Atmospheric Blocking in a General Circulation Model

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Abstract

Diagnostics from ECHAM climate simulations are presented in this thesis. The focus is on a flow phenomenon in the midlatitudes of the Northern Hemisphere that challenges the simulation capabilities of atmospheric GCMs. With their partly irregular incidence, blocking highs contribute significantly to the variability in a frequency range that is positioned between the variability due to cyclones and due to climate fluctuations. The distinctive feature of the present study is the availability of multi-year integrations of high resolution (T106 spectral resolution, corresponding to a maximal grid spacing of 125 km).

Blocking-like flow situations in sequences of 500 hPa geopotential heights are identified by a meridional gradient criterion. It is found that the ECHAM models (mark 3 and 4) capture the distinct annual cycles of the blocking activity in the two main blocking sectors of the Northern Hemisphere. However, the models simulate too few blocked days in winter. This underestimation is attributed to model biases of the high resolution flow which is too zonal particularly over the northeastern Atlantic. Despite a larger zonal bias, the T106 model is generally superior to the T42 resolution in terms of blocking frequencies. Individual blocking cases in the high resolution simulations share many features of real events. Multiple incursions of low-latitude air, advected by strong southerlies, and homogenization of low potential vorticity air inside the blocking high are displayed in the sequences of maps from four cases distributed over the seasons.

Perpetual January simulations (T42) with gradually reduced orography have been conducted. The impact of the GWD is similar to truncating the mountain heights. Removal of the mountains leads to a weaker Asian anticyclone and to an intensification of the Icelandic low. The more uniform zonal winds in the no-mountain simulations give less blocking days. Orography above a certain threshold seems sufficient to trigger large scale Rossby wave breaking resulting in dipole blocks. The truncated orography excites almost exclusively broad ridge events.

Finally, the North Atlantic oscillation index and blocking frequencies have been extracted from a Twentieth Century integration to document the role of internally generated interannual and decadal atmospheric variability.
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Zusammenfassung

Simulationen mit dem Klimamodell ECHAM werden in dieser Arbeit vorgestellt. Die Diagnostik konzentriert sich auf ein Phänomen der nordhemisphärischen Strömung in den mittleren Breiten, welches hohe Anforderungen an die Simulationsfähigkeiten atmosphärischer GCMs stellt. Mit ihrem teilweise unregelmäßigen Auftreten liefern blockierende Hochdruckgebiete einen wesentlichen Beitrag zur Variabilität in einem Frequenzbereich, der zwischen der baroklin verursachten Variabilität und klimatischen Schwankungen liegt. Eine Besonderheit der vorliegenden Studie ist die Verfügbarkeit mehrjähriger, hochaufgelöster Integrationen mit spektraler Auflösung T106, was einer maximalen Maschenweite von 125 km entspricht.


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1 Introduction

Numerical models of the atmosphere were among the first applications of the computer. Weather prediction and climate simulation are still considered as first class consumers of today's computing power. A broad stream of research develops ever more comprehensive models to meet the complexity of the climate system, and the public demand for reliable estimates of anthropogenic climate change. The claim of these global climate models is to represent as many as possible of the relevant processes as good as possible. That simpler models or even iterations of the logistic equation may be more appropriate to study the time evolution of the system is one sting to climate modelling. The recent IPCC report reflects this by stressing the uncertainties which arise from the nonlinear behaviour of the atmosphere-ocean system.

A classical subject of dynamical meteorology is taken up in this thesis that may well be generated almost exclusively by internal dynamics within the troposphere of the middle latitudes. Namely, the tendency of the atmosphere to enter at recurrent positions, from time to time, into a blocked state is investigated with a general circulation model (GCM). The aim of diagnosing output from such a complex tool is to learn more about GCM behaviour, as well as to better understand blocking processes. With a time scale of about seven days, blocking anticyclones and the 'unusual' weather associated with them may lead to severe drought for the region below and very cold conditions on the eastern flank and to the south of the blocking high. Especially this is true when the block is reinforced and persists for several weeks. Such fluctuations obscure the perception of a gradual or abrupt change of climatic conditions. In any case, it would be useful to know more about possible influences on the blocking activity, and on the low-frequency variability in general.

1.1 An outline of some previous investigations

It is known for a long time that warm anticyclones exert a steering effect on weather systems. The first specific description of blocking was given by Garriott (1904) in a review on long-range weather forecasts:

> When the great continental high area extends westward over west-central Europe and the British Isles it checks the succession of North Atlantic storms, and finally affects the rate of progression of high and low areas over the United States.

This sentence suggests that blocking highs originate from the Asian anticyclone, slowly propagate westward while slowing down and deflecting North Atlantic disturbances. Whereas the westward drift of the blocking action is observed for many cases, the dynamically active regions over the oceans are considered to play an important role for the initiation and maintenance nowadays.
Another long known feature of the general circulation in the middle latitudes of the northern hemisphere is the tendency to fluctuate between a predominantly zonal and a more meridional flow pattern, a behaviour which has been termed the 'index cycle' (e.g., Namias and Clapp 1951). The phenomenon was extensively studied when upper air soundings became available after World War II. Berggren et al. (1949) proposed to explain the breakdown of the westerly flow by analogy to the hydraulic jump. A thorough study of the blocking action and its effect upon European climate was given by the work of Rex (1950, 1951). According to Rex, a blocking case must exhibit the following characteristics:

a) the basic westerly current must split into two branches,
b) each branch current must transport an appreciable mass,
c) the double-jet system must extend over at least 45° of longitude,
d) a sharp transition from zonal type flow upstream to meridional type downstream must be observed across the current split, and
e) the pattern must persist with recognizable continuity for at least ten days.

Many subsequent studies have used this definition, often with a less restrictive duration requirement.

Namias (1964) suggested that the enhanced blocking activity during 1958 to 1960 was due to sea-surface temperature anomalies and cold-season snow cover over Scandinavia. Whether the ocean plays an active role or is merely an integrator of the atmospheric 'noise' is, however, still under debate (e.g., Latif and Barnett 1994; Lau 1997).

The higher predictability during blocking periods and its use for medium range forecasts was investigated by Bengtsson (1981) in a case study with numerical weather prediction models of varying resolutions. Input also came from theory by the finding of multiple stable solutions of the barotropic vorticity equation (Charney and DeVore 1979) and its application to blocking (Charney et al. 1981). In addition to this global approach, McWilliams (1980) applied the existence of localised solutions, called modons, to blocking. The renewed interest in blocking triggered a number of observational studies. A subjective catalogue of blocking situations during the period 1945–1977 was composed by Treidl et al. (1981). Lejenäs and Økland (1983) defined an index to search for blocked flow in the NMC analyses of the period 1950–1979 in an objective manner. ‘Subjective’ and ‘objective’ are used here merely to distinguish between a catalogue either established by a computer program or one based on inspection of charts ‘by hand’. A somewhat more general approach was followed by Dole and Gordon (1983) who analysed geopotential height anomalies of either sign. The aim of these studies was mainly to collect empirical facts and statistical properties of the phenomenon, and not much was said about the mechanisms. As for blocking theories, ‘anything goes’ (Feyerabend 1983) seems to describe the situation best. The explanations either stress the circumpolar or the local aspect more, and the variability is either thought of as being internally generated or triggered by surface anomalies. A survey of the causes of the low-frequency variability in general is given, e.g.,
in Wallace and Blackmon (1983), while a review of theories of blocking is presented by Bengtsson (1979). The short notes under the following headings introduce seven notions considered relevant for blocking.

**Rossby wave dispersion**

In the simplest case of a uniform basic zonal mean flow $U$, the dispersion relation for a plane Rossby wave with total wavenumber $K = (k^2 + l^2)^{1/2}$ is

$$\omega = kU - \frac{k\beta}{K^2},$$  \hspace{1cm} (1.1)

where $\beta$ is the variation of the Coriolis parameter with latitude. The zonal phase speed, $\omega/k$, of Rossby waves is always westward relative to the basic flow. Long waves become stationary relative to the ground for

$$K_s = \left(\frac{\beta}{U}\right)^{1/2}. \hspace{1cm} (1.2)$$

The zonal group velocity for a barotropic Rossby wave is

$$u_g = \frac{\partial \omega}{\partial k} = U + \frac{\beta(k^2 - l^2)}{K^4},$$  \hspace{1cm} (1.3)

which gives, evaluated at the stationary wave number $k_s = (K_s^2 - l^2)^{1/2}$,

$$u_g = 2U \frac{k_s^2}{K_s^2}. \hspace{1cm} (1.4)$$

Energy propagation from a stationary Rossby wave is thus confined to the downstream region. The situation becomes more complicated when the basic flow is no longer uniform. The same formulas, but with $\beta$ replaced by $\beta - U_{yy}$ apply for flows with variations in the $y$-direction. The curvature term $U_{yy}$ is particularly large for zonal mean wind profiles $U(y)$ with a double jet and a minimum in between.

Yeh (1949) investigated the energy propagation through dispersive waves in four atmospheric models and found that upstream propagation is possible with a free upper surface only. Furthermore he established that the lifetime of a solitary wave increases strongly with latitude.

**Amplification and breaking of planetary waves**

Tung and Lindzen (1979) suggested that planetary wave amplitudes grow strongly during some winters when the topographic and thermal forcings interfere more constructively than during other winters. Saturation and wave breaking limit the resonant growth.

**Inverse energy cascade**

Imagine an incompressible fluid in two dimensions with energy containing eddies confined to a narrow band of wave numbers. Inertial interactions will spread the energy to adjacent
wave numbers. As enstrophy, i.e. half the squared vorticity, must also be conserved, any spreading towards larger wavenumbers (smaller eddies) is accompanied by a shift of the center of the energy spectrum to smaller wavenumbers (larger scales). This fact leads to the accumulation of energy in large eddies while the vorticity is strained into thin filaments (Rhines 1979). Examples of such behaviour of geostrophic turbulence are found in the Gulf Stream, and in midlatitude atmospheric dynamics. Tanaka (1991) undertook numerical experiments to demonstrate the role of the upscale energy cascade from synoptic disturbances for blocking formations. He finds that nonlinear wave-wave interactions cause wavenumber one to grow. Topographic forcing at wavenumber two plays a catalytic role for the feeding of wavenumber one by the synoptic waves. A particular situation of upscale energy transfer is mentioned next.

*The eddy straining mechanism*

Figure 1.1, taken from Shutts (1983), illustrates the straining of eddies in a pre-existing split jetstream encompassing a blocking dipole. The travelling disturbances are severely deformed and subsequentially split in two. In the course of this process the synoptic scale depressions loose parts of their energy to the larger scale blocking field. Anticyclonic vorticity forcing occurs in the northward branch of the jet. Green (1977) put forward transient eddy forcing arguments to explain the summer blocking and drought of 1976. This case was further investigated with potential vorticity diagnostics by Illari (1984).

*Modons*

Both the barotropic and the quasi-geostrophic vorticity equations possess exact localized solutions that very much resemble the observed pattern of blocks of the dipole type. While these solutions called modons (from 'Mid-Ocean Dynamics experiment') or solitary eddy solutions are very appealing they do not account for time dependent processes, and it is not
clear whether they are stable under dissipative perturbations (McWilliams 1980; Haines and Marshall 1987). Still, the existence of modons helps to understand the longevity of some dipole blockings, namely the dispersion is balanced by nonlinear interactions which maintain the coherent structure over a long time. A really long-lived anticyclonic vortex, the Great Red Spot, has been observed since Galilei's time on Jupiter where the ratio of the external Rossby radius \( r_R = (gH)^{1/2}/f \) to the planetary radius is smaller than on Earth. More promising for the initiation phase is the following point.

**Three-dimensional instability**

Frederiksen (1982) examined the instability characteristics in a spherical quasigeostrophic two-layer model and found that the fastest growing mode in the less unstable case (his case 2a) has a large-scale dipole structure with maximum amplitude in the Pacific. In this way, he was able to establish a theory of both cyclogenesis and blocking onset.

**Regime-like behaviour of the atmosphere**

The notion that the atmospheric evolution does not visit all regions of the phase space with equal probability is both very natural and abstract. Several authors have investigated whether regimes may be identified in the long term weather, and have related regime transitions with blocking (Reinhold and Pierrehumbert 1982; Hansen and Sutera 1986; Marshall and Molteni 1993). Often the problem of these theories is to determine whether the truncation of the abstract phase space of the atmosphere to a tractable size does not generate spurious regimes but preserves the structure of the attractor.

Apparently, a number of theoretical concepts to explain blocking in idealized settings coexist. But also for weather and climate simulations, the phenomenon is a notorious challenge. Tibaldi and Molteni (1990) found that the ECMWF model almost consistently missed blocking onsets by forecast day 3 to 4 in the winter forecasts from 1980–1987. More recently, Anderson (1993) investigated the climatology of blocking in the NMC numerical forecast model and observed a drop of the simulated blocking frequency at lead times near ten days in extended forecasts. For greater lead times, when the transition between observed and model climates had finished, the number of blocks increased. The blocking performance is also assessed for climate simulations as one element of the diagnostic evaluation (Blackmon et al. 1986; D'Andrea et al. 1996). Since deviations from the climatological means are high during blocking episodes, a poor skill in the blocking distribution impairs the credibility of a global model on the regional scale. GCMs are also used as a surrogate data source to study blocking processes under a controllable model environment (Mullen 1986). The present study assesses the blocking performance of the simulation tool ECHAM, installed in Switzerland in 1992. It also presents results from a series of no-mountain experiments.
1.2 Objectives of the present study

The following questions have motivated the work:

- What is the skill of a comprehensive atmospheric general circulation model to simulate blocking-like flows?
- How realistic are developments of individual cases? In particular: Does the blocking process profit from an increase in horizontal resolution of the simulations?
- Does the examination of blocking processes provide specific hints on model deficiencies (e.g., GWD-parameterization)?
- Assumed that the answer to the first two questions is reasonably positive, how can GCM generated data be used to test existing concurrent blocking theories and help to answer the question whether blocking is more dependent on global conditions (amplitudes of planetary scale waves, zonal mean flow) or can be understood as a local phenomenon induced and maintained by forcing through transient eddies and self-interactions?

These questions have been addressed with the ECHAM-3 and ECHAM-4 GCMs developed at the Max-Planck-Institute for Meteorology in Hamburg. The principal objective of this thesis, which is to diagnose the impact of high resolution on blocking in the simulations performed at the Swiss Center for Scientific Computing, is studied by mean flow diagnostics (Section 3), with a blocking index (Section 4), by case studies (Section 5), and by comparison with the blocking frequencies from lower resolution experiments (Section 6). Additional focal points became important during the course of the work, namely:

- processes during the onset phase,
- the role of orography for blocking incidence, and
- interannual and interdecadal variations of the blocking frequency.

Possible onset mechanisms are discussed in view of the selected cases (Section 5.6). The role of the large scale orography is studied by reducing, respectively completely removing the topography in perpetual January simulations (Section 7). Finally, a multi-decadal experiment is briefly examined focusing on the North Atlantic oscillation and blocking incidence (Section 8).
2 Model and data

2.1 The ECHAM models

The main data source of this study consists of output from the atmospheric general circulation models ECHAM-3 and ECHAM-4. These powerful tools were developed at the Max-Planck-Institute for Meteorology, Hamburg, and are widely used for climate simulations either with prescribed sea surface temperatures or coupled to an ocean model. Although they belong to the same family of spectral models, the two versions differ in many aspects. They also show distinct differences, e.g., in cloud radiative forcing from their (grand-) father, the ECMWF weather prediction model. A summary of the main aspects of the ECHAM-3 and the ECHAM-4 code is given in Table 2.1. The models are fully described in two technical reports (DKRZ 1992; Roeckner et al. 1996).

Apart from the overall balance of the model, a few parts are more directly relevant to the blocking performance. Namely, the representation of the resolved and unresolved orography is important as it shapes the planetary scale flow which may be more or less favourable for blocking. ECHAM uses mean elevations, calculated from a high resolution U.S. Navy dataset which are spectrally fitted to the respective horizontal resolution. The model topography displays 'Gibbs ripples' close to steep mountains (e.g., Holzer 1996 for a discussion on how to reduce these oscillations). The effect of gravity waves excited by subgrid scale orography is parametrized on the basis of a saturation hypothesis to calculate the vertical stress profile. The scheme which is the same in the ECHAM-4 as in the ECHAM-3 model is described in the appendix.

Another parameterization that may affect blocking is the horizontal diffusion. The purpose of this routine is to prevent the accumulation of energy at the truncation wavenumber. The scheme, however, also influences the kinetic energy spectrum at longer wavelengths where energy is also transferred from the smaller eddies to the larger ones. The two formulations, a higher order linear scheme in the ECHAM-4 model and another scale selective method for ECHAM-3 are described in the appendix. The horizontal diffusion constants are tuned in both cases with the kinetic energy spectrum of the ECMWF analyses. The matter is discussed further in Koshyk and Boer (1995).
Table 2.1: The ECHAM models (* as in ECHAM-3)

<table>
<thead>
<tr>
<th></th>
<th>ECHAM-3</th>
<th>ECHAM-4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Prognostic variables</td>
<td>Vorticity, divergence, temperature, log surface pressure, water vapour, cloud water</td>
<td>* plus turbulent kinetic energy and chemical tracers</td>
</tr>
<tr>
<td>Horizontal advection</td>
<td>Spectral transform with triangular truncation at wave number 42 or 106. Non-linear and diabatic terms are calculated on a Gaussian grid (2.8° or 1.1° lat/lon)</td>
<td>* but semi-Lagrangian advection of water vapour and cloud water</td>
</tr>
<tr>
<td>Vertical representation</td>
<td>Hybrid coordinate system (σ → p with increasing height) 19 levels. Top at 10 hPa</td>
<td>*</td>
</tr>
<tr>
<td>Time integration</td>
<td>Semi-implicit. Leap-frog with time filter. Timestep: 24 min (T42), 12 min (T106)</td>
<td>*</td>
</tr>
<tr>
<td>Orography</td>
<td>Mean</td>
<td>*</td>
</tr>
<tr>
<td>Radiation</td>
<td>Two stream approximation. 4 solar and 6 terrestrial intervals. Absorbers are CO₂, O₃, aerosols (all prescribed) and water vapour</td>
<td>2 solar intervals. Additional absorbers (CH₄, N₂O, CFCs). Modifications in water vapour continuum absorption. Cloud optical properties dependent on effective drop size</td>
</tr>
<tr>
<td>Surface albedo</td>
<td>Climatological, except over water in all forms</td>
<td>New background albedo</td>
</tr>
<tr>
<td>Land surface processes</td>
<td>5-level heat diffusion into the ground, one layer infiltration/runoff scheme</td>
<td>* but bucket spatially variable</td>
</tr>
<tr>
<td>Surface fluxes</td>
<td>Monin-Obukhov similarity theory with exchange coefficients dependent on roughness length and &quot;moist&quot; Richardson number. Corrections for weak winds</td>
<td>*</td>
</tr>
<tr>
<td>Vertical diffusion</td>
<td>Eddy viscosity dependent on mixing length</td>
<td>Higher order closure. Diffusion coefficients dependent on turbulent kinetic energy</td>
</tr>
<tr>
<td>Horizontal diffusion</td>
<td>Scale dependent</td>
<td>Linear higher-order scheme</td>
</tr>
<tr>
<td>Mountain drag</td>
<td>Gravity wave drag parameterization</td>
<td>*</td>
</tr>
<tr>
<td>Cumulus convection</td>
<td>Mass flux scheme</td>
<td>* with modified closure based on buoyancy and inclusion of entrainment</td>
</tr>
<tr>
<td>Clouds and precipitation</td>
<td>Prognostic equation for cloud water</td>
<td>* additionally, detrained convective cloud water is included in the stratiform cloud scheme</td>
</tr>
</tbody>
</table>
2.2 Model experiments

