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Article

Simulating a severe windstorm in complex terrain

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Abstract

The severe windstorm that hit Iceland on 1 February 2002 is analyzed using high-resolution numerical simulations, conventional observations at the ground and satellite images. The windstorm and the great mesoscale variability in the observed wind are reproduced by the numerical simulations, with increasing accuracy as the horizontal resolution is increased, stepwise from 9 km to 1 km. At a horizontal resolution of 333 m the flow pattern is realistic, but the quantitative improvement is not clear. The strongest surface winds are found in localized downslope windstorms below steep and amplified gravity waves which presumably break in a reverse (negative) vertical wind shear at middle tropospheric levels. Surface winds are in general slightly overestimated and the model performs worst at locations where subgrid topography is expected to be of importance. The overestimating of the simulated surface wind speed is greatest immediately downstream and upstream of steep mountains. The surface winds are only moderately affected by the parameterization of surface friction and the magnitude of the downslope windstorms shows some sensitivity to the distance to the next downstream mountain. The study indicates that the turbulence is overestimated immediately upstream of mountains at 1 km horizontal resolution.

Zusammenfassung

Das starke Sturmereignis vom 1. Februar 2002 auf Island wird mit hoch auflösenden numerischen Simulationen, konventionellen Bodenbeobachtungen und Satellitenbildern untersucht. Das Windfeld und seine mesoskalige Variabilität werden durch die numerischen Simulationen, deren Genauigkeit mit der schrittweise von 9 km auf 1 km erhöhten räumlichen Auflösung steigt, wiedergegeben. Bei der höchsten räumlichen Auflösung von 333 m ist das Strömungsmuster zwar realistisch, eine weitere quantitative Verbesserung aber unklar. Die stärksten bodennahen Windgeschwindigkeiten werden in örtlich eng begrenzten Fallwinden hinter Gebirgszügen unterhalb von steilen und verstärkten Schwerewellen, die vermutlich in einer invertierten vertikalen Windscherung in der mittleren Troposphäre brechen, gefunden. Die Bodenwinde werden im Allgemeinen leicht überschätzt, wobei die schlechtesten Ergebnisse in Gebieten auftreten, in denen ein Einfluss der kleinskaligen, vom Rechengitter nicht aufgelösten Orographie erwartet werden muss. Diese Überschätzung ist am stärksten unmittelbar vor und hinter steilen Bergen. Die Parametrisierung der Bodenreibung hat nur mäßigen Einfluss auf die bodennahe Windgeschwindigkeit, die leeseitigen Fallwinde zeigen aber eine gewisse Abhängigkeit vom Abstand zur nächstfolgenden Bergkette. Die Untersuchung legt den Schluss nahe, dass in der 1 km-Auflösung die Stärke der Turbulenz unmittelbar stromauf von Bergen überschätzt wird.

1 Introduction

Severe windstorms are one of the big threats of weather to vegetation, infrastructure and lives. The greatest danger in such storms is related to fluctuations in the wind speed at periods as short as a few seconds, otherwise known as wind gusts. In extreme windstorms in complex terrain, the gust strength may easily exceed twice the 10-minute mean wind speed at 10 metres above ground, (e.g. DURRAN, 1990; GRØNÅS and SANDVIK, 1999). At weaker winds, gusts are far weaker and more similar to the mean wind (e.g. NAESS et al., 2000; ÁGÚSTSSON and ÓLAFSSON, 2004b).

Gustiness is a manifestation of atmospheric turbulence, which is primarily found close to the surface of the earth, i.e. in the atmospheric boundary layer. Here, turbulent motion arises due to the low static stability and high vertical wind shear caused by surface friction. Aloft, turbulence is also found, for example in regions of wind shear near the tropospheric and stratospheric jets as well as in deep convective cells. Of greater interest in the context of this study is the turbulence produced by large amplitude gravity (buyoancy) waves, which may form in a stably stratified atmosphere above mountainous regions. The turbulence is produced aloft, either in regions of high wind shear or due to local convective instability where the waves break, and may reach down to the surface of the earth, accompanied by strong wind gusts. The gustiness has been suggested to be associated with wave breaking (CLARK and FARLEY, 1984) but also with Kelvin-Helmholtz instability (SCINOCCA and PELTIER, 1989; PELTIER and SCINOCCA, 1990). A

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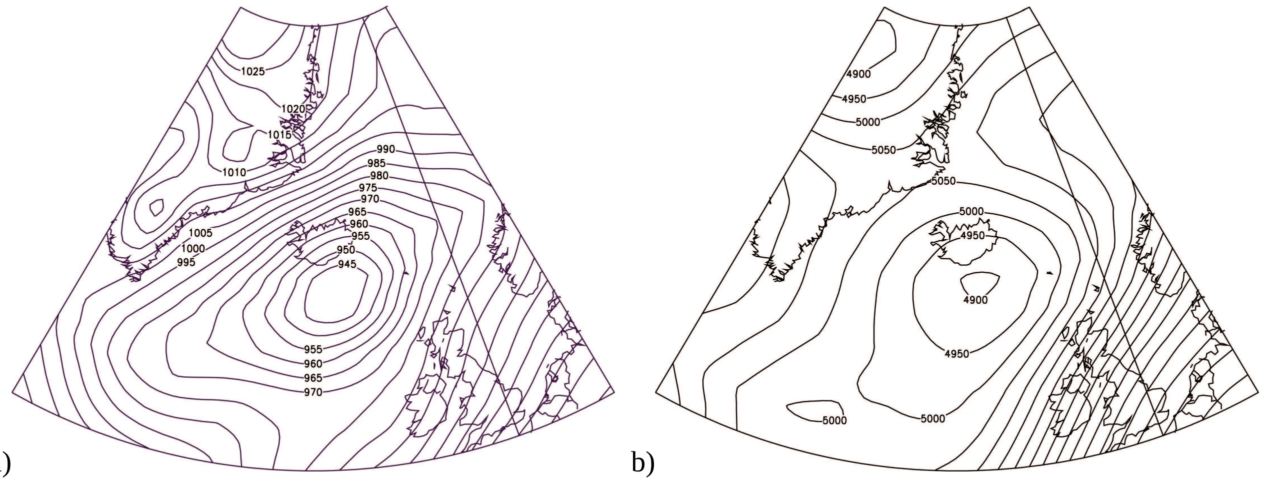

a) b)

Figure 1: a) Mean sea level pressure with intervals of 5 hPa at 00 UTC on 2 February 2002. b) Geopotential height at 500 hPa [m] with intervals of 50 m at 00 UTC on 2 February 2002. Data from NCEP/NCAR, acquired through NOAA/CDC.

general description of a downslope windstorm in laminar flow below turbulent flow has been presented by SMITH (1985). There are many studies of breaking gravity waves in the troposphere. DOYLE et al. (2005) describe an event where a large-amplitude wave breaks in a strong southwesterly flow over South-Greenland. Intense wave breaking has in fact also been observed and reproduced by a numerical model in an easterly flow over the same area when on 6 December 2005, conditions for generation, vertical propagation and breaking of gravity waves were very favourable (ÓLAFSSON and ÁGÚSTSSON, 2006). Perhaps the best known events, related to gravity wave activity, are the Boulder downslope windstorms (e.g. CLARK et al., 1994), where the strongest gusts can easily exceed 50 m/s and the mean wind speeds are as great as 25 m/s (DURRAN, 1990).

