The Impact of Sea Spray Evaporation in a Numerical Weather Prediction Model

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ABSTRACT

Strong winds above the sea lead to large amounts of spray in the lowest part of the boundary layer. Through their evaporation, spray droplets influence the exchange of sensible and latent heat between sea and air. In this study, the impact of spray on the atmosphere is investigated using the numerical weather prediction high-resolution limited-area model HiRLAM. The effect of spray is taken into account via modified surface heat fluxes. The impact on forecasts is then investigated in two case studies of intense midlatitude storms. A general consequence of spray is a significantly cooler and moister surface layer. Indirect effects are reported as well, such as enhanced precipitation, and associated midlevel latent heat release. The second case study, which deals with a rapidly deepening depression, indicates that including spray leads to a marginal intensification of the depression.

1. Introduction

The surface fluxes over the sea are important boundary conditions for atmospheric models. In many studies, the sensitivity of such models to the parameterization of surface momentum and heat fluxes has been investigated (e.g., Miller et al. 1992; Beljaars 1995). Whereas at moderate wind speeds the surface heat fluxes above the sea are reasonably known, during storm conditions a complicating factor is present in the form of sea spray. Through the evaporation of spray droplets, the fluxes might be significantly modified, which can, in turn, have effects on the atmosphere.

Especially, tropical cyclones might be sensitive to sea spray, not only because of the extreme wind speeds involved, but also because of the importance of the air–sea fluxes in their dynamics. Fairall et al. (1994) claimed that, without taking into account evaporating spray droplets, the boundary layer of a modeled tropical cyclone evolves in an unrealistic manner. Kepert et al. (1999) and Bao et al. (2000) investigated the impact of spray on the development of a simulated hurricane using a coupled atmosphere–ocean–wave model. They found that the hurricane intensity can substantially increase. Wang et al. (2001) reported a moderate enhancement of the final intensity of a modeled tropical cyclone by spray. Effects on extratropical storms have received less attention in the literature.

These studies rest on fairly uncertain assumptions concerning the role of evaporating sea spray in air–sea exchange. Andreas and DeCosmo (1999) found that spray has a large impact on the surface heat fluxes during high wind conditions and even in wind speed ranges below 20 m s^{-1}. Hasse (1992) argued that sea spray has a negligible effect on the fluxes, after which Andreas (1994) wrote a rebuttal to his arguments. Emanuel (1995) initially stated that spray cannot affect the air–sea enthalpy transfer but later changed his opinion (Andreas and Emanuel 1999).

In the many field experiments measuring sensible and latent heat fluxes over the ocean (e.g., Large and Pond 1982; DeCosmo et al. 1996; Fairall et al. 1996; Eymard et al. 1999), hardly any wind speed dependence of the neutral heat exchange coefficients was found. This contrasts with the expectation that evaporating spray leads to deviations of these exchange coefficients. Andreas et al. (1995) and Kepert et al. (1999) gave several possible explanations as to why no wind speed dependence has been observed. The first option is that not enough spray is produced to significantly modify the fluxes, at least not at wind speeds below 18 m s^{-1}, at which the above experiments were performed. Additionally, negative feedbacks (cooling and moistening of the surface layer) might considerably limit the effect of spray evaporation. Finally, it is possible that the heat fluxes are influenced but that the height of the measurements was, in general, too low to observe this. That is, the measurements were within the layer where the droplets were evaporating.

Due to the violent conditions above the sea at high wind speeds, performing accurate direct measurements of spray is a complicated task. Hence, considerable uncertainty is present concerning the amount of spray that is produced from the oceans. One can distinguish be-

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tween droplets produced by bubbles (film and jet droplets) and droplets that are directly formed from breaking wave crests (splash and spume droplets). Estimates of the former have been obtained by combining measurements of bubble spectra and laboratory observations of bubble bursting. Estimates of the latter have been deduced from spray measurements at some height above the water. Both methods have large errors [see Mestayer et al. (1996) for a discussion]. Andreas (1998) collected the various generation functions reported in literature and showed that estimates of spray volume fluxes vary by around 6 orders of magnitude. Nevertheless, he claimed that, on theoretical grounds, most of those estimates can be rejected. Consequently, the uncertainty is much reduced but still present.

