The impact of Greenland on cyclone evolution in the North Atlantic

By J. E. KRISTJÁNSSON* and H. McINNES
University of Oslo, Norway

(Received 6 November 1997; revised 1 February 1999)

SUMMARY

In order to understand better the role of Greenland’s orography in determining the position and strength of the Icelandic low, we have carried out a number of experiments using a numerical weather prediction model. In those experiments, Greenland’s orography was modified, and the impact on cyclone developments over the North Atlantic was investigated. We have focused on one fairly characteristic winter case from January 1995, but another one from January 1993 gave qualitatively similar results.

We have found evidence that the deepening of baroclinic cyclones near Iceland is hampered by the presence of Greenland’s orography. This has been shown to be related to the inability of cold air from the north to cross over Greenland, leading to a distortion of the thermal field associated with the disturbance and a halting of the progression of the cold front. In the January 1995 case, a characteristic secondary cyclone forming between Greenland and Iceland was shown to be entirely orographic, in that it was absent in runs where Greenland’s orography was removed. The results are considered in the context of recent theoretical studies of flow splitting in stratified flow impinging on high mountains.

KEYWORDS: Cyclone evolution Greenland Orographic effects

1. INTRODUCTION

The ‘Icelandic low’ is a prominent feature of the surface pressure pattern of the northern hemisphere. It is clearly seen, for instance, in monthly-mean surface pressure maps, particularly those for wintertime (see, for example, Fig. 6.18 in a paper by Hartmann (1994)). A priori, there may be several causes for the Icelandic low that we would wish to address in the present study. Its being considerably stronger in winter than in summer suggests that it owes its existence partly to thermal forcing, since in winter the North Atlantic Ocean serves as a reservoir of heat, while very cold surface air is being produced over Greenland, North America and Siberia. But, why is the low located SW of Iceland, rather than elsewhere in the North Atlantic? Strong gradients of sea surface temperature (SST) in the region where the warm North Atlantic Drift and the cold East Greenland Current are adjacent are probably relevant. One might also seek an explanation based on the upper-level orographically forced trough east of the Rocky Mountains. This trough, together with land–sea temperature-contrasts, gives rise to frequent cyclogenesis off the east coast of North America. These cyclones very often find their way north-eastwards, along the strong SST gradients, towards Iceland, contributing to the signature of a low there. But, Held (1983) showed, through a comparison of general circulation model (GCM) experiments with and without orography, that the Icelandic low was in fact much weaker with orography than without it. Due to the low resolution (R15) used in Held’s investigation, it was mainly the effect of the Rocky Mountains that was isolated. By contrast, it seems likely that the orography of Greenland may play an important role in explaining the position and persistence of the Icelandic low.

It is instructive to compare two figures (not shown) from a study by Petterssen (1956) of northern hemisphere cyclone activity. One (13.6.2) shows pronounced maxima in the frequency of cyclone centres in winter south-west of Iceland, south-west of Greenland and over the Gulf of Genoa. The other (13.6.1) shows the frequency of cyclogenesis for the same season: the maximum in the Icelandic area is much less pronounced and is surpassed in many regions, such as the Gulf of Genoa and the region downstream of the

* Corresponding author: Department of Geophysics, University of Oslo, PO Box 1022, Blindern, N-0315 Oslo, Norway.
Rocky Mountains. Furthermore, Blackmon et al. (1977), who calculated the 500 hPa variability and subjected it to frequency filtering, found that the Icelandic low was virtually absent in the dataset showing periods of 2.5 to 6 days, whereas it showed very distinctly in the low-pass-filtered data, with variability on timescales greater than 10 days. Ayrault et al. (1995) investigated variability on timescales shorter than 1.5 days in the North Atlantic, using the weather regimes of Vautard (1990), and showed that in the zonal regime there is a maximum in variability south-east of Iceland, whereas their corresponding figure for the 2 to 6 day variability shows maxima near 50°N, south of Greenland. From these sources, it seems that the Icelandic low, as seen on monthly averaged surface maps, is caused by a combination of quasi-stationary lows and rapidly moving, well developed systems.