Data gained from larger projects as well as from experiments specifically performed for the present study are used. The primary source for the first sections are two high resolution experiments with climatologically prescribed sea ice and sea surface temperatures (SSTs) covering five and ten years, respectively, which were performed at the Swiss Center for Scientific Computing (SCSC), Manno, with ECHAM-3 and ECHAM-4. The collaboration with the Max-Planck-Institute makes available a whole matrix of further experiments for comparison. Table 2.2 lists the experiments and names them according to the SST (C for climatological mean, S for scenario, A for AMIP, and G for GISST), the resolution (H stands for high and M for medium resolution), and the ECHAM model version.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>SST</th>
<th>resolution</th>
<th>period</th>
<th>sampling</th>
<th>section</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH3</td>
<td>clim. mean 1979–1988</td>
<td>T106</td>
<td>5 y</td>
<td>12 h</td>
<td>3–5,9</td>
</tr>
<tr>
<td>CH4</td>
<td>&quot;</td>
<td>T106</td>
<td>10 y</td>
<td>6 h</td>
<td>3–5,9</td>
</tr>
<tr>
<td>CM3</td>
<td>&quot;</td>
<td>T42</td>
<td>30 y</td>
<td>12 h</td>
<td>6,9</td>
</tr>
<tr>
<td>CM4</td>
<td>&quot;</td>
<td>T42</td>
<td>10 y</td>
<td>12 h</td>
<td>6,9</td>
</tr>
<tr>
<td>SH3</td>
<td>ECHAM1/LSG, 2045–2055</td>
<td>T106</td>
<td>5 y</td>
<td>12 h</td>
<td>4,9</td>
</tr>
<tr>
<td>AM4</td>
<td>AMIP</td>
<td>T42</td>
<td>1979–1994</td>
<td>12 h</td>
<td>6,9</td>
</tr>
<tr>
<td>AH4</td>
<td>&quot;</td>
<td>T106</td>
<td>1979–1988</td>
<td>6 h</td>
<td>4,6,9</td>
</tr>
<tr>
<td>CM</td>
<td>perpetual January</td>
<td>T42</td>
<td>20 mon</td>
<td>12 h</td>
<td>7,9</td>
</tr>
<tr>
<td>NGWD</td>
<td>no GWD</td>
<td>T42</td>
<td>5 mon</td>
<td>12 h</td>
<td>7,9</td>
</tr>
<tr>
<td>RM</td>
<td>reduced mountains</td>
<td>T42</td>
<td>20 mon</td>
<td>12 h</td>
<td>7,9</td>
</tr>
<tr>
<td>NM</td>
<td>no mountains</td>
<td>T42</td>
<td>20 mon</td>
<td>12 h</td>
<td>7,9</td>
</tr>
<tr>
<td>GM4</td>
<td>GISST 1903–1995</td>
<td>T42</td>
<td>1903–1995</td>
<td>12 h</td>
<td>8,9</td>
</tr>
</tbody>
</table>

The following numbers give an idea on the computational and archival requirements. One month of an ECHAM-4/T106 simulation requires ~ 8 CPU hours on one NEC/SX-3 processor and deposits 1.5 Giga bytes of GRIB compressed raw data in the storage archive. The data for the whole ten year experiment archived with an output interval of 6 hours amounts to 0.2 Tera bytes. For T42, the memory requirement is ~ 6 times smaller while the CPU time on the above mentioned vector machine is one tenth of the high resolution simulation. The doubled time step at T42 is not fully reflected by the computational time requirements as the T106 model vectorizes better.

Together with the MPI model, the ‘afterburner’ was installed at SCSC. This post-processing program reads the GRIB compressed raw data, does certain conversion, selection, averaging and/or interpolation tasks according to the specified namelist and writes the output in binary format. The data flow is depicted very schematically in Fig. 2.1. On the one hand, data is selected from the huge raw output but also expanded to be in a
format that does not rely on GRIB conversion software. For studying time sequences of three dimensional fields the above ad hoc data handling heavily uses the links between the computing centers in Manno or Hamburg and the working place where the graphics is done. A file containing, for example, gridded 6-hourly T106 geopotential fields for one month at one level is 24 Mega bytes large (which is four times as much as the amount when using spectral coefficients).

![Data flow scheme](image)

**Figure 2.1: Data flow scheme.**

## 2.3 Observational data

Model validation in general is of primary importance for the advancement of climate modelling. With more and more sophisticated models, however, the demands for global observational data are very high and the comparison of simulations with the 'observed truth' becomes subtle. For the large scale flow fields, weather services provide analyses of good quality.

Monthly mean global ECMWF analyses of the period 1981–1990, kindly provided by Dr. Klaus Arpe from MPI, are used in the following section for the validation of some large scale flow fields. This data is of T42 horizontal resolution. At the time of writing, the ECMWF reanalyses, performed with a frozen forecast model at T106 resolution, are about to be finished, and are assumed to be of higher quality.

Daily 500 hPa height fields of the period 1950–1988 derived from the U.S. National Meteorological Center (NMC) analyses of the northern hemisphere are used in Section 4 to determine the observed climatology of blocking. This data source was obtained from Dr. Michael Christoph on a regular $5^\circ \times 5^\circ$ latitude/longitude grid.
3 Mean flow of the high resolution experiments

Before discussing methods and results of the blocking statistics it is useful to know the model climate and some of its systematic errors. The basic validation of the ECHAM models is presented in Roeckner et al. (1992; 1996) for the standard T42 version. Selected aspects of the T106 high-resolution experiments have been investigated before, namely the hydrological cycle (Arpe et al. 1994), the radiative fluxes (Wild et al. 1995), the frequency of hurricanes (Bengtsson et al. 1995) and the mass balance of ice-sheets (Ohmura et al. 1996). Here, the T106 simulations with climatological SSTs are compared among each other and with the ECMWF climatology of the ten year period 1981-90. Shown are seasonal means over 5/10 years for ECHAM-3/4.

The work of May and Bengtsson (1996) on the ENSO effects upon the midlatitude circulation should be kept in mind when comparing simulations with a fixed yearly SST cycle with the analyses of the eighties that saw strong ENSO events. The use of inhomogeneous analyses for the following evaluation is considered as a minor weakness since only primary variables such as temperature or winds are compared.

3.1 Zonal means

The zonal mean temperature errors of the two model versions with respect to the ECMWF analyses are shown in Fig. 3.1 for the winter and summer seasons. In neither simulation are the lower tropospheric differences much larger than 2 K, except over the Arctic summer in ECHAM-4 where a warm bias of about 5 K is present around 850 hPa over the north pole. Larger differences occur at higher levels where the air is less dense. Both the ECHAM-3 and the ECHAM-4 simulation are too cold in the polar lowermost stratosphere. This cold bias is somewhat stronger during the respective summer season but prevails throughout the year and is also found in other experiments with lower resolution and with different models (Boer et al. 1991). A possible cause for this deficiency is the insufficient vertical resolution of the stratosphere.

The most significant change between the two models is a substantial cooling of the upper troposphere and lower stratosphere in ECHAM-4 compared to the older model. Annual mean temperatures at the 150 hPa level are up to 10 K cooler in the new model than they are in ECHAM-3 (Fig. 3.2). This cooling almost completely removes the warm bias in the upper tropical troposphere which is characteristic of the ECHAM-3 model but enhances the above mentioned cold bias over the poles. A second important difference in the temperature structure of the two model versions is the warmer north polar troposphere in ECHAM-4 compared to ECHAM-3.

Candidates from the numerous model changes to explain this cooling aloft as well as the lower level polar warming are the radiation scheme and the parameterization of cloud
properties. The detailed work of Wild et al. (1996) has shown that the ECHAM-4 atmosphere is less transparent to shortwave radiation than the ECHAM-3 atmosphere. The deficit at the surface is nearly compensated by enhanced downwelling longwave radiation. Furthermore, the longwave cooling in the upper troposphere is underestimated by the ECHAM-3 radiation scheme (Roeckner et al. 1992). These facts partly explain the large cooling aloft. Enhanced shortwave absorption and multiple reflection, on the other hand, may be the cause for the Arctic warming.

To illustrate that the many modifications introduced into ECHAM-4 have led to major changes in the model climatology, cloud cover is also shown (Fig. 3.3). Although the vertical distribution of the cloud fraction is not a typical validation quantity since detailed observations are hard to obtain, it helps to understand what is going on in the model. The cloud fraction is higher in ECHAM-4 than in ECHAM-3 almost everywhere. Particularly near the tropopause, clouds are 10% less abundant in ECHAM-3 than in ECHAM-4 which simulates a sharper defined (and colder) tropopause.

Of more relevance to blocking is the zonal wind which is discussed next. Concentrating first on the northern hemisphere, Fig. 3.4 shows a well closed winter jet which, however, is too
Figure 3.2: Latitude-height distribution of the annual mean temperature difference between the CH4 and the CH3 simulation. The contour interval is 2 K. Contours at ±1 K are also drawn.

Figure 3.3: Zonal mean cloud fraction of the CH3 (top) and the CH4 simulation (bottom) for DJF (left) and JJA (right) respectively. The contour interval is 5%.

The simulated zonal wind in the southern hemisphere is less satisfactory. During the summer (DJF), the simulated jet position is almost 10° too far south and the surface
winds are too strong. In winter (JJA), the subtropical jet is $\sim 10 \text{ m s}^{-1}$ stronger in the models than in the analyses. Furthermore, the CH3 experiment shows an unrealistic splitting of the jet structure which is improved in ECHAM-4. Generally, the changes from ECHAM-3 to ECHAM-4 in the u-wind are small, despite the large changes in the vertical temperature distribution. A beneficial equatorward shift together with a slightly detrimental acceleration of the northern hemisphere jet in winter may be noted.

### 3.2 Northern hemisphere winter climate

Maps of mean sea level pressure give a summary view of the simulated climate. Averaged only in time, this field is less robust than the zonal means shown in the previous section. Still, a few systematic errors become apparent in Fig. 3.5 which compares the simulated pressure with the ECMWF analyses. The most serious discrepancy of the models is a positive bias of 5–10 hPa over southern Europe. Also, the Asian anticyclone and the subtropical high over the western U.S. are too strong. The Aleutian low is successfully simulated in strength but somewhat too small in the ECHAM-3 simulation. In the North Atlantic, the models capture the low pressure near Iceland. ECHAM-3 more than ECHAM-4 simulates a too deep Icelandic low. Together with the positive bias over Spain, this results in excessive westerly flow across the Atlantic and into northwestern Europe for both models.
The mean flow in the middle troposphere is contoured by the 500 hPa height in Fig. 3.6. Both simulations with fixed SSTs capture the quasi-stationary planetary waves of the northern hemisphere winter with considerable skill. The troughs over the east coasts of Asia and North America as well as the ridges over the west coasts of North America and Europe are well positioned. The ridging in the latter two regions, however, is too weak. The ECHAM-3 model in particular has problems to simulate the broad east Asian trough and misses the northward bending over the northeast Pacific almost completely resulting in large systematic errors on the order of 150 gpm. The simulation in this region not only profits from the inclusion of SST variability due to ENSO (not shown) but also from a more careful parameterization of the radiative properties of clouds in ECHAM-4, which displays much reduced systematical errors in the Pacific sector. The connection between cloud microphysics and circulation is put forward in Kiehl (1994), where the reduction of the effective drop radius over land induced a stronger ridge off the northwest American
coast in the CCM2. In the ECHAM-4 model, the drop size is controlled by specifying different number concentrations for maritime and continental clouds, which also gives smaller drops over land.

Over the Atlantic, too strong zonal flow is apparent in both simulations. The exaggerated subtropical high over Spain noted already on the sea level pressure map is also apparent at 500 hPa. While the ECHAM-3 has maximal errors near the Greenwich meridian, directly above the pressure bias, the error pattern in the ECHAM-4 simulation is shifted to the west with a maximal meridional error gradient at 45°W. The observed zonalization is a common problem of high resolution numerical weather and climate simulations (Arpe and Klinker 1986) that is still present despite the parameterization of the drag due to subgrid scale orography. To increase the momentum sink by mountains further one can introduce an 'envelope orography.' The global addition of \((2 \text{ subgrid scale oro. var})^{1/2}\) to the mean terrain heights, however, leads to elevated plateaux that represent unrealistic heat sources or may accumulate snow so that this approach is not adopted for ECHAM. The zonalization may also be due to erroneous momentum transports (too strong horizontal tilt of waves). It is known from the T42 model that the momentum convergence of the transient eddies is too strong over central Europe (Roeckner et al. 1996).
Figure 3.6: 500 hPa geopotential height for the boreal winter; ECMWF analyses (top), CH3 (left) and CH4 (right) simulation together with the systematic errors (bottom). The contour interval is 80 gpm for the plain fields and 20 gpm for the systematic errors.
The strength of the polar and subtropical jet streams are shown by the zonal wind component at 200 hPa in Fig. 3.7. The zonal wind maximum to the east of the Asian coast is shifted a few degrees to the north in the CH3 simulation compared to the ECMWF analyses of the boreal winters 1981–1990. The Asian jet in the ECHAM-4 experiment is perfectly positioned, and the maximum over northeastern Africa is also better captured in the CH4 than in the CH3 simulation. The maximal zonal flow speed of the CH4 experiment (77 m s\(^{-1}\)) is slightly stronger than the value of the analyses (72 m s\(^{-1}\)), which may be attributed to ENSO events influencing the Pacific jet in the analyses.

![Figure 3.7: 200 hPa zonal wind during boreal winter for ECMWF analyses (top center), CH3 (bottom left) and CH4 (right) simulation. The contour interval is 5 m s\(^{-1}\).](image)

The east American jet, on the other hand, is stronger in the CH3 than in the CH4 simulation, which is closer to the analyses. The most serious discrepancy between the simulations and the analysed zonal winds is the lack of sufficient deceleration over the Atlantic. While the observed 30 m s\(^{-1}\) isotach ends in the middle of the Atlantic, the simulated lines extend towards northwestern Europe and reach Scotland. This common systematic error of both
model versions is somewhat less pronounced at lower resolution. A possible explanation is the tendency of the T106 simulations to generate deep lows which occlude too slowly. In any case, the above mentioned positive bias of the convergence of westerly momentum in this region accelerates the flow.

3.3 Seasonal aspects

The large continental land masses of the northern hemisphere lead to a strong seasonality of the general circulation. Fig. 3.8 displays the 200 hPa zonal wind of the CH4 simulation for the summer and the two equinoctial seasons.

Figure 3.8: 200 hPa zonal wind for JJA (top left), MAM (bottom left) and SON (bottom right) of the CH4 simulation. Unlike the other maps, which are bounded by 20°N, the summer winds are displayed for the entire hemisphere together with the observations (top right). The contour interval is 5 m s⁻¹. Easterly winds are dashed.
In summer, the east Asian jet is weak. The zonal wind maxima are located over Asia near 90°E and over North America. The simulated zonal wind over North America reaches more than 30 m s\(^{-1}\) while it is only 20 m s\(^{-1}\) in the analyses. The entire hemisphere, starting at the equator, is plotted for the summer winds to display also the tropical easterlies. The high resolution simulation captures the maximum over India with remarkable skill.

During spring and autumn, the east Asian jet is of equal strength in both seasons. The position of the core is at least 5° more to the north in autumn than in spring. A second maximum over North Africa is very pronounced in spring. The zonal winds over eastern America are comparatively weak. Interestingly, there are midlatitude regions where the zonal wind remains on the order of 20 m s\(^{-1}\) throughout the year. One such region is central and northeastern Europe. The zonal wind for JJA and DJF (Figs. 3.7 and 3.8) is also fairly constant between 15 and 20 m s\(^{-1}\) for the analyses in this region. The length scale of a stationary Rossby wave can be estimated from this wind speed by inserting the half-wavelength \(L = \pi/K\) in Eq. 1.2:

\[
L_s = \pi \left(\frac{U}{\beta}\right)^{1/2} \sim 4000 \text{ km}.
\] (3.1)

A flow speed of 18 m s\(^{-1}\), and \(\phi = 60^\circ\) in \(\beta = 2\Omega \cos \phi / a\) have been inserted to obtain the order of magnitude estimate of Eq. 3.1. The stationary flow features that may occur in the above mentioned quasi-constant wind regions are thus expected to extend over about 4000 km, or 72° longitude at 60°N.

The differences of July minus January temperatures at 500 hPa in Fig. 3.9 give an impression of the annual cycle in different regions. The largest summer/winter temperature contrasts in the mid-troposphere occur over the eastern parts of the northern hemispheric continents, especially over eastern Siberia around 50°N, while they are roughly half of that over the eastern parts of the Pacific and Atlantic oceans. Both models capture the gross features of the temperature contrasts but with an eastern Siberian maximum which is too far north and with a problem of the tilting in the two blocking sectors. The southwest-northeast tilting seems too strong in the models. This would imply a too strong poleward momentum transport in the northeast Atlantic and an acceleration of the westerly flow, consistent with the zonalization already noted for winter above.

As a preliminary for the next section, the annual cycle of a zonal flow index is discussed. The index is calculated as the difference of the geopotential at 40°N and 60°N from monthly mean climatologies of the 500 hPa height and measures the mean strength of the geostrophic zonal flow in a fixed band of the middle latitudes. Fig. 3.10 displays this gradient for the entire hemisphere as well as for four sectors separately. The sectorial decomposition is given in the figure caption. The huge Asian continent is represented by a 120° sector and North America by a 60° sector. The blocking regions over the Atlantic and the Pacific are each 90° wide.
The two lines in Fig. 3.10 display the observed annual cycle calculated for the period 1950 – 1979 and for the ten AMIP years 1979 – 1988 from the NMC analyses. After a strong increase of the hemispheric index in August and September, maximal zonal flow prevails during October and November followed by a rather steady decline until the minimum is reached in June. The monotoneous decrease is not a robust feature of the northern hemispheric index as can be seen from the secondary minimum in February for the ten year period 1979–1988. February shows by far the largest difference between the two observed sample periods which were chosen to match the Lejenäss and Økland (1983) paper and the averaging period of the experiments. The ‘interdecadal variability’ in the individual sectors is of course larger than that of the hemispheric index. In particular, the variance is large during the cold seasons in the Atlantic and Pacific sectors. The index values of the period 1979 to 1988, however, stay within the interannual variability, the shaded band in Fig. 3.10, which is determined by the standard deviation of the thirty monthly means of the period 1950 to 1979. A characteristic of the Pacific sector, which is especially marked in the eighties, is the midwinter suppression of the zonal flow in the specified latitude band. This weakening is caused on the one hand by the southward shift of the jet stream and is reinforced on the other hand by the midwinter peak of the North Pacific blocking activity. Nakamura (1992) has shown that the baroclinic wave activity over the Pacific is suppressed at the jet stream level during midwinter when the Asian jet is strongest. The suppression and a secondary maximum in April is also apparent albeit much weaker in the Asian sector.
Figure 3.10: Annual march of the zonal flow index $Z_{40^\circ N} - Z_{60^\circ N}$, repeated twice for better survey, for NMC data of 1950–1979 (thick line), for the period 1979–1988 (thin solid), CH3 (dash dotted) and CH4 simulation (dashed). The shaded band indicates the interannual variability ($\pm 1\sigma$) of the observed index of 1950–1979. From top to bottom: average over the northern hemisphere, Atlantic sector $60^\circ W - 30^\circ E$, Pacific sector $150^\circ E - 120^\circ W$, Asian sector $30^\circ E - 150^\circ E$ and North American sector $120^\circ W - 60^\circ W$. 
The broken lines in Fig. 3.10 display the simulated annual cycle of the zonal index for the two high resolution simulations CH3 and CH4 respectively. The hemispheric index is generally too high during the whole year which is another manifestation of the zonalization problem of the T106 simulations. With a yearly mean overestimation of 27 gpm, ECHAM-4 does a considerably better job than ECHAM-3 which overshoots by 51 gpm. Much larger discrepancies occur in the two blocking sectors. Both models have a large bias to excessive zonal flow lying well outside the observed interannual variability during winter and spring in the Atlantic sector. From

\[ u_g = -\frac{g}{f} \frac{\partial Z}{\partial \phi}, \]  

(3.2)

a geopotential height difference of 100 gpm over 20° latitude centered at 50°N corresponds to 4.1 m s\(^{-1}\). Thus the geostrophic zonal wind error during midwinter over, say, the British Isles is on the order of 8 m s\(^{-1}\) near the equivalent barotropic level. Over the North Pacific, ECHAM-4 captures the midwinter suppression with remarkable skill but the zonalization is again present during October and November. In ECHAM-3, the index is too high during the whole year. ECHAM-4 is also superior to ECHAM-3 over the Asian continent where both models simulate a realistic seasonal march which is characterized by a sharp acceleration around September. Finally, the cycle in the North American sector is once more distinct in displaying a gentle increase from May to December followed by a steeper decrease in early spring. This deceleration is retarded in both simulations by one month.

