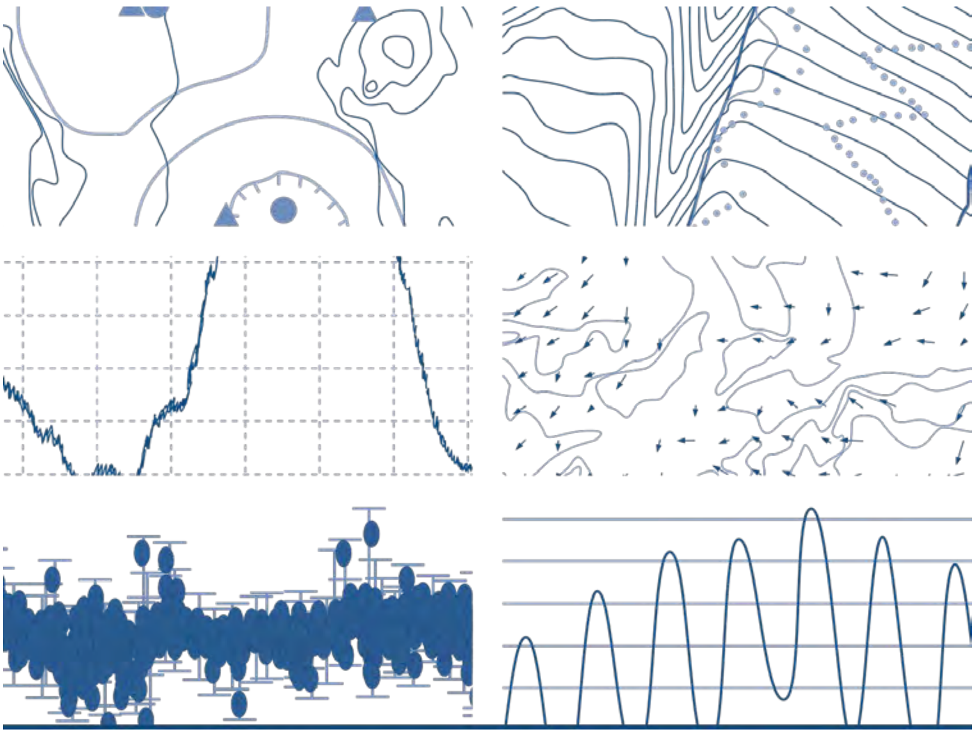
Icelandic
Meteorological Office

Veðurstofa
Íslands



Statistical Correction of WRF Mesoscale Model Simulations of Surface Wind over Iceland based on Station Data

Nikolai Nawri
Guðrún Nína Petersen
Halldór Björnsson
Kristján Jónasson

Report

VÍ 2012-011

Icelandic
Meteorological Office

Statistical Correction of WRF Mesoscale Model Simulations of Surface Wind over Iceland based on Station Data

Nikolai Nawri, Icelandic Met Office
Guðrún Nína Petersen, Icelandic Met Office
Halldór Björnsson, Icelandic Met Office
Kristján Jónasson, University of Iceland

Report

Veðurstofa Íslands
Bústaðavegur 7–9
150 Reykjavík

+354 522 60 00
vedur@vedur.is

VÍ 2012-011
ISSN 1670-8261

Keypage

Veðurstofa Íslands

Report no.:	Date:	ISSN:	Public ☒ Restricted ☐
VÍ 2012-011	September 2012	1670-8261	Provision:

Report title / including subtitle Statistical Correction of WRF Mesoscale Model Simulations of Surface Wind over Iceland based on Station Data	**No. of copies:** 20 **Pages:** 26 **Managing director:** Jórunn Harðardóttir
Authors: Nikolai Nawri Guðrún Nína Petersen Halldór Björnsson Kristján Jónasson	**Project manager:** Halldór Björnsson **Project number:** 5813-0-0004
Project phase:	**Case number:** 2012-333

Report contracted for:

IceWind Project

Prepared in cooperation with:

Summary:

In this study, local station data are used to statistically correct Weather Research and Forecasting (WRF) Model fields of surface wind speed. Separately for three categories of the geostrophic wind speed in each month, the correction is applied through linear transformation of model data at each grid-point, whereby the rescaling factor and offset are determined at station locations by comparison with measurements, such that locally modelled averages and standard deviations are equal to the measured values. Differences between corrected model surface wind speeds and interpolated measurements are positively correlated with terrain elevation and mean curvature, and negatively correlated with terrain sheltering. They are therefore related to the model terrain in a qualitatively realistic fashion.

Keywords: Surface wind speed and direction, effects of large-scale circulation, statistical correction, weather research and forecasting, model station data, gridded surface measurements, Iceland	**Managing director's signature:** Jórunn Harðardóttir **Project manager's signature:** **Reviewed by:**

Contents

List of Figures

1 Introduction

This study is a continuation of previous work (Nawri et al., 2012b), in which the performance of a particular setup (Rögnvaldsson et al., 2007, 2011) of the Weather Research and Forecasting (WRF) Model (Skamarock et al., 2008) with regard to the surface wind conditions over Iceland was analysed. In comparison with local and interpolated station data it was determined that WRF model simulations show systematic biases in surface wind speed across the land area of Iceland. In this study, local station data is therefore used to correct model fields of surface wind speed.

The statistical correction follows a methodology which is similar to that employed for the development of empirical terrain models, discussed in Nawri et al. (2012a). At first, measurements and model data are projected to mean sea level, based on a linear least squares regression model. At station locations, correction factors and offsets are determined for a linear transformation of the model data, such that local model averages and standard deviations are equal to the statistical properties of the measurements. The local correction factors and offsets are then horizontally interpolated onto the model grid and applied to the monthly model fields projected to mean sea level. Using the linear model, the corrected fields are then projected back onto the elevated terrain. In principle, the same methodology can be applied to other variables, for which a sufficient amount of station data is available, and meaningful averages and standard deviations can be calculated (which excludes wind direction). However, in this study, only surface wind speed is discussed.

As in the previous study, measured and modelled surface wind conditions are analysed for individual months, and for different geostrophic wind conditions at 850 hPa, as determined by the European Centre for Medium-Range Weather Forecasts (ECMWF) operational analyses. The largest differences in modelled surface wind speed for different directions of the geostrophic wind are found at upper leeside slopes of the highest terrain (Nawri et al., 2012b). No surface data is available at these locations to test whether the windward (leeside) decrease (increase) in wind speed over the given terrain, and under the given boundary-layer conditions is realistic. Since these effects are not captured by the current network of surface stations, for the statistical correction of WRF model results discussed here, geostrophic wind directions are ignored. However, measured surface wind speeds do show a systematic dependence on the intensity of the geostrophic wind (Nawri et al., 2012a). Geostrophic wind speeds at 850 hPa are therefore taken into account for the correction of monthly WRF model surface wind speeds.

2 Data

All data sources for this study are described in Nawri et al. (2012b). Here, however, in contrast with the previous studies, small-scale variability of measured surface wind speed, which is not properly represented by WRF model simulations on the 3-km grid, is eliminated by distance-weighted horizontal averaging of locally measured values. Due to the wide range in station elevations of almost 900 m, prior to horizontal averaging, local measurements of surface wind speed are projected to mean sea level in the manner described in Nawri et al. (2012a), using linear vertical terrain gradients, $c_s(l,m)$, determined separately for each month, $m = 1,\dots,12$, as well as for three categories,

$l = 1,2,3$, of the geostrophic wind speed at 850 hPa, with limits at the 30th and 65th percentile within each month. At the i-th station location, the smoothed surface wind speed at time t is then given by

$$\tilde{O}(t,i) = \frac{\sum_j \exp(-\lambda r_{ij})\,(O_l(t,j) - c_s(l(t),m(t))\,(z_s(j)+h))}{\sum_j \exp(-\lambda r_{ij})} + c_s(l(t),m(t))\,(z_s(i)+h)\ , \quad (1)$$

where $O_l(t,j)$ are the locally measured speeds, $z_s(j)$ is the j-th station elevation, and h is the reference height of measurements. Differences in sensor height are accounted for by correcting all surface measurements to $h = 10$ mAGL (Nawri et al., 2012b). Additionally, r_{ij} is the distance[1] between the i-th and the j-th station, and $\lambda = \log(2)/r_h$, with half-width r_h. Horizontal distances to the nearest station vary between 1 and 50 km, with an average of 15 km, which is chosen as the half-width of the exponential weighting function. The sums run over all station locations for which data are available at time t, including the i-th. However, due to the large distances between some stations, missing data at the i-th station are not filled in by regional averaging. The smoothed time-series are rescaled,

$$O(t,i) = \alpha(t)\tilde{O}(t,i)\ , \quad (2)$$

such that at each time-step the total kinetic energy of the measured surface wind at all stations is preserved. Ignoring differences in air density between different station locations at any given time, the factor $\alpha(t)$ is then given by

$$\alpha(t) = \sqrt{\frac{\sum_j O_l(t,j)^2)}{\sum_j \tilde{O}(t,j)^2)}}\ . \quad (3)$$

In the following, "measured" surface wind speeds, or "measurements," will always be referring to the horizontally averaged values, rather than the actual local measurements. Also, the empirical terrain models used for comparison, and developed as described in Nawri et al. (2012a), are based on the regionally averaged data.