In the present study, we seek to cast light on the influence of the Greenland orography on the Icelandic low. Among other things, we wish to explore the suggestion by Scorer (1988) that the Iceland low should really be termed the Greenland lee-low. Such a quasi-stationary lee-cyclone would be caused by persistent westerly flow crossing the mountains of Greenland and undergoing vortex stretching on the lee (eastern) side. Rather than resort to idealized flow calculations, the relevance of which might be difficult to demonstrate, we have used a numerical weather prediction (NWP) model, with which we have performed several controlled experiments for two selected cases of characteristic wintertime North Atlantic cyclone activity. In these experiments, the orographic heights over Greenland are either enhanced or reduced compared to the more realistic orography of a control run. In the case that we shall focus on in this report, the cyclone of interest forms in a baroclinic zone lying just along Greenland’s south-east coast. Hence, the cyclone formation is directly affected by Greenland. We investigate how both the cyclone formation and the subsequent development are affected by imposed changes to the Greenland orography. As a second case, we investigated one of the deepest North Atlantic cyclones this century (see, for example, Burt (1993)), but the main results from it were quite similar to those of the first case so we show results only from the first case. One motivation for undertaking the present study was that cyclonic developments near Greenland and Iceland are often poorly captured by NWP models (Ólafsson 1998). The resulting errors can have a large impact over Iceland in short-range forecasts, and over continental Europe in the course of a few days. Hence, it is of great importance for weather forecasting in this region to understand how Greenland affects the impinging airflow. We also expect the results of simulations with reduced orographic heights over Greenland to be of relevance for GCM simulations with coarse resolution, since in those models Greenland’s orography is not properly resolved.

In the following section, we review some existing theories dealing with airflow over mountains. Section 3 describes the model configuration used for the experiments that follow. The simulations of our selected case are described in section 4. Finally, a summary and conclusions are given in section 5.

2. Theoretical Considerations

Various theoretical approaches have been reported in the literature to explain the effect of mountains on atmospheric flow. Charney and Eliassen (1949) showed that, on the largest horizontal scale, of order $10^4$ km, topographic Rossby-waves seem to explain the existence of the major 500 hPa troughs in the lee of the Himalayas and the Rocky Mountains satisfactorily. However, as we shall see in section 4, the airflow over Greenland is on too small a horizontal scale for this theory to be applicable. Rossby waves are stationary only when $U = \beta/k^2$, where $U$ is the wind speed, $\beta$ the
change in Coriolis parameter with latitude, and \( k \) the zonal wave-number. This implies a wavelength \( L = 2\pi (U/\beta)^{1/2} \). With the low value of the \( \beta \)-parameter in the latitude of Greenland and observed wind speeds of \( U \approx 20 \text{ m s}^{-1} \), \( L \approx 9600 \text{ km} \), which is totally unrealistic for such a narrow mountain.

A seemingly more useful approach to understanding the influence of Greenland on the airflow is the regime diagram of Smith (1989). He uses the non-dimensional quantity \( Nh/U \), where \( N \) is the Brunt–Väisälä frequency and \( h \) the mountain height. With the aid of this parameter and the mountain aspect ratio, \( r \), (the ratio of the length scales in the across flow and along flow directions), four different flow-regimes are defined. The aspect ratio of Greenland depends on the flow direction. With north-westerly flow over southern Greenland, as in our study (section 4), we may estimate a value for \( r \) of about two. According to Smith, this means that flow splitting would be expected for \( Nh/U > 1.2 \) or so, giving rise to strong lee-effects, whereas, for smaller values of \( Nh/U \), one would expect the airflow to be able to cross over the mountain. We have found the average value of \( Nh/U \) to be considerably larger than the critical value most of the time, making the observed strong lee-effects theoretically likely.

