</td>
<td>-4.4</td>
<td>31.0</td>
<td>70.0</td>
<td>7.8</td>
<td>60.3</td>
<td>-7.3</td>
</tr>
<tr>
<td>Europe</td>
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<td>21.8</td>
<td>35.8</td>
<td>38.5</td>
<td>13.9</td>
<td>37.5</td>
<td>35.6</td>
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</tr>
<tr>
<td>Alps</td>
<td>-35.4</td>
<td>4.8</td>
<td>-25.5</td>
<td>3.7</td>
<td>-2.2</td>
<td>-28.9</td>
<td>15.9</td>
<td>-8.6</td>
<td>54.3</td>
<td>2.9</td>
</tr>
</tbody>
</table>

warm fronts or in the warm sector of extratropical cyclones, even during the cold season. These convective cells are likely generated by the lifting of low-level air by frontal motions until the air becomes saturated and unstable in the midtroposphere. The fact that the Grell scheme does not generate convective precipitation in the cold season suggests that its closure, based on available buoyant energy production, is unable to capture this process.

January precipitation tends to slightly increase with model resolution, both over land and ocean. In the Kuo experiments, this is mostly due to an increase in convective precipitation, while in the Grell experiments, the contribution of resolvable-scale precipitation is smaller. In July the increase in precipitation with resolution is more pronounced than in January, both for convective- and resolvable-scale precipitation. Total July simulated precipitation increases by about 35% over land and 50–60% over ocean between the 200 km and the 50 km, both in the KI and in the GE model configurations.

For continental-scale averaged precipitation, the effect of resolution change is greater than that of topography, as the precipitation amounts in the high-resolution/coarse-topography runs are closer to those of the high-resolution runs rather than those of the corresponding coarse-resolution runs. If we, however, restrict our attention to the Alpine region, the effect of topography becomes greater than that of resolution in all cases except the July runs in the KI configuration. The results of Figs. 5a–f thus indicate that the direct effect of resolution alone can strongly modify the precipitation amounts averaged over regions of continental size, while topography mostly contributes to modifying the spatial redistribution of precipitation and is thus especially important only over subregions characterized by pronounced topographical features. A more extended discussion of topographic effects on precipitation is given in section 4c(1).

Also of interest in Figs. 5a–f is the result that the model sensitivity to the precipitation parameterization used (e.g., KI versus GE configurations) is of the same order of, or greater than, the sensitivity to topography and horizontal resolution, at least in the summer simulations. This occurs at all resolutions and mostly over land, when convective activity is at its maximum.

Comparison of model and observed precipitation indicates that both in January and July the model overpredicts precipitation except at the lowest resolutions over the Alpine region in July (see Table 3). Because

Table 4. January and July precipitation threat scores for different precipitation thresholds $T_r$.

<table>
<thead>
<tr>
<th>$T_r$ (mm day$^{-1}$)</th>
<th>KI-200</th>
<th>KI-100/200</th>
<th>KI-100</th>
<th>KI-50/200</th>
<th>KI-50</th>
<th>GE-200</th>
<th>GE-100/200</th>
<th>GE-100</th>
<th>GE-50/200</th>
<th>GE-50</th>
</tr>
</thead>
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<tr>
<td>January</td>
<td></td>
<td></td>
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<tr>
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<td>0.95</td>
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<td>0.70</td>
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</tr>
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<td>0.11</td>
<td>0.18</td>
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<tr>
<td>July</td>
<td></td>
<td></td>
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<tr>
<td>0.1</td>
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<td>0.95</td>
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<td>0.43</td>
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<td>0.24</td>
<td>0.24</td>
<td>0.25</td>
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</table>
Fig. 5. Histograms of total simulated precipitation for various experiments. (a) January 1991, land points; (b) July 1991, land points; (c) January 1991, ocean points; (d) July 1991, ocean points; (e) January 1991, Alpine region; and (f) July 1991, Alpine region. Darker areas indicate the amount of convective precipitation; lighter areas indicate resolvable-scale precipitation. Units are millimeters per day.
precipitation tends to increase with model resolution, the largest biases are found for the 50-km runs, especially when using the GE configuration. Note that this result is different from that of Giorgi and Marinucci (1991) and Giorgi et al. (1993a,b), who mostly found an underprediction of precipitation for the simulation of January and June 1979 over Europe (with model gridpoint spacing of 70 km). This suggests that to some extent the model biases depend on the simulated period and perhaps on the characteristics of the analysis fields (in particular water vapor) used to drive the model. A much longer simulation time is required to fully assess the model performance, which is beyond the purpose of the present study.