3 Original Model Fields

Horizontal model fields of surface wind speed are separated by month, as well as into three categories of the geostrophic wind speed at 850 hPa, with limits at the 30th and 65th percentile for each month. The ensemble average fields of WRF modelled surface wind speed for each of the three geostrophic wind speed categories in January and July are shown in Figure 1. In winter and summer, modelled surface wind speeds increase more systematically with increasing intensity of large-scale pressure gradients than the interpolated measurements of surface wind speed based on the one-parameter empirical terrain model (see Figure 16 of Nawri et al. (2012a)). This is primarily due to low modelled surface wind speeds under the influence of weak large-scale forcing imposed by ECMWF operational analyses at the model boundaries. This connection between modelled surface winds and operational analyses, in comparison with measurements, is discussed in more detail in Nawri et al. (2012b).

[1] All horizontal distances in this study are calculated along great circles on a spherical Earth at mean sea level.

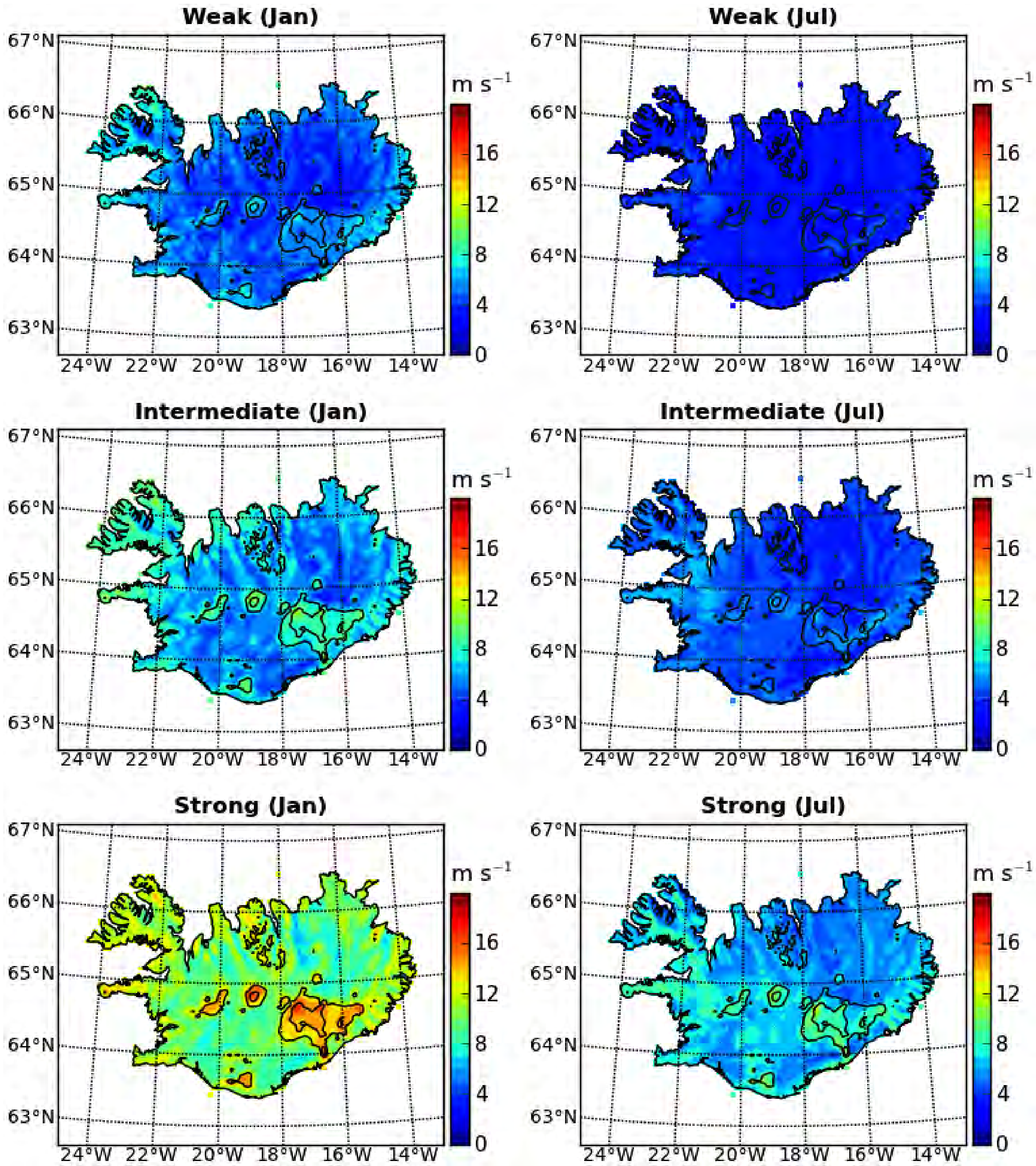


Figure 1. Average WRF model surface wind speeds in January and July for different intensities of the geostrophic wind.

The differences in surface wind speed between the WRF model and the terrain model are shown in Figure 2.[2] In winter, modelled surface wind speeds along the coast are higher relative to interpolated measurements than in the interior of the island. Also, modelled wind speeds increase relative to interpolated measurements with increasing geostrophic wind speeds. For weak geostrophic winds, approximately the same surface wind speeds exist in both datasets along the coast, but with lower modelled wind speeds in the interior, by about 6 m s^{-1} at intermediate elevations, and up to 9 m s^{-1} at the highest terrain points. With increasing intensity of the geostrophic wind speed, a large increase relative to measurements in modelled surface wind speeds is found, especially at low elevations in the interior. For strong geostrophic winds, modelled surface wind speeds in the interior are approximately equal to interpolated measurements, with 6 – 9 m s^{-1} higher modelled surface wind speeds along the coast. In summer, the magnitudes of differences are smaller, both spatially and for different intensities of the geostrophic wind. As in winter, modelled wind speeds increase relative to interpolated measurements, particularly in the interior. For strong geostrophic winds in summer, differences are consistently positive, with 2 – 3 m s^{-1} higher modelled wind speeds in regions with a good coverage of observational data.

Due to these systematic biases of WRF model results, local station data can be used to statistically correct simulations. Since biases differ significantly for different intensities of the geostrophic wind in different months, the statistical correction is applied to individual datasets for each category of the geostrophic wind speed in each month.

[2] These results differ from those discussed in Nawri et al. (2012b), since here individual fields of surface wind speed are not separated into different sectors of geostrophic wind directions.