The most serious weaknesses of the above theory for explaining the observed features are that the theory is inviscid and irrotational. Ólafsson and Bougeault (1997) studied airflows with different values of \( Nh/U \) when Coriolis and frictional forces were included. They found that friction strongly suppressed wave breaking, and that the combined effects of rotation and friction led to an extension of the range of usefulness of linear theory. Hence, the qualitative results of Smith should still largely apply in flows with rotation and friction. Another limitation of the theory of Smith (1989) is that, in the real atmosphere, both \( N \) and \( U \) can vary significantly with height, as indeed in our case, so that the flow may have the characteristics of more than one regime at the same time. Durran (1986) showed that sharp vertical gradients in \( N \) can have a dramatic effect on wave breaking, but, in the present study, we are more concerned with flow splitting. The vortex formation associated with flow splitting has been investigated in detail by Schär and Durran (1997), and some of their results for \( Nh/U = 3 \) appear to be relevant for our study.

Much of the literature on airflow over mountains deals with lee cyclogenesis, for example in the wake of the Alps. Many of these studies, Pichler et al. (1990) for example, explain lee cyclogenesis through conservation of potential vorticity (PV). Typically, a positive PV anomaly, propagating in the upper troposphere, interacts with the mountain in such a way that a cyclone is created on the downwind side while an anticyclone occurs on the upwind side. This would seem to indicate an interaction extending throughout much of the troposphere. On the other hand, some authors, for example Ólafsson and Bougeault (1996) and Tauffer (1990), have emphasized barrier effects and thermal effects in the lowest part of the troposphere. Tauffer (1990) suggests a scenario where, as an upper-level trough moves through, the mountain causes a retardation of the cold air, and a deformation of the low-level cold front. This retardation slows the movement of the trough and creates favourable conditions for lee-cyclone formation. Aeberharder and Schär (1998) found that when stratified air flow was directed across the Alps or Pyrenees, flow splitting caused streamers of high PV downstream. These streamers were found to contribute significantly to the evolution of lee cyclones. Thorpe et al. (1993) investigated westerly flow parallel to the Alps, and found that the Alps were then a significant source of PV to the south, and that this result depended crucially on friction effects. Although these studies may be relevant to
Greenland, in our present study strong cyclogenesis occurs even when the orography is removed, and a secondary cyclone behaves like a lee cyclone.

3. MODEL CONFIGURATION

We have used a version of the regional model NORLAM, which was used for operational weather forecasting in Norway until June 1996. It is a hydrostatic, primitive-equation grid-point model, with about 50 km horizontal resolution in an Arakawa D-grid, with 18 sigma levels in the vertical, and a rigid lid at 100 hPa. The time-integration procedure of Bratseth (1983) allows time steps of 90 s to be used. The physical parametrizations are described in detail by Nordeng and Rasmussen (1992). As main components, they include a Kuo scheme for moist convection, vertical diffusion based on \( K \)-theory, a simple soil parametrization due to Deardorff (1978), and simplified schemes for short-wave and long-wave radiation.

Figure 1 shows the 121 \( \times \) 97 \( \times \) 18 grid-point integration domain. Figure 2 shows the standard model orography with model orographic heights over Greenland well in excess of 3000 m over a large area, mainly in the eastern part of the interior. We shall refer to runs using this orography as CONTROL in what follows. Simulations where standard orographic heights over Greenland were multiplied by 0.5 will be termed HALFGREEN, simulations with Greenland's orography removed NOGREEN, and simulations in which all standard orographic heights over Greenland have been doubled will be termed DOUBLEGREEN.

Initial data for the simulations were obtained by interpolations to the model’s sigma surfaces from analyses on pressure surfaces produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). Clearly, the ECMWF analyses are affected by the presence of Greenland’s orography. In the runs with modified orography, we use those same ECMWF analyses, but perform a new initialization, as well as dry adiabatic adjustment and humidity adjustment. As a result, the model fields are adjusted to the new orography in a smooth manner, that does not create significant noise.