In addition, in the comparison of model and observed data, two points should be considered. First, in the station dataset there is a prevalence of low elevation and valley stations. Second, it has been pointed out that gauge measurements can underestimate precipitation by up to 20%-30%, especially at high elevations under strong wind conditions (e.g., Sevruk 1975, 1989). Elevation corrections designed to account for this effect have been developed for annual precipitation amounts. However, for our monthly January and July datasets we did not include any elevation correction. For these two reasons, precipitation in the observed dataset is likely underestimated, especially in mountainous areas. Also to note is that the most pronounced overprediction of July precipitation occurred over the Iberian Peninsula and accounted for a substantial portion of the overall positive bias in the GE runs.

The threat scores are close to 1 for the lowest threshold, 0.1 mm day$^{-1}$ (see Table 4). This is simply an indication that both in the model and in the observations most of Europe received some precipitation during the simulated months. The scores then decrease as the precipitation threshold increases and are generally higher in the summer than in the winter simulations. The lowest threat scores are found for the 200 km runs at all thresholds, an indication that at this lowest resolution the precipitation spatial patterns are less accurately resolved. Although the differences in scores between various experiments are not large, the January runs at 100-km resolution and the July runs at 50-km resolution show the highest scores, and in almost all cases the use of finer topography enhances the scores. The July threat scores of Table 4, are in the same range as those found by Giorgi and Marinucci (1991) for their 1979 simulations over Europe. The January scores are, however, somewhat lower than those of Giorgi and Marinucci (1991) over Europe and Giorgi and Bates (1989) over the western United States, likely a reflection of the January model precipitation overestimate in the present experiments.

A final aspect of interest in the results of Figs. 5a–f is that in July the same boundary conditions can support precipitation amounts that vary by up to 60% over land and 50% over ocean between the KI-200 and GE-50 simulations. This is mainly due to two effects: 1) changes in the model efficiency to precipitate moisture advected into the region and 2) changes in model efficiency to locally precipitate evaporated water. If we restrict our attention to land areas, the comparison of the precipitation data in Figs. 5a–f and the latent heat flux data of Table 5 shows that the relative precipitation changes between experiments are substantially greater than the changes in surface evaporation, so that the effect of local water recycling is relatively small. This implies that the atmospheric water vapor amounts in the domain are lower for the runs with greater precipitation. We compared the model water vapor amounts with the water vapor in the driving analysis and found that the lower boundary layer was drier in the model than in the analysis for the higher-resolution (and higher precipitation) experiments.

2) Topographic forcing of precipitation

Figures 6a–d show plots of precipitation difference as a function of topography difference between the 50- and 200-km runs and between the 50- and 50/200-km runs for the KI and GE model configurations. Both for January and July and in both the KI and GE model configurations the importance of the topographic forcing is illustrated by the positive slope of the curves in Figs. 6a–d.

It is interesting to compare the topography/precipitation curves of Figs. 6a–d for the 50- versus 200-km experiments (continuous curves) and the 50- versus 50/200-km experiments (dashed curves). The latter isolate the effects of topographic forcing, while the former is the result of the combined effects of topography and resolution. In July, although the two curves are displaced from each other as a result of the resolution change, they show similar monotonic, topographically induced trends (see Figs. 6b,d). In January, however, the continuous curves tend to asymptote to a near-horizontal trend after a threshold in topography difference is reached. This suggests that in January, when precipitation is mostly generated at the resolvable scale, the topographic effect is somewhat masked by the resolution effect. The comparison of the 100-, 200-, and 100/200-km experiments showed results similar to those of Figs. 6a–d.

The effect of topography on precipitation is further illustrated by Figs. 7a–b, which show precipitation as a function of elevation for the observing stations and the 50-km model runs. The observation curves were produced by averaging precipitation from all stations within given elevation intervals, while the model curves were produced with a similar procedure using the simulated data at the model grid points. Note that the averages at the highest elevation intervals were obtained from only a few stations and grid points. In addition, as discussed in section 4c(1), because of the absence of a gauge correction, the observed high-ele-
Table 5. Simulated total cloud fractional cover (CLOUD), surface net solar (SOL), net infrared (IR), latent heat (LH), and sensible heat (SH) fluxes over land for a subset of experiments at different resolutions. The fractional cloud cover is in percent and the surface fluxes are in watts per square meter (positive upward for IR, SH, and LH; positive downward for SOL).