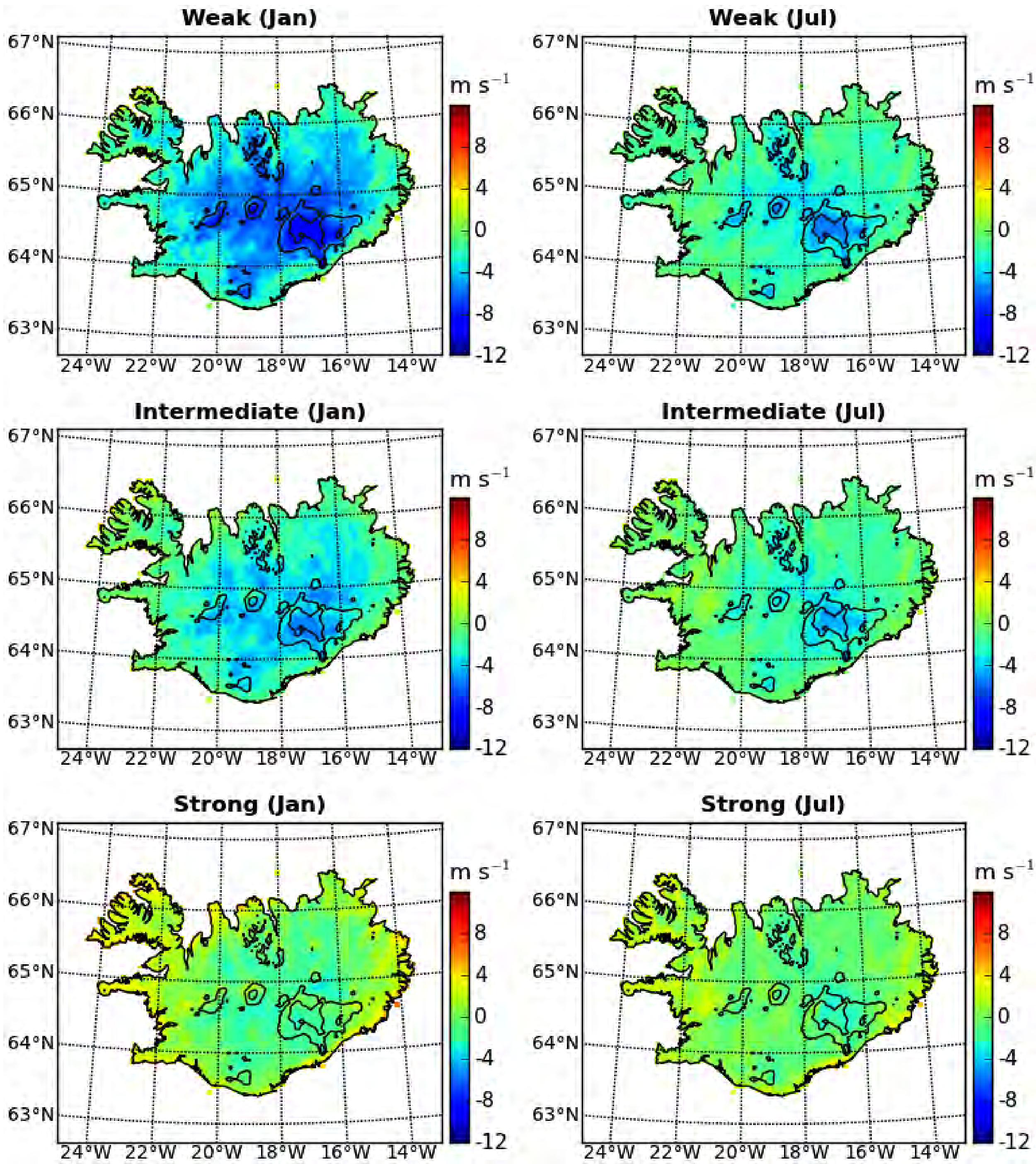


Figure 2. Differences between average WRF model and empirical terrain model surface wind speeds in January and July for different intensities of the geostrophic wind.

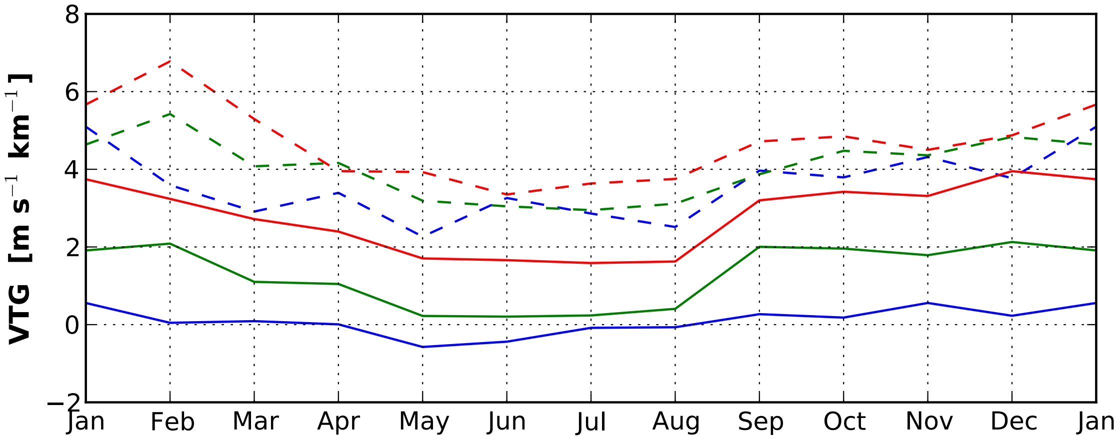


Figure 3. Monthly vertical terrain gradients (VTGs) of surface wind speed based on measurements (dashed lines) and WRF model simulations (solid lines) for weak (blue lines), intermediate (green lines), and strong (red lines) geostrophic winds.

4 Statistical Correction of Model Fields

The deviations of model results from actual measurements, rather than from terrain models, which by themselves contain uncertainties beyond measurement errors, are necessarily only determined at station locations. For the elimination of model biases, local measures of differences between model and station data need to be interpolated horizontally onto the model grid. Due to the strong dependence of surface wind speed on terrain elevation, and to account for differences between model and actual orography, effects of terrain elevation need to be removed from measured and modelled data, before determining local differences and horizontally interpolating between them. In a simple parameterised fashion, this is done by projecting measured and modelled wind speeds to mean sea level, following the methodology described in Nawri et al. (2012a) and Nawri et al. (2012b).

Central to this projection are vertical gradients of wind speed, calculated along the surface of elevated terrain. The measured and modelled vertical terrain gradients, for each of the three categories of geostrophic wind speed in each month, are shown in Figure 3. Throughout the year, measured terrain gradients are systematically larger than those derived from model results. This is due to the fact that, as seen in the previous section, modelled wind speeds along the coast tend to be higher than the measured values, but lower in the interior of the island. With terrain elevation generally increasing towards the interior, vertical terrain gradients of modelled surface wind speed are therefore reduced by the effects of excessive model surface roughness over much of the land area of the island, with the exception of the glaciers (Nawri et al., 2012b). Generally, terrain gradients increase with the intensity of the geostrophic wind, especially for modelled surface wind speeds. This is true either for different categories of geostrophic wind speed in individual months, or for the same category throughout the year, as the limits between categories are higher in winter than in summer. The reason for this is a stronger effect of large-scale forcing at higher elevations than below, due to the reduced land area at or above that level. This reduces the effect of surface drag on momentum transfer from the overlying flow, and increases the wind shear between high and low elevations. Monthly terrain gradients are generally positive throughout the year. The only exception are negative values in May and June for modelled surface wind speeds under the influence of weak geostrophic winds. For model simulations under those large-scale conditions, the negative effect of surface roughness on surface wind speed, increasing towards the interior of the island, outweighs the positive effects from increased terrain elevation.

Given these terrain gradients, measured surface wind speeds, $O(l,m,i)$, for a given category, $l = 1,2,3$, of the geostrophic wind speed in each month, $m = 1,\ldots,12$, at the i-th station location can be projected to mean sea level such that the reference values are given by

$$O_0(l,m,i) = O(l,m,i) - c_s(l,m)\,(z_s(i)+h) \ . \tag{4}$$

Here, the vertical terrain gradients, $c_s(l,m)$, are based on regionally averaged station data, rather than actual local measurements, as described in Section 2. For modelled surface wind speeds, $S(l,m,i)$, the reference values at mean sea level are given by

$$S_0(l,m,i) = S(l,m,i) - c(l,m)\,(z(i)+h) \ , \tag{5}$$

where $z(i)$ is the model terrain elevation at the i-th station location, and $c(l,m)$ is the vertical terrain gradient based on model data. WRF model surface wind speeds are valid at the same height $h = 10$ m above ground as the measurements. The full model reference fields of surface wind speed projected to mean sea level are given by

$$s_0(l,m,x,y) = s(l,m,x,y,z(x,y)+h) - c(l,m)\,(z(x,y)+h) \ , \tag{6}$$

where $s(l,m,x,y,z(x,y)+h)$ is the original field of wind speed at 10 m above the model terrain elevation, $z = z(x,y)$.