Below, we shall show several maps depicting mean sea-level pressure (MSLP), as well as 1000 hPa geopotential height (from now on, as is customary, mostly referred to as 'height') and 1000-500 hPa thickness maps. Over orography, these fields may be not exactly fictitious but certainly extrapolations relying heavily on convention, since surface pressure may be as low as 700 hPa over the highest mountains of Greenland in CONTROL, and even lower in DOUBLEGREEN. In these cases, the heights ZM at the 'missing' pressure levels are obtained by the following procedure. First, a vertical extrapolation from the lowest sigma-level is carried out, assuming an atmospheric lapse rate of 6 K km\(^{-1}\). Then, these fields are corrected by computing the vorticity (\( \zeta \)) at the lowest sigma-level in a subdomain around the mountain and then solving the Poisson equation \( \nabla^2 \psi = \zeta \), \( \psi \) being the non-divergent streamfunction, which is related to geopotential height through the geostrophic approximation \( Z = f_0/g \cdot \psi \) (\( f_0 \) being the Coriolis parameter at a reference latitude and \( g \) denoting the acceleration under gravity). Wind fields below the surface are then found by solving the equations \( u_\psi = -\partial \psi / \partial y \) and \( v_\psi = -\partial \psi / \partial x \).

4. SIMULATIONS OF THE 10–13 JANUARY 1995 CYCLONE

All the simulations start from 00 UTC 12 January 1995. In Fig. 1(a), we show the surface chart for twelve hours earlier, and in Fig. 1(b) that for twelve hours later. At 12 UTC 11 January (Fig. 1(a)), there was a broad baroclinic zone to the SW of Greenland,
Figure 1. Model analysis of 1000 hPa geopotential height (m) (solid lines) and 500–1000 hPa thickness (m) (pecked lines): (a) 12 UTC 11 January 1995; (b) 12 UTC 12 January 1995.
and a surface disturbance associated with this zone, with its centre located at 64°W, 55°N over the west coast of Greenland. The upper-level flow was south-westerly, and the whole baroclinic zone was being advected north-eastwards. However, Greenland’s orography exerted a strong impact on the surface cyclone as it approached Greenland. So that, rather than simply being advected with the background flow, the surface cyclone split into two, one near the west coast of Greenland that gradually weakened, and another along the east coast that can just be discerned in Fig. 1(a), but emerged as the main cyclone 24 hours later (Fig. 1(b)). According to Hsu (1987), this kind of splitting is quite characteristic of the area.

Figure 3(a) shows that, twelve hours after the beginning of the simulation, the CONTROL run had already started to diverge from the analysis (Fig. 1(b)), but still showed the same main features, i.e., a 983 hPa low near Greenland’s south-east coast at 65°N, 39°W, with a trough extending towards the north-east, located on the cold side of a broad baroclinic zone that extends all the way from Newfoundland towards Jan Mayen. During the NOGREEN run (Fig. 3(b)), the situation was already very different at this early stage. The low pressure centre was located about 650 km farther north-east, at 70°N, 31°W, with a trough extending towards the north-west into the interior of Greenland. This means that, although the upper-level conditions were almost the same in both cases, in CONTROL Greenland’s orography retarded the progression of the low. Interestingly, the surface pressure patterns in the HALFGREEN run (Fig. 3(c)) were roughly midway between those of CONTROL and NOGREEN, with a low pressure centre located at 68°N, 35°W. In DOUBLEGREEN (Fig. 3(d)), on the other hand, the
retardation of the cyclone was enhanced, with a stronger and more southerly position of the secondary cyclone than in CONTROL, as well as a pronounced troughing towards the NE.