The potential use of the above diagnostic for assessing the blocking performance of a model simulation from monthly means alone will be taken up in later sections.
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4 Climatology of blocking frequencies

Blocked flow is defined by the index of Lejenås and Økland (1983). A discussion of this method is given and the blocking activities from the high resolution experiments listed in Table 2.2 are compared among each other and with analysis data. Although the T106 experiments provide only short samples to make up blocking climatologies, the blocking frequencies show whether the simulated blocks, if at all, occurred in the right sectors or if the model has preferred blocking locations differing from the observed regions.

4.1 Blocking index

As there is no generally accepted theory of blocking, the phenomenon may be defined in a number of ways. Traditionally, weather charts have been examined for characteristic features to build a catalogue of blocking events that allowed the extraction of statistical results (Rex 1950; Treidl et al. 1981). To circumvent the subjectivity and the laborious task of the visual inspection of a large number of charts, so-called objective methods have been devised that are suitable to scan through digitized observational or synthetical data. These objective methods can be roughly divided in two classes. Either the algorithm searches for persistent anomalies of the 500 hPa geopotential height or some blocking flow feature is translated into a criterion for the gridded data field. The latter approach is adopted by Lejenås and Økland (1983), Tibaldi and Molteni (1990), and also by Kaas and Branstator (1993), while persistent anomalies are used by Dole and Gordon (1983), and many others. Specifically, Sausen et al. (1995) and May (1994) have studied the blocking performance of the ECHAM-3 model in terms of persistent anomalies with different thresholds. The major drawback of the persistent anomaly methods is that they select situations that are not always found to be blocked by synoptical standards. One such erroneous pattern consists of a broad ridge originating from a subtropical high that is shifted to the north of its usual position. To complement the above mentioned studies, blocking is defined as an anomalous flow pattern in this thesis.

Before entering into more detail, some alternative attempts to define blocking are mentioned. Schilling (1986) formed a blocking number with the zonal mean eddy kinetic energy and other 'circumpolar energetic indices' and tried to classify blocking cases. The locality and individuality of blocking events made firm conclusions difficult. Another non-local approach is taken in studies on regime-like behaviour of the hemispheric flow (e.g., chapter 6 of Ghil 1987). Blocking is then associated with either a meta-stable equilibrium state that is close to resonance with the topography (Charney and DeVore 1979) or with the high amplitude mode of the wavenumber 2–4 planetary waves (Hansen et al. 1993). Another interesting definition uses the projection on a precalculated anomaly pattern as a blocking index (Liu 1994). Finally, blocks may also be characterized by the different slope of the function $q(\psi)$, where $q$ is the potential vorticity and $\psi$ the streamfunction, inside and away from the coherent structure (Butchart et al. 1989).
I have chosen to apply the simple method of Lejenäs and Økland (1983, hereafter LO) as it is simple, efficient, and has been tested for the northern hemisphere with a long record of observations. Based on the classical characteristics of Rex (1950) and synoptic experience, LO proposed to use the criterion

\[ \tilde{I}(\lambda) < 0, \quad I(\lambda) = Z(\lambda, 40^\circ N) - Z(\lambda, 60^\circ N), \]  

(4.1)

where \( \lambda \) is longitude, \( Z \) the 500 hPa geopotential height and the tilde denotes an average over 30 degrees longitude. A longitude at a particular day is thus blocked in the sense of LO if the geostrophic flow in the latitude band 40 to 60°N is easterly. The longitudinal averaging excludes small scale features from being counted as blocks. Still, large cutoff lows near 40°N may be counted as blocks. Tibaldi et al. (1990, 1995) have added a second gradient criterion to the north and modified the index further in order to remedy this drawback. As cutoff lows have much in common with blocking highs, I prefer to use the original LO index which is less restrictive and involves a minimum of parameters.

Figure 4.1: Hovmöller diagram of NMC analyses from 1983. Days fulfilling the Lejenäs and Økland criterion are marked black. Large clusters are blocking candidates.
A convenient way to identify blocking events from a long sequence of daily 500 hPa data is to plot the blocked days in a longitude-time diagram. As an example, the days of the year 1983 that fulfill the above criterion of a change in sign of the meridional geopotential gradient are displayed in Fig. 4.1. Potential blocking events show up as extended clusters in the Hovmöller diagram. Two major blocking episodes may be noted in the Atlantic sector for 1983. The strong blocking case in February has been extensively studied by Shutts (1986) and others. The somewhat weaker event in November has influenced the predictive skill of the ECMWF forecasts as described by Simmons (1986). In the Pacific sector, a few less persistent candidates occurred during the first four months. A stronger retrogressing case started in December.

The LO criterion itself does not contain a persistency requirement. Usually, an event is considered as a true blocking case if the pattern persists for at least 5 days. Five days is at the low end of the range encountered in the literature of minimal duration for an event. Rex's originally proposed ten days are often found too long to yield a useful blocking catalogue. For comparison with the LO paper, no duration check is done in the calculation of the observed and simulated blocking frequencies in this section. The long lasting cases contribute dominantly to the activity anyway.

Fig. 4.2 shows blocking frequencies from the thirty Januaries 1950–1979 as a function of longitude. The full line is for the original criterion while the broken lines display the effect of a latitudinal shift of the criterion by ±5 degrees. The blocking activity over large parts of Asia and North America is vanishingly small so that the Atlantic-European peak is well separated from the Pacific center. A local minimum is found at 30°W over the Atlantic which separates Greenland blocking from cases that develop further to the east.

![Figure 4.2: Blocking frequencies according to the Lejenäs and Økland index for the 30 Januaries of the period 1950–1979 (full line). Index shifted five degrees to the south (dashed) and to the north (dash-dotted).](image)

All days that fulfill the LO criterion at the longitudes 50°W and 15°E are collected in composites in Fig. 4.3. The mean 500 hPa flow patterns display strong blocking ridges
tilting backwards for the Greenland region. The pattern over northwestern Europe consists of a ridge which is tilted from the southwest to northeast and an accompanying trough over southeastern Europe.

Figure 4.3: Composite 500 hPa chart made up from all January days of the NMC data with $Z_{40^\circ N} - Z_{60^\circ N} < 0$ at $50^\circ W$ (left) and $15^\circ E$ (right).

When shifted northwards five degrees, the criterion indicates much higher blocking frequencies in the Pacific sector as well as over Greenland where peak values are twice as large as for the standard index. Furthermore, the separation of the Atlantic-European sector into two peaks is more pronounced for the latitude band $45^\circ - 65^\circ N$ than for the original latitudes.

Fig. 4.4 displays the composite patterns for days that are blocked at $180^\circ E$ for the original criterion and for the northward shifted index. The pattern, a strong backward tilted ridge over Alaska, is nearly the same for both latitude bands. With 35% of all the January days, the blocking ridge over Bering street is quite frequent a flow pattern over the North Pacific.

4.2 Longitudinal distribution of the four seasons

The LO blocking activity for the four high resolution experiments listed in Table 2.2 is discussed in this section. The blocking frequencies are calculated from twice-daily geopotential data by extracting the gradient between the 45nd and the 27nd Gaussian latitude, averaged over $29.25^\circ$ longitude. The two Gaussian latitudes correspond to $39.81$ and $59.99^\circ N$, respectively. This procedure comes close to the original LO blocking definition. The results, displayed as seasonal percentages of blocked half-days, are given in Fig. 4.5 for the ECHAM-3 control simulation CH3 and the $2 \times CO_2$ time slice SH3. The latter experiment used the SST anomalies of the ECHAM1/LSG scenario (Cubasch et al. 1992)
at the time of equivalent CO₂ doubling for its boundary conditions. Also shown is the observed blocking activity which has been calculated from the 5×5 degrees gridded NMC data for the period 1950 to 1979 together with a range of the interannual variability. The interannual standard deviation of each season is calculated from the interannual variances of the corresponding three months.

In winter, the observed frequencies are of similar magnitude for the two main blocking regions of the northern hemisphere. The ECHAM-3 model severely underestimates the activity in the Pacific sector. Deficiencies in this sector have already been noted in the time mean flow (e.g., Fig. 3.6). In the Atlantic-European sector the performance of the control simulation CH3 is better. However, almost no blocks occurred to the west of the Greenwich meridian and the peak is at 30°E, which is 20 degrees too far east compared to the observations. The excessive winter flow over the Atlantic prevents the formation of blocks in this region and shifts the preferred location for blocking incidence to the east.

The spring season displays a still strong peak in the Euro-Atlantic sector and a reduced blocking activity over the North Pacific. The model successfully captures the main peak over Europe but has too few blocked days over the Atlantic. In summer, the LO method indicates modest blocking activity in both regions. The simulated frequency in summer is generally smaller than the observed. Blocking frequencies in the Pacific sector are nearly zero during autumn while the activity in the Euro-Atlantic sector already starts to increase slightly. The CH3 simulation captures this restart in the European sector as well as the absence of blocking over the North Pacific in autumn.

Summarizing, the annual cycles described above are captured by the ECHAM-3 model with relatively high skill (compared to other models) but the systematic errors noted
before prevent more realistic blocking frequencies in the Pacific sector and over the North Atlantic ocean.

The decrease of blocking frequencies which may be noted for the 2×CO₂ simulation during the winter half year has to be taken with low confidence as the integrations extend over five years only. A general decrease of the blocking frequency under doubled carbon dioxide concentrations is also reported by the IPCC (Kattenberg et al. 1995). An explanation for the hypothetical decrease of the blocking activity over the Atlantic in winter may involve the thermohaline circulation, which will likely become weaker in the coming decades. In any case, the projected temperature changes in the North American-Atlantic sector are such that the west-east temperature gradient in winter decreases (the continent experiences a stronger warming than the North Atlantic, where no warming occurs). This leads to reduced southerlies, which decreases the probability of blocking incidence.

Figure 4.6 displays the blocking frequency for the ECHAM-4 climatological and the AMIP SST experiment. Here, AMIP means that the actual march of the SST of 1979 to 1988, the standard period for the atmospheric model intercomparison project, is used as boundary forcing. The comparison of the CH4 and the AH4 experiment gives an indication of the effect of SST variability on blocking. Observational NMC analyses of the years 1979 to
Figure 4.6: Seasonal blocking frequencies according to the Lejenäs and Økland index for the period 1979–1988 (full line), the CH4 simulation (dash-dotted) and the AH4 simulation (dotted).

1988 are used as a reference in Fig. 4.6 to make the comparison as fair as possible.

Generally, ECHAM-4 does a better job in the Pacific sector than ECHAM-3. The winter blocking activity is still underestimated in the CH4 simulation but reaches the observed level in the eastern Pacific when interannual SST variability is included. Blocking days in the other seasons are less frequent than 5% which is successfully simulated by both the CH4 and the AH4 experiment.

In the Euro-Atlantic sector, the CH4 simulation severely underestimates the blocking frequency in winter. The systematic errors in the time averaged fields over the North Atlantic prevent a realistic level of the blocking activity in this sector. Interestingly, the simulation recovers from this deficiency in spring when the model is able to capture the main peak near 20°E for both experimental setups. A few blocking cases in the AH4 simulation lead to an overestimation during autumn.

To comment on the effect of SST variability on the blocking activity is difficult for at least two reasons. First, the high resolution atmospheric simulations cover only two cycles of the primary atmosphere-ocean oscillation in the equatorial Pacific, and secondly, SST anomalies in other regions may also have an influence. Still, the inclusion of SST variability seems to be beneficial to the blocking performance. A similar comparison for the T42 model will be given in Section 6.
The blocking frequencies of the CH3 and CH4 simulations shown in Fig. 4.5 and Fig. 4.6 may be compared with the mean zonal index of Fig. 3.10. Over the North Pacific, the overestimation of the mean index is consistent with the poor blocking performance of the CH3 simulation. The zonalization of the circulation over the Atlantic in winter (Fig. 3.10) suppresses the blocking frequencies over the Atlantic for both simulations. CH3 seems to be better than CH4 over Europe, but a decomposition of the 10 years of the CH4 experiment in two samples shows (Fig. 4.7) that almost as many blocking days occurred in the first five winters of the CH4 simulation as in the five winters of the CH3 simulation.

Figure 4.7: Seasonal blocking frequencies for the first five years of the CH4 simulation (full line), and the last five years (dashed). Blocking days are also drawn for the CH3 simulation (dash-dotted), and the NMC analyses of the period 1979–1988 (dotted).

4.3 Blocking frequencies calculated from anomalies

The LO blocking index uses absolute fields and is sensitive to changes in the mean climate. Systematic model biases can result in large discrepancies of the blocking frequencies as shown in Figs. 4.5 and 4.6. To overcome these limitations, anomalies may be calculated by subtracting the climatology from the daily fields. By doing so for the model data and the analyses, the zonalization of the simulations is subtracted, and the blocking frequencies can be defined properly. This is undertaken for the winter months of the CH4, CH3, and SH3 simulations in this subsection. Anomalies are calculated according to

\[ Z_a = (Z - \overline{Z}) \frac{\sin 45^\circ}{\sin \phi}, \]

where \( Z \) is the instantaneous 500 hPa geopotential, and \( \overline{Z} \) the corresponding long term monthly mean of the simulation or the analyses. The scaling by the latitudinal factor is included to obtain streamfunction-like anomalies.
A blocking criterion can be formulated based on the same latitude band as the LO index:

\[ Z_a(\lambda, 60^\circ N) - Z_a(\lambda, 40^\circ N) > M \]  

(4.3)

Different values of the threshold parameter \( M \) have been tested. Fig. 4.8 displays the fraction of blocked winter days of the CH4 experiment for \( M = 150 \) gpm and \( M = 250 \) gpm. The percentages from the NMC analyses of the thirty winters 1949/50–1978/79 and the ten AMIP winters are drawn for the same thresholds in Fig. 4.8. While \( M = 150 \) gpm results in a background activity of about 10%, \( M = 250 \) gpm seems to be an appropriate threshold for winter anomalies. The analyses show a maximum in the Atlantic sector that is shifted about 45° to the west in the anomaly method compared to the plain LO method. The maximum in the Pacific sector is located further east compared to the LO peak. The simulated frequencies are underestimated over the Atlantic but exceed the values of the analyses further eastwards over Eurasia. Much smaller differences between the model and the analyses are also seen in the Pacific sector for the anomaly gradient method than for the original LO index. But still, blocking days around the dateline are underestimated by the model.

![Figure 4.8](image)

Figure 4.8: Fraction of DJF days where the anomaly at 60°N is more than 150gpm (left), respectively 250gpm (right) higher than at 40°N. Anomalies from ten winters of the CH4 simulation (dashed), NMC analyses of the period 1979/80–1988/89 (thick solid), and the thirty winters 1949/50–1978/79 (thin solid) are included.

Figure 4.9 shows the blocking days according to Eq. 4.3 for the CH3 and the SH3 experiments together with the winter climatology of the period 1950 to 1979. The lack of blocking days over the Atlantic apparent in Fig. 4.5 for the LO method is mitigated to an underestimation by about one quarter when using anomalies. The same happens to the blocking activity in the Pacific sector. The simulated frequencies of the CH3 experiment have a second maximum over Eurasia, where the analyses show a continuous decline. The zonalization of the high resolution flow seems to produce this eastern maximum in the CH3 and CH4 simulations. The scenario run SH3 displays smaller blocking frequencies than the control experiment over the Atlantic and in the Pacific sector, but slightly larger ones near the Eurasian maximum. The explanation for the hypothetical decrease of the winter blocking activity in the Atlantic sector, given after Fig. 4.5, may still apply for the
anomalies, which have the main peak of blocking frequencies further to the west compared with LO method.

Figure 4.9: As Fig. 4.8 with $M=250$ gpm, but for CH3 (dash-dotted) and SH3 (dotted) simulations. Frequencies from the NMC analyses of the winters 1949/50–1978/79 are indicated by the solid line.

4.4 Blocking signatures

The poor winter blocking frequency of the CH4 simulation is further examined in this subsection by decomposing all DJF days into blocked and non-blocked days. The analysis is restricted to the Euro-Atlantic sector. In Fig. 4.10, the blocking criterion of equation 4.1 is applied with longitudinal averaging over 30° centered at 14.6°E. Only 32 days out of 900 winter days in the CH4 simulation, respectively 111 days for the NMC analyses of the period 1979–1988 are counted as blocked. The composite 500 hPa height charts of the non-blocked days, 792 for the NMC analyses, respectively 868 days for the CH4 experiment, are very similar to the mean flow charts in Fig. 3.6. The zonalization over the eastern North Atlantic and the positive bias over Spain are present in the non-blocked days. The simulated blocked ensemble mean in the bottom right panel of Fig. 4.10 displays a blocking ridge-trough system over Europe, which is quite similar to the observed pattern. A positive model bias on the order of 80 gpm is also notable over Spain for the blocked ensemble means. This systematic error is thus independent of the blocking performance of the model.

The differences between the blocked and non-blocked means give the so-called blocking signatures (D’Andrea et al. 1996). Blocking signatures for the NMC analyses of the thirty winters 1950–1979 and the ten AMIP winters 1979–1988, as well as for the CH4 and CH3 simulation are displayed in Fig. 4.11. The observed blocking signatures consist of a localized blocking dipole with the dominant positive center over southern Scandinavia. The strength of the ten winter blocking high exceeds 300 gpm, which is 40 gpm more than the value for the period 1950–1979. Furthermore, the larger sample produces a more localized field than the ten winter pattern. The cancellation away from the blocking
Figure 4.10: Composite 500 hPa chart of non-blocked (top) and blocked (bottom) winter days for the NMC analyses of the period 1979–1988 (left) and the CH4 simulation (right). A day is considered as blocked if the geopotential gradient between 40 and 60°N, averaged over 0 to 30°E, is negative. With this criterion, 111 (792) days are (non-) blocked for the NMC analyses, respectively 32 (868) days for the CH4 simulation.