For a given correlation, $C(l,m,i)$, between measured and modelled wind speeds projected to mean sea level, a lower bound, $2\,(1-C(l,m,i))$, exists for normalised mean squared deviation (NMSD),

$$N\left[O_0(l,m,i),S_0(l,m,i)\right] = \frac{E\left[(S_0(l,m,i)-O_0(l,m,i))^2\right]}{D[S_0(l,m,i)]\,D[O_0(l,m,i)]} \ , \tag{7}$$

which is attained if and only if modelled averages and standard deviations are equal to the measured values (Nawri et al., 2012b). Within each dataset corresponding to a specific category of the geostrophic wind speed in a given month, locally a reduction in NMSD between measurements and model data to that minimum value can be achieved by transforming model data such that the transformed values are given by

$$\tilde{S}_0(l,m,i) = \frac{D[O_0(l,m,i)]}{D[S_0(l,m,i)]}\,(S_0(l,m,i) - E[S_0(l,m,i)]) + E[O_0(l,m,i)] \tag{8}$$

$$= r(l,m,i)\,S_0(l,m,i) + q(l,m,i) \ , \tag{9}$$

where averages and standard deviations of any set of values are denoted by $E[\,]$ and $D[\,]$, respectively. The local non-dimensional rescaling factor is defined as

$$r(l,m,i) = \frac{D[O_0(l,m,i)]}{D[S_0(l,m,i)]} \ , \tag{10}$$

and the local offset, in units of wind speed, is defined as the difference between the measured and rescaled model average

$$q(l,m,i) = E[O_0(l,m,i)] - \frac{D[O_0(l,m,i)]}{D[S_0(l,m,i)]} E[S_0(l,m,i)] \; . \tag{11}$$

It is therefore different than it would be for the correction of biases in model averages only, not taking into account biases in standard deviations. For each category of the geostrophic wind speed in each month, the transformed model data, $\tilde{S}_0(l,m,i)$, then has the same average and standard deviation as the measured data at the same location, whereas correlation is unaffected by the linear transformation.

Since the wind statistics at station locations are already known from measurements, this methodology is only beneficial, if the correction can be applied to the entire model fields. The rescaling factors and offsets determined locally are therefore interpolated onto the model grid through distance-weighted horizontal averaging, whereby the half-width of the exponential shape function, at each grid-point, is set equal to the distance to the nearest station (see Nawri et al. (2012a) for a detailed description of this interpolation technique). The local values $r(l,m,i)$ and $q(l,m,i)$ are then mapped onto the horizontal fields $\tilde{r}(l,m,x,y)$ and $\tilde{q}(l,m,x,y)$, respectively.

Before applying the described methodology, the improvement or otherwise of model results needs to be established, away from station locations from which measurements were included in determining local correction factors and offsets. This is done analogously to the test of empirical terrain models (Nawri et al., 2012a). For each individual station, the local measurements are omitted from the calculation of correction factors and offsets. The correction factors and offsets determined at the other stations are then interpolated to the omitted station location, and used to correct the interpolated WRF time-series there.

The differences between corrected and original model data, in comparison with the differences between measurements and original model data, are shown in Figure 4. The two differences, by definition, would be identical if local station data were used for the correction. However, even if local station data are not used, the effects of the correction are similar to the original differences. The main deviations between predicted and necessary changes exist for the largest positive original differences, for which the necessary changes are under-predicted by the statistical correction. Those locations are all on elevated terrain in the interior of the island (Figure 2), in a region with relatively large distances between neighbouring stations.

Normalised mean squared deviations between averages of measured and corrected model surface wind speeds are larger in winter than in summer, and larger for weak than for strong geostrophic winds. At the low end, NMSD = 0.09 for strong geostrophic winds in July, and at the high end, NMSD = 0.36 for weak geostrophic winds in January. For comparison, the same test was done for a statistical correction that only takes into account averages, i.e., by setting the rescaling factor equal to one, and horizontally interpolating local differences between measured and modelled averages to determine the required offset. Normalised mean squared deviations between averages of measured and corrected model surface wind speeds are consistently higher than for the correction taking into account standard deviations, with NMSD varying between 0.21 and 0.54. Correcting for differences in standard deviations, in addition to averages, therefore improves the correction of averages away from station locations.

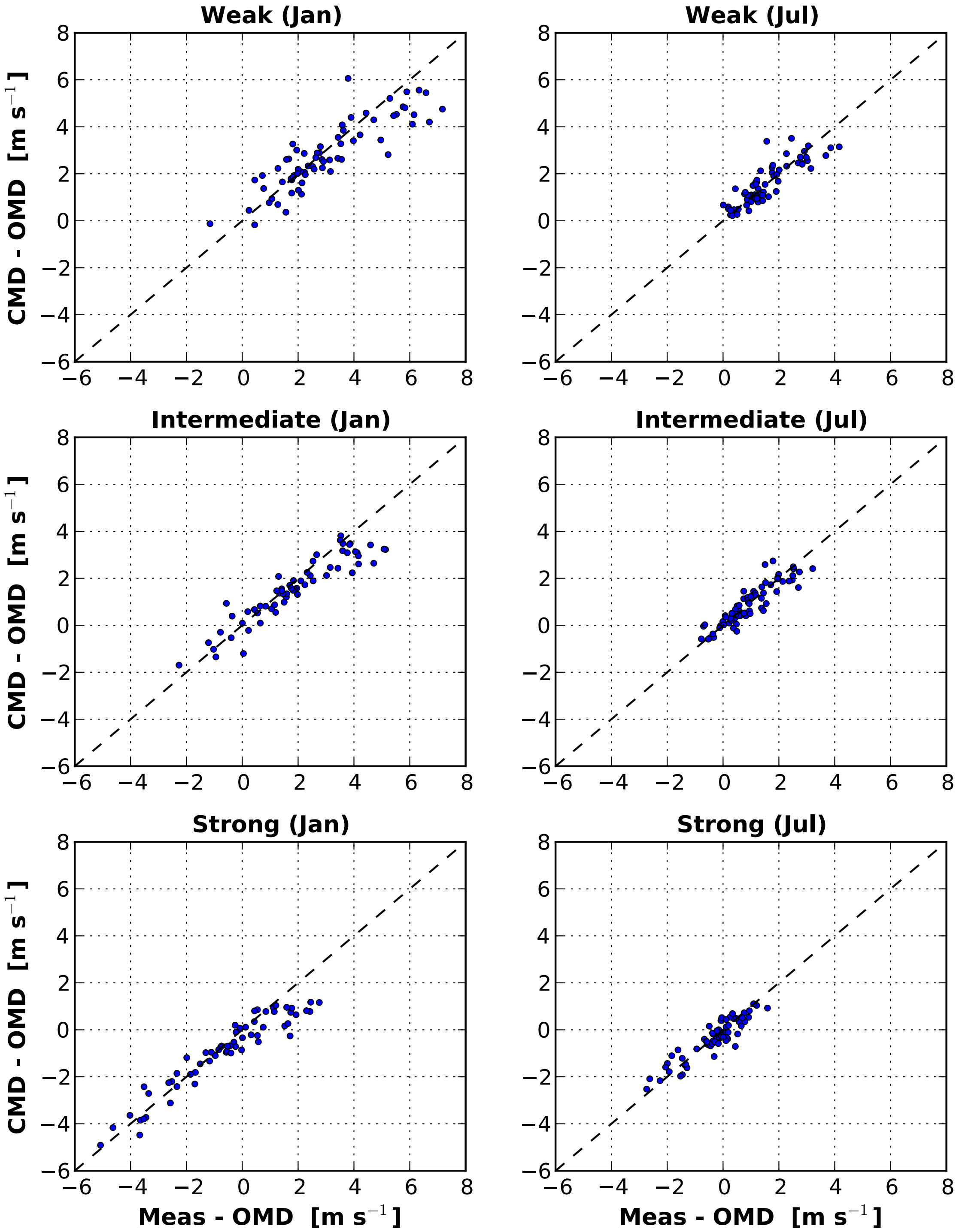


Figure 4. Differences in surface wind speed at station locations between measurements and original WRF model data (OMD), as well as between corrected (CMD) and original WRF model data in January and July for different intensities of the geostrophic wind.