Twenty-four hours later, the main baroclinic low had developed further and was just about to reach its mature stage (Fig. 4). The differences between the four runs that were seen in Fig. 3 are even more pronounced in Fig. 4. In CONTROL (Fig. 4(a)) there are now two low centres, the baroclinic low ("low A") near 72°N, and a quasi-stationary residual low ("low B") in the Denmark Strait, see Tables 1 and 2. Low A has moved north-eastwards with the upper-level flow, while low B is located only about 300 km ENE of the position 24 hours earlier. Low A is clearly connected with the frontal systems that have partly occluded near 75°N, 10°E (Fig. 4(a)). Low B, on the other hand, has no such connection to the ongoing baroclinic development, and, as will be seen, is related to orographic forcing by Greenland. This proposition is supported by the corresponding surface chart from NOGREEN (Fig. 4(b)), in which low B is completely absent, while low A is in much the same place as in CONTROL (Fig. 4(a)), with a central pressure that is almost 8 hPa lower than in CONTROL (Table 1). The surface chart from HALFGREEN (Fig. 4(c)) has features in common with both CONTROL and NOGREEN, e.g. low B is absent, but there is a distinct surface pressure trough extending south-westwards from low A, and the position of the cold front on the coast of northern Norway is similar to that in CONTROL, whereas in NOGREEN it was located some 150 km farther east. Conversely, in Fig. 4(d), from DOUBLEGREEN, low B lags further behind than in CONTROL, and there is a marked high pressure ridge between lows A and B. These results are analogous to those of Ólafsson and Bougeault (1996) (Fig. 8), who found that the lee vortex forming in inviscid, irrotational flow, in cases where \( N h / U \) was relatively large, was located increasingly farther to the right as \( N h / U \) was increased, as seen from the onslope flow, which is north-westerly over southern Greenland in this case (see Fig. 5).

Figure 6 shows differences between the surface pressure in CONTROL and NOGREEN on 12 and 13 January 1995. A distinct dipole pattern is apparent, becoming gradually more prominent with time. Thirty-six hours after the beginning of the simulation (+36 h) (Fig. 6(c)) CONTROL has pressure in an area near 65°N, 37°W 25–30 hPa lower than NOGREEN and higher near 75°N, 22°W. This dipole is reminiscent of that found by Schär and Durran (1997) in their \( N h / U = 3 \) case, but clearly in ours the flow patterns are more complex (see Fig. 5) than in such theoretical studies. The negative values are associated with the residual low B, which is evidently an orographic feature. The positive values centred over north-east Greenland in Fig. 6(a) represent the well-known ‘Greenland high’, caused by intense cooling at the surface. In Fig. 6(b) the positive values are shifted north-eastwards and strengthened. This shift is probably the result of strong onslope winds in this region (see Fig. 5). In addition to this dipole, Fig. 6(c) shows large positive values extending eastwards along the cold front associated with cyclone A. This is because low A is significantly deeper in NOGREEN than in CONTROL, and the frontal systems are correspondingly more developed. In order to understand possible causes for this, and for the shift in frontal position mentioned in the previous paragraph, compare (Fig. 7(a,b)) the wind and thermal fields at sigma-level 15 (\( \sigma_{15} = 0.866 \)), in the two extreme simulations, NOGREEN and DOUBLEGREEN. There are large differences in the wind fields in the two cases. Whereas in NOGREEN (Fig. 7(a)) the wind field over southern Greenland is from WNW, blowing perpendicular to the isotherms and so causing strong cold advection, there is a marked barrier effect in DOUBLEGREEN. In this run the wind field over southern Greenland is very weak (Fig. 7(b)), while there is a strong westerly jet just south of Greenland. After the cold
Figure 3. Model-simulated 1000 hPa geopotential height (m) and 500–1000 hPa thickness (m) (pecked lines) at 12 UTC 12 January 1995 (+12 h): (a) Control; (b) NOGREEN; (c) HAlFGREEN; (d) DOUGBLEGREEN.
Figure 4. Model-simulated 1000 hPa geopotential height (m) (solid lines) and 500-1000 hPa thickness (m) (pecked lines) at 12 UTC 13 January 1995: (a) CONTROL; (b) NOGREEN; (c) HALFGREEN; (d) DOUBLGREEN.
TABLE 1. SURFACE PRESSURE AND GEOPOTENTIAL HEIGHT AT CENTRE OF THE LOW AT DIFFERENT TIMES IN THE VARIOUS RUNS: BAROCLINIC LOW A, 10–13 JANUARY 1995