Signatures is still smaller in the two simulations, which display large negative values over Alaska (CH4), respectively over the Hudson Bay (CH3). On the one hand, it is known that the ECHAM models 'like' PNA-like wave trains (Roeckner et al. 1996), and on the other hand, the blocking ensembles are made of 32 and 22 days, only. The simulated blocking signatures itself are stronger than the observed ones. A deep negative center of -160 gpm over Italy is simulated to the south of the positive center exceeding 320 gpm in the CH4 run. The signature of the CH3 experiment is shifted to the east and displays a weaker low.
Figure 4.11: DJF European blocking signatures for NMC analyses of the period 1979–1988 (top left, 111 blocked days), analyses from 1950–1979 (bottom left, 315 days), CH4 simulation (top right, 32 days), and CH3 simulation (bottom right, 22 days).
Fig. 4.12 displays the observed and the CH4 simulated blocking signatures for the spring season, when the model simulates a realistic number of blocked days. The model pattern is as localised as the observed signature but more zonally elongated in shape and about 50 gpm stronger.

![Figure 4.12: MAM European blocking signatures for NMC analyses of the period 1979–1988 (left, 95 blocked days) and for CH4 simulation (right, 105 days).](image)

### 4.5 Composites

To display the typical blocking patterns found in the model, blocking events of the CH3 simulation which persisted for at least five days have been collected in two composites. Table 4.5 gives the onset dates and the durations for the cases that make up the two composites. The strongest blocking event of the CH3 simulation, which is described in Ohmura et al. (1996), occurred slightly eastward of 30°E, and is thus not included in the composite. The 500 hPa height patterns are displayed in Fig. 4.13 for the Euro-Atlantic sector and the Pacific sector, respectively. A split-flow type blocking signature extending across 60° longitudes is simulated in the Euro-Atlantic sector with the anticyclone centered over the North Sea. Troughs encompass the high from both sides, and almost join over Italy. Far away, the composite chart displays a strong trough over eastern Asia, and an unusual trough exit over the Pacific. The Pacific pattern, on the other hand, consists of a strong, backward tilting blocking ridge over the northeast Pacific. The flow is largely parallel to the Rocky Mountains, and is very intense further downstream over the Atlantic and western Europe. Both blocking signatures are very strong and not much cancellation was introduced by the superposition of cases from different seasons, indicating that all the events developed at more or less the same locations within the prescribed 60° sectors.
Table 4.1: Start dates (day, month, model year) and durations (in days) of the cases that make up the European-Atlantic and the Pacific composite, respectively. Only cases with a minimum duration of five days within the sector limits $30^\circ W - 30^\circ E$ (EU-ATL), respectively $170^\circ E - 130^\circ W$ (PAC) are extracted from the 5 1/2 year sample provided by the CH3 simulation.

<table>
<thead>
<tr>
<th>EU-ATL</th>
<th>PAC</th>
</tr>
</thead>
<tbody>
<tr>
<td>start</td>
<td>duration</td>
</tr>
<tr>
<td>17. 3. 2</td>
<td>9</td>
</tr>
<tr>
<td>9. 4. 2</td>
<td>6</td>
</tr>
<tr>
<td>20. 5. 2</td>
<td>9</td>
</tr>
<tr>
<td>4. 6. 2</td>
<td>8</td>
</tr>
<tr>
<td>30. 10. 2</td>
<td>6</td>
</tr>
<tr>
<td>1. 12. 3</td>
<td>7</td>
</tr>
<tr>
<td>16. 12. 4</td>
<td>6</td>
</tr>
<tr>
<td>11. 4. 5</td>
<td>9</td>
</tr>
<tr>
<td></td>
<td>60</td>
</tr>
</tbody>
</table>

Figure 4.13: 500 hPa composite chart for Euro-Atlantic cases (left), and Pacific cases (right). The contour interval is 40 gpm.
5 Case studies

5.1 Objectives and methodology

The full value of the present high resolution climate simulations emerges when sequences of
atmospheric states are examined. At T106, the operational resolution of weather forecasts
at ECMWF during 1985–1991, the evolution of lows and highs over the oceans and their
alteration by continental elevations is captured with relatively fine detail. It was already
found in the pioneering case study of Bengtsson (1981) that a high resolution is essential
for blocking simulations, and weather forecasts in general.

A few blocking episodes are selected to show that very realistic developments occurred
in the CH3/4 experiments. Attempts have been made to choose archetypal cases, still
the selection of individual events from the available sample is subjective. A classification
beyond geographical position, namely North Pacific, Greenland, Atlantic-European,
Northern Russia blocking, or the degree of cutting off, i.e. Rex or dipole block, Ω-flow,
strong ridge, has not become evident from the inspection of numerous simulated and real
weather charts. In the case of cyclones, great knowledge exists about their life cycles, and
it is found that the barotropic shear of the ambient flow is decisive on the type of develop¬
ment (Davies et al. 1991). The situation is less clear for blocking flows which are of larger
scale than the transient systems, and which themselves influence the planetary scale flow.
Lateral shear certainly is important also for blocking dynamics, e.g., for the eddy straining
mechanism described in the Introduction but additionally nonlinear processes like Rossby
wave breaking have to be considered. Prior to overturning, the planetary waves have to
grow by one or the other of the proposed amplification mechanisms. A tentative answer
to the initiation question is tried in the last section. Before, the diagnostic tools which
are used to describe the cases are briefly discussed. Four Atlantic-European events have
been selected for a description, one in late autumn, a winter case, a spring case, and a
summer episode. A first specification of these cases is given by the Hovmoller diagrams
of the geopotential height gradient in Fig. 5.1, and in Table 5.1, which displays the start
and end dates, the maximum strength of the blocking highs, and their positions.

As for the methodology, traditional stereographic projections of the 500 hPa geopotential
height of the northern hemisphere are used to describe the synoptics. It is said (Mak
1991) that it suffices to diagnose the 500 hPa level, since barotropic processes are of primary
importance during an atmospheric blocking episode. Besides the 500 hPa level, which gives
a representative view of the block within the troposphere, surface and upper level features
are also considered, in particular for the European-Atlantic sector. The objective for doing
so is to identify whether locking of upper level into surface features (e.g., Davies et al.
1991) is essential during the onset phase of a blocking event. The surface development is
depicted by mean sea level charts, and also by a suitably selected potential temperature
isoline at 700 hPa to provide a separation line between cold and warm air masses. The
situation near the tropopause is viewed by wind vectors and potential vorticity on the
Table 5.1: Start and end dates of the cases that are described in the following sections. All cases, except case 3 which is from the CH3 simulation, are taken from the CH4 simulation. The year is not indicated in the blocking periods as it is irrelevant for the discussion. The third column gives the maximum reduced surface pressure of the blocking high. The number of days it took the high to develop since the onset is added in parentheses. The position of the maximum pressure is given in the fourth column.

<table>
<thead>
<tr>
<th>Case</th>
<th>period</th>
<th>max. strength</th>
<th>position</th>
<th>remark</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>22. Nov. - 4. Dec.</td>
<td>1044 hPa (8d)</td>
<td>55°N, 0°W</td>
<td>diffuse onset</td>
</tr>
<tr>
<td>2</td>
<td>11. - 19. Feb.</td>
<td>1062 hPa (3d)</td>
<td>62°N, 10°E</td>
<td>distorted vortex</td>
</tr>
<tr>
<td>3</td>
<td>17. - 25. Mar.</td>
<td>1053 hPa (2d)</td>
<td>57°N, 15°E</td>
<td>sequential blocking</td>
</tr>
<tr>
<td>4</td>
<td>7. - 15. Jun.</td>
<td>1031 hPa (6d)</td>
<td>57°N, 15°W</td>
<td></td>
</tr>
</tbody>
</table>

Figure 5.1: Hovmöller diagrams of $Z(\lambda, 40^\circ N) - Z(\lambda, 60^\circ N)$ for the four selected cases. Negative values of the index are emphasized. The contour interval is 100 gpm. Time runs downward covering November 1 to December 10 for case 1, February 1 to March 10 for case 2, March 1 to 30 for case 3, and June 1 to 30 for case 4.
250 hPa level. The following isobaric approximation is used for the potential vorticity (PV henceforth) near the tropopause level:

\[
PV_{250} \approx \left\{ f + \zeta + \frac{\partial v}{\partial p} \frac{\partial \theta}{\partial x} - \frac{\partial u}{\partial p} \frac{\partial \theta}{\partial y} \right\} \left( -g \frac{\partial \theta}{\partial p} \right).
\] (5.1)

This is Ertel's PV formulated on \( p \)-surfaces and not on \( \theta \)-surfaces. The actual calculation is done with the 300 and 200 hPa wind components, the 250 hPa relative vorticity \( \zeta \), and potential temperature \( \theta \). As a derived quantity, the PV field shows rich structures and deep gradients, which provide the essence of atmospheric dynamics. When evaluated on isentropes, the PV can not change within an isentropic surface in the absence of friction and diabatic heating. A thorough discussion of PV is given in the milestone by Hoskins et al. (1985), where the unifying concept is also illustrated with Rossby wave breaking and cutoff systems.

To justify the use of an isobaric surrogate of PV instead of the full isentropic PV, an IPV chart has been prepared for comparison in Fig. 5.2 for one example situation. Larger discrepancies may be noted south of 40°N, where the isentropes start to slope towards the surface with decreasing latitude. The gross features, namely the elongated streamer with a rollup vortex at its end over the western Mediterranean and the trailing trough over the Atlantic, are however remarkably similar on the 250 hPa surface compared with the 315 K isentropic picture, except for the different plotting programs used for the two panels. The subsequent development is shown in section 5.4.
Figure 5.2: 250 hPa isobaric chart of wind vectors and potential vorticity (top) versus 315 K isentropic chart (bottom) of the same situation. Contours at $-1, -0.5, 0, 0.5, 1, 1.5, 2, 3, \ldots$ for potential vorticity with shading above $2 \text{PV units}$, where $1 \text{ unit} = 10^{-6} \text{ m}^2 \text{s}^{-1} \text{ K kg}^{-1}$. The $1 \text{PV unit}$ isoline and values below are drawn thicker in the top panel. The plotting program unfortunately does not fill regions that cross the boundaries and considers also the Greenwich meridian as one. The magnitude of the maximum wind vector is given in the lower right part of each panel.
5.2 A late autumn Atlantic blocking event

Synoptic description

The model weather preceding November 9, the first day shown in the sequence of states from year 3 of the CH4 simulation, is characterized by the formation of a surface high over central Europe. Although the anticyclone is strong, with mean sea level pressures of up to 1045 hPa, it is not truly blocking as it is centered at 45°N, slightly too far south to check Atlantic depressions. Still, the strongly developing cyclone west of northern Scandinavia, as well as the small cutoff over the Atlantic visible on the 500 hPa chart in Fig. 5.3 is affected by the anticyclone. The circumpolar flow on the 9th is relatively distorted compared with the previous days mainly because of two deep troughs over the Sea of Okhotsk and eastern North America, and also due to the developing large cutoff low over southeastern Europe. Three days later, a part of the European anticyclone has moved northwestward over Ireland and a weak dipole pattern is present over the eastern Atlantic. November 15 is characterized by a wavenumber 5 pattern, wherein the trough-ridge system in the Atlantic-European sector is blocked by the anticyclone, which is pushed to the east in the following days by the wave. Again three days later, the ridge is situated over northeastern Europe. Large meridional excursions of air parcels may be inferred from the height chart of November 18. On November 21, a ridge over southern Greenland helps to cut off the stagnant air south of Ireland in the following days. Another cutoff low joins the primary low on November 24, while the high to the north of Scotland is intensifying. A well developed Ω-flow characterizes the situation on November 27. This pattern persists during the next five days. On December 3 the blocking ridge is extended in the zonal direction, but still resists the approaching trough to the east of Newfoundland for the following two days. By December 6, however, the high pressure cell has weakened and is superseded by a large Atlantic cyclone in the following days.

To identify a clear-cut beginning is not obvious from the synoptic development just described, and it is instructive to check in the Hovmöller diagram when the blocking event begins and when it terminates. The largest connected region with negative index values in Fig. 5.1 commences on November 22, when the low apparent in Fig. 5.3 on November 21 is cut off from the westerly flow. The preceding weather situation is marginally recognized as blocked by the index in the two short negative blobs on November 13 and 20, respectively. The onset of this particular case takes several days and seems to be initiated by the advection on the eastern side of several cold air troughs of warm air that merges with the already existing high over central Europe to form the blocking high. The high remains blocking for 12.5 days according to the LO index displayed in Fig. 5.1.

Surface and upper level features

The same three day interval sequence as for the 500 hPa height maps is shown for the mean sea level pressure and the 250 hPa PV field over the Atlantic-European sector in
Figure 5.3: 500 hPa geopotential heights for November 9, 12, 15, ... of CH4 model year 3. Contours every 80 gpm. Continued on next page.
Fig. 5.4. Both the surface and the upper level charts display a lot of synoptic activity during this blocking period, which is not a particularly stationary one. The blocking high reaches maximum strength around November 30. The 700 hPa potential temperature on that date shows a poleward fold of warm air with the 295 K isoline bending northward at 40°N, 25°W and reaching eastern Greenland. The corresponding upper level chart displays a meridionally elongated PV region where values exceed 8 units. Strong northward winds ahead of this structure transport low latitude air into the blocking region which is characterized by a large region where the PV values are quasi-homogeneously below 0.5 units. Three days later, on December 3, the low PV region has split in two pieces. The larger portion is situated over the North Sea and grows again but also moves back southeastwards. On December 6, the last situation shown in Fig. 5.4, the warm air fold is nearly smoothed out and the anticyclone over Europe is fading.
Figure 5.4: Mean sea level pressure (left) and potential vorticity together with wind vectors on the 250 hPa surface (right) for November 9, 12, 15,…to December 6 of CH4 model year 3. Contours every 5 hPa on the pressure maps with the 1000 hPa isoline and pressures above 1030 hPa emphasized. The 295 K potential temperature contour at 700 hPa is also drawn (heavy broken). Contours at $-1, -0.5, 0, 0.5, 1, 1.5, 2, 3,…$ for potential vorticity with shading above 2 PV units. The maximum wind vector is drawn heavy and its magnitude is given in the lower right part of each panel. Continued on next page.
Prior to the peak of the block, a number of troughs cross the Atlantic and advect low PV air to higher latitudes. Incursions of this kind are visible for November 12, 15, and in particular November 18 on the upper level charts. The mean sea level distribution on November 21, the date just prior to the onset of the main blocking case according to the LO index, displays the anticyclone receded over Russia and an occluding cyclone vertically aligned with a cyclonic PV feature to the southwest of the British Isles. This large cutoff together with the trough-ridge system between 60°W and 30°W on November 21 leads to the final formation of the block. On November 24, the blocking high is established over northern Scandinavia and migrates slowly southwards in the next twelve days.
5.3 A European winter case

Synoptic description

Case number two occurred in midwinter of year 8 in the CH4 simulation. A sequence of 500 hPa geopotential height charts is shown in Fig. 5.5 in two day intervals. Unusually strong flow across the North Pole and a large trough over the Middle East are present on February 9. A further prominent feature of the same map is the intense gradient over Newfoundland. Two days later, the trough is in the middle of the Atlantic and is compressed in the zonal direction. The same happens to the ridge at 15°W. The zonal compression and stretching in the meridional direction is at an advanced stage on February 13, 0 UTC. Trough merging takes place over the Atlantic during this day. At the same time, the compressed ridge overturns and becomes a separated high pressure cell. The trough over the Middle East, on the other hand, has developed into a large cutoff low which moves slowly westwards to become the cyclonic cell of the blocking dipole. A dipolar flow configuration is present for the next five days between February 14 and 18. During this period, the westerlies split up near 50°N, 30°W, flow around the high-low system in two branches, and join some 90° longitude further east. On February 19, the dipole is about to disintegrate. The high retreats eastwards and remains blocking over Russia until March 5 (not shown, except in Fig. 5.1).

Surface and upper level features

The sea level pressure distribution prior to February 9, the first instance shown in Fig. 5.6, was characterized by the merging of an anticyclone over Central Europe with a polar high. The resulting high pressure complex is seen to stretch over the eastern Atlantic, Europe, turning northwards and extending up to the Barents Sea on February 9. Two days later, the zonally elongated high over the eastern Atlantic is meridionally elongated. The 300 K potential temperature isoline displays a large wedge of warm air in the sector 30°W to 0°W at the 700 hPa level. Coherent southward and northward transports over about 30° latitude can also be seen near the tropopause level. Cyclonic PV regions with values of up to 9 units flank the ridge on both sides on February 11. Again two days later, the blocking high is present in full strength with a central pressure of 1062 hPa over Finland. This high value is not unrealistic for Scandinavia, which experiences pressures in excess of 1050 hPa every now and then (e.g., in November 1993 or during March 1996). The highest pressure ever measured in Helsinki is 1065.7 hPa, on 22 January 1907 (P. Räisänen, personal communication).

Besides the dominating anticyclone, a large low over the Balkans with a central pressure of 1005 hPa, and a meridionally elongated low over the Atlantic are present on the sea level pressure chart of February 13. A large region of low PV values bows from the subtropics to the Barents Sea on the upper level chart of the same date. A strong positive PV feature is located on the inner side of the low PV bow over northern Germany, where the PV
Figure 5.5: 500 hPa geopotential heights for February 9, 11, 13, to February 19 of CH4 model year 8. Contours every 80 gpm.
Figure 5.6: As Fig. 5.4 but for February 9, 11, 13, and 15 of CH4 model year 8. The 290 K isoline of the 700 hPa potential temperature is drawn heavy broken in the left hand panels. Continued on next page.
Figure 5.6: Continuation from previous page. February 15, 17, 19, 21, and 23.
attains values as high as 11 units. Large cyclonic PV regions are also found to the north of the blocking region. Furthermore, very intensive advection with wind speeds exceeding 85 m s\(^{-1}\) takes place over the western Atlantic. This positive vorticity advection results in a vortex at 35°N, 15°W on February 15. With a maximum of 14 PV units, this vortex is very strong and interacts with the large, zonally elongated cyclonic feature to the north.

Further north, a quasi-homogeneous low PV and low windspeed region marks the block. On February 17, the large scale mixing effect on lower tropospheric temperatures by the block is evident. The 700 hPa potential temperature over Scandinavia, e.g., is larger than that over Spain. Two days later, the 300 K isoline runs more zonally and the blocking high over Scandinavia has weakened. The surface low over southern Europe and an Atlantic low pressure wedge have merged to form a 1000 hPa cyclone over France on February 19. Aloft, the main vortex has moved eastward 30° in two days while another disturbance reached western Europe. This stage terminates the blocking event over Europe. As indicated in Fig. 5.1, the episode finds a continuation over Russia that persists until March 5. The Scandinavian blocking high retreats eastwards, absorbs parts of a subtropical high over southwestern Europe, and becomes stationary near 55°N, 45°E.

**Distortion of the stratospheric vortex**

Having illustrated the very intensive tropospheric dynamics during a particular wintertime blocking event, it is interesting to briefly look into the lower stratosphere. The objective is to check whether this particular midwinter blocking was accompanied by a sudden warming. In a major warming event, the polar stratospheric temperatures increase spectacularly by several tens of degrees Celsius in a few days. The westerly circumpolar flow in the stratosphere breaks down completely and does not recover by the end of the winter. As discussed in Tung and Lindzen (1979, and earlier references therein), simultaneous blocking seems to be necessary for stratospheric warmings. No systematic search for major warmings has been done for the model data.