In recent years, an increasing number of studies have focussed on simulations of airflow at very small scales ($\sim$1 km) in complex terrain. This increase is primarily based on improved nonhydrostatic models available for high-resolution numerical simulations, and the more readily available and powerful computational resources. The joint use of high-resolution simulations and observations of wind can improve the description of the local wind climate, as was the case in Iceland in e.g. ÓLAFSSON and RÖGNVALDSSON (2004); ÓLAFSSON et al. (2006a). In these studies, an atmospheric model and observations gave valuable information regarding conditions for wet snow icing and drifting snow in locations for planned powerlines and a new road in Northeast-Iceland. High-resolution mesoscale modeling is also a valuable tool to gain a better understanding of extreme events related to orographic flow. One such extreme mountain weather event occured in North-Norway on 12 October 1996 where 30 m high power pylons broke in a narrow and steep-sided valley. Very high-resolution simulations gave a relatively good description of the event where mean wind speeds were estimated to be as high as 25 m/s and the gusts approx. twice as strong (GRØNÅS and SANDVIK, 1999). In Iceland, there have been succesful simulations of extreme mountain weather events, e.g. ÓLAFSSON (1998), where a strong, localized, downslope windstorm on the Snæfellsnes peninsula is associated with breaking mountain waves aloft. In another extreme windstorm in Iceland, strong surface winds in complex terrain in East-Iceland were presumably caused by gravity wave activity aloft and the interaction of the flow with orography (ÓLAFSSON and ÁGÚSTSSON, 2004; ÁGÚSTSSON, 2004). The greatest damage during the storm, including a near fatal incident, was caused by wind gusts ($\sim$50 m/s) which exceeded twice the mean wind speed. There have been successful attempts to predict gusts during the windstorm (ÁGÚSTSSON and ÓLAFSSON, 2004a), as well as during other such windstorms in complex terrain e.g. BELŬSIĆ and KLAIĆ (2004).

Here, conventional observations of wind at the ground, numerical simulations and satellite images are used to analyze a severe windstorm which hit Iceland on 1 February 2002. The storm is of particular interest because of the large mesoscale variability in the observed winds in the complex terrain of Northwest-Iceland. The strong winds and heavy snowfall disrupted transport on ground and in the air, caused damage to structures and led to several avalanches.[1]

The results of the high-resolution simulations on which this study is based, have already proven valuable for the development of an operational forecasting system based on high resolution (3 and 9 km) atmospheric simulations over Iceland (ÓLAFSSON et al., 2006b). The system has proven to be a valuable addition to the available tools of the forecasters, but fore-

[1]Rafmagnstruflanir og viðbúnaður vegna snjóflóða, 3. Feb. 2002. Morgunblaðið (The Morning Paper), Reykjavík, Iceland, p. 2.

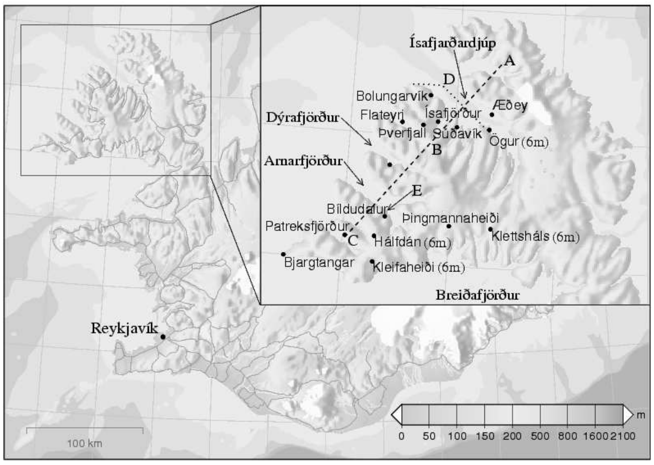

Figure 2: Topography and locations of chosen weather stations in Northwest-Iceland. Scale shows height above sea level.

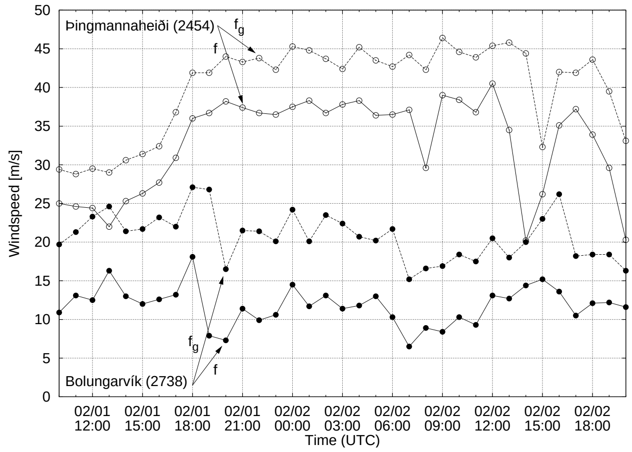

Figure 3: Observed mean winds, f [m/s], and gusts, f_g [m/s], at Þingmannaheiði and Bolungarvík in Northwest-Iceland.

casts made in Iceland have previously been based on numerical models running at much lower resolution (20–50 km). As a part of the development of this new system, the storm studied here was simulated using four different parameterizations of the planetary boundary layer processes (PBL schemes) as well as two different moisture physics schemes. The authors are not aware of any literature describing an extensive study which compares a number of different PBL schemes in high-resolution simulations. Many authors have studied individual PBL schemes (e.g. BURK and THOMPSON, 1989; JANJIĆ, 1994; HONG and PAN, 1996), while others have compared two different schemes (e.g. SHAFRAN et al., 2000; RÖGNVALDSSON and ÓLAFSSON, 2002) as well as different parameterizations of e.g. moisture and convection (DENG and STAUFFER, 2006).

In the following section, the available observations and the synoptic situation for the storm are discussed. Section 3 describes the atmospheric model while the following section shows the results of the numerical simulations. The results are discussed in section 5 and the final section gives a summary of the study and concluding remarks.

2 The storm of 1–2 February 2002

2.1 The synoptic situation

At 00 UTC on 2 February 2002, a deep surface low was located at the south coast of Iceland and a high was over Greenland (Fig. 1a). There was a large surface pressure gradient over Iceland and the Denmark strait, and a strong northeasterly windstorm over Northwest-Iceland.

At middle and upper tropospheric levels, the winds were much weaker as is implied by the isobaric surfaces (Fig. 1b). There was in other words a considerable reverse (negative) vertical wind shear in the lower troposphere.

2.2 Surface observations

In Northwest-Iceland, there are a number of weather stations with readily available observations. Most of the stations used here belong to Veðurstofa Íslands[2] while a few stations belong to Vegagerðin and Siglingastofnun.[3] All the stations are located in complex terrain and approximately half of them are located at high altitudes (Fig. 2).