<table>
<thead>
<tr>
<th>Run</th>
<th>Field</th>
<th>+12 h</th>
<th>+24 h</th>
<th>+36 h</th>
</tr>
</thead>
<tbody>
<tr>
<td>CONTROL</td>
<td>MSLP</td>
<td>983 hPa</td>
<td>972 hPa</td>
<td>964 hPa</td>
</tr>
<tr>
<td>CONTROL</td>
<td>700 hPa</td>
<td>–</td>
<td>2507 m</td>
<td>2392 m</td>
</tr>
<tr>
<td>CONTROL</td>
<td>500 hPa</td>
<td>–</td>
<td>–</td>
<td>4764 m</td>
</tr>
<tr>
<td>HALFGREEN</td>
<td>MSLP</td>
<td>983 hPa</td>
<td>964 hPa</td>
<td>963 hPa</td>
</tr>
<tr>
<td>HALFGREEN</td>
<td>700 hPa</td>
<td>–</td>
<td>2428 m</td>
<td>2338 m</td>
</tr>
<tr>
<td>HALFGREEN</td>
<td>500 hPa</td>
<td>4911 m</td>
<td>4795 m</td>
<td>4694 m</td>
</tr>
<tr>
<td>NOGREEN</td>
<td>MSLP</td>
<td>981 hPa</td>
<td>960 hPa</td>
<td>956 hPa</td>
</tr>
<tr>
<td>NOGREEN</td>
<td>700 hPa</td>
<td>2565 m</td>
<td>2430 m</td>
<td>2367 m</td>
</tr>
<tr>
<td>NOGREEN</td>
<td>500 hPa</td>
<td>4899 m</td>
<td>4811 m</td>
<td>4748 m</td>
</tr>
<tr>
<td>DOUBLEGREEN</td>
<td>MSLP</td>
<td>983 hPa</td>
<td>979 hPa</td>
<td>967 hPa</td>
</tr>
<tr>
<td>DOUBLEGREEN</td>
<td>700 hPa</td>
<td>–</td>
<td>2587 m</td>
<td>2473 m</td>
</tr>
<tr>
<td>DOUBLEGREEN</td>
<td>500 hPa</td>
<td>–</td>
<td>–</td>
<td>4862 m</td>
</tr>
</tbody>
</table>

TABLE 2. SURFACE PRESSURE AND GEOPOTENTIAL HEIGHT AT CENTRE OF THE LOW AT DIFFERENT TIMES IN THE VARIOUS RUNS: OROGRAPHIC LOW B, 10–13 JANUARY 1995