Where measurements leave many questions unanswered, modelers have tried to fill this gap. Detailed calculations of spray droplets and their interaction with the fields of temperature and humidity through evaporation were performed by Rouault et al. (1991), Edson and Fairall (1994), Edson et al. (1996), Kepert et al. (1999), and Van Eijk et al. (2001), for example. Apart from these relatively complicated models, simplified approaches have been applied, mainly to assess the impact of spray evaporation on heat and moisture fluxes (e.g., Andreas 1992; Fairall et al. 1994; Makin 1998). The simplifications made in these models seem to be justified considering that many inputs, in particular the amount of spray production, are poorly known. Meanwhile, they have the advantage of a low computational cost and, hence, the possibility to include them in atmospheric models.

The goal of the present study is to assess the sensitivity of the atmosphere to sea spray evaporation over the midlatitude oceans. This is done using the numerical weather prediction (NWP) high-resolution limited-area model HiRLAM, which contains a comprehensive physics parameterization package for vertical diffusion, radiation, condensation, precipitation, and surface processes. The impact of spray is included in the form of modified air–sea surface heat fluxes, based on the bulk parameterization of Fairall et al. (1994). We present two case studies of intense midlatitude storms and analyze the effects of spray evaporation on the simulations. Although the wind speeds are not as extreme as in tropical cyclones, still effects of spray are anticipated.

The remainder of this paper is organized as follows. In section 2, the setup of the numerical simulations is outlined. This includes a description of the HiRLAM model and the inclusion of the effects of spray in it. The results of the case studies with HiRLAM are presented in section 3, along with a discussion. Finally, concluding remarks are given in section 4.

2. Model description

This section deals with the HiRLAM model, the parameterization of surface fluxes, and the effect of sea spray on these fluxes.

a. The NWP model

In this study we use the NWP model HiRLAM, a limited-area, hydrostatic, gridpoint model, developed in a cooperation of several European meteorological institutes. We run the model on a domain covering Europe and the North Atlantic, with a horizontal resolution of 55 km and with 31 layers in the vertical. The lowest model level is located approximately 30 m above the surface. The model is driven by the European Centre for Medium-Range Weather Forecasts (ECMWF) global model; that is, lateral boundary conditions are taken from ECMWF analyses.

We use HiRLAM version 4.8.1, with the exception that the nonlocal, first-order turbulence closure scheme of Holtslag and Boville (1993) is employed for modeling vertical diffusion. The parameterization of the surface fluxes above the sea is described below. Important for determining these fluxes is the sea surface temperature (SST). The SST is analyzed using coarse-resolution satellite data in combination with buoy and ship measurements and is frozen during forecasts. For further detailed information on HiRLAM, the reader is referred to Källén (1996).

b. Surface fluxes

In HiRLAM, the surface fluxes above the sea are calculated from mean model parameters using Monin–Obukhov similarity theory. The resulting bulk formulations for the turbulent fluxes of momentum τ, sensible heat H_s, and latent heat H_l are

$$\tau = \rho_a C_D U_{z_l}^2, \tag{1}$$

$$H_s = \rho_a c_{pa} C_H U_{z_l}(\theta_0 - \theta_{z_l}), \tag{2}$$

$$H_l = \rho_a L_v C_E U_{z_l}(q_0 - q_{z_l}). \tag{3}$$

Here U is the mean horizontal wind speed, θ the potential temperature, q the specific humidity, ρ_a the density of air, c_{pa} the specific heat of moist air at constant pressure, and L_v the latent heat of evaporation of water. The heat fluxes are defined positive in the upward direction. The subscript z_l denotes the lowest model level, while 0 refers to the water surface. The exchange coefficients $C_i(i = D, H, E)$ are determined from their neutral counterparts C_{iN} with stability functions from Louis (1979).