Observations of the 10-minute mean wind speed and wind direction are available at 10-minute intervals from all of the stations. The wind is either observed at 10 m or at the top of a 6-m mast, raised approx. 1 m above its immediate surroundings (Hálfdán, Kleifaheiði, Klettsháls and Ögur stations of Vegagerðin, as seen in Fig. 2). This supposedly leads to observed winds being somewhat weaker at the stations of Vegagerðin. However, a subjective criteria for these observation sites is to detect maximum winds, while Veðurstofa Íslands seeks to place weather stations where they can provide data which is representative for a large area.

Most stations are equipped with Young anemometers, while the mountain station at Þverfjall is equipped with a heavier, heated, Hydro-Tech anemometer which is prone to overestimating the mean winds and underestimating wind gusts. All data has been checked for errors at Veðurstofa Íslands, but no systematic corrections have been made of wind data.

Observations of a localized windstorm at Þingmannaheiði, and weak winds at Bolungarvík, where there were indeed no indications of the storm, show the large mesoscale variability in the winds during the windstorm (Fig. 3).

[2]The Icelandic Meteorological Office
[3]The Public Roads Administration and The Maritime Administration

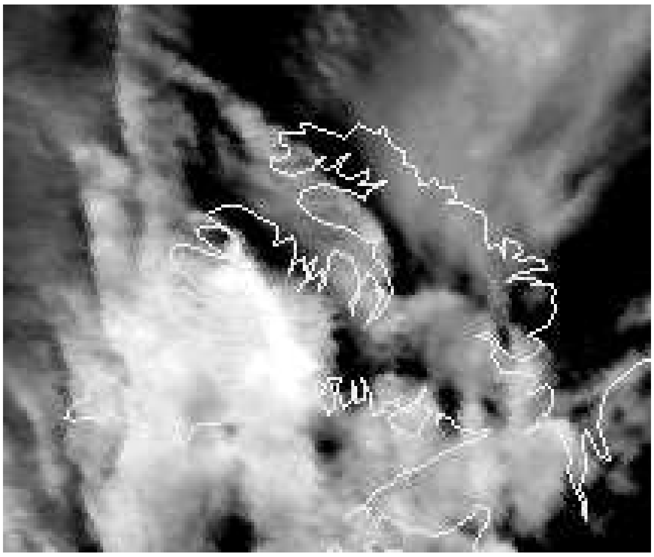

Figure 4: Satellite image (IR) valid at 03:52 UTC on 2 February 2002. The image is from a NOAA satellite, and was aquired through the Satellite Receiving Station of Dundee University in Scotland.

2.3 Satellite observations

Satellite images of Northwest-Iceland (Fig. 4), taken during the windstorm, show cloud bands running perpendicular to the wind direction, indicating gravity waves aloft. The signature of the waves is for example particularly clear in the vicinity of Æðey. It should be noted that the waves extend far to the west, to an area where there are no mountains below.

3 The atmospheric model

The storm is simulated with the nonhydrostatic mesoscale atmospheric model, MM5 (GRELL et al., 1994). The modeling system has a wide range of applications in meso- and even microscale meteorology, and is highly applicable for both operational weather forecasting as well as theoretical studies.

The operational analysis from the European Centre for Medium-Range Weather Forecasts (ECMWF) are used to initialize and force the model at its boundaries. The model was run with a horizontal resolution of 9, 3 and 1 km, and up to 333 m in one sensitivity test, with respectively 90 x 95, 166 x 202, 163 x 190 and 100 x 100 gridpoints in the nested domains (see Fig. 5 for domain locations). To ensure that the development of the storm was well captured and the results not compromised by spin-up, the simulations were started at noon on 1 February 2002, or 12 hours before the storm reached its maximum. The following 24 hours were simulated, i.e. until noon on 2 February 2002. Each nested domain was initiated 3 hours after its mother domain, with the 333 m domain starting last at 21 UTC. Forty vertical σ-layers were used, terrain following at lower levels but flatter towards the model top. Previous studies of flow over Iceland have shown the 10-m winds and the winds at the lowest sigma level to be far too weak in windstorms. Here, the simulated surface wind is therefore taken from the second lowest σ-layer (approx. 40 m above the ground).

As this study was a part of preparations for operational weather forecasting in Iceland using a limited area model, the storm was simulated using a setup where the choice of moisture physics and planetary boundary layer schemes was varied, while in some cases atmospheric radiation (Rad) was ignored. The relevant moisture schemes are the relatively simple scheme of DUDHIA (1989) which includes cloud and rain water as well as simple ice phases, and the more complex scheme of REISNER et al. (1998) which also includes mixed phases, cloud ice, snow and graupel. Four different PBL-schemes were used to investigate the different impact of surface friction on the simulated wind, the MRF scheme (HONG and PAN, 1996), the ETA scheme (JANJIĆ, 1990, 1994), the Burk-Thompson (B&T) scheme (BURK and THOMPSON, 1989) and the Gayno-Seaman (G&S) scheme (BALLARD et al., 1991; SHAFRAN et al., 2000). Results shown here are found with the ETA and Reisner schemes, with atmospheric radiation taken into account. For further information on the setup of the simulation of the storm, the reader is referred to ÁGÚSTSSON and ÓLAFSSON (2004a) while more extensive information is given in ÁGÚSTSSON (2004).

4 Numerical simulations

4.1 The simulated surface flow

At a resolution lower than 1 km, the complex topography in Northwest-Iceland is in general poorly resolved. As a consequence, the simulated surface flow at the coarsest resolution (9 km) has little spatial detail and does not appear to capture much of local effects in Northwest-Iceland (Fig. 5). The simulated flow does however give an overview of the large-scale flow. There are strong winds to the north and west of Iceland. The flow over Iceland is by far strongest in the northwest and in the central highlands. The strong winds in the southeast are associated with high mountains in the region.

At a resolution of 3 km, the simulated surface windfield gives a much more detailed wind pattern, with the same large scale features as seen at coarser resolution (Fig. 6). The resolution is however still too low to resolve the complex topography, and consequently, anticipated details of the flow in Northwest-Iceland are still missing. The large scale flow (3 and 9 km resolution) has been verified by comparison with ground based observations spread throughout Iceland and partly with satellite based observations of sea surface winds (not shown).