<table>
<thead>
<tr>
<th>Run</th>
<th>Field</th>
<th>+12 h</th>
<th>+24 h</th>
<th>+36 h</th>
</tr>
</thead>
<tbody>
<tr>
<td>CONTROL</td>
<td>MSLP</td>
<td>983 hPa</td>
<td>978 hPa</td>
<td>974 hPa</td>
</tr>
<tr>
<td>CONTROL</td>
<td>700 hPa</td>
<td>–</td>
<td>2502 m</td>
<td>2471 m</td>
</tr>
<tr>
<td>CONTROL</td>
<td>500 hPa</td>
<td>–</td>
<td>–</td>
<td>4810 m</td>
</tr>
<tr>
<td>HALFGREEN</td>
<td>MSLP</td>
<td>983 hPa</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>HALFGREEN</td>
<td>700 hPa</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>HALFGREEN</td>
<td>500 hPa</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>NOGREEN</td>
<td>MSLP</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>NOGREEN</td>
<td>700 hPa</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>NOGREEN</td>
<td>500 hPa</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>DOUBLEGREEN</td>
<td>MSLP</td>
<td>983 hPa</td>
<td>974 hPa</td>
<td>973 hPa</td>
</tr>
<tr>
<td>DOUBLEGREEN</td>
<td>700 hPa</td>
<td>2551 m</td>
<td>2476 m</td>
<td>2478 m</td>
</tr>
<tr>
<td>DOUBLEGREEN</td>
<td>500 hPa</td>
<td>4917 m</td>
<td>4843 m</td>
<td>4826 m</td>
</tr>
</tbody>
</table>

air rounds southern Greenland, it is diverted around the (lee) vortex off south-eastern Greenland (see Figs. 3(d) and 4(d)), becoming south-westerly, thus weakening the cold air advection in this area. As a result, the thermal gradient on the cold front to the east and north-east of Iceland is weakened. Consequently we can expect the baroclinic energy conversion to be reduced.

Another signature of the barrier effect of Greenland in Fig. 7(b) is a strong north-easterly jet near 75°N, 18°W, whereas in NOGREEN (Fig. 7(a)) the wind there is easterly, blowing slightly across the isotherms and hence causing warm advection ahead of the approaching warm front.

The weakening of the baroclinic low in runs with orography can also be understood by viewing the distortion of the thickness field by low B. Figure 8 shows 700 hPa heights and 500–1000 hPa thicknesses (the ‘thermal field’) from runs NOGREEN and DOUBLEGREEN. The thermal field has a well developed ‘wave structure’ in the NOGREEN run (Fig. 8(a)), reminiscent of idealized theoretical studies of baroclinic waves (for example that by Thornroft et al. (1993)), while in DOUBLEGREEN (Fig. 8(b)), the wave crest is almost absent and the cold front is rather weak. There is also a positive feedback effect related to the lee cyclone: in DOUBLEGREEN (Fig. 8(b)), assuming quasi-geostrophic flow, low B causes warm advection behind the cold front.
west and north-west of Iceland. The same signature, although somewhat weaker, is also found in CONTROL (not shown). It is interesting to note from Table 1 that while the baroclinic cyclone is 19 hPa weaker in DOUBLEGREEN than in NOGREEN at +24 h (corresponding to Figs. 7 and 8), this difference is reduced to 11 hPa at +36 h. A similar reduction between +24 h and +36 h is found at 700 hPa. A plausible explanation is that at +36 h the baroclinic wave has come away from the retarding influence of low B (see Fig. 4(d)) and is therefore able to realize more of its deepening potential.

Following Smith (1989), we have estimated the degree of flow splitting through the non-dimensional parameter $Nh/U$. We have computed this quantity in the area west of Greenland at different stages of the CONTROL integration and found that it was always large enough for flow splitting and corresponding lee effects to be expected (see section 2). Hence, we suggest that the dipole seen in Fig. 5 is largely caused by lee effects downwind of the north-westerly flow over southern Greenland (see Fig. 6). We further suggest that the north-easterly winds over eastern Greenland, near 70°N, also contribute to the dipole, by tending to create an onslope high and a lee low, and note that once the dipole is formed it will tend to maintain itself by sustaining the onslope flow near 70°N, 22°W.