Figure 5.7 displays the geopotential at the 50 hPa level on the first and last day of the discussed period. Also shown is the mean over the particular February along with the ten winter average of the CH4 simulation. The lower stratospheric flow on February 9 is not quite circumpolar but has moved its minimum center over northern Canada, indicating a pool of cold arctic air in that region. Anomalously high geopotential values are centered over the Sea of Okhotsk. Ten days later, the flow is more zonally symmetric, except above Scandinavia where the westerlies are weaker. The February mean flow, shown in the lower left panel, is more elongated in a wavenumber two pattern than the total winter ensemble average.

To check whether the ultra-long waves propagating into the stratosphere have caused a sudden warming, the 50 hPa temperature variation with latitude and time is displayed in Fig. 5.8. The zonally averaged temperature has not changed considerably over the blocking
Figure 5.7: 50 hPa geopotential for February 9 and February 19 (top). Monthly mean 50 hPa heights for the particular February (bottom left) and averaged over the 10 winter seasons of the CH4 simulation (bottom right). The contour interval is 200 gpm.

period described above (days 8–18 in Fig. 5.8). A minor warming, with no reversal of the zonal flow, has occurred towards the end of the month and during the first days of March.
5.4 Double blocking in spring

Synoptic description

The third case considered is taken from the CH3 simulation and occurred in March of year 2. The 500 hPa height on March 15, the first day in Fig. 5.9, shows the rollup vortex already encountered in the introductory section. Strong southwesterlies flow between Iceland and Scotland. In the Pacific sector, a blocking ridge, which has been established six days earlier, persists throughout the blocking event in the European sector. The double, or rather sequential blocking situation is also displayed in the Hovmöller diagram in Fig. 5.1. On March 17, a blocking dipole configuration, where the high is still connected with the subtropics, is established over western Europe. Two days later, the anticyclonic cell has settled and become the dominant part of the dipole. The northern branch of the jet is very intense on this day. Again two days later, on March 21, the main circumpolar flow attains a dumb-bell shape due to the two blocking highs which are simultaneously present.
Figure 5.9: 500 hPa geopotential heights for March 15, 17, 19, 21, 23, and 25 of CH3 model year 2. Contours every 80 gpm.
over the northeastern Pacific and Scandinavia. On March 23 and 25, blocking action is still apparent in the 500 hPa charts. However, a strong cyclone develops off Newfoundland which terminates the blocking event.

Surface and upper level features

The surface chart of March 15 shows some similarity with the first chart of the winter case in the previous section. Namely, a large anticyclonic complex covers central Europe and continues over Russia. Over the Atlantic, however, a deep cyclone with central pressure of 955 hPa near Iceland makes the initiation of this case particularly rapid. On March 17, the blocking high is already present in full strength. The intense advection from the south has led to air of less than 0.5 PV units over Iceland. The vortex on the southern edge of the plotting region has absorbed parts of the positive PV streamer, which is now oriented to the east-northeast. The block is at its mature stage during the following couple of days. Large regions north of 60° show PV values below one on March 19. Anticyclonic flow encompasses this extended area as can be seen from the 250 hPa wind vectors. The surface chart again displays a low pressure system over the Atlantic giving up its energy to the blocking field as it is meridionally stretched. Another low in the Barents sea leads to maximum wind speeds of 82.4 ms⁻¹ in the polar branch of the jet. Two days later, on March 21, the surface high has weakened by 10 hPa, and the polar low has moved out of the plotting region. The 700 hPa temperature displays low temperatures over southern Europe due to cold air advection on the southeastern flank of the blocking high. On March 23, the last day shown in Fig. 5.10, the cold air blob has migrated further west. The anticyclone is still present with mean sea level pressure of about 1035 hPa. Aloft, the winds are very weak above the high, but the low PV region in the north has shrunk. Although the low PV air is regenerated by the other low PV blob wherein negative values are reached, the block comes to an end three days later as a deep low takes over from the Atlantic.

Vertical sections

To illustrate the vertical structure of the blocking situation, vertical sections of potential temperature and the wind components are displayed in Fig. 5.12 for a north-south section and a west-east section through the blocking configuration. The sections are averages over the eight day period March 17–24 when the Euro-Atlantic block is in its mature phase. The mean sea level pressure and 500 hPa geopotential height averaged over this period is shown in Fig. 5.11, which also indicates the position of the sections by heavy lines. The average 500 hPa geopotential chart displays the blocking high centered over the North Sea, and also the blocking ridge which is present at the same time in the Pacific sector.

The west-east section at 56.6°N in Fig. 5.12 displays a warm anticyclone extending through the entire troposphere from 30°W to 40°E. The tropopause in the blocking region is lifted by approximately 30 hPa. Moreover, intense meridional winds just below the tropopause
Figure 5.10: As Fig. 5.4 but for March 15, 17, 19, 21, 23, and March 25 of CH3 model year 2. The 295K isoline of the 700 hPa potential temperature is drawn heavy broken in the left hand panels. Continued on next page.
Figure 5.10: Continuation from previous page.
characterize the mature block. The strength exceeds 40 m s\(^{-1}\) for the southerlies in the west, whereas the northerlies to the east of the blocking high are below 30 m s\(^{-1}\) for the chosen latitude of the section. The zonal winds in the south-north section on the meridian at 4.5°E display two westerly maxima. With over 50 m s\(^{-1}\), the northern branch of the jet at 75°N is stronger than the southern branch at 25°N. Maximal easterly flow of 20 m s\(^{-1}\) occurs between the split jet at the 300 hPa level. Larger variations of the tropopause height, and stronger winds occur on individual days, as exemplified in Fig. 5.13 for March 17 and 18, when the block is at a mature stage. Crum and Stevens (1988) show cross sections of an observed case that are similar to the ones in Fig. 5.12.

Figure 5.12: West-east section at 56.6°N (left) and south-north section at 4.5°E (right) for the same period as in Fig. 5.11. The thin solid lines are isentropes and the thicker lines the wind component normal to the section. Northerly and easterly winds, respectively, are dashed.
Figure 5.13: As Fig. 5.12 but for 12 UTC of March 17, 18, and 19 (from top to bottom).
Vertical sections are also displayed in Fig. 5.14 for the Pacific sector averaged over a slightly different period than that shown in Fig. 5.12 to capture the mature phase in that region. In contrast to the Atlantic-European blocking high, the blocking ridge over the northeast Pacific displays stronger northerlies than southerlies in the west-east section. An intensification of the northern branch of the jet is visible in the north-south section, whereas easterlies are absent above the 500 hPa level.

Figure 5.14: Cross sections for the ten day period March 11–20 over the North Pacific. West-east section at 56.6°N (left) and south-north section at 140°W (right). Contours as in Fig. 5.12.

Figure 5.15: As Fig. 5.14 but for March 11, 12 UTC.
5.5 A summer episode

Synoptic description

A model blocking situation that occurred in June over the eastern Atlantic is described as a last example. The meridional temperature contrasts in summer are much smaller than during the cold seasons, and the 500 hPa geopotential consequently displays a smaller gradient towards the pole. Also, the jet aloft is weaker and more to the north than during winter. The situation at the 500 hPa level is displayed in Fig. 5.16 in three day intervals for the selected case. A cold air advancement towards southern Europe initiates the episode on June 3. Three days later, the summer jet splits in two branches near 50°N, 30°W. Soon afterwards, the LO blocking index turns negative on June 7 (Fig. 5.1).

Figure 5.16: 500 hPa geopotential heights for June 3, 6, 9, and 12 of CH4 model year 7. The contour interval is 80 gpm. Continued on next page.
Figure 5.16: Continuation from previous page. June 15, 18, 21, 24, 27, and June 30.
On June 9, a weak dipole is present with the high to the northwest of the low over the Gulf of Genoa. Flow in the form of an $\Omega$ around the blocking high may be noted three days later. The anticyclonic cell then moves over the British Isles around June 15. Troughs flank the high on both sides. The trough to the east intensifies by June 18, whereas the high weakens and is more zonally elongated on that date. By June 21, when the event has terminated according to the LO index, the blocking action has travelled westwards to form a blocking ridge over Greenland. This ridge exhibits some persistence, more in time than in position, until the end of the month.

**Surface and upper level features**

That the dynamics is also quite active during a summertime blocking is displayed by the PV charts in Fig. 5.17. The first panels display a low over western Europe at the surface and the southward advancement of high PV air at the 250 hPa level. June 6 is characterized by a low to the southern tip of Greenland and accompanying southwesterly advection with wind speeds of nearly 60 m s$^{-1}$. Three days later, low PV values are found over the North Atlantic, while a number of cyclonic PV anomalies are around in the south. On June 12, the cyclonic PV features have joined to a large scale structure, which intensifies the southern branch of the jet. Also the low PV region has homogenized and extended in size. The surface anticyclone is at its mature stage with central pressure above 1030 hPa. A large scale rollup process takes place in the following three days. The supply of subtropical air has more or less stopped on June 15, and the low PV air slowly circulates around the British Isles taking the form of an embryo. Three days later, a very intense trough has developed on the eastern flank of the blocking high, which is shifted westwards. A cyclonic anomaly of 11 PV units rapidly moves into the continent from the north. The remaining vortex is still strong three days later on June 21 over southeastern Europe. Another trough is advancing from the north on the same date, while the anticyclone has moved further west and got somewhat weaker. Six days later, however, the high is restrengthened again near Ireland on June 27. In the sequel, the high retreats to the subtropics and the blocking episode comes to an end.
Figure 5.17: As Fig. 5.4 but for June 3, 6, 9, ..., to June 30 of model year 7. The 300 K isoline of the 700 hPa potential temperature is drawn heavy broken in the left hand panels. Continued on next page.
Figure 5.17: Continuation from previous page.
5.6 A closer look at the initiation phase

As mentioned in the introductory section, a key question is what causes the amplification of the planetary waves prior to blocking onset. A number of processes creating blocking have been proposed but none of the following mechanisms applies to all block onsets. The antecedent growth may alternatively be due to:

a) external forcing (Tung and Lindzen 1979; Charney et al. 1981),
b) barotropic conversions of kinetic energy (Simmons et al. 1983),
c) baroclinic processes (MacVean 1985; Colucci and Alberta 1996), or
d) three-dimensional instability (Frederiksen 1982).

In reality, a combination of the above mechanisms may create blocks, with one or the other process being dominant for certain cases. To interpret the onset phase in GCM
simulated cases is almost as difficult as for observations. A useful tool to diagnose the barotropic mean flow feedbacks of eddies is the extended Eliassen-Palm flux, defined by the horizontal velocity correlations $\overline{u'v'} - \overline{u}^2$ and $-\overline{u'u'}$ (Hoskins et al. 1983). The study concentrates here on the items c) and a).

**Rapid cyclogenesis and cold air outbreaks as precursors**

Colucci and Alberta (1996) have shown that explosive cyclogenesis 1–3 days prior to the onset is favorable but neither necessary nor sufficient for cold-season blockings. This inconclusive situation is also found in the model cases. Still, cyclogenesis is always found upstream and prior to the block, and often an individual cyclone intensifies rapidly and may become a large cyclonic complex, e.g., as in case 3 (Fig. 5.10). In all four cases, however, pronounced cold frontal troughs developed during the onset phase. It seems as though the potential energy of cold air is needed to initiate the large scale mixing of air masses during blocking events.

A more coherent picture of the time evolution of the onset of case 1, which is displayed in three day intervals in Section 5.2, is given in Figs. 5.18 and 5.19. The two figures show the sea level pressure distribution and the situation near the tropopause during two incursions of low latitude (low PV) air towards Iceland. Westward propagation of an anticyclone, which is already blocking synoptically despite the positive Lejenäs-Økland index, is displayed by the pressure charts in Fig. 5.18 covering the period November 9, 18 UTC, to November 11, 6 UTC, in 12-hourly intervals. Rapid cyclogenesis occurs at the western edge of the plotting region to the north of Newfoundland. The cyclone deepens from 990 hPa on November 10, 6 UTC, by 15 hPa in the following 12 hours, thus fulfilling the deepening rate for 'explosive' cyclogenesis (24 hPa in one day at 60°N, according to Konrad and Colucci 1988) during this half-day. The second incursion shown in Fig. 5.19 occurs further east and reaches well into the polar latitudes. Frontal structures and strong meridional shears are visible on November 18 over the Atlantic and western Europe. The onset of the main event of this blocking episode occurs 4 days later between November 21 and 24, when a cutoff low develops over the Gulf of Biscay accompanied by another incursion of low PV air on the upstream side (c.f. Figs. 5.3 and 5.4).
Figure 5.18: Incursion of low latitude air during November 9, 18 UTC to November 11, 6 UTC. Representation as in Fig. 5.4.
Figure 5.19: Another incursion of low latitude air during November 17, 6 UTC to November 18, 18 UTC. Representation as in Fig. 5.4.
Shifts of the jet streams and orographical forcing

One way in which the external forcing may vary is by latitudinal shifts of the jet streams flowing over different orographical profiles. Da Silva and Lindzen (1993) have shown that a southward shift of the jet leads to stronger orographical forcing and larger amplitudes of the stationary eddies. As suggested by Tanaka (1991), and further elaborated on in Section 7, the amplified planetary waves then play a catalytic role for blocking initiation by baroclinic developments.

Five day mean profiles of the zonal wind are displayed in Fig. 5.20 for the four months of the blocking cases 1 to 4. The main subtropical jet stream maximum near 30°N shows no latitudinal shifts for case 2 (in February) and 3 (in March). If the zonal wind is averaged only over North America (not shown), the differences in the wind profiles between the pentads are naturally larger (particularly for cases 2 and 3) but they are not very systematic. Two blocked pentads of case 1, which have been emphasized by thicker lines in Fig. 5.20, display a relatively strong subtropical jet near 40°N. This position is typical for autumn, and it is difficult to judge whether the large subtropical variations of case 1 are normal during the transition season or are linked with the blocking occurrence. Wind and surface stress vectors in Fig. 5.21 give some further illustration of the situation during November 16–20 (the thin dash-dotted line in Fig. 5.20), respectively November 21–25 (the thick dash-dot-dot-dot-dotted line). The summer profiles of case 4 scatter on the order of 10 ms⁻¹ for all latitudes. Case 3 displays the most pronounced departures from the climatological zonal mean wind during the blocked pentads in the middle latitudes. A second jet maximum of 20 ms⁻¹ at 70°N and a local minimum of 10 ms⁻¹ at 55°N characterize the two blocked latitudinal wind profiles. The blocking anticyclones simultaneously present in the Pacific and Atlantic sectors contribute to this form of the zonal mean profile. Fig. 5.22 shows 250 hPa wind and surface stress vectors for the pentad prior to block onset and the five day mean fields during onset. The surface stress fluctuates very strongly in space and in time.

No conclusive answer to the onset question can be given from the above very limited analysis. The cases presented in this chapter, however, allow to estimate the realism with which the T106 simulations capture many features of observed events by comparing them with the pertinent studies (e.g., Bengtsson 1981; Illari 1984; Shutts 1986; Hoskins and Sardeshmukh 1987; Crum and Stevens 1988) or by inspecting weather charts for blocking periods (e.g., November 1993, or March 1996). The warm core anticyclone, replenished by vigorous inflows of subtropical air, the lifted tropopause, the quasi-uniform low potential vorticity values in the blocking cell, and the formation of cold cutoffs to the south of the warm air pool are tropospheric features reproduced realistically in the climate model. However, as nothing systematic has been noted in the baroclinic developments, it seems as though the planetary scale background is decisive for the onset and the stationarity of blockings.
Figure 5.20: Five day mean profiles of the zonally averaged 250 hPa zonal wind for case 1 (middle left), case 2 (middle right), case 3 (bottom left), and case 4 (bottom right). The ‘blocked’ pentads are emphasized and the sequence of linestyles is solid (first five days of the respective month), dotted (second pentad), dashed, dash-dotted, dash-dot-dot-dotted, long-dashed. Also shown in the top panel are seasonal reference profiles from the CH4 simulation for DJF (solid), MAM (dash-dotted), JJA (dashed), and SON (dotted).
Figure 5.21: Five day means of 250 hPa wind vectors and isotachs (left) and surface stress vectors (right) for days 16–20 (top), respectively days 21–25 (bottom) during the November of case 1. Isotachs are drawn every 10 m s\(^{-1}\), and isolines of the surface stress magnitude every 250 mN m\(^{-2}\). The numbers at the bottom right edge give the magnitude of the longest vector, which is drawn thicker in each panel.
Figure 5.22: As Fig. 5.21, but for days 6–10 (top), respectively days 11–15 (bottom) during March of case 3.
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6 Sensitivity to model resolution

One of the major concerns in climate modelling is the separation of explicitly resolved and parametrized processes. Reducing the horizontal mesh spacing of a global model from, say, 300 km to 100 km hardly introduces new processes that have to be modeled explicitly but allows to represent the continental land masses and the circulation features with more spatial detail. In fact, the T106 simulations are done with the same model as the T42 runs, except that the boundary and initial files are adapted to the respective resolution. Furthermore, a few schemes for subgrid scale processes contain parameters that have to be adjusted to the resolution. Two key control knobs are the rain efficiency parameter of the large scale precipitation scheme and the constants of the horizontal diffusion. The tuning is done against the observed outgoing longwave radiation and the kinetic energy spectrum, respectively. Low or medium resolution may harmonize better with the various parameterizations than high resolution so that the climatic means do not universally converge to the atmospheric analyses with increasing horizontal resolution (e.g., Williamson et al. 1995). The optimal resolution for a particular application is thus determined in a heuristic manner, and is constrained in practice by the available computing power and the envisaged simulation period.

Blocking is a good example to elaborate more on the resolution question. On the one hand, experience from numerical weather prediction tells that high horizontal resolution is essential for successful blocking forecasts (Bengtsson 1981; Tibaldi and Ji 1983; Simmons et al. 1989; Tracton 1990) but on the other hand, it is also known that highly truncated models are capable to sustain a relatively realistic blocking activity (Liu and Opsteegh 1995; Tibaldi et al. 1995). Often however, the satisfying blocking performance of the simpler models is accompanied by large systematic errors, e.g., the quasigeostrophic model of Marshall and Molteni (1993) which is used in the two above mentioned studies simulates a too strong ridge over the North Atlantic. Also, the perpetual January GCM experiment of Blackmon et al. (1986) gave encouraging results which are probably due to the large zonal wind error of the low R15 resolution favouring blocking incidence.

If the eddy straining mechanism is a valid theory then high horizontal resolution should be a positive factor for blocking simulations as the enhanced resolution captures the transient eddies and the nonlinear transfer of energy to the blocking background field better. How well this idea is realized in the ECHAM simulations is studied in this section. After a brief discussion of the systematic errors, the effect of the increased horizontal resolution on the blocking frequencies is examined.

6.1 Systematic errors of the midtropospheric geopotential

Figure 6.1 compares the systematic errors of the winter 500 hPa height fields for the T42 simulations with the previously discussed T106 results. The top left panel displays a
five winter mean from the CM3 simulation while the bottom left panel shows the much reduced errors in the ECHAM-4 simulation CM4. The high resolution errors of the CH3 and CH4 simulations shown on the right hand side are generally somewhat larger and the zonalization is more pronounced in the T106 experiments.