<table>
<thead>
<tr>
<th></th>
<th>KI-200</th>
<th>KI-100</th>
<th>KI-50</th>
<th>GE-200</th>
<th>GE-100</th>
<th>GE-50</th>
</tr>
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<tbody>
<tr>
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<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>CLOUD</td>
<td>72</td>
<td>65</td>
<td>44</td>
<td>57</td>
<td>48</td>
<td>40</td>
</tr>
<tr>
<td>SOL</td>
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<td>47.2</td>
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</tr>
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<td>-10.6</td>
<td>-12.1</td>
<td>-13.1</td>
</tr>
<tr>
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<td>10.1</td>
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<td>9.3</td>
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<tr>
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<tr>
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<td>57.5</td>
<td>51.7</td>
<td>60.1</td>
<td>61.9</td>
</tr>
<tr>
<td>LH</td>
<td>78.0</td>
<td>80.6</td>
<td>83.5</td>
<td>89.6</td>
<td>96.4</td>
<td>102.8</td>
</tr>
</tbody>
</table>

viation precipitation data may be somewhat underestimated.

In January and especially in July, observed precipitation tends to increase with elevation until it reaches a maximum and then decreases at higher elevations. This can be expected in view of the fact that little water vapor is available at high altitudes. Both in the model and in the observations the elevation dependency of precipitation is more pronounced, and the precipitation maximum occurs at higher elevations in summer than in winter. However, the elevation effect in the model is more pronounced (i.e., the curves have steeper slopes) than in the observations. In addition, in summer the model maximum occurs at lower elevations than observed, although it is difficult to compare simulated and observed high-elevation summer data, since the observations reach much higher elevations than attained by the model topography. In summary, the model at least qualitatively reproduces some observed trends, although significant discrepancies with observations still exist. Comparison of results with different model configurations indicates that the GE configuration shows a more pronounced sensitivity to topography than the KI configuration, especially in the summer experiments, likely due to the production of buoyant energy by high-elevation heating.

3) PRECIPITATION FREQUENCIES AND INTENSITY DISTRIBUTIONS

An aspect of precipitation climatology that is important for impact assessments is the frequency and intensity distribution of daily precipitation events, which can be as important, or even more important, than average precipitation. Analyses of daily precipitation statistics in models and station data generally indicate that models tend to overpredict precipitation frequencies and the number of light precipitation events (e.g., Mears et al. 1995). This has been attributed in part to the model parameterizations and in part to the fact that the model precipitation is more representative of an average over the grid box size rather than a point value. If the latter effect is predominant, the simulated precipitation intensity distributions should be sensitive to the model resolution.

With our set of experiments we can test the sensitivity of precipitation frequencies and intensity distributions to both model parameterizations and model resolution. Figures 8a–d show daily precipitation intensity distributions over land grid points for the different experiments. These were produced by calculating the frequency of occurrence of daily precipitation events within given intervals of intensity divided by the total number of daily rain events. Table 6 shows the precipitation frequencies, that is, the total number of daily rain events (i.e., with precipitation greater than 0.1 mm day$^{-1}$) divided by the total number of gridpoint days for a subset of experiments.

Comparing precipitation frequencies for different resolution experiments shows that over land the frequencies increase with resolution, both in January and July. This indicates that local supersaturation in winter and cumulus cloud activation in summer tend to occur more often as the resolution increases. This result is consistent with the general increase of precipitation with resolution shown in Figs. 5e–f. Exceptions to this trend are the January frequencies over ocean, which tend to decrease with increasing resolution in response to an increase in the convective component of precipitation. Of interest is the result that in winter the explicit scheme of the GE configuration tends to produce lower frequencies than the implicit scheme of the KI configuration. This can be expected, because the use of a cloud water threshold for precipitation generation tends to inhibit the occurrence of drizzle events characteristics of implicit schemes that are solely based on super-
saturation. The results of Table 6 also indicate that resolution is not the primary cause of the precipitation frequency overestimate found by Mearns et al. (1995).