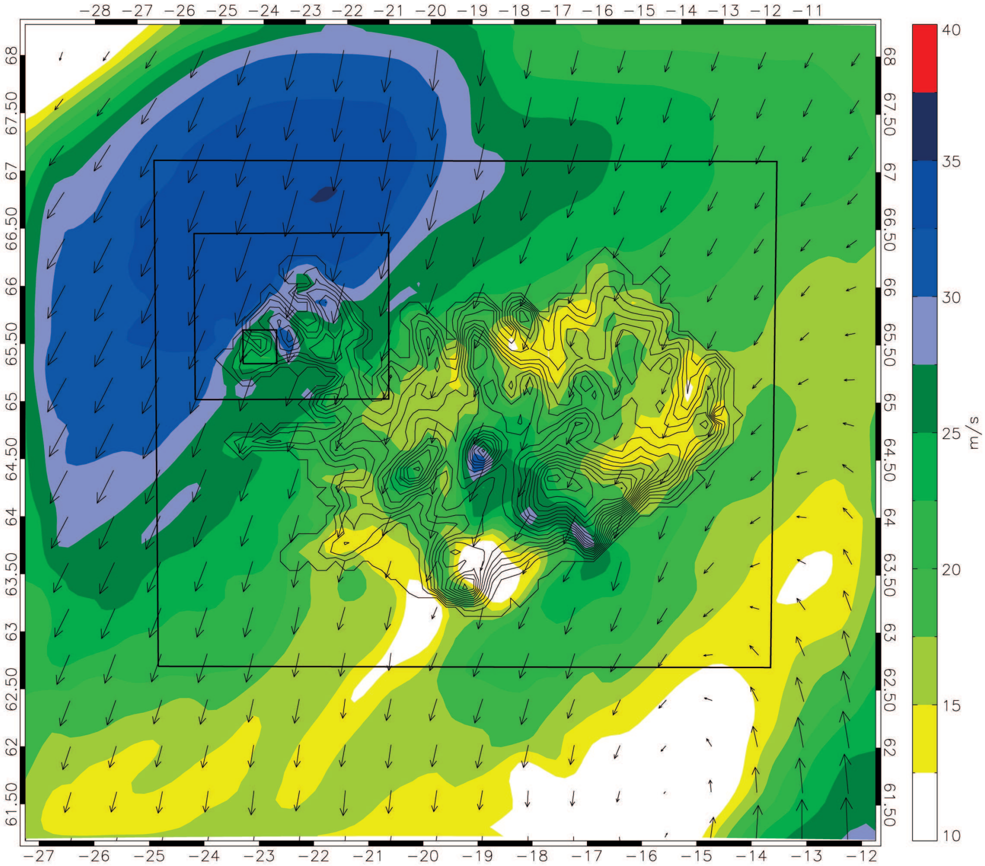

Figure 5: Topography, with contours every 100 m, and simulated surface wind [m/s] in a subdomain at low resolution (9 km) at 00 UTC on 2 February 2002. The Figure shows the locations of the numerical domains with resolutions of 3 km, 1 km and 333 m.

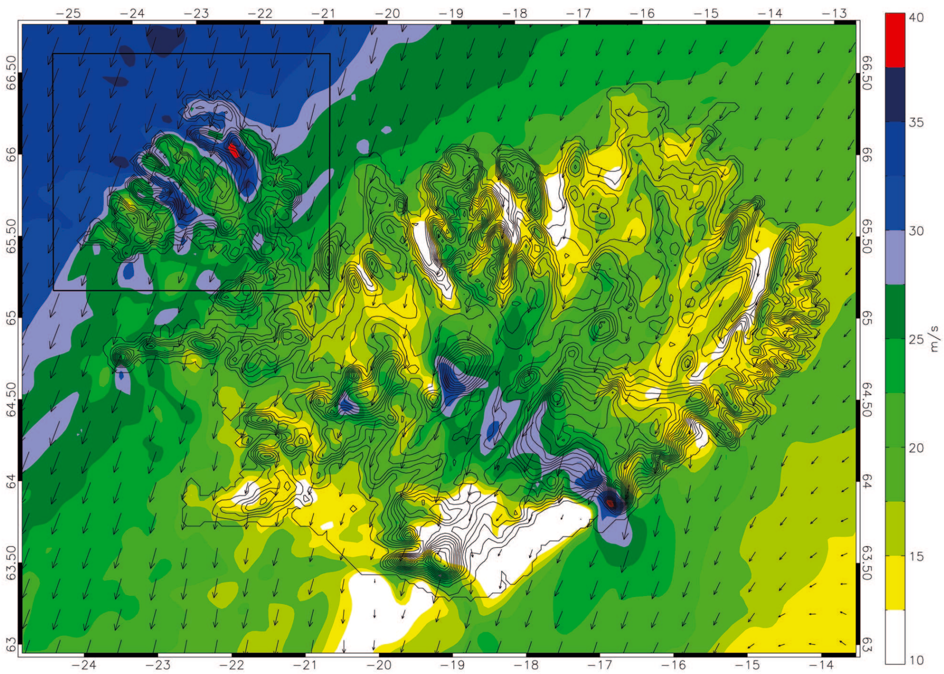

Figure 6: Topography, with contours every 100 m, and simulated surface wind [m/s] in a subdomain at medium resolution (3 km) at 00 UTC on 2 February 2002. The Figure shows the location of the domain with a resolution of 1 km.

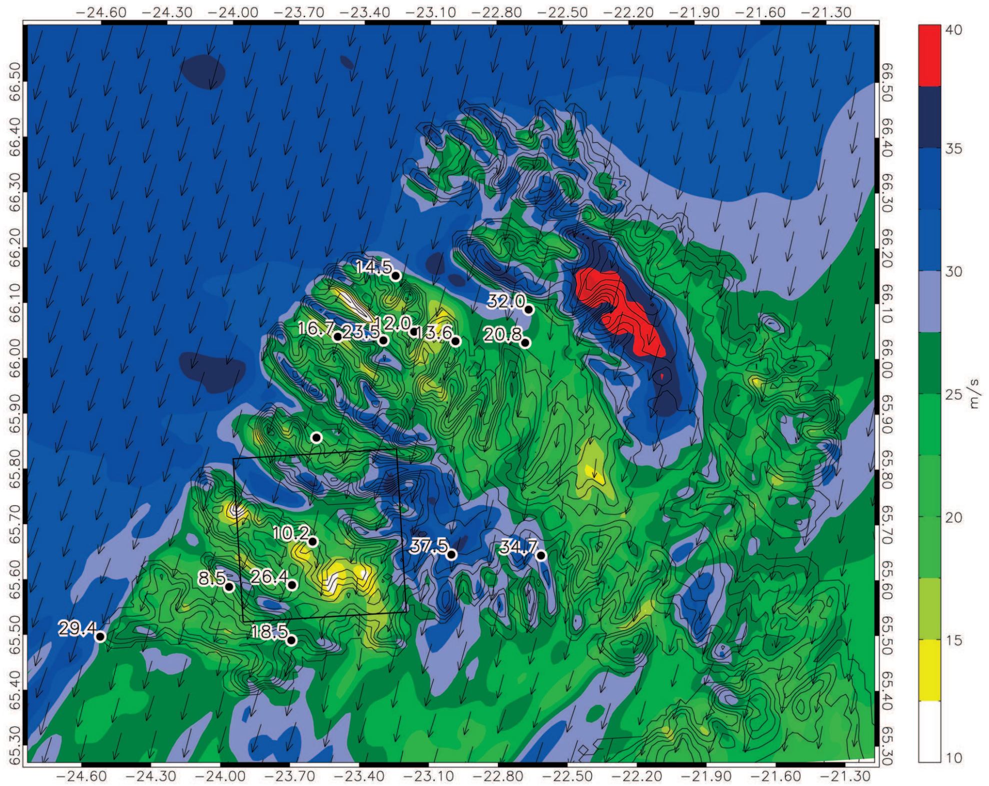

Figure 7: Topography, with contours every 100 m, and simulated surface wind [m/s] in a subdomain at high resolution (1 km) at 00 UTC on 2 February 2002. The Figure shows wind observations [m/s] and the location of the domain with a resolution of 333 m.