Figure 6.1: Systematic error of the 500 hPa geopotential height field in winter for the CM3 (top left) the CH3 (top right), CM4 (bottom left) and CH4 (bottom right) simulation. The contour interval is 20 gpm.

Two regions show particularly obstinate systematic errors. A relatively large positive bias is found in all four simulations in the Pacific sector where the model has problems to simulate the broad Asian trough. A comparison with the work of May and Bengtsson (1996) shows that the error pattern of the present climatological SST simulations with the ECHAM-3 model has much in common with the departures from normal conditions during the La Niña events. In other words, a large fraction of the systematic error is due to the neglect of El Niño events. The other problematic region is found downstream of
the Atlantic storm track over southwestern Europe, except for the CH4 simulation where
the overestimation has a maximum off Newfoundland. Neither horizontal resolution, to
the contrary, nor revised parameterizations are able to remove this error.

Figure 6.2 displays the winter stationary waves of the northern hemisphere for the
ECMWF analyses, the CM4, and the CH4 simulation. The stationary eddies are ob-
tained by subtracting the zonal mean from the time averaged 500 hPa geopotential height
field. Fig. 6.2 shows that the east Asian trough is deep enough in the CM4 simulation and
even overestimated in the CH4 experiment. The northeast American trough, on the other
hand, is underestimated by both model simulations. The ridge over the Rocky Mountains
is captured in strength and only a slight underestimation over Alaska may be noted for
the simulations. The ridging over Scandinavia is also underestimated by the model, in
particular by the T106 version which, however, has a positive bias further south. The
secondary ridge in the vicinity of the Ural is in accordance with the analyses.

An illustration of the excitation of the stationary waves by the orography is provided in
Fig. 6.3 which displays the wave amplitude together with the topographical profile as a
function of longitude. The profile at 45°N shows a close coincidence with observations
of the simulated troughs downstream of the central Asian mountains whereas this trough
is too deep in the high resolution simulation at 51°N. Interestingly, the T42 truncated
orography is at least as high as the T106 spectral mountains at 45°N (Tien Shan and
Altai mountains). The Rocky Mountains, on the other hand, are higher at T106 than at
T42 resolution in the 45°N profile. Still, the ridge and trough system over North America
is somewhat weaker for the higher resolution, and also underestimated compared with
the analyses. The ridge over the North Atlantic is shifted to the west and too strong for
both latitudes in the CH4 simulation. In summary, the mean orography together with the
gravity wave drag parameterization results in realistic stationary waves which show only
minor, but intriguing, deviations from the observations.

The resolution dependent behaviour of the midlatitude westerlies throughout the year is
Figure 6.3: 500hPa stationary waves at 51°N (top) and 45°N (bottom) in winter for the ECMWF analyses (solid line), the CM4 (dashed) and the CH4 simulation (dotted). The T42 orographical heights are indicated by shading. The fine dotted line is the T106 orographical profile.

depicted in Fig. 6.4 by the zonal flow index introduced previously. The seasonal cycles of three ECHAM-3 simulations with the different resolutions T21, T42 and T106 are shown for the northern hemisphere and the two blocking sectors. The medium resolution is generally closer to the NMC analyses than the high resolution simulation, except during midwinter in the Pacific sector. The low resolution experiment is included to show that even T21 resolution, corresponding to a grid mesh of about 600 km, performs quite good for time mean quantities.
Figure 6.4: Annual march of the zonal flow index $Z_{40^\circ N} - Z_{60^\circ N}$, repeated twice for better survey, for NMC data of 1950–1979 (solid line), and for the three different resolutions T106 (dash-dotted), T42 (dashed), and T21 (dotted) of ECHAM-3 simulations. The shaded band indicates the interannual variability ($\pm 1\sigma$) of the observed index of 1950–1979. From top to bottom: average over the northern hemisphere, Atlantic sector 60° W – 30° E, and Pacific sector 150° E – 120° W.

6.2 T42 versus T106 blocking frequencies

Three comparisons of T42 with T106 experiments with an equivalent setup are available. The first pair is the long control integration CM3 and the five years of the CH3 climatological SST simulation. Fig. 6.5 displays the seasonal blocking activities of the two ECHAM-3 simulations together with the observed frequencies. The Euro-Atlantic sector will be considered only in the following as the ECHAM-3 simulations without interannual SST variability have large systematic errors and suppressed blocking activity over the North Pacific. Blocked conditions are more frequent for the high resolution simulation than the standard resolution run throughout the year in the European sector. While the T42 experiment simulates about 70% of the observed blocking activity during winter and spring, the T106 model reaches the observed level, albeit shifted to the east in winter and severely underestimated over the Atlantic to the west of the Greenwich meridian. Still, the T106 simulation captures the occurrence of anomalous flow situations better than the standard model despite the larger zonal bias at high resolution apparent in Fig. 6.4.
Figure 6.5: Seasonal blocking frequencies according to the Lejenäs and Økland index for the CH3 (full line) and the CM3 simulation (dash-dotted). The observations of the period 1950–1979 are indicated by dots.

Two comparisons are possible for the ECHAM-4 model. One pair of experiments uses climatologically prescribed SSTs, and the other couple AMIP SSTs. The comparison of the CM4 with the CH4 experiment in Fig. 6.6 shows that, at least in the two available model samples, the T106 does not have a big advantage over the T42 simulation in the new model. With SST variability, shown in Fig. 6.7, the blocking frequencies come closer to the observations in both sectors, and for both resolutions. A few autumn cases distributed over the whole AMIP period 1979 to 1988 resulted in the fall peak in the European sector of the AH4 simulation. Otherwise this experiment shows a fine performance.
Figure 6.6: Seasonal blocking frequencies for the CH4 (full line) and the CM4 simulation (dash-dotted). The observations of the period 1950 to 1979 are indicated by dots.

Figure 6.7: Seasonal blocking frequencies for the AH4 (full line) and the AM4 simulation (dash-dotted). The observations of the period 1979 to 1988 are indicated by dots.
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7 Orographical forcing

The purpose of this chapter is twofold. First the effect of the gravity wave drag (GWD) parameterization on the January model climate is briefly discussed. Then the effect of removing the terrain heights on the blocking activity is investigated. Three experiments with reduced orographical forcing have been performed with the ECHAM-4/T42 model under radiative conditions of January. A perpetual January integration with the standard mean orography was also done for comparison. In a first simulation, the topography was truncated above 500 m. A second true ‘no-mountain’ experiment was done with completely flat continents. To separate the effect of the GWD parameterization, the model was run for six months with standard orography but no drag due to subgrid scale terrain variations.

7.1 The effect of the gravity wave drag scheme

Simulating the drag due to subgrid scale orography is a difficult task for every GCM, except for low resolution models where error compensation makes it favorable to apply no GWD parameterization. The GWD scheme applied in ECHAM-4 is a modified form of that described in Miller et al. (1989). It consists of a specification of the low-level drag which may propagate upward under stable conditions. A local Richardson number is computed to determine the level where the drag is reduced due to wave breaking (c.f. appendix B). A short integration of six perpetual Januaries was performed with the GWD switched-off in order to get a quantitative idea on the importance of this parameterization on the ECHAM-4 large scale flow.

The impact of the GWD parameterization on the northern hemispheric pressure distribution is shown in Fig. 7.1. The maps to the left display mean sea level pressures and monthly standard deviations from twenty perpetual Januaries of the control simulation CM, while the right hand panels display the sea level pressure averaged over the last five Januaries of the ‘no-GWD’ experiment and the pressure differences CM−NGWD. The centers of action apparent in the pressure distribution respond to the GWD by intensification in the Asian-Pacific sector but by weakening over the Atlantic. Without GWD the Asian anticyclone is weaker by some 5 hPa than in the standard model, where the high exceeds 1040 hPa. Likewise, with a central pressure of 997 hPa, the Aleutian low is stronger in the control simulation than in the NGWD experiment. The Icelandic low, on the other hand, is deeper by 10 hPa when no GWD is operating. As the Azores high is also slightly intensifed, the flow over the northeast Atlantic is much stronger in the NGWD simulation than in the standard model. The pressure differences CM−NGWD in Fig. 7.1 are characterized by a pressure increase due to the GWD parameterization of up to 14 hPa around the north polar cap. A pressure drop of 8 hPa occurs over the North Pacific and a weaker one over the Mediterranean. The differences are of the same order of magnitude or larger as the January mean standard deviations for large areas of the middle to high
latitudes. The monthly standard deviation at every grid point is

$$\sigma = \left( \frac{1}{N} \sum_{i=1}^{N} (p_i - \bar{p})^2 \right)^{1/2}, \quad (7.1)$$

where $p_i$ is the monthly mean pressure of January $i$, and $N = 20$. The intermonthly variance is largest over the Bering Sea and the polar Atlantic, i.e. north of the two main blocking regions.

Figure 7.1: January mean sea level pressure for the CM (left) and the NGWD (right) simulation, together with the monthly standard deviation of the CM run (bottom left) and the pressure difference CM – NGWD (bottom right). The contour interval for the standard deviations and the differences is 2 hPa.

The effect of the GWD parameterization at the 200 hPa level is displayed by the geopotential height contours and the temperature difference in Fig. 7.2. The GWD parameterization alleviates the zonalization apparent in the upper tropospheric heights of the NGWD
simulation and leads to more realistic horizontal shears in the trough-ridge transition regions. The large temperature increase of up to 8 K in the polar lower stratosphere indicates that the GWD parameterization is successful in disrupting the polar vortex which prevails in the NGWD experiment. The impact of the GWD parameterization on the zonal mean winds and temperatures are shown together with the no-mountain simulations in Fig. 7.4 and Fig. 7.5.

In conclusion, the GWD parameterization is essential in shaping the northern hemispheric mean flow. The present results are similar to the impact found with a simpler parameterization by McFarlane (1987).

Figure 7.2: January geopotential height at the 200hPa level for the CM (left) and the NGWD (right) simulation. The height difference map CM – NGWD (bottom right) is displayed aside the temperature differences (bottom left) for the same level. The contour interval for the differences is 1 K and 40 gpm for the temperature and the height field, respectively.
7.2 No-mountain experiments

The effects of large-scale orography on the atmospheric circulation and climate in general have previously been investigated by Bolin (1950), Manabe and Terpstra (1974), Smith (1979), Held (1983), Jacqmin and Lindzen (1985), Nigam et al. (1988), Kutzbach et al. (1989), Broccoli and Manabe (1992), and others. One general conclusion from these comprehensive studies is that orography, in addition to thermal land-sea contrasts, shapes the stationary planetary waves of the winter troposphere. Much expertise on the representation of mountain barriers in numerical models has also been gained at weather forecast centers (Jarraud et al. 1988). Wallace et al. (1983) found that the introduction of an 'envelope' orography reduced the errors of the winter forecasts. The reduction seems to be largest in the two preferred regions of atmospheric blocking over the northeastern parts of the Atlantic and the Pacific oceans (c.f. Fig. 3 of Tibaldi 1986).

The link between the orography and blocking is threefold. Firstly, mountains represent large-scale roughness elements which decelerate the zonal flow. Reduced westerlies, however, favour the development of cutoffs and westward propagation or stationarity, as is indicated in equation 1.3. Secondly, mountain ranges are a source for planetary waves. The importance of amplified stationary waves for blocking may be inferred from the southern hemisphere, where the orographical forcing is much weaker. Blocks still occur in the more zonal conditions of the southern hemispheric midlatitudes but they are less frequent and they do not persist as long as their northern hemispheric counterparts (Tibaldi et al. 1994). Thirdly, the role of the mountains may be less direct and more of a 'catalytic' nature (Hansen and Sutera 1995). Performing idealized GCM experiments with a single mountain, Yu and Hartmann (1995) recently found a strongly nonlinear dependence of the low-frequency variability on mountain height. More specifically, Mullen (1989, 1994) showed that orography 'greatly aids the blocking process in the model but is not crucial for the existence of model blocks.' The objective in what follows is to test this conclusion which is based on low-resolution experiments (R15) with a finer general circulation model, the ECHAM-4/T42. The remainder of this section is organized as follows. After a description of the experimental design, the changes of basic climatological fields in response to the removal of mountains are examined. Then, two methods to find blocks and their number of occurrence are discussed, followed by a summary.

Figure 7.3 shows the model coastlines and the orography, which is determined by the mean elevation of 10' data within each 2.8° x 2.8° box of the corresponding Gaussian grid. At this horizontal resolution, the Rocky Mountains for example reach heights above 2000 m, while the highest Alpine peaks are smoothed to 900 m.

Three different experiments are described in the following: a control perpetual January run (CM), a simulation with reduced orographical forcing (RM), and one with completely flat topography (NM). In the RM experiment, topographical heights above 500 m are truncated, the subgridscale orographic variances and the roughness length is held below
Figure 7.3: Topography and land-sea distribution of the T42 model (top). CM–RM elevation difference in the stereographic projection used for the following plots (bottom left). The contour interval is 500 m. Contours at 100, 500, and 1000 cm of the surface roughness length are also shown in the bottom right panel.
specified values, and the truncated orography Fourier transformed forward and backward to avoid edges. To remove the effects of mountains completely, the topography is set to zero everywhere, the gravity wave parameterization switched off, and only the vegetation part of the roughness length retained in the NM integration. The total surface roughness length shown in Fig. 7.3,
\[ z_0 = \left( z_{0,\text{veg}}^2 + z_{0,\text{oro}}^2 \right)^{1/2}, \]
consists of a vegetation and an orographic part (discarding the urban term). The 'oro' part, which is determined by the subgrid scale orographic variance and the number of significant ridges in the grid box, is dominant in the mountainous regions, where it attains values above 10 m, while the roughness length due to vegetation does not exceed 1 m for the midlatitude forest regions (Claussen et al. 1994).

All integrations start from the same initial atmospheric state and are run for 660 model days under radiative conditions characteristic of January. During the first 60 days, which are discarded for the analysis, the model adjusts to the respective boundary conditions. The sea surface temperature and sea ice distributions are fixed at January climatological values of the period 1979–1988. They are the same for each experiment and there is no adaption to a different no-mountain climate. Soil moisture and snow cover are not held constant but are determined by the model atmosphere. To justify this approach, a brief comparison of the CM run with another simulation including the seasonal cycle was undertaken. No significant drift in the general circulation occurs during the 22 perpetual Januaries. The accumulation of snow and the corresponding increase in the surface albedo is considered to have a negligible effect on the upper air flow compared to the effect caused by removing mountain ranges.

7.2.1 Comparison of the model climatologies

In this subsection a few aspects of the mean states and the storm tracks of the simulations with and without mountains are compared. The focus is on the atmospheric circulation of the northern hemisphere. No attention is given to diabatic processes, although the large scale precipitation and condensational heating distributions are strongly affected by orography.

Zonal means

Figure 7.4 shows the latitude-height distributions of the zonal wind component. The difference between the CM simulation and ECMWF analyses for the winters 79/80–88/89 is small in the troposphere except near the equatorial and subtropical tropopause. These too strong westerlies are also present in seasonal cycle simulations and are not attributable to the perpetual January conditions alone. The changes in the RM run, in which only the high mountains are removed, are small throughout the troposphere and lower stratosphere but one recognizes the typical pattern of systematic errors with a poleward shift of the
jet in the northern hemisphere. Relatively minor changes in zonal mean fields were also found in other no-mountain experiments, e.g., by Lindzen (1986) or Mullen (1989). With completely flat continents, however, the surface winds intensify markedly, the jet core broadens, and the westerlies poleward of 40°N become stronger. The effect on the zonal wind in the NM experiment is more than twice of that of the RM run. The impact in the no-GWD simulation, also shown in Fig. 7.4 is very similar to the RM simulation up to about 300 hPa in the northern hemisphere. The lack of gravity wave drag results in larger wind changes than reducing the mountain heights in the lower stratosphere. This is consistent with the strong cooling of the Arctic lower stratosphere induced by switching...
off the GWD (Fig. 7.5). The warming due to the subgrid-scale orographic wave drag parameterization has a maximum of 14 K over the pole at 50 hPa, which is the uppermost level shown. This value may be compared with the 8 K at the 100 hPa level found by McFarlane (1987) in T21 simulations with the Canadian Climate Centre model. While this warming is considerably smaller than the one found in the present ECHAM-4/T42 simulations, the changes in the zonal mean winds are more similar in the two models.

Figure 7.5 also shows the temperature impact due to the mountains. The presence of the orography warms the model atmosphere by 20 K at 50 hPa over the pole and cools the zonal mean tropospheric temperatures by up to 4 K in the midlatitudes of the northern hemisphere. The completely flat continents have thus a considerably larger impact on the zonal means than the GWD parameterization alone.

Sea level pressure

The main features of the sea level pressure distribution are captured realistically in the simulation with mountains (Fig. 7.6). The Aleutian and Icelandic lows are well positioned and their minimum pressures agree with observations when taking into account the considerable interannual variability in both observed and modeled fields. The Siberian high is very strong with central pressure above 1040 hPa. Pressure over the Iberian peninsula is also higher than observed values by approximately 5 hPa.

With flat continents, shown in the lower right panel of Fig. 7.6, the Asian anticyclone is weaker, has retreated equatorwards and assimilated to the other two high pressure cells on the 35°N latitude circle. The Icelandic low is much deeper and becomes the dominant feature of the pressure distribution. The Aleutian low, on the other hand, weakens in the absence of mountains. While the deepening of the Icelandic low is a consequence of the enhanced heat transport due to transient eddies in a circulation with much smaller stationary waves than the actual winter circulation, the behaviour of the Aleutian low needs a different explanation. The presence of the Tibetan Plateau leads to a southward shift and a concentration of the jet over Asia and most of the Pacific as can be seen from the 200 hPa zonal winds in the global maps of Fig. 7.7. Thermodynamics require that
the more intense jet is accompanied by an enhancement of the baroclinicity. Together with the outflow of cold air on the eastern flank of the Siberian high over the warm sea off the Chinese coast, enhanced cyclogenesis is induced in the mountain case. Still, the asymmetry between the Aleutian and Icelandic lows in the NM pressure pattern is remarkable. Previous no-mountain experiments with lower resolution models (Tokioka and Noda 1986; Ruddiman and Kutzbach 1989) showed similar changes but not such a strong response as is simulated by ECHAM-4/T42. Even with reduced mountain forcing, the lower left panel of Fig. 7.6, the changes in sea level pressure are larger than those reported in older no-mountain studies. The global mean difference in sea level pressure between the CM and the NM (and also the RM) integration is added as a constant to the NM integration for a better comparison in Fig. 7.6. The reduction to sea level leads to higher pressure values in the mountain case which has the same total atmospheric mass as the flat experiment.
Figure 7.7: Zonal wind at the 200 hPa level for the CM simulation (top) and the NM simulation (bottom). Topography above 1000 m is indicated by shading. Contour intervals are 8 m s$^{-1}$.

**Geopotential height**

The geopotential height field at the 500 hPa level (Fig. 7.8) of the mountain experiment shows the well developed winter stationary waves with the two major troughs downstream of the Rocky Mountains and the Tibetan Plateau. Further downstream, ridging over Alaska and the Atlantic can be seen. The root-mean-square deviations of the geopotential height, filtered to retain fluctuations in the band 2.5 to 6 days according to Blackmon (1976) are indicated by shading on the same figure. The model captures this measure of the synoptic scale disturbance activity with remarkable skill in the two major storm track regions (c.f. Fig. 23 of Roeckner et al. (1996) which displays the bandpass filtered
standard deviations of the ECMWF analyses for the winters 1980–1992). Both the Pacific and in particular the Atlantic storm track bend northward at their exits, following the relatively strong ridges of the mean geopotential height.