Compared with the errors of empirical terrain models (Nawri et al., 2012a), the largest differences between predicted and necessary changes seen in Figure 4 show a more systematic spatial pattern, which can be used to adjust rescaling factors and offsets. The local error in rescaling factor, $\delta(l,m,i) = \bar{r}(l,m,i) - r(l,m,i)$, is defined as the difference between the test value interpolated onto the omitted station location, $\bar{r}(l,m,i)$, and the actual rescaling factor, as determined in comparison with local measurements. With local errors interpolated onto the model grid in the same fashion as the rescaling factor, $\delta(l,m,x,y)$, the adjusted field of the rescaling factor is defined as

$$r(l,m,x,y) = \tilde{r}(l,m,x,y) - f(d)\,\delta(l,m,x,y)\ , \qquad (12)$$

where d is the horizontal distance of grid-point (x,y) from the nearest station location. The weighting function $f(d) = 1 - \exp(-d/d_0)$, with $d_0 > 0$, is zero for $d = 0$, and positive otherwise. For grid-points close to any station location, $r(l,m,x,y) \approx \tilde{r}(l,m,x,y)$, since there, $\tilde{r}(l,m,x,y)$ is accurately determined from measurements. With increasing distance from the nearest station, the relative magnitude of the correction term increases at a rate determined by d_0, with a maximum value of the weighting function, $f(d_{\max}) = f_{\max}$, for d equal to half the maximum distance between stations, $d_{\max} = 25$ km. In this study, $d_0 = -d_{\max}/\ln(1 - f_{\max})$ is determined such that $f_{\max} = 0.5$. The adjustment of offset is done analogously.

The original and adjusted fields of rescaling factor and offset for intermediate geostrophic winds in January are shown in Figure 5. Since only land-based measurements are available, the correction of model fields is only applied to land grid-points. Over the ocean, the rescaling factor is set to one, and the offset to zero. These values are omitted from the figure. Rescaling factor and offset maintain similar spatial patterns throughout the year, with smaller magnitudes in summer than in winter (not shown). Unlike offset, the rescaling factor has a simple interpretation individually, since standard deviation scales in the same way as time-series. As such, rescaling is always positive. Throughout the year, it is larger in the interior of the island than along the coast, and decreases with increasing intensity of the geostrophic wind (not shown). For strong geostrophic winds, the highest values in the interior are equal to about one. Rescaling is lowest in the southwest and eastern coastal regions, where throughout the year, values are consistently below one. Due to the negative bias of average surface wind speeds in the interior, especially with weak large-scale forcing (as discussed in the previous section), variability of a positive variable such as wind speed is necessarily reduced. For weak geostrophic winds, this results in a reduction of standard deviations by a factor of less than 1/2, or inversely, in correction factors of greater than 2. On the other hand, along the coast and under the influence of strong geostrophic winds, modelled variability, as measured by standard deviation, is too large by up to a factor of 2. Interpolated errors of the rescaling factor are mostly positive, and largest at the maxima of the original rescaling factors, i.e., in the interior of the island. The effects of the adjustment are therefore primarily to increase rescaling around the maximum values, rather than affecting the maximum values themselves, which are close to station locations. Unlike the rescaling factor, offset is not systematically related to the intensity of the geostrophic wind. It also does not show a consistent qualitative difference between the interior of the island and along the coast. However, consistently low values are found in the Westfjords, whereas large positive values are consistently found in the southwest. Offset is positive throughout most of the island, with significant negative values at some inland locations. Offset errors tend to have the same sign as the original values. Adjustment therefore tends to increase the magnitude of extreme values.

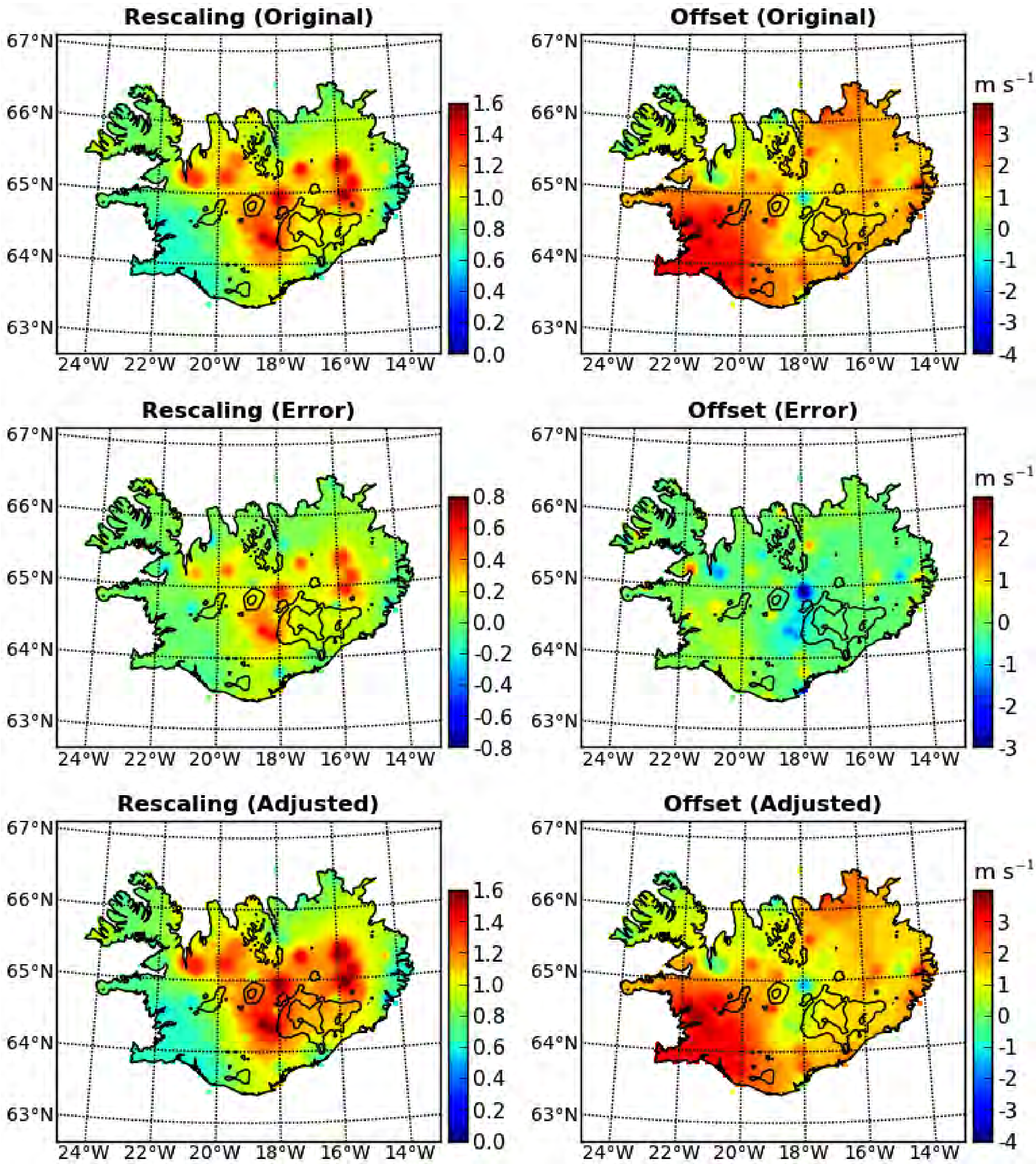


Figure 5. Interpolated original and adjusted correction factors and offsets for surface wind speed in January for intermediate geostrophic winds. See Section 4 for a definition of errors.

5 Corrected Model Fields

The linear transformation of model data, discussed in the previous section, eliminates horizontal biases in local averages and standard deviations at mean sea level. By projection back onto the elevated terrain, the vertical dependence of surface wind speeds can be corrected as well by using the terrain gradients determined from measurements, rather than those determined from the original model fields. The corrected surface wind speed on the WRF model orography for each category of the geostrophic wind speed in each month is then given by

$$\begin{aligned} s_c(l,m,x,y,z(x,y)) = r(l,m,x,y)\,(s(l,m,x,y,z(x,y)) - c(l,m)\,(z(x,y)+h)) \\ +q(l,m,x,y) + c_s(l,m)\,(z(x,y)+h)\ . \end{aligned} \tag{13}$$

For the three categories of the geostrophic wind speed in January and July, differences of the corrected from the original (Figure 1) model fields of surface wind speed are shown in Figure 6. These differences are qualitatively similar, though not identical, to the inverse of the differences between the original model fields and the empirical terrain model that were shown in Figure 2. The overall effects of the statistical correction primarily are to increase model wind speeds, especially in the interior and under conditions of weak geostrophic winds. Negative correction only occurs along the coast and with strong large-scale forcing in winter.