As expected, increased horizontal resolution has the greatest impact in regions of steep and narrow mountains, as in Northwest-Iceland. Several areas with locally enhanced, or reduced wind speeds, that were not present at a resolution of 3 km, appear at a resolution of 1 km (Fig. 7). At most locations where observations are available, e.g. Æðey, Þingmannaheiði and Hálfdán (cf. Figs. 7, 8a and 8c), the strong simulated winds are confirmed by observations. There are however locations, e.g. at Patreksfjörður (Fig. 8b), where the weak observed winds are greatly overestimated throughout the storm.

There is an overall improvement in the quality of the simulated winds as the horizontal resolution is increased from 9 to 3 km, and also, from 3 to 1 km (Fig. 9).

The sensitivity of the simulated surface wind field to increased horizontal resolution was tested further using a grid size of 333 m for a domain covering an approx. 30 km x 30 km area around Bíldudalur (Figs. 10–11).

The representation of the topography near Bíldudalur improves considerably when the resolution is increased from 3 km to 1 km. This is not surprising as the most important features in the topography, e.g. the fjords, have a characteristic width on the order of 10 km or less. The improvement is not as dramatic when the resolution is increased from 1 km to 333 m. There are however significant differences in the overall simulated windfield. Both lee-side and upstream sheltering are more clearly defined at 333 m than at 1 km (cf. Figs. 7 and 10).

4.2 Breaking waves

A NE-SW oriented (along the large scale wind) section along the line from A to C in Fig. 2, reveals large amplitude internal gravity waves over the mountains (Fig. 12). In these waves, the maximum wind speed is found where the air is descending, while in the ascending part of the wave, the wind is much weaker. Æðey is below the descending part of the steepest wave, while Bíldudalur is below the ascending part of another wave, further downstream. The waves break in a region of reverse vertical wind shear in the ambient flow close to 600 hPa. There is statically unstable air and strong turbulence where the waves break.

The distribution of turbulent kinetic energy (TKE), e.g. at 600 hPa (Fig. 13), indicates breaking of the waves in several areas in the region. Very strong winds are indeed observed in Æðey, which is below the first breaking wave, and at Þingmannaheiði (Fig. 8a), which is close to the eastern limits of the southernmost wave breaking region. Ground observations are not available from elsewhere below breaking waves.

4.3 Modified orography

The sensitivity of the gravity waves, represented by the maximum wind speed close to the ground, to the orography was investigated in a simulation where the width of the fjord Ísafjarðardjúp was reduced by extending the

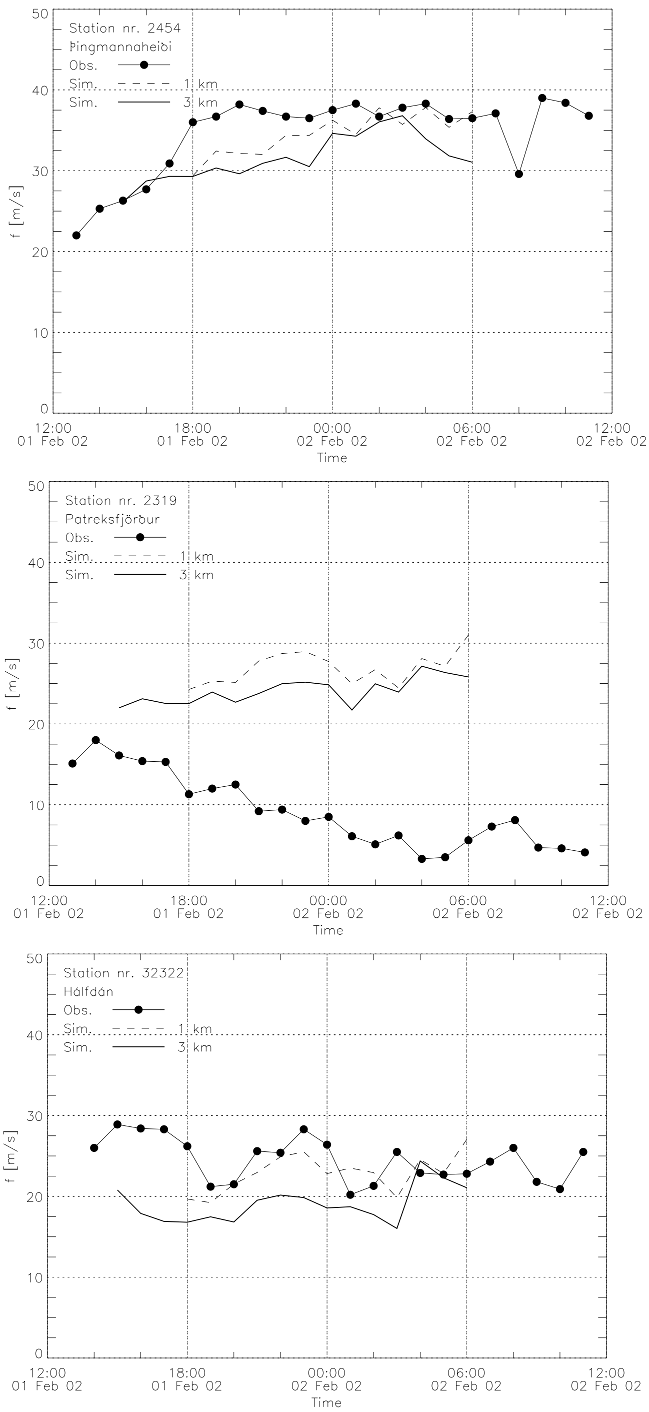

Figure 8: Observed and simulated surface winds [m/s] at a resolution of 1 and 3 km at: a) Þingmannaheiði, b) Patreksfjörður, c) Hálfdán.

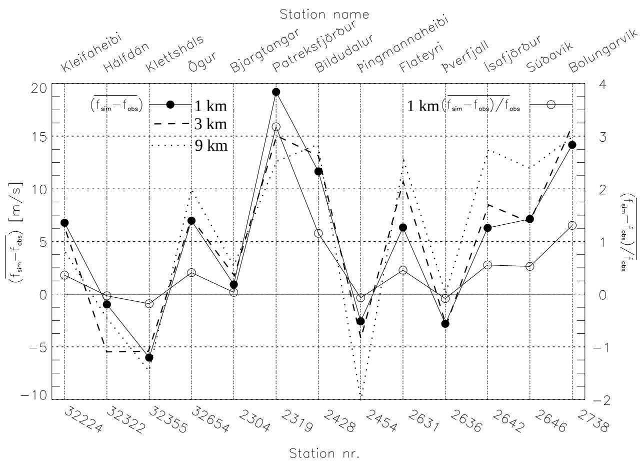

Figure 9: Mean difference of simulated and observed surface winds.

mountains on the southern side into the middle of the fjord (dotted line, D, in Fig. 2).