Figure 7.8: Geopotential height at 500 hPa; rms deviations of bandpass filtered fluctuations indicated by shading starting at 20 m and darkening every 10 m. Placement as in Fig. 7.6.

The flow in the NM integration is much more zonally symmetric with a wavenumber one departure from this symmetry. Also the rms of bandpass filtered heights show much reduced longitudinal variation. Only the Atlantic storm track stands out against the uninterrupted band of high synoptic scale activity around 50°N. The Pacific storm track, which is anyhow weaker during midwinter, has disappeared by the removal of the mountains. This opposite behavior in the two sectors is consistent with the discussion on the sea level pressure in the previous subsection, except that the Atlantic storm track is not shifted southwestwards by the presence of orography as is the case for the Icelandic low. The stationary planetary waves are still present under reduced mountain forcing but with smaller amplitude than in the CM simulation (Fig. 7.8). The Pacific and Atlantic storm tracks are
still separated. In summary, the variability on the subweekly time scale is weaker in the CM than both the NM and RM simulation, indicating that the poleward heat transport is redistributed partly from the transient to the quasi-stationary eddies by the presence of mountains.

The difference of geopotential height at 200 hPa (Fig. 7.9) shows higher values over the polar cap for the CM and RM simulation than for the NM experiment, even after subtracting the global mean difference of the two respective fields. The 200 hPa level has been chosen for better comparison with the work of Blackmon et al. (1987). The pattern at the 500 hPa level is the same but the differences are smaller in the mid-troposphere than near the tropopause. The anomaly of the above mentioned work has the same order of magnitude as the present RM experiment. The impact in the true NM simulation is much stronger, especially over the North Atlantic ocean where geopotential heights rise by up to 400 gpm. Also, a strong wavenumber three component is apparent in Fig. 7.9. Whether this response is due to the European mountains or is simply a far distance effect of the two major orographical obstacles in the midlatitudes of the northern hemisphere is not clear. The large surface roughness length of 16 m in the Alpine region of the T42 model (Fig. 7.3) gives some credence to the first hypothesis while experiments by Yu and Hartmann (1995) with a single mountain support the second view, as a weak secondary trough was simulated some 180° downstream of the main trough.

### 7.2.2 Blocking activity

Figure 7.10 shows Hovmöller diagrams of blocking candidates for the three experiments while the percentage of blocked flow is summarized in Fig. 7.11. The observed blocking...
frequency as a function of longitude is added in Fig. 7.11 for comparison. The values are computed from daily NMC analyses, interpolated to a $5^\circ \times 5^\circ$ grid, yielding almost the identical curve as was obtained originally by Lejenås and Økland (1983) from Fourier coefficients of the geopotential height data. For the model data, archived twice daily, the following index is used:

$$I(\lambda) = \tilde{Z}(\lambda, 40.5^\circ N) - \tilde{Z}(\lambda, 60.0^\circ N), \quad (7.3)$$

where the tilde denotes a centered average over 11 longitudes ($28.125^\circ$) of the Gaussian grid on which the 500 hPa geopotential height $Z$ is given.

The fact that the ECHAM models have reasonable skill in reproducing the observed blocking distribution is confirmed in the present perpetual January experiment. With control mountains, the two blocking regions of the northern hemisphere are well captured. The activity in the Pacific sector matches the observed values while too few model events occur in the Euro-Atlantic region. Also, the separation near $90^\circ E$ of the two sectors in the ECHAM-4 simulation is not as marked as in the observations. In contrast to the described experiment, an older ECHAM-3 simulation with fixed SSTs showed a strong underestimation in the Pacific sector but performed somewhat better in the European area. Whether this is a sampling problem or attributable to the changes in the model is not known.
Figure 7.11: Blocking frequency as a function of longitude for the NMC analyses of January 1950–1979 (thin solid), the CM (solid line), the RM (dashed), and the NM (dash-dotted) simulation.

Table 7.1: Blocking events of the three perpetual January experiments each 600 days long. An event is counted if the respective sector is ‘blocked’ in the sense of Lejenäs and Økland for at least five days without interruption.

<table>
<thead>
<tr>
<th>Pacific sector, 150°E–120°W</th>
<th>Atlantic sector, 60°W–30°E</th>
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<tr>
<td>events</td>
<td>avg. duration (days)</td>
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<td>CM</td>
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</tr>
<tr>
<td>RM</td>
<td>13</td>
</tr>
<tr>
<td>NM</td>
<td>3</td>
</tr>
</tbody>
</table>

In the absence of topography, no major blocking event occurs during the 20 Januaries (Figs. 7.10, 7.11). The longest event, in the last month of the NM run, lasts for 10 days but is migrating eastward by approximately 10° per day. With reduced topographical forcing, the blocking frequencies as determined by the original Lejenäs and Økland criterion are nearly zero in the European-Atlantic sector whereas a few strong blocking episodes occurred over the North Pacific. Table 7.1 compiles the number of events with a persistency greater than 5 days and their mean duration for the three experiments separated by sector.

The discussion will be continued after the next subsection where the occurrence of positive persistent anomalies is examined. The application of two different blocking criteria allows a more reliable reasoning and gives information on the particularities of each method.

7.2.3 Persistent anomalies

Dole and Gordon (1983) studied persistent anomalies of the height field in general and associated blocking with the cases that exhibited a positive anomaly exceeding some threshold
value (~150 m) for at least 5, 10, ... days. In addition to the well-known regions of blocking, they found northern Russia to be a preferred location for winter blocks. May (1994) investigated the intraseasonal variability of the ECHAM-3 model and found good agreement in the number of positive persistent anomalies (PPAs in the following) in both fixed SST and variable SST simulations compared to ECMWF analyses. He applied a threshold value dependent on the season and required a lower persistency limit of 9 days. Fig. 7.12 shows PPAs that exceed 100 m without an interruption for at least 9 days for comparison with the above-mentioned work. The anomalies are calculated by subtracting the grid point mean value of the 20 Januaries and multiplying the anomalies by $\sin 45^\circ / \sin \phi$. This latitudinal scaling removes the high latitude bias of the geopotential height anomalies. No time filter is applied to the data, which may be justified by Dole and Gordon's (1983) observation that lowpass filtering hardly affects the statistics of PPAs (it does so for the negative anomalies). Both maxima in the eastern Pacific and eastern Atlantic of Fig. 7.12 indicate that approximately 12% of the days are 'blocked'. Two minor maxima are located over northern Russia and the Baffin Bay. Only the first maximum is also present in the observations (c.f. Fig. 3 of Dole and Gordon 1983) which show small numbers of PPAs over the northeastern parts of Canada.

Figure 7.12: Positive persistent anomalies (100 m, 9 days) in percentages. Placement as in Fig. 7.6. The contour interval is 2%.
The RM and NM distributions of 100 m, 9 day anomalies (Figs. 7.12) are quite different from one another. Compared with the CM simulation, the occurrence of PPAs increases in the Pacific sector and remains almost constant over the North Atlantic for reduced mountain forcing. With completely flat continents, the activity in both blocking regions has weakened while the third maximum over Siberia has become stronger, resulting in a more zonal distribution of PPAs.

Also shown are PPAs that exceed 250 m for at least five days (Fig. 7.13). With such a large threshold, which was chosen for a comparison with the work of Sausen et al. (1995), the number of PPAs is not only reduced but also shifted westwards in the Atlantic sector. Comparing the three experiments, it is found that the RM simulation experienced the largest number of large amplitude anomalies while fewer cases occurred in the control and the no-mountain sample.

Figure 7.13: As Fig. 7.12 but with (250m, 5 days) as selection criteria and a contour interval of 1%.

Discussion of results of the two methods

Considering only the results of the NM simulation analysed by the Lejenäls and Økland (LO) index, one would conclude that mountains are necessary for the formation of blocks.
in both sectors. This conclusion, however, needs some elaboration when also the RM experiment is examined and when the PPAs are taken into consideration as a second measure of the blocking activity. The situation is further complicated by the fact that changing the threshold and persistency of the PPA method within reasonable limits may lead to qualitatively different answers. In the Atlantic sector, for example, the activity decreases by a complete removal of the mountains when looking at 100 m, 9 day anomalies but increases for the parameters 250 m, 5 days. Inspecting consecutive five day mean charts and also the timeseries at selected points (not shown), this behaviour can be tracked down to one prolonged episode that is marked by a strong, broad ridge over the North Atlantic ocean. Fig. 7.14 shows a ten day window of this event. The first pentad is characterized by a broad ridge that is not detected by the LO index. The mean of the following five days, however, indicates that the wave is about to overturn leading to eastward winds to the south of the blocking anticyclone which is recognized by the LO method (c.f. Fig. 7.11, days 259 – 262). This case also illustrates that it is the individual type of development that determines whether the pattern is captured by one of the methods. By construction, the LO index finds ‘omega’ and dipole blocks within the specified latitude band but misses strong ridges that do not overturn. These latter flow patterns contribute to the PPA counts in a dominant way.

Fig. 7.15 compares the two methods for the CM simulation. Only cases fulfilling the LO criterion for at least five days at a fixed longitude are considered for the LO blocking frequencies. The PPA frequencies as a function of longitude are obtained by averaging the 200 m, 5 day anomalies in the latitude band 46 to 66°N, which is emphasized in Figure 7.16. In the European-Atlantic region, the PPA maximum is at 40°W while the maximum of the LO method is near the Greenwich meridian. The shift of the two maxima is in the
Figure 7.15: Comparison of the two methods. Persistent anomalies (200 m, 5 days) in the latitude band indicated in Fig. 7.16 (full line) versus Lejenäs and Økland’s criterion with additional persistency requirement of 5 days (broken line).

other direction for the Pacific sector. Taking into account the different sensitivity of the two chosen methods with respect to the blocking type, it may be concluded for the European-Atlantic sector that strong blocking ridges are favored over the North Atlantic ocean, while dipole blocks are typical for European blocks with the high pressure cell over Scandinavia. Coming back to the RM run, the enhanced PPA activity is due to the more frequent occurrence of broad ridges in both sectors.

7.2.4 Summary

Continental elevations have been removed in perpetual January experiments with a GCM that has relatively small systematic errors compared to previous studies of the same type. A first simulation with reduced topographical forcing, where all mountains above 500 m
are truncated, shows moderate changes in basic climatological fields. The response in the distributions of the zonal wind, sea level pressure and geopotential height, however, is very strong in a second true no-mountain experiment. The bandpass filtered variance increases in the RM as well as in the NM run. A concentration into one single storm track over the Atlantic ocean together with a very deep Icelandic low is found in the NM simulation. The Aleutian low on the other hand intensifies in the presence of topography.

The experiments were performed to get a better understanding of atmospheric blocking processes and to check whether blocking-like flows develop even when the planetary scale stationary waves are small. Applying the Lejenås and Økland index, it is found that no long lasting (more than six days) event occurs in the twenty Januaries of the NM experiment. With truncated orography, the blocking activity is reduced in the European-Atlantic sector but not over the North Pacific ocean. To test the robustness of these results, positive persistent anomalies are also examined. This method is less sensitive to the type of development but the patterns may differ from what is synoptically recognized as blocking. The greater percentages in the RM and the small reduction for the NM simulation compared to the control experiment are due to broad ridges, which are counted as persistent anomalies. These conclusions are consistent with the earlier findings of Mullen (1989) that orographic forcing greatly aids the blocking process but is not necessary for blocking to occur.

Finally, it is important to keep in mind the limitations inherent in the experimental setup. Bigger sample sizes from longer runs would give a firmer statistical basis. Also, it would be desirable to perform integrations including the seasonal cycle and experiments where individual mountain ranges are removed to study the impact of the Rocky Mountains or Greenland separately. Such model experiments would help to better understand the characteristic annual marches of the blocking activity and the partly different mechanisms responsible for the breakdown of the westerly flow in the Euro-Atlantic and the Pacific sector.
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8 A multidecadal experiment

The simulated North Atlantic oscillation (NAO) index and blocking frequencies from a twentieth century integration are briefly analysed in this section. In this type of experiment, sea-ice cover and sea surface temperatures are updated monthly according to the GISST observational data set (Rayner et al. 1996). A history of greenhouse gases is also included but other forcings (e.g., aerosols) are not altered so that they are represented only indirectly by the SST.

8.1 The North Atlantic oscillation

The time series of the mean sea level pressure at two grid points corresponding to Lisbon and Stykkisholmur, Iceland, respectively, are taken as a starting point. Pressure values averaged over the four months December (of the previous year) to March of each year are displayed in Fig. 8.1. The correlation between pressure anomalies in Portugal and Iceland is larger in winter than for the other seasons. In winter it is also slightly larger for these two locations than for the pair Ponta Delgada, Azores and Akureyri, Iceland (see Hurrell and van Loon (1997) and references therein for a description of the NAO). The correlation coefficient for the January to March season is -0.75 in the observations. This figure compares favourably with the simulated correlation of -0.76 for the December to March season. The NAO seesaw is thus well captured in the model.

![Figure 8.1: Time series of the winter (defined as DJFM) mean sea level pressure for the grid points corresponding to Lisbon and Stykkisholmur, respectively.](image)

Not reproduced by the model, however, is the decadal variation of the normalized NAO index displayed in Fig. 8.2, which shows the time series of the observed NAO index versus
Figure 8.2: *Observed versus simulated NAO index of 1904–1994. Fluctuations shorter than 4 years are filtered out in the heavy lines.*

The simulated one for the period 1904 to 1994. The index is defined as in Hurrell (1995):

\[ I_{NAO} = \frac{p_L}{\sigma_L} - \frac{p_S}{\sigma_S} \quad (8.1) \]

where \( p_L \) is the December through March sea level pressure anomaly at Lisbon of each year, and \( \sigma_L \) is the long term mean standard deviation (1864 to 1994 for the observations) at this station. The subscript 'S' stands for Stykkisholmur. The standard deviations of the simulation for the period 1904 to 1994 are \( \sigma_L = 3.1 \) hPa, and \( \sigma_S = 4.6 \) hPa for the grid points corresponding to Lisbon and Stykkisholmur. While the observed index has a pronounced upward trend starting in the seventies, such a trend is lacking from the simulation which displays the strongest break in the fifties and sixties. This break,
which is also visible in the timeseries of Fig. 8.1 as a change from relatively moderate variability to stronger fluctuations in the Icelandic pressure series, may be attributed to the inhomogeneity of the GISST2 data set. In particular, no variability in the ice margin is included in this ‘perfect’ boundary data set prior to 1949. The main point to be made is not the inhomogeneity of the underlying sea ice cover and SSTs but the absence of any significant correlation (0.15) between the simulated and observed time series, and in particular the missing of the upward trend towards the end of the period considered. The fact that the atmospheric GCM, driven by a perfect history of SSTs, does not reproduce the observed variability in the North Atlantic region reinforces the view that the atmospheric dynamics in middle latitudes sustains a high level of internal variability which is little affected by SST anomalies even on a decadal scale. It would be interesting to perform an ensemble of simulations with slightly changed initial states to assess the regional potential predictability (May and Bengtsson 1996), and to determine the strength of the ‘atmospheric bridge’ linking SST variations in the tropical and extratropical oceans (Lau 1997).

Interestingly, the winter 1995/96 had a very low NAO index. What will happen to the NAO index in the coming decades, and will the stronger SST forcing in coupled scenario simulations have a more deterministic effect on the NAO index are two naturally arising questions of relevance for the future climatic conditions in Europe. In the recent past, the North Atlantic surface temperature increased by some 0.4 K from 1920 to 1940 without influencing the NAO in a clear manner. To anticipate no marked changes for the coming century by analogy, however, is not appropriate since the nonlinear behaviour makes surprises likely.

Figure 8.3: High index (> 1) minus low index (< −1) sea level pressure difference. The contour increment is 2 hPa.
To give an idea of the climatological effects of the NAO phases, composites are formed from the winters with an index larger than one in absolute value. The high minus low index winter pressure pattern, shown in Fig. 8.3, is very similar to the observed difference pattern displayed in Hurrell (1995), with the exception that the model gives higher pressure in the North Pacific during high NAO index winters than during low index winters. This balance between the Icelandic and Aleutian lows is not in the observations. It seems as though the link between the North Pacific and the North Atlantic is too strong in the ECHAM-4 model (c.f. Table 2 of Roeckner et al. 1996). The pattern of Fig. 8.3 is also similar to the pressure difference NGWD – CM in Fig. 7.1. Whether this similarity is due to subgrid scale orography at a particular location (e.g., northeastern Canada and southern Greenland) or is due to shifts in the zonal mean wind (Fig. 7.4) needs further investigation. In any case, the GWD reduces the zonalization over the Atlantic and leads to a lower NAO index pattern.

The North Atlantic dipole pattern which is well simulated results in increased advection (of moisture) towards Scandinavia and dry conditions over the Iberian Peninsula during high index winters. Less clear are the effects of extreme NAO winters on the Alpine region which lies either on the southern edge of enhanced storm activity during high NAO winters or on the northern border of increased rainfall during the low phases. As will be discussed below, negative index winters can be associated with enhanced blocking action and cold conditions while the advection of temperate maritime air for positive NAO winters implies relatively warm conditions for the Alpine region.

The pressure and 500 hPa mean flow patterns for winters with an index larger than ±2 in Fig. 8.4 show a much reduced and westward shifted Icelandic low for the negative phase of the NAO. In all but two of the 15 winters with a NAO index less than -2, at least one blocking event occurred during the respective four months. In the remaining two winters, a diffluent 500 hPa flow was noticeable in the respective monthly means, similar to the mean negative NAO geopotential pattern displayed in Fig. 8.4.
Figure 8.4: Averaged pressure (top) and 500hPa geopotential pattern for winters (DJFM) with a NAO index greater than 2 (left), respectively less than -2 (right).
8.2 Variations of the blocking activity

A first diagnosis of blocking in the multidecadal experiment GM4 is displayed in Fig. 8.5, which shows the seasonal blocking frequencies in the same manner as in previous sections. Additionally, the simulated interannual variability of the seasonal activity is represented by the shaded band. The percentages from the NMC analyses and the model of the period 1950 to 1979 are also indicated by the broken lines in the same figure.

![Figure 8.5: Seasonal blocking frequencies according to the Lejenäs and Økland index for the entire GM4 simulation period 1903–1994 (full line), and the subperiod 1950–1979 (dash dotted). The dashed lines are percentages from the NMC analyses of the years 1950–1979. The interannual variability is indicated by shading (dotted lines) for the model (for the analyses).](image)

The T42 model driven by the observed SST of the period 1903 to 1994 generally has less blocking days than the NMC analyses of the years 1950 to 1979. The underestimation is more pronounced in winter and summer than during the transitional seasons. Differences between the simulated frequencies of the subperiod 1950 to 1979 and those of the entire experiment are largest over the Atlantic in spring (2% at 45°W, which is equivalent to approximately 2 days per season). The differences are on the order of 1% in both sectors for winter. The interannual variability, defined by one standard deviation of the seasonal frequencies, comes close to the observed variance for each of the four seasons.