The corrected WRF model fields are not identical to the corresponding terrain models on the same orography, since horizontal gradients of original and corrected model fields are dynamically motivated, rather than determined by horizontal averaging, as in the case of terrain models. This raises the important question about what realistic information is gained from the corrected model fields, in addition to local and interpolated surface measurements. For the three categories of the geostrophic wind speed in January and July, differences of the corrected model fields of surface wind speed from the corresponding terrain models are shown in Figure 7. Although measured wind speeds are not exactly reproduced by either the terrain models or the corrected WRF model fields, differences between the two near station locations are negligible compared with inter-station differences. In winter, large differences with magnitudes of at least 4 m s^{-1} are exclusively positive, and are situated on intermediate and high terrain elevations. The largest differences are found on the northern and western slopes of Hofsjökull, as well as along the slopes of Vatnajökull. In summer, the magnitude of differences is reduced by about a factor of 2. However, the most significant differences are also positive, and concentrated in places with complex terrain and steep slopes.

This suggests a more systematic investigation of the relationship between measured and modelled surface wind speeds, including the differences between them, and terrain characteristics based on the WRF model orography. In addition to elevation, two measures of terrain variability are considered here: mean curvature and sheltering.

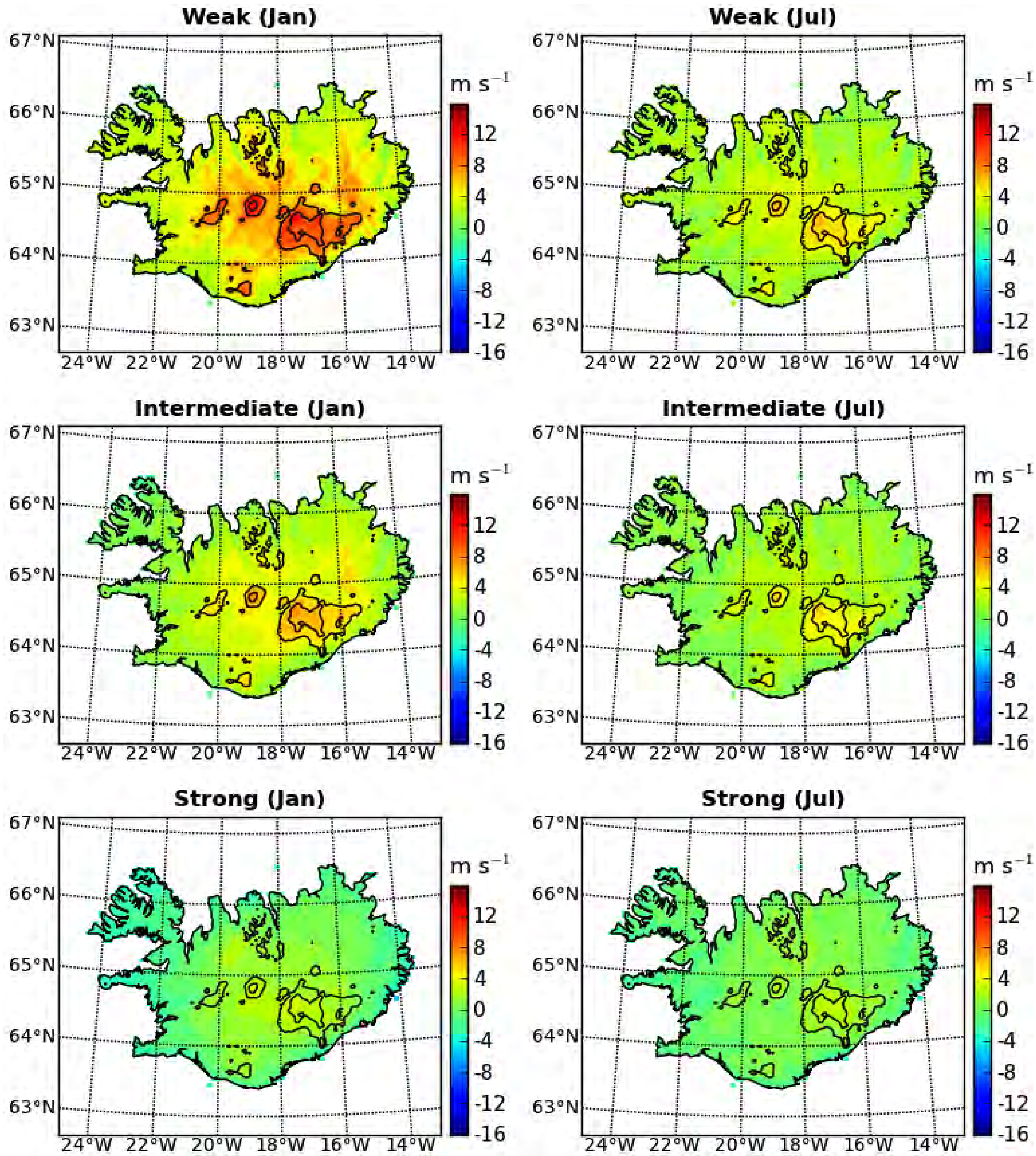


Figure 6. Differences between average corrected and original WRF model surface wind speeds in January and July for different intensities of the geostrophic wind.

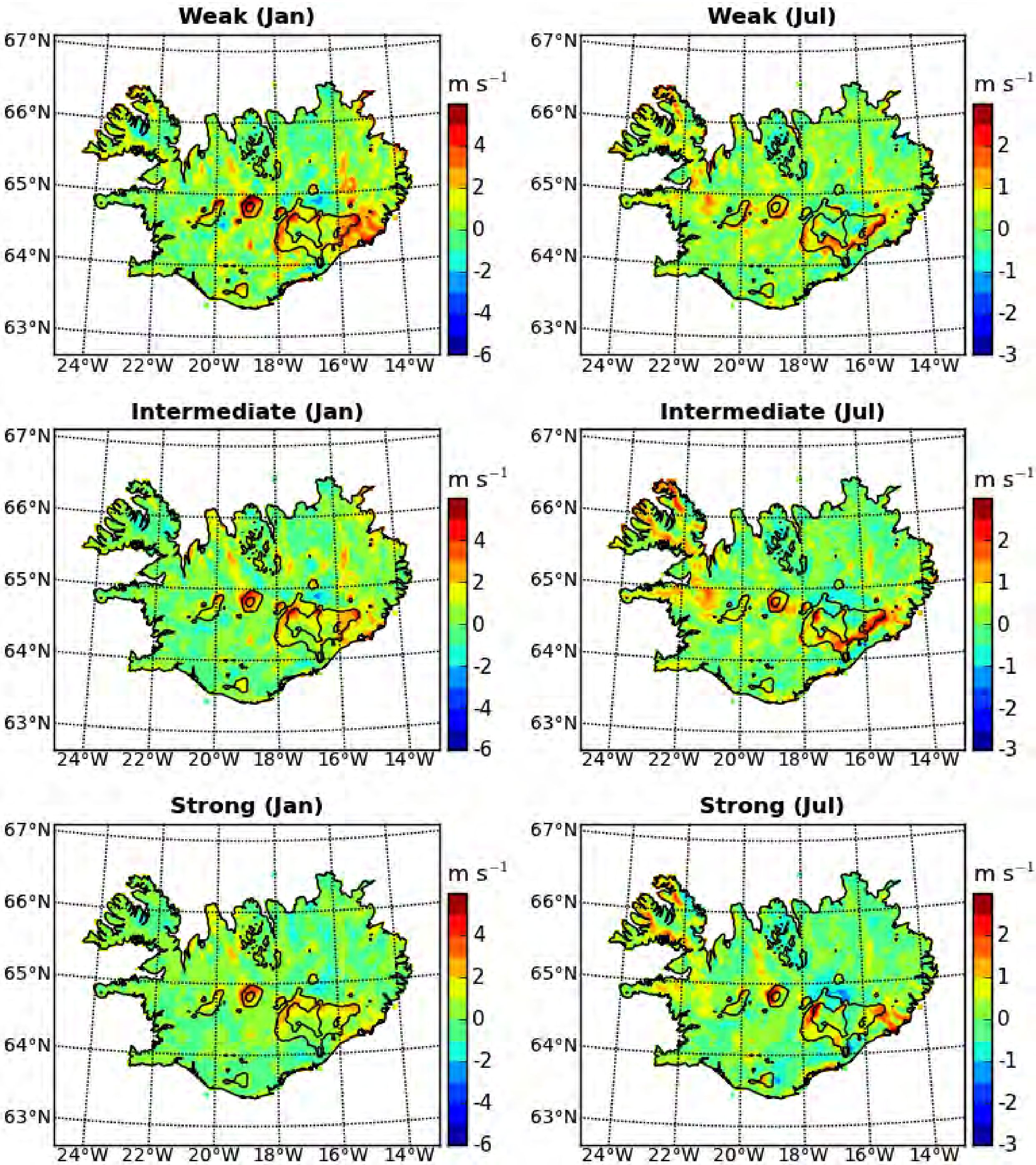


Figure 7. Differences between average corrected WRF model and empirical terrain model surface wind speeds in January and July for different intensities of the geostrophic wind. Note the difference in wind speed scale between January and July.