The maximum wind speeds in the breaking wave over Ísafjarðardjúp are slightly greater, and they are found at a slightly lower altitude in the simulation which uses true orography than in the sensitivity run (cf. Figs. 14a and 14b). The largest difference in wind speeds is found at the surface. In the sensitivity run (narrow fjord), the maximum surface wind speeds do not exceed 33 m/s, while with true orography, the maximum wind speeds at the surface are approx. 37 m/s.

4.4 Sensitivity to parameterization

The mean difference between simulated and observed surface winds at a horizontal resolution of 1 km is calculated for simulations using four different PBL-schemes, two different moisture schemes and inclusion/exclusion of radiation processes (Fig. 15).

All the PBL schemes give somewhat similar results for the simulated wind speeds, i.e. similar RMS-errors. However, the MRF scheme simulates the observed wind direction significantly worse than the other schemes (not shown). The inclusion of radiation processes (Rad) had nearly no impact on the simulated wind field, while a more sophisticated moisture scheme (Reisner) had a large positive impact on the performance of the model.

5 Discussion

5.1 Parameterizations

The quantitative comparisons between different model setups (i.e. Fig. 15) indicates that surface winds are best reproduced with the ETA and Reisner schemes. In fact, the succesful HRAS-system (ÓLAFSSON et al., 2006b) uses this setup. The better performance of the simulations using the Reisner scheme may be a result of the more accurately simulated precipitation. The simulations may describe better the increased atmospheric stability at low levels due to release of latent heat and possibly also to the evaporation of precipitation and the subsequent cooling of the air at the surface. An increased stability at very low levels will presumably lead to weaker surface winds as surface friction is more efficient at decelerating the surface airflow in more stably stratified boundary layers. An extensive discussion on the sensitivity to parameterization is given in ÁGÚSTSSON (2004).

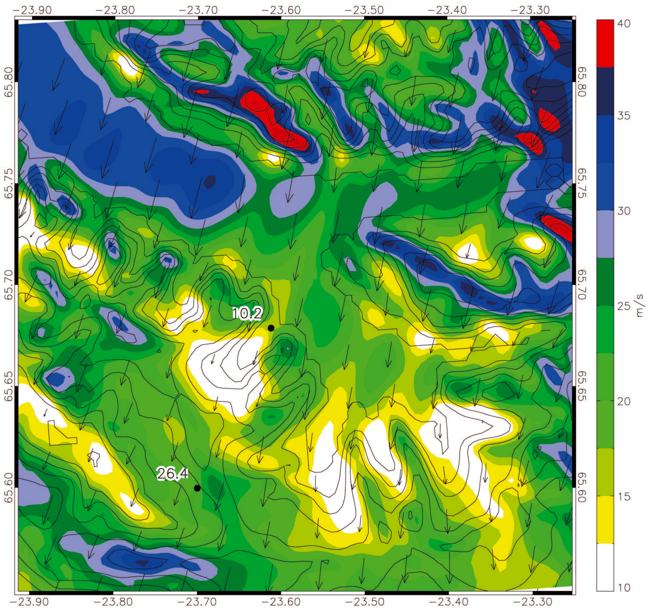

Figure 10: Topography, with contours every 100 m, and simulated surface wind [m/s] in a subdomain at very high resolution (333 m) at 00 UTC on 2 February 2002. The Figure shows wind observations [m/s].

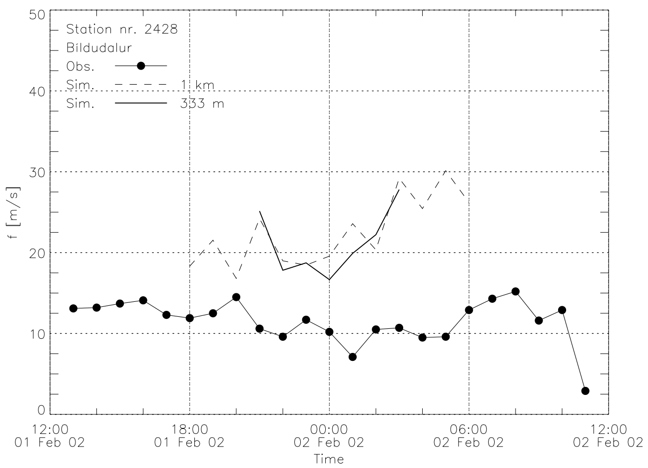

Figure 11: Observed and simulated surface winds [m/s] at a resolution of 1 km and 333 m at Bíldudalur.

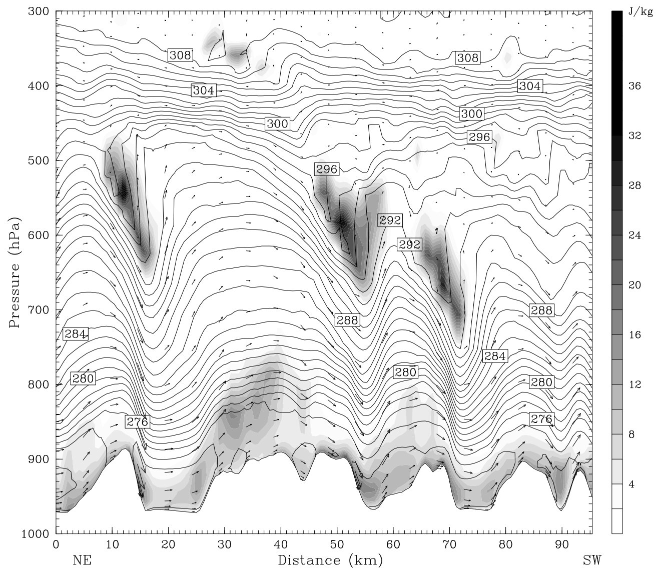

Figure 12: Section from A to C in Fig. 2, at 00 UTC on 2 February 2002. Topography, wind vectors (max. vector is 50 m/s), potential temperature [K] isolines with intervals of 1 K and TKE (shaded) at a resolution of 1 km.

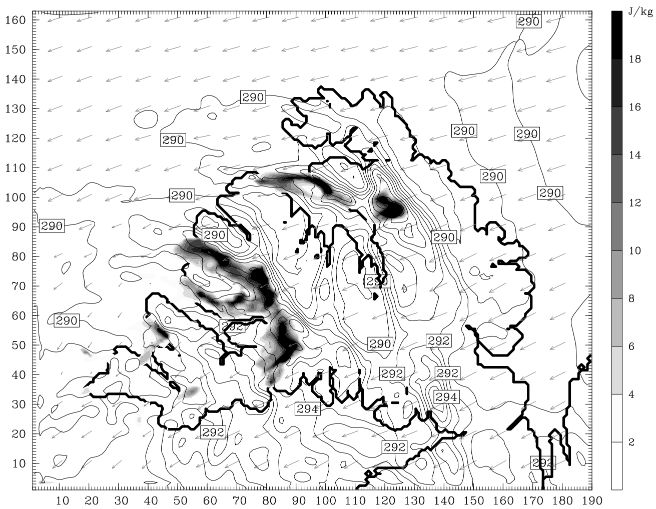

Figure 13: Coastline (bold), wind vectors (max. vector is 30 m/s), potential temperature [K] isolines with intervals of 1 K and TKE (shaded) at a resolution of 1 km at 600 hPa on 2 February 2002 at 00 UTC.