The intensity distributions show a well-defined peak at the interval 1–2.5 mm day$^{-1}$, which is more pronounced in summer than in winter. In winter, a larger percentage of light events occurs than in summer. These trends are to be expected in view of the different nature of the summer (local convective) and winter (large scale) precipitation processes. The effect of resolution is evident in Figs. 8a–d. In both model configurations, and especially in summer, the percentage of events with intensities greater than 5 mm day$^{-1}$ increases with resolution and the peak of the distribution becomes less pronounced. From the discussion of Mearns et al. (1995), who indicate that models tend to underestimate the occurrence of medium to heavy precipitation events and overestimate the occurrence of light events, the higher-resolution results appear in the direction of increased realism and suggest that resolution may account for a significant fraction of the model–observation intensity discrepancies.

d. Cloudiness and surface energy budget

By affecting precipitation, model resolution can modify the surface energy and water budgets. In addition, since cloudiness is calculated from threshold rel-
ative humidity values (see section 2a), it is also likely to depend on model resolution. Table 5 shows the simulated total cloudiness and surface fluxes of sensible heat, latent heat, and net infrared and solar radiation averaged over land areas in the interior of the domain for a subset of different resolution experiments.

All the quantities presented in Fig. 8 show monotonic variations with resolution. Cloudiness decreases significantly with increasing resolution, both in the KI and GE configurations. We qualitatively compared the model average cloud amounts with the International Satellite Cloud Climatology Program cloudiness data (Hurrell and Campbell 1992) for the two simulated months and found that the simulated average cloudiness was closer to the observed values in the 50-km runs than in the 200-km runs (not shown).

There are two main processes that may affect the cloudiness sensitivity to resolution in the model. The first is associated with precipitation. As precipitation increases, the atmospheric water vapor decreases (since the influx from the boundaries is approximately the same in all runs for a given month); hence, the atmospheric relative humidities, and therefore the cloudiness, decrease. This process appears more important in the July than in the January simulations, since precipitation does not change much between different January experiments. The second process that affects cloudiness is associated with the fact that non-convective clouds are formed when a relative humidity threshold (80% in the KI and 100% in the GE configuration) is exceeded and that the maximum fractional cover attainable at a grid point is constant (100% when gridpoint saturation is reached). Intuitively, it can be expected that these lower and upper limits should be scale dependent and that in particular both the lower threshold for cloud formation and the maximum cloud cover should increase as the gridpoint spacing decreases. Lack of such scale dependency could contribute to the sensitivity of simulated cloudiness to resolution.

As cloudiness decreases with increasing resolution, the absorbed solar flux and net infrared loss at the surface increase. In summer, the increase in solar absorption exceeds the increase in infrared loss, so that surface temperature and sensible and latent heat fluxes also tend to increase with resolution. Conversely, in winter the increase in infrared radiative loss exceeds the increase in solar absorption, so that surface temperature as well as (upward) sensible and latent heat fluxes decrease with resolution. In summer the sensitivity of the surface fluxes to the resolution tested is of the order of 5–15 W m\(^{-2}\) for latent and sensible heat flux, 17–22 W m\(^{-2}\) for the infrared loss, and 40–45 W m\(^{-2}\) for solar radiation absorbed. In winter, the sensitivity to resolution is of the order of a few watts per square meter for sensible and latent heat flux and a few tens of a watt per square meter for infrared and sensible heat flux. Significant differences in cloudiness and surface fluxes (up to a few tens of a watt per square meter) are also found between the KI and GE configurations. The ranges of variations in surface fluxes shown in Table 5 are significant and, in fact, they are comparable to those that can be produced by different types of vegetation (Seth et al. 1994).

The surface temperatures showed a sensitivity to resolution of about 1\(^\circ\)–2\(^\circ\)C. For the subset of experiments of Table 5, Table 7 reports the model versus station observation surface air temperature biases over the European region and the Alpine subregion. These were calculated by interpolating the model temperatures to the station locations and applying a constant-lapse-rate correction (6.5 K km\(^{-1}\)) to account for differences be-
between the model and the station elevations. The biases are generally small, less than 1°C, both for summer and winter. Note that the observed temperatures in Table 7 are taken at an anemometer height of a few meters, while the model surface air temperatures are given by the temperatures at the lowest atmospheric model level, which is at a height of approximately 20 m. Previous experience has shown that while this discrepancy can affect the comparison of model and observed maximum and minimum daily temperatures, its effect on average daily temperatures is generally small.