Next, the blocking events with a minimum duration of five days are taken into account. An event is counted if the respective sector is blocked according to the Lejenäs and Økland
(1983) criterion for at least five days. A particular sector is considered as blocked if any longitude within the sector width of 90° has a negative index. To compare the timeseries of blocking with the NAO index, only the Atlantic sector during December through March is considered in a first step. Fig. 8.6 displays the accumulated duration of blocking events versus the NAO index for the observations of the period 1950 to 1988 and for the entire twentieth century simulation. The total number of blocked days of each DJFM season has been normalized by the standard deviation from the mean. The standard deviation is 10.7 days for the NMC analyses and 7.8 days for the model. The time series of the NAO index and the normalized blocking activity are anticorrelated to some degree. The correlation coefficient of the plain timeseries is $-0.43$ for the observations and $-0.42$ for the model.

Figure 8.6: Time series of the normalized number of blocked days in the Atlantic sector during DJFM versus NAO index (thin line) for the observations (top) and the GM4 simulation (bottom).
From the simple statistics presented so far, it seems that not all low NAO index winters are characterized by blocking episodes. The simulation for DJFM 1923/24, for example, has a NAO index of −4.3 but the LO method indicates no blocking event for this season. The mean sea level pressure pattern for these particular months displays an Icelandic low that is shifted to the west together with a relatively weak high to the south that does not extend over Spain. Also shown in Fig. 8.7 is the mean 500 hPa flow which displays strong diffluence over the eastern Atlantic.

A closer look at the individual monthly means shows that January contributes dominantly to the anomalous pattern. An omega flow pattern around a blocking high over Iceland characterizes the 500 hPa geopotential height field shown in Fig. 8.8. High pressure over northeastern Greenland prevails at the ground. The inability of the LO method to detect a blocking ridge that has a strictly south-north axis orientation is responsible in this particular case for the lack of blocking days despite the strongly negative NAO index. The low correlation noted above may thus be explained by the limitations inherent to the method used to find blocks. This caveat may be kept in mind for the following discussion of the annual blocking frequencies.

The number of simulated blocking events per year and their cumulated duration are shown in Fig. 8.9 for the Atlantic and Pacific sector. To minimize the aliasing introduced by events that begin in one year and end in the next, the year is defined as starting in the previous September when the observed blocking activity is minimal and ending in August. The number of events per year determined in this manner varies from zero to seven for both sectors. Similarly, when the cases are multiplied by their durations to obtain the total number of blocked days, the model shows large year to year variations in the range 0 to 60 days per year. These results are roughly in agreement with the observations which are
discussed in, e.g., Lejenās (1995). The latter study, however, determines blocking events from pressure maps to extend the time period to the early twentieth century and stratifies the results according to the four seasons so that a direct comparison is difficult.

In Fig. 8.10 the simulated number of blocked days per year is compared with the observed number for the period 1950–1988. With an average of 20.5 days per year, the model underestimates the observed number of blocked days by 37% in the Atlantic sector. The underestimation of the blocking activity is only 15% in the Pacific region where the model gives 17.8 blocked days while 20.9 days are blocked in the observations of the period 1950–1988. The mean duration of the blocking events is well reproduced by the model as can be seen in Table 8.1 which gives the average number of events per year as well as their mean durations for the model and the NMC observations. The numbers for the subperiod 1950–1988 when observations are available for comparison are nearly the same as for the whole simulation period.

Table 8.1: Number and average duration of simulated and observed blocking events. An event is counted if the respective sector is 'blocked' in the sense of Lejenās and Økland for at least five days without interruption.

<table>
<thead>
<tr>
<th></th>
<th>PAC (150°E – 120°W)</th>
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<td>events</td>
<td>avg. duration (days)</td>
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</tr>
<tr>
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</tr>
<tr>
<td>obs. (1950-88)</td>
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<td>8.22</td>
</tr>
</tbody>
</table>
Figure 8.9: Time series of the simulated annual blocking activity for the Atlantic sector (top) and the Pacific sector (bottom). The lower lines indicate the blocking events per year while the upper lines give the cumulated duration of the cases in days divided by five, the minimum duration of an event to be considered.
Figure 8.10: Time series of the simulated (dashed) and the observed (solid) annual blocking activity. Fluctuations shorter than 4 years are filtered out in the heavy lines.
To conclude this brief discussion of secular variations of the blocking activity, the autocorrelations for the two sectors are displayed in Fig. 8.11. The functions are calculated from the annual number of blocked days for lags up to eight years. The autocorrelation functions for both sectors drop close to zero for a one year lag indicating no persistency of the annual blocking frequencies. In contrary, the model has a tendency in the Atlantic sector for alternating behaviour; a blocking-rich year is often followed by a year with few blocked days. This behaviour is not present in the observations. Finally, the blocking frequencies for both the model and the observations are much less persistent than SST anomalies in either sector and it seems that the ocean surface temperature is a very weak controlling factor of the blocking activity (but remember Fig. 4.6). This conclusion is in line with the results of Liu and Opsteegh (1995), obtained with a quasi-geostrophic model.

Figure 8.11: Autocorrelation functions of the annual blocking activity.
9 Conclusions

Aspects of the ECHAM-3/4 high resolution experiments have been diagnosed in this thesis. The focus has been on a particular phenomenon of the midlatitudinal flow in the northern hemisphere that challenges the simulation capabilities of atmospheric GCMs.

**Model biases and blocking performance**

The ECHAM models capture the distinct annual cycles of the blocking activity in the two main blocking sectors of the northern hemisphere with some skill. Whereas a pronounced midwinter maximum is observed for the Pacific sector, blocking in the Atlantic sector is more evenly distributed throughout the year with a maximum in spring. Compared to the NMC analyses, much too few blocked days are simulated during the winter season for both models in both sectors. Only the ECHAM-3/T106 simulation attains nearly the observed winter blocking frequency in the Euro-Atlantic sector, albeit shifted to the east. A high pressure bias over Spain and too zonal flow over the northeastern Atlantic suppress the simulated blocking activity in this sector. Furthermore, the secondary maximum present in the observed January blocking frequencies near the longitude of Greenland is weakly simulated by ECHAM, and not overestimated as in other models (D'Andrea et al. 1996).

A considerable part of the severe underestimation in the North Pacific is attributable to the lack of ENSO variability in the simulations with climatologically prescribed SSTs as simulations with variable observed SSTs clearly improve on this deficiency. Figure 4.6 shows also an impact from ENSO on Europe. Moreover, the sophisticated parameterizations of the ECHAM-4 model result in a generally improved performance over the Pacific compared to ECHAM-3.

**Effect of horizontal resolution**

The present T106 simulations confirm that horizontal resolution can only partly cure biases of GCM simulations (e.g., Williamson et al. 1995). Despite a larger overestimation of the geopotential height gradient in midlatitudes, the T106 model is generally superior to the T42 resolution in terms of blocking frequencies. This gives strong support to the view that the interaction of the synoptic-scale eddies with the large scale flow is an important ingredient for any blocking theory. Case studies of four blocking events covering all seasons show model developments in great detail. The simulations reproduce many features of observed cases realistically (Sect. 5.6). Multiple incursions of low-latitude air, advected by strong southerlies, take place during the evolution of the blocking high, which is characterized by a lifted tropopause and quasi-uniform low potential vorticity. Cold air outbreaks support the development already during the onset stage, which may take 2 to 7 days. A derived quantity like potential vorticity profits much from the high resolution which can better represent steep gradients than the standard T42 truncation. Judging subjectively, the T106 model develops shear lines and cuts off structures near the tropopause at least as vigorously as the atmosphere.
Role of orography

Perpetual January simulations (T42) with no GWD, with truncated, and with completely removed orography have been conducted. The impact of the GWD parameterization in the northern hemisphere, e.g., on the zonal wind is found to be about half as large as the effect of setting the topographical heights to zero. The no-mountain simulations show that the quasi-stationary background flow strongly influences the blocking incidence. Significantly less blocking days occur in the more uniform zonal winds in the no-mountain experiments. Orography above a certain threshold seems sufficient to trigger large scale Rossby wave breaking resulting in dipole blocks. The truncated orography excites almost exclusively broad ridge events. Tanaka (1991) has shown that orography plays a catalytic role for blocking initiation, supported by baroclinicity. This is confirmed by the present study. The no-mountain experiments have also highlighted that different mechanisms are important in the Pacific and Atlantic sectors. A strong deepening of the Icelandic low but a weakening of the Pacific storm track are observed upon removal of the mountains.

Blocking index or robustness of results

The blocking definition of Lejenäs and Økland (1983) is used in the present study, except in Sect. 4.3 and in Sect. 7.2.3. This criterion is easy to implement and has been tested with analysis data. Using absolute geopotential height fields at fixed latitudes, however, makes the method sensitive to changes in the climatology. The simulated biases of the blocking activity are considerably reduced when a criterion is applied to the anomalies instead of the plain geopotential heights (Sect. 4.3). A geographical distribution of the 'blocking activity' is obtained in Sect. 7.2.3, where positive persistent anomalies are discussed. This method selects somewhat other flow situations than the LO index.

Predictability?

The examination of a Twentieth Century simulation with 'perfect' boundary conditions provided by the GISST data has shown that the extratropical atmosphere exhibits large interannual variability which is hardly affected by the SST anomalies of this century (except for ENSO signals which are discernible from the 'noise'). It is found that the simulated North Atlantic oscillation index is very weakly correlated with the observed time series of the period 1904 to 1994. Blocking occurrence, however, anticorrelates with the NAO index in winter in the simulations and in the observations.

It is suggested that the projected temperature changes in the North-American/Atlantic region, which result in a smaller longitudinal temperature contrast in winter, lead to a reduced blocking activity in the Atlantic sector in the coming decades (Sect. 4.2).

Given the nonlinearity of a phenomenon that is influenced by many agents (zonal mean flow, Rossby waves, individual cyclones, surface conditions) it is not surprising that blocking mitigates the atmospheric predictability of the first and second kind. An enhanced predictability due to blocking may only be expected on the time scale of a blocking high
(10 days). For shorter forecast times, the transition to blocking is difficult to capture. On longer time scales, all kind of nonlinearities influence the blocking climatology and affect the predictability of the second kind.

**Outlook**

The present study has touched many open questions that relate either to GCM behaviour or to lack of understanding of the phenomenon itself (causes for the initiation, the annual cycles of the blocking activity). Subsequent research may be undertaken in the following directions:

- further case studies with more powerful diagnostics (e.g., 'E-vectors')
- blocking analysis for the southern hemisphere
- further twentieth century integrations to analyse the predictability of, e.g., the NAO and blocking in an ensemble of simulations
- analysis of blocking in coupled simulations
- revision of the horizontal diffusion parameterization to better capture the energy cascade of geostrophic turbulence (Koshyk and Boer 1995)
- revision of the orographical drag in the model

The last two points aim at reducing the zonal bias of the ECHAM high resolution model over northwestern Europe in winter. A further mitigation would make the envisaged climate change simulations for the Alpine region more trustful.
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References


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A Horizontal diffusion in the ECHAM models

The lateral diffusion of any prognostic variable $X$ is treated as an implicit term in the time-stepping scheme of both model versions. Formally, an equation for the spectral component with zonal and total wavenumber $m$ and $n$, respectively, may be written

$$\frac{X_n^m(t + \Delta t) - X_n^m(t - \Delta t)}{2\Delta t} = \ldots - \Lambda_n X_n^m(t + \Delta t), \quad (A.1)$$

where the dots indicate all other terms except the horizontal diffusion. Solving for $X_n^m(t + \Delta t)$ shows that the diffusion is included by multiplying the dissipation-free result with the factor $(1 + 2\Delta t\Lambda_n)^{-1}$. A particular scheme is defined by specifying how fast $\Lambda_n$, the inverse damping time, increases with $n$.

ECHAM-3

Following Laursen and Eliasen (1989), the ECHAM-3 scheme damps only waves above a cut-off wavenumber $n_*$ by taking $\Lambda_n$ of the form

$$\Lambda_n = k_X L_n, \quad \text{where} \quad L_n = \begin{cases} (n - n_*)^\alpha & \text{for } n > n_* \\ 0 & \text{for } n \leq n_* \end{cases} \quad (A.2)$$

with $\alpha = 4$ and $n_* = 15$ (30) for T42 (T106) truncation. The diffusion coefficient $k_X$ varies for different variables and is enhanced strongly in the five uppermost levels.

ECHAM-4

Linear diffusion of the form $K \nabla^2 \phi$ (MacVean 1983) is applied to the vorticity, divergence and temperature tendencies in ECHAM-4 which treats the moist prognostic variables in a Lagrangian manner. The Laplacian is a simple multiplication in spectral space giving

$$\Lambda_n = K(n(n + 1)\alpha^{-2})^\theta. \quad (A.3)$$

Specification of the damping time $\tau = 1/\Lambda_n$ at the largest total wavenumber $n$ retained in the spectral truncation determines the diffusion constant $K$. Both the order of the scheme and the e-folding time $\tau$ are dependent on the horizontal resolution. A $9h \nabla^{10}$ diffusion is chosen for T42 while the T106 truncation uses $3h \nabla^8$. The diffusivity is enhanced by decreasing the order in the six (T42), respectively five (T106) uppermost model levels, which act as a sponge to arriving waves.
B  The GWD parameterization in ECHAM

Gravity waves excited by subgrid scale terrain variations influence the momentum budget according to
\[
\left( \frac{\partial v}{\partial t} \right)_{gw} = -g \frac{\partial \tau_{gw}}{\partial p},
\]
where \(v\) is the horizontal wind vector and \(\tau_{gw}\) the wave stress (Miller et al. 1989). The scheme, which is a modified form of that proposed by Boer et al. (1984), and by Palmer et al. (1986) consists of two parts:

(i) the parametric form for \(\tau_{gw}\) which determines the longitude/latitude distribution of the surface stress and

(ii) the modelling of the dynamical processes which determine the vertical distribution.

The modification includes an additional term that accounts for high-drag situations when resonant trapping of waves occurs. A low-level Froude number is calculated to this end,
\[
Fr = \frac{\sqrt{\text{var}}}{V_L},
\]
where \(V_L\) is the average wind speed of the lowest three levels and \(N_L\) the low-level Brunt-Väisälä frequency. The direction of \(V_L\) determines which component of the four available subgrid scale variances \(\text{var}\) is chosen. The directional variances are computed from an U.S. Navy dataset containing global surface information for 10' x 10' lat/lon areas. With the additional term \(\tau_{Fr}\) which is added when the flow is supercritical, i.e. for \(Fr > Fr_c = 2\), the scheme can be written as
\[
\tau_{gw}(p) = \tau_w(p) + \tau_{Fr}(p)
\]
\[
\tau_w(p) = \begin{cases} 
\frac{\tau_w(p_s)}{p - \frac{p_s - p}{p - p'}} \beta \tau_w(p_s) f(p) & \text{for } p > p' \\
\beta \tau_w(p_s) f(p) & \text{for } p < p'
\end{cases}
\]
\[
\tau_{Fr}(p) = \tau_{Fr}(p_s) \frac{p - \frac{p_z}{p} z}{p - p_z} \quad \text{if } Fr > Fr_c
\]
where \(f(p)\) describes the vertical stress profile, computed as shown in (ii) below. The parameter \(\beta\) controls the ratio of low to high-level drag (currently \(\beta = 0.3\)). The choice of \(p'\) determines the depth of the subcritical low-level drag (currently \(p' = 0.8p_s\)).

The top level for the supercritical drag, \(z_c\), is formally given by the equation
\[
\int_0^{z_c} \frac{N(z) U(z)}{U(z)} \, dz = \frac{3\pi}{2}.
\]
Typical values of \(z_c\) are around 3 to 5 km but much larger values do occur.
(i) The surface stress is given by

\[ \tau_{gw}(p_s) = K \rho_L v_L N_L \var^* , \quad (B.7) \]

where \( K \) is a tunable parameter. Currently, in ECHAM-4 \( K = 2.5 \cdot 10^{-5} \text{m}^{-1} \) for T42 and \( 3 \cdot 10^{-5} \text{m}^{-1} \) for T106 resolution. To avoid unrealistically large stresses near major mountain ranges a cut-off is introduced for the variance

\[ \var^* = \min \left( \var, \left( \frac{F_r v_L}{N_L} \right)^2 \right) . \quad (B.8) \]

Under supercritical conditions the following term is also needed

\[ \tau_{Fr}(p_s) = K L \rho_L \frac{v_L^3}{N_L} (F_r - F_r c)^2 f_2 \quad \text{if} \quad F_r > F_r c . \quad (B.9) \]

Here, \( K_L \) is chosen to be four times larger than \( K \) and \( f_2 \) is an orographic anisotropy function ranging from 0 to 1. \( f_2 = 0 \) if all four directional variances are equal; \( f_2 \to 1 \) for a terrain with mountain ridges and valleys oriented predominantly in one direction.

(ii) The vertical structure of \( \tau_{gw} \) is calculated by constructing a local wave Richardson number which attempts to describe the onset of turbulence due to the gravity waves becoming convectively unstable or encountering critical layers.

This wave Richardson number can be written in the form

\[ \tilde{R} = \frac{\tilde{R}(1 - \alpha)}{(1 + \alpha \delta z^2/U^2)} , \quad \text{where} \quad \alpha = N/|\delta z|/U . \quad (B.10) \]

\( \tilde{R} \) is the Richardson number of the basic flow, \( \delta z \) represents the amplitude of the wave and \( U \) is the projection of \( \mathbf{v} \) in the direction of \( \tau_{gw} \).

It is assumed in regions of instability, \( \tilde{R} < \tilde{R}_{crit} \), that the wave amplitude is decreased until \( \tilde{R} = \tilde{R}(\delta z) = \tilde{R}_{crit} = 1/4 \), i.e. until marginal stability is achieved. If \( \tilde{R} > \tilde{R}_{crit} \) then the value of stress is retained and computation proceeds to the next higher level.
C Acronyms

AMIP           Atmospheric Model Intercomparison Project
CCMx          Community Climate Model (NCAR), version x
CPU           Central Processing Unit
DJF           December, January, February
DKRZ          Deutsches KlimaRechenZentrum
ECHAM-x       European Center model with HAMburg physics, version x
ECMWF         European Center for Medium range Weather Forecasts
ENSO          El Niño/Southern Oscillation
GCM           General Circulation Model
GISST         Global sea-Ice and Sea-Surface Temperature data set
GRIB          GRid In Binary
GWD           Gravity Wave Drag
IPCC          Intergovernmental Panel on Climate Change
IPV           Isentropic Potential Vorticity
JJA           June, July, August
MAM           March, April, May
MPI           Max-Planck-Institut für Meteorologie, Hamburg
NAO           North Atlantic Oscillation
NCAR          National Center for Atmospheric Research
NMC           National Meteorological Center (now NCEP)
PNA           Pacific/North-American teleconnection pattern
PPA           Positive Persistent Anomaly
PV            Potential Vorticity
R15           Rhomboidal truncation at total wavenumber 15
SCSC          Swiss Center for Scientific Computing
SON           September, October, November
SST           Sea Surface Temperature
T106          Triangular truncation at total wavenumber 106
Curriculum vitae

1966 Born on October 2 in Münsterlingen (TG), Switzerland
1973–1981 Primary and secondary schools in Bottighofen and Kreuzlingen
1981–1985 Kantonsschule Kreuzlingen, Matura typus B
1986 Military service; England traveller
1986–1991 Study of physics at the Swiss Federal Institute of Technology in Zürich
1991 Diploma thesis in particle physics on ‘Fermion number violation in the O(3) sigma model’ under the supervision of Dr. Andreas Wipf and Prof. Dr. Ch. Schmid
1992–1997 Ph.D. student at the Institute for Geography, ETH Zürich, in the climatology group of Prof. Dr. Atsumu Ohmura; Project: global climate simulations with general circulation models
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