Mean curvature of any two-dimensional surface, $Z = Z(x,y)$, embedded in three-dimensional space is defined as (e.g., Spivak, 1999, Chap. 3)

$$H = -\frac{\left(1+\left(\frac{\partial Z}{\partial x}\right)^2\right)\frac{\partial^2 Z}{\partial y^2} - 2\frac{\partial Z}{\partial x}\frac{\partial Z}{\partial y}\frac{\partial^2 Z}{\partial x \partial y} + \left(1+\left(\frac{\partial Z}{\partial y}\right)^2\right)\frac{\partial^2 Z}{\partial x^2}}{2\left(1+\left(\frac{\partial Z}{\partial x}\right)^2+\left(\frac{\partial Z}{\partial y}\right)^2\right)^{3/2}} . \tag{14}$$

With the normal vector pointing upwards, mean curvature is negative (positive) over a concave (convex) surface. As a result, with negative extrema at valley bottoms, and positive extrema on summits and along ridges, regardless of absolute terrain elevation, mean curvature can be expected to be positively correlated with low-level wind speeds.

Sheltering is defined here as the difference between the highest terrain elevation within a radius of 30 km and the local elevation. As a measure of the magnitude of terrain blocking, sheltering can be expected to be negatively correlated with low-level wind speeds.

Average surface wind speed as a function of mean terrain curvature, based on the empirical terrain model, as well as on the original and corrected WRF model simulations, for the three categories of the geostrophic wind speed in January and July are shown in Figure 8. Throughout the year, and for all intensities of the geostrophic wind, terrain model and corrected WRF model surface wind speeds show a systematic near-linear dependence on terrain curvature. The same is true for the original WRF model wind speed under the influence of strong geostrophic winds. However, with weak and intermediate geostrophic winds, the dependence of surface wind speed on terrain curvature is less well defined. Similarly, the results for sheltering (not shown) confirm the initial expectation of a systematic decrease in average surface wind speed with increasing blocking effects.

Having established the general dependence of measured and modelled surface winds on mean terrain curvature and sheltering, in addition to the previously known dependence on terrain elevation, the dependence of differences between modelled surface wind speeds and interpolated measurements on those terrain characteristics can be investigated, focusing on regions between station locations. For this it is necessary to determine changes in model elevation and other terrain characteristics within these intervening regions, relative to the surrounding station locations. Therefore, WRF model terrain characteristics at station locations are interpolated onto the model grid through distance-weighted horizontal averaging, analogous to the interpolation of the correction factors and offsets. These background fields of terrain characteristics are then subtracted from the actual model fields. By definition, the resulting perturbation terrain characteristics are small at those grid-points nearest station locations, but potentially increase in magnitude between stations. If differences between the corrected WRF model fields of surface wind speed and the corresponding terrain model are dynamically motivated, these differences are expected to be positively correlated with perturbation terrain elevation and mean curvature, and negatively correlated with perturbation terrain sheltering.

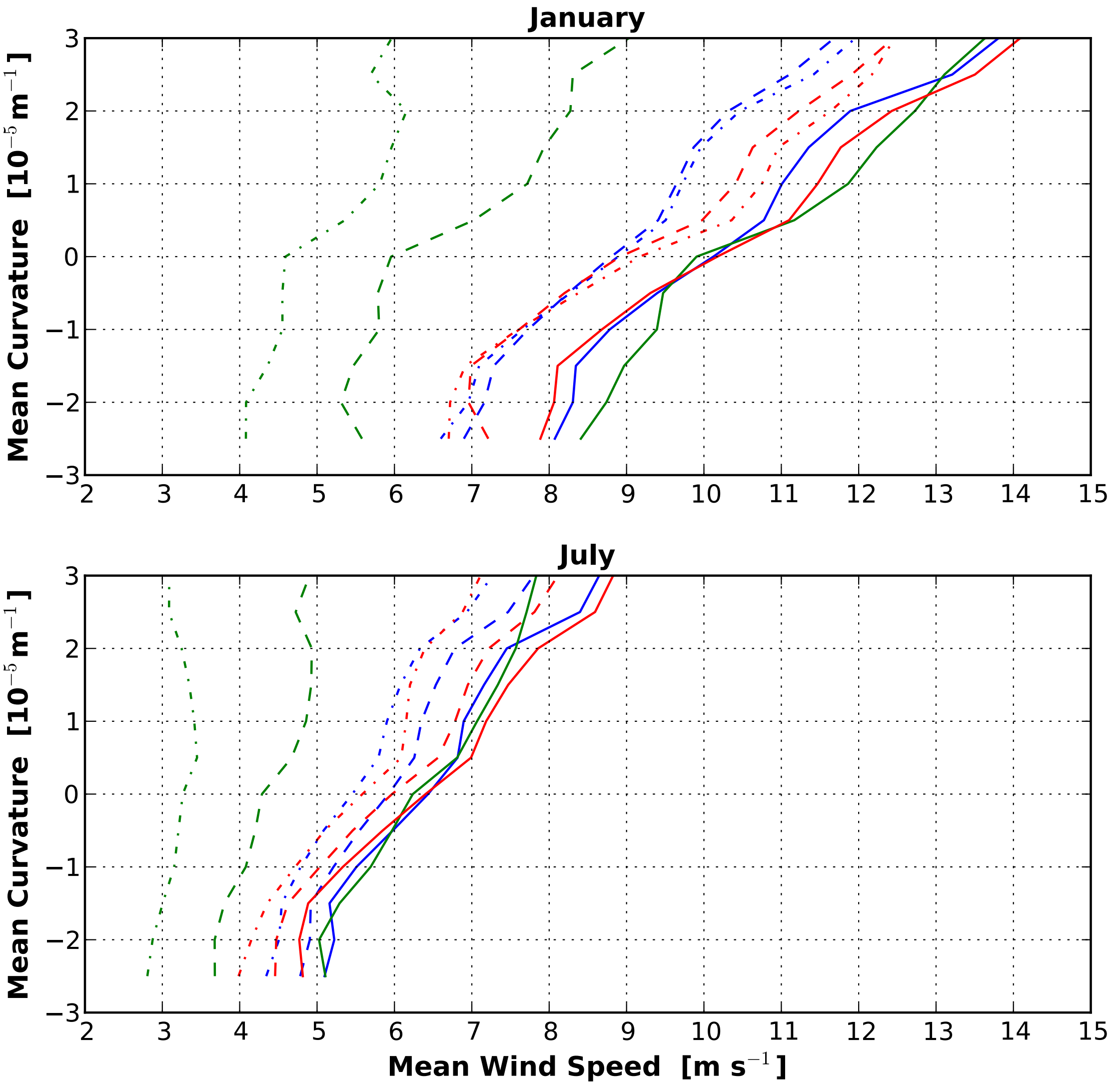


Figure 8. Average surface wind speed as a function of mean terrain curvature based on measurements (blue lines), original WRF model simulations (green lines), and corrected WRF model simulations (red lines) for weak (dash-dotted lines), intermediate (dashed lines), and strong (solid lines) geostrophic winds.