5.2 Error analysis

The largest underestimation of wind speed in Fig. 9 is at Klettsháls. In fact, this underestimation is relatively small when compared to the wind speed. At Bjargtangar, Hálfdán, Þingmannaheiði and Þverfjall, the wind speed is quite accurately simulated and the moderate discrepancy between observed and simulated winds can presumably be explained by subgrid orography not represented at the current resolution of the atmospheric model. Also, the error at Þverfjall may partly be explained by the possible overspeeding of the anemometer.

The greatest overestimation of the wind speed is at Patreksfjörður, Bíldudalur and Bolungarvík, where the error is up to three times the observed wind speed (Fig. 9). At these locations, the observed mean wind shows no indications of the passage of the storm; winds at Bíldudalur and Bolungarvík (Figs. 3 and 11) are relatively constant at approx. 12 m/s, while the winds at Patreksfjörður decrease throughout the storm (Fig. 8b). In fact, the quality of the simulated wind speed at Patreksfjörður increases with decreased resolution. Mean winds are also overestimated at Kleifaheiði, Ögur, Flateyri, Ísafjörður and Súðavík, but to a far smaller extent than at Patreksfjörður, as the mean error at these stations is less than half the observed wind speed. The large error at Patreksfjörður appears to be re-

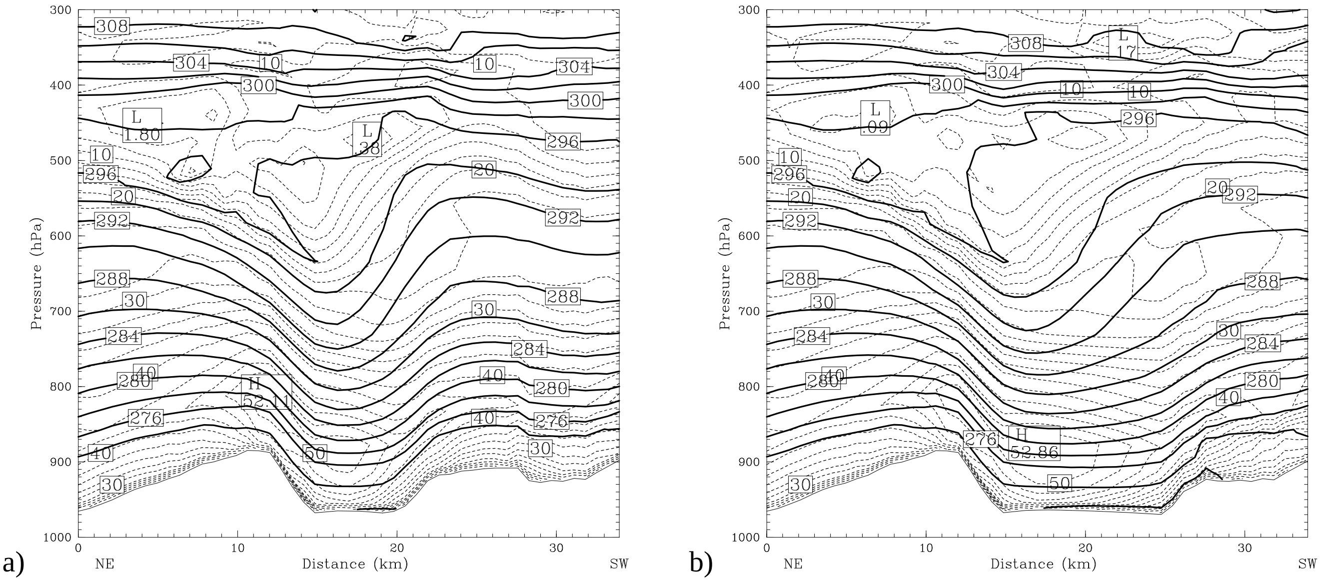

Figure 14: Section from A to B in Fig. 2, at 00 UTC on 2 February 2002. a) modified topography, potential temperature isolines [K] (solid) with intervals of 2 K and wind speed [m/s] isolines (dotted) with intervals of 2.5 m/s at a resolution of 1 km. b) true topography.

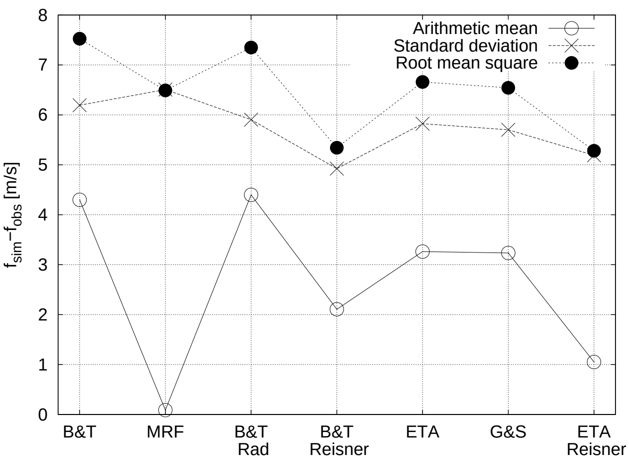

Figure 15: Mean difference of simulated and observed surface winds at thirteen stations in Northwest-Iceland for different simulation setups and a resolution of 1 km.

lated to a boundary-layer separation not being captured correctly by the model. Similarily to Patreksfjörður, the Flateyri station is located immediately downstream of a steep mountain and yet, there are much stronger winds at Flateyri than at Patreksfjörður. The Flateyri winds are well reproduced at a resolution of 1 km. Comparing these two stations, Flateyri and Patreksfjörður, indicates that the lee-side sheltering is presumably quite sensitive to details in the shape of the topography.

Weak winds, and no indications of the passage of the storm, are observed at Bíldudalur, Bolungarvík, Ísafjörður and Súðavík. All four of these stations are located upstream of steep and high mountains (Fig. 2). Climatic data from Bíldudalur (Fig. 16) indicates that northeasterly winds are in fact common but however never strong. The complete absence of northeasterly windstorms at Bíldudalur confirms the topographic nature of the Bíldudalur sheltering. The flow pattern in Fig. 10 indicates indeed that the flow is being decelerated by the downstream mountains. This observed deceleration may be a case of boundary layer blocking, where the surface winds are blocked, while the airflow at higher levels is not (CHEN and SMITH, 1987).

At a horizontal resolution of 333 m, the flow field is indeed realistic. There are well-defined areas of weak winds upstream of the mountains, but there is only marginal improvement in the simulated wind speed at Bíldudalur. This slight improvement is presumably associated with the steepness and the height of the downstream mountains being better represented in the model (Fig. 17). Steeper mountains are indeed more efficient at decelerating airflow than mountains with more gentle slopes (e.g. BAUER et al., 2000; MAYR and GOHM, 2000) The model does in fact reproduce significant deceleration of the flow at a resolution of 333 m and consequently, the large error in the simulated wind at Bíldudalur may only be a moderate error in the extension of the area of upstream deceleration. However, the fact that the wind speed at Bíldudalur is still grossly overestimated by the model, even when the topography is quite correct, indicates that the model is overestimating the vertical mixing at the highest resolutions. This is not surprising, since the parts of the parameterized turbulence can be expected to be resolved and consequently double-counted (e.g. DENG and STAUFFER, 2006).