5. Summary and discussion

In this paper we have discussed a series of numerical experiments with a regional climate model aimed at investigating the model sensitivity to resolution and topographic forcing under different configurations. The primary result we found was that the continental-scale average precipitation was more sensitive to resolution than to topographic forcing. The effect of topography was more important in spatially redistributing precipitation, and its role was dominant in subregions of the European continent, such as the Alpine region, characterized by complex topographical features. Other climatological variables examined in this work, such as precipitation frequencies and intensity distributions, cloudiness, and the surface energy and water fluxes were sensitive to model resolution. In addition, simulated precipitation was sensitive to the parameter-
TABLE 6. Daily precipitation frequency over land and ocean areas for a subset of experiments at different resolutions.

<table>
<thead>
<tr>
<th></th>
<th>KI-200</th>
<th>KI-100</th>
<th>KI-50</th>
<th>GE-200</th>
<th>GE-100</th>
<th>GE-50</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>January</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land</td>
<td>62%</td>
<td>66.1%</td>
<td>65.0%</td>
<td>50.5%</td>
<td>57.2%</td>
<td>61.7%</td>
</tr>
<tr>
<td>Ocean</td>
<td>78.8%</td>
<td>77.9%</td>
<td>72.2%</td>
<td>68.2%</td>
<td>68.4%</td>
<td>65.7%</td>
</tr>
<tr>
<td></td>
<td>July</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land</td>
<td>50.1%</td>
<td>52.6%</td>
<td>55.9%</td>
<td>54.4%</td>
<td>57.9%</td>
<td>62.5%</td>
</tr>
<tr>
<td>Ocean</td>
<td>40.1%</td>
<td>45.3%</td>
<td>43.6%</td>
<td>36.4%</td>
<td>41.0%</td>
<td>46.0%</td>
</tr>
</tbody>
</table>

izations used, especially during the summer over land areas.

Several factors in the precipitation parameterizations may contribute to the model sensitivity to resolution found in this study. For example, the closures adopted by the parameterizations used in this work, based on moisture convergence and buoyant energy release by clouds, may behave differently at different resolutions. The use of threshold values is also likely to enhance the model sensitivity to resolution. Thresholds are, for example, used in the convergence criterion of the Kuo scheme and in the formation of large-scale cloud and precipitation. Finally, key parameters, such as the precipitation efficiency, are possibly scale dependent.

A further source of resolution dependency of precipitation is provided by the simulated scales of motions involved in the process, regardless of the surface forcings. From our experiments, it is difficult to assess the importance of this effect, but the fact that the measures of the model skill in simulating synoptic weather patterns did not vary substantially between experiments employing different resolutions suggests that, at least for our experiments, the sole effect of finer scales of motion was not dominant.

Our results have important implications for surface climate simulation, especially at the regional scale, be they produced by nested models, variable resolution global models, or high-resolution GCMs. They indicate that, when increasing resolution, the effects of physical forcings (e.g., better representation of topography and coastlines) may be masked by the direct sensitivity of the model parameterizations to resolution itself, at least on the continental scale. The parameterizations we tested here are specific to our model but they are similar in design and complexity to those used in current global and regional climate models, so that our results are more general in nature.

The issue is then of how to correct for the direct model sensitivity to spatial resolution when a model is designed to be used on a wide range of spatial scales. First, the model sensitivity needs to be assessed with controlled experiments of the type we have performed here. Different strategies can then be followed to minimize the effects of model resolution. The simplest approach is to return the model at different resolutions by modifying suitable key parameters. This would be the most useful approach when a nested climate model employs the same parameterizations as a driving GCM, or when a GCM is used on a wide range of resolutions (such as in time slice mode). An alternative approach is to use different parameterizations at different resolutions. This is the main reason underlying the use of different schemes in a driving GCM and a nested RegCM: each scheme is developed and tested for a specific resolution range. The third, most general but also most challenging approach is to design a scheme that would give internally consistent results on a wide range of resolutions. This would be especially useful, if not necessary, when using variable resolution models in which the gridpoint spacing varies continuously rather than abruptly over scale ranges of up to one order of magnitude.

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REFERENCES