As shown in Figure 9 for the wind conditions in January, contrary to these expectations, the original differences between WRF model surface wind speeds and interpolated measurements are negatively correlated with perturbation terrain elevation, especially with weak geostrophic winds. However, the corrected WRF model wind speeds do show the positive correlation with perturbation elevation, that is expected if deviations of model results from interpolated measurements are realistically influenced by the model orography. For the original model wind speeds, these dynamical terrain-related effects are masked by the excessive model surface roughness over much of the land area of the island (Nawri et al., 2012b). Similarly, after spurious effects of surface roughness are removed from the WRF model fields, the differences between corrected model surface wind speeds and interpolated measurements are positively correlated with perturbation terrain curvature, and negatively correlated with perturbation terrain sheltering (not shown). At least qualitatively, deviations of corrected model results from interpolated measurements away from station locations are therefore dynamically motivated, in a manner that is realistic with respect to the model terrain.

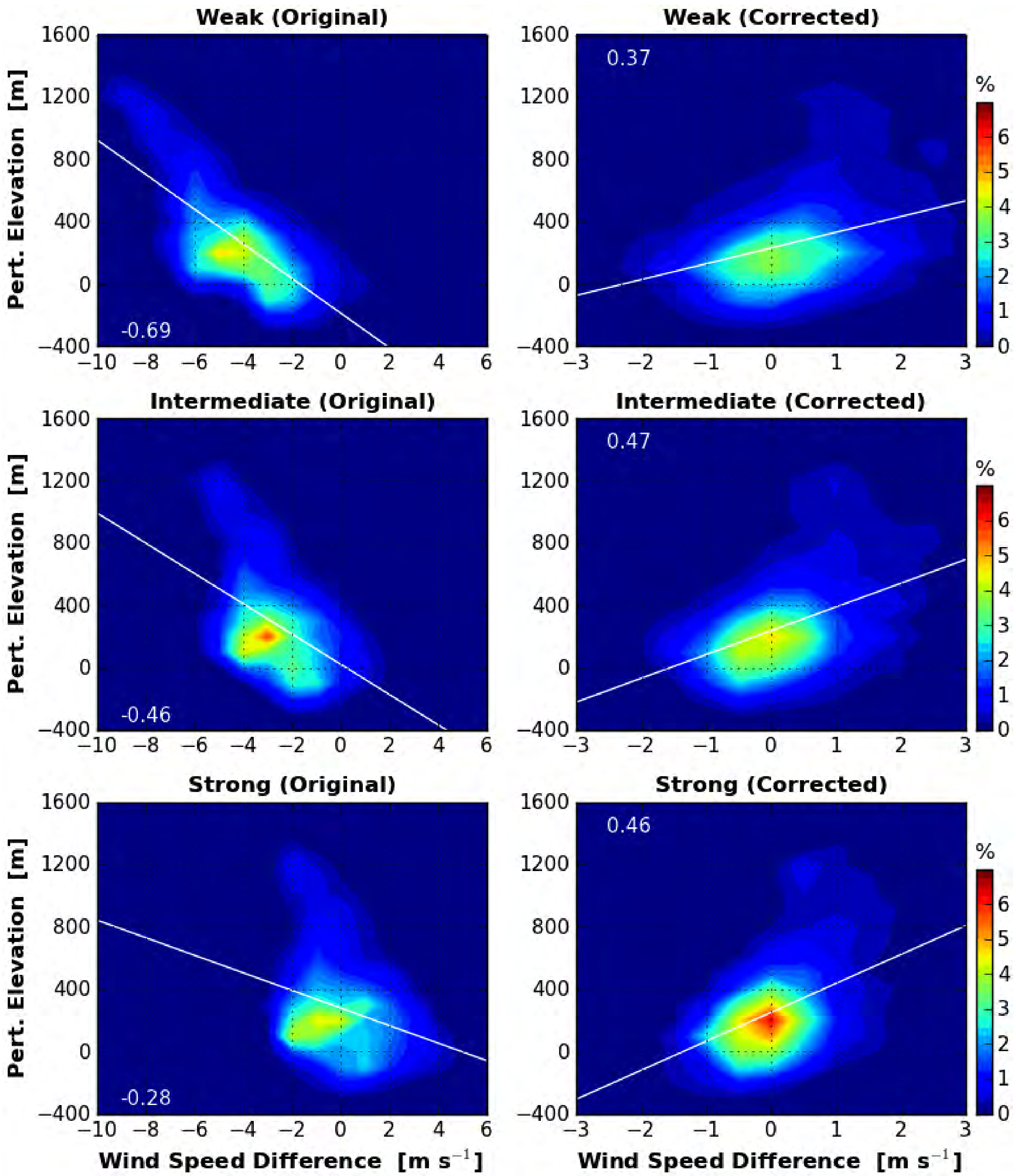


Figure 9. Joint histograms of perturbation terrain elevation and differences between average WRF model and empirical terrain model surface wind speeds in January for different intensities of the geostrophic wind.

6 Summary

Previously (Nawri et al., 2012b), in comparison with local and interpolated station data, it has been established that WRF model simulations show systematic biases in surface wind speed across the land area of Iceland. In this study, local station data are therefore used to statistically correct model fields of surface wind speed.

Separately for three categories of the geostrophic wind speed in each month, the correction is applied through linear transformation of model data at each grid-point, whereby the rescaling factor and offset are determined at station locations by comparison with measurements, such that locally modelled averages and standard deviations are equal to the measured values. The local values of rescaling factor and offset are then interpolated onto the model grid through distance-weighted horizontal averaging. To take into account the strong dependence of surface wind speed on terrain elevation, as well as differences between model and actual orography, prior to correction, measured and modelled wind speeds are linearly projected to mean sea level. Tests of the correction methodology indicate a generally good agreement of corrected model fields with measurements away from station locations, at which data were used for correction.

By definition, differences between corrected WRF model fields and interpolated measurements on the model orography are small in the vicinity of station locations. Between stations, these differences are due to the fact that horizontal gradients of original and corrected model fields are dynamically motivated, rather than determined by horizontal averaging, as in the case of terrain models. Differences between corrected model surface wind speeds and interpolated measurements are related to local differences in model terrain characteristics relative to the surrounding station locations. Specifically, differences in surface wind speed are positively correlated with perturbation terrain elevation and mean curvature, and negatively correlated with perturbation terrain sheltering. They are therefore related to the model terrain in a qualitatively realistic fashion.

Given the current network of surface weather stations, and in combination with measurements, there are some benefits from WRF mesoscale modelling for establishing past climatological surface wind conditions over Iceland, especially in regions between surface weather stations, where the terrain varies on a scale that is adequately represented by the model.

The statistical correction discussed here indirectly removes effects on low-level winds of excessive model surface roughness over much of the land area of the island. In an ongoing study, this is done by reverse modelling, using simple boundary-layer parameterisations.

References

Nawri, N., Björnsson, H., Jónasson, K., and Petersen, G. N. (2012a). Empirical terrain models for surface wind and air temperature over Iceland. Report VÍ 2012-009, Icelandic Meteorological Office, Reykjavik, Iceland.

Nawri, N., Björnsson, H., Jónasson, K., and Petersen, G. N. (2012b). Evaluation of WRF mesoscale model simulations of surface wind over Iceland. Report VÍ 2012-010, Icelandic Meteorological Office, Reykjavik, Iceland.

Rögnvaldsson, Ó., Ágústsson, H., Einarsson, E. M., Ólafsson, H., Björnsson, H., and Sveinsson, Ó. G. B. (2007). Stöðuskýrsla vegna fyrsta árs RÁV verkefnisins. Technical report, Reiknistofa í veðurfræði, Reykjavik, Iceland.

Rögnvaldsson, Ó., Ágústsson, H., and Ólafsson, H. (2011). Aflræn niðurkvörðun veðurs innan LOKS verkefnisins. Technical report, Reiknistofa í veðurfræði, Reykjavik, Iceland.

Skamarock, W. C., Klemp, J. B., Dudhia, J., Gill, D. O., Barker, D. M., Duda, M. G., Huang, X.-Y., Wang, W., and Powers, J. G. (2008). A description of the Advanced Research WRF Version 3. NCAR Technical Note NCAR/TN-475+STR, National Center for Atmospheric Research, Boulder, Colorado, USA.

Spivak, M. (1999). *A Comprehensive Introduction to Differential Geometry, Volume Three, 3rd Edition*. Publish or Perish, Houston, TX, USA, 314 pp.