5.3 Wave activity

The local intensification of the windstorm appears to be associated with gravity waves aloft. The waves are observed by satellite and simulated numerically. They

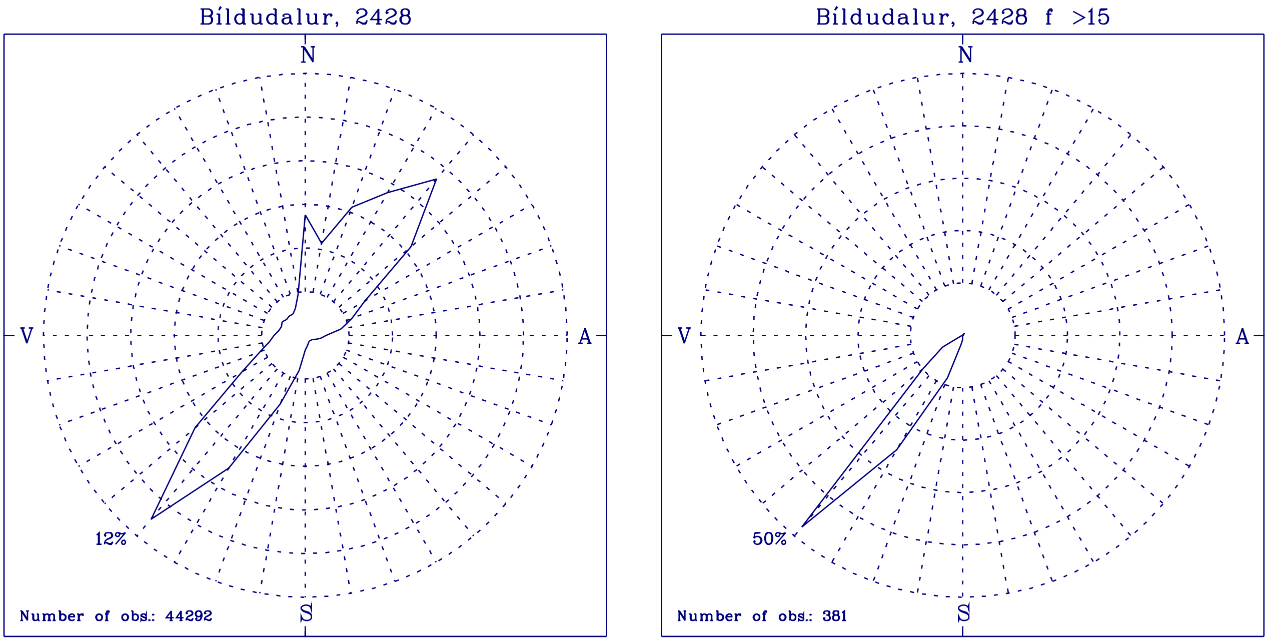

Figure 16: Frequency of wind direction at Bíldudalur for all wind speeds (left) and winds greater than 15 m/s (right). The ratio of maximum number of observations is approx. 1/25. Data from 1998–2003.

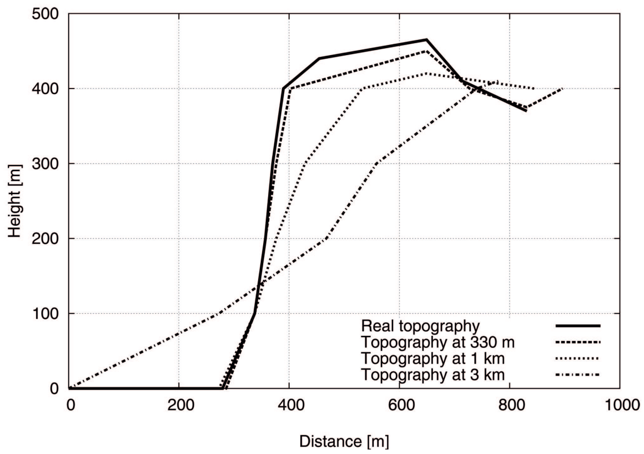

Figure 17: Terrain slope near Bíldudalur at location marked E in Fig. 2. The height is found from the interpolated contours at the respective resolution.

are steep and amplified, but break in a reverse vertical wind shear in the background flow at a level close to 600 hPa. The wave breaking is reminiscent of the downslope windstorm model proposed by Smith (1985). There are indications, e.g. in the complex θ-surfaces in Fig. 13, that the wave energy is partly transported away from the regions of the strongest wave motion above the mountains, in a direction perpendicular to the flow, i.e. to the northwest. This is possibly to some extent related to the clockwise veering of the winds above the boundary layer. The wave signature is in fact weaker at coarser resolutions (not shown), emphazising the need for high horizontal resolution when predicting the atmospheric turbulence associated with waves in this region.

Tests with modified topography show that if the next mountain downstream is only 6, but not 12 km, downstream of the downslope windstorm, the magnitude of the surface windstorm is reduced. Calculating the natural wavelength of the gravity waves by averaging the Brunt-Väisälä frequency and the wind speed separately in the lower part of the troposphere (Fig. 14a) gives 10–20 km and the dampening of the windstorm may be explained by the fjord being too narrow to allow for a full development of the gravity wave. Another way of explaining the reduction in the wind speed in the case of a narrow fjord is that the positive pressure anomaly created by the downstream mountain is reducing the acceleration in the downslope flow further upstream. This result calls for further tests of the impact of surrounding topography on downslope flow. Such a study may explain why downslope windstorms are never observed downstream of some mountains that may seem ideal for creating such storms, e.g. in very narrow fjords.

6 Summary and conclusions

In this paper, a violent windstorm that hit the complex terrain of Northwest-Iceland has been studied and simulated with various numerical configurations. Great spatial variability in wind speed is observed and the strongest surface winds are found below amplified and probably breaking gravity waves in the middle troposphere. Numerical simulations reproduce the windstorm and the variability in wind speed with increasing accuracy as horizontal resolution is increased. The greatest errors in the simulations are an overestimation of the wind speed at locations immediately downstream and immediately upstream of a steep mountain. At both places the flow can be expected to be very sensitive to details in the topography, and in the upstream case strong deceleration is simulated very close to the observation site. The fact that the model simulates stronger winds than ever observed on the upstream side of the mountains indicates that the turbulence may indeed be overestimated at resolutions greater than 1 km. The surface winds during the windstorm are only moderately affected by the method of representing surface friction, and the magnitude of the downslope windstorms shows some sensitivity to the distance to the next downstream mountain. This study indicates very strongly that every increase in horizontal resolution in steps from 9 km to 1 km and even beyond 1 km improves the representation of the local variability of winds in strong windstorms in complex terrain. Simulations at such high resolutions can be expected to be very beneficial for local weather forecasts.

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