In the simulations described so far, the mountains of Greenland have been represented by several different model orographies. In addition to these simulations, some sensitivity experiments were also carried out. Although none gave results significantly different from CONTROL, they will be reviewed briefly. In one simulation, the surface of Greenland was changed from ice to tundra. The effect was very small at the surface,
Figure 6. Difference in mean sea level pressure (5 hPa intervals); CONTROL minus NOGREEN: (a) 12 UTC 12 January 1995 (+12 h); (b) 00 UTC 13 January 1995 (+24 h); (c) 12 UTC 13 January 1995 (+36 h).
Figure 7. Simulated winds and potential temperature at sigma level 15 ($\sigma_1 = 0.866$) at 00 UTC 13 January 1995 (+24 h): (a) NOGREEN; (b) DOUBLGREEN.
Figure 8. Simulated 700 hPa geopotential height (m) and 500–1000 hPa thickness (m) at 00 UTC 13 January 1995 (+24 h): (a) NOGREEN; (b) DOUBLEGREEN.
but the height at 500 hPa rose by a few metres, as part of a slight weakening of the cold air dome over Greenland. This experiment shows that, at least on the short timescales considered here, orographic height is a much more important parameter than surface type. In another simulation, the height of the orography of Iceland was set to zero. This had a negligible influence on the cyclone evolution and pressure distribution in the North Atlantic. Simulations were also carried out beginning 24 hours later than those presented so far. A comparison between simulations CONTROL, NOGREEN, HALFGREEN and DOUBLEGREEN showed the two periods yielded exactly the same qualitative results.

5. CONCLUDING REMARKS

A deep baroclinic winter cyclone in the vicinity of Iceland has been investigated using a NWP model. Sensitivity tests have been carried out, in which the orography of Greenland was modified, in order to understand its role in cyclone development in the case examined and for the 'Iceland low' in general. Cyclonic development took place near the east coast of Greenland. As a baroclinic low deepened and moved north-eastwards, a secondary, quasi-barotropic low, very characteristic for this area, formed behind the baroclinic low and between Iceland and Greenland. In simulations where the orography of Greenland was removed, this 'residual low' disappeared, as did an anticyclone over eastern Greenland. This dipole has similarities to those found in theoretical studies of flow splitting in stratified flow impinging on high mountains (for example, by Schär and Durrán (1997) and Ólafsson and Bougeault (1996)).

Investigations of the wind field around Greenland in the different simulations revealed that the conditions were indeed favourable for flow splitting in the lower troposphere during this event, suggesting that the characteristic secondary cyclone is a lee cyclone. We further suggest that the anticyclonic part of the dipole is caused partly by a combination of thermal effects through cold-air outflow from the Greenland plateau and the occurrence of strong onshore winds and cold-air damming.

In a simulation with the orography of Greenland removed, the baroclinic low became deeper and more intense than in a control simulation. Furthermore, a doubling of the orographic height yielded an even weaker baroclinic low than in the control run, and halving the orographic heights gave a result about midway between that of the control simulation and the simulation without orography. In order to explain the relationship between orography and strength of the baroclinic lows, we have investigated the flow and thermal fields at various levels around Greenland. The results suggest that the weakening of the baroclinic low is a result of the orographic deflection of cold air to the rear of the cyclone. This weakens the cold advection behind the cyclone, and consequently the baroclinicity. This effect is further enhanced by warm air advection ahead of the lee cyclone.

Results analogous to those described in the present study were found in another case (not shown). In that run of the model, the surface of Greenland was set at sea level. Even so, a secondary cyclone occurred.

These findings should help understanding of flow patterns in the North Atlantic. They indicate, for example, that models with inadequate resolution of Greenland's orography may be expected to underestimate the frequent secondary depression between Greenland and Iceland, and, at the same time, overestimate the strength of baroclinic lows which pass through the area. Such errors may well have implications for weather forecasts for the European continent a few days later. We hope this work will stimulate others to investigate the challenging and important problems set by the orographic effects of Greenland.
ACKNOWLEDGEMENTS

The authors would like to thank two anonymous reviewers for most helpful comments on the manuscript.

REFERENCES

Charney, J. G. and Eliassen, A. 1949 A numerical method for predicting the perturbations of the middle latitude westerlies. Tellus, 1, 38–54