The neutral exchange coefficients are given by

$$C_{DN} = \frac{\kappa^2}{\ln^2(z_l/z_{0m})}, \tag{4}$$

$$C_{HN} = \frac{\kappa^2}{\ln(z_l/z_{0m})\ \ln(z_l/z_{0t})}, \tag{5}$$

$$C_{EN} = \frac{\kappa^2}{\ln(z_l/z_{0m})\ \ln(z_l/z_{0q})}, \tag{6}$$

where $\kappa = 0.4$ is the von Kármán constant, while z_{0m},

z_{0t}, and z_{0q} are the roughness lengths for momentum, temperature, and humidity, respectively. In HiRLAM, z_{0m} is calculated from the Charnock relation,

$$z_{0m} = \alpha_c \frac{u_*^2}{g}, \tag{7}$$

where α_c is the Charnock constant, u_* the friction velocity, and g the acceleration due to gravity. The value $\alpha_c = 0.014$ is used over open sea, whereas $\alpha_c = 0.032$ is employed at grid points with a nonzero land cover fraction, that is, in coastal zones. The roughness lengths for sensible and latent heat are taken from Garratt (1992, p. 102):

$$\ln(z_{0m}/z_{0t}) = 2.48\mathrm{Re}_*^{1/4} - 2, \tag{8}$$

$$\ln(z_{0m}/z_{0q}) = 2.28\mathrm{Re}_*^{1/4} - 2. \tag{9}$$

The roughness Reynolds number is defined as $\mathrm{Re}_* = z_{0m}u_*/\nu$, where ν is the viscosity of air. For low wind speeds, smooth surface and free convection regimes are included in the surface flux parameterization. More details on the implementation can be found in Woetmann-Nielsen (1998).

c. The effect of spray on the surface heat fluxes

The parameterizations of the surface heat fluxes in the previous section are based on empirical knowledge, consisting of measurements taken during many field campaigns. Measurements in conditions with wind speeds exceeding 20 m s^{-1} are rare. At those higher wind speeds, the amount of spray increases rapidly, so that its evaporation can become important. Several studies have tried to quantify the effect of sea spray on the air–sea heat fluxes (e.g., Andreas 1992, 1998; Fairall et al. 1994; Makin 1998). Although these studies estimate different magnitudes of the spray-mediated fluxes, they all indicate substantial spray fluxes at high wind speeds.

To include spray evaporation in a numerical weather prediction model, a parameterization of the contribution of spray to the surface heat fluxes is required. Such a parameterization was provided by Fairall et al. (1994). In their model, they assumed that the concentration of spray is uniform in the so-called droplet evaporation layer and negligible above. The amount of spray was derived from the Andreas (1992) source function. They also assumed that the droplets stay in the air long enough to reach thermal equilibrium with their surroundings, that is, to fully exchange sensible heat.

Using these assumptions and a number of other simplifications, Fairall et al. (1994) derived the following explicit expression for the spray-induced sensible heat flux Q_s and latent heat flux Q_l:

$$Q_s = \alpha \cdot 6.4 \times 10^{-8} U_{10}^{3.4} \gamma \rho_a c_{pa} (T_0 - T_{z_l}), \tag{10}$$

$$Q_l = \alpha \cdot 7.2 \times 10^{-9} U_{10}^{5.4} \gamma B(T_{z_l}) \rho_a L_v (q_s(T_{z_l}) - q_{z_l}), \tag{11}$$

where U_{10} is the wind speed at 10-m height, T the temperature, and q_s the saturation specific humidity. The function γ corrects for the difference between the height of the lowest model level, where the temperature and humidity of the air are given, and the depth of the droplet evaporation layer, estimated as $z_{\mathrm{DEL}} = 0.015U_{10}^2$:

$$\gamma = 1 - \frac{C_E^{1/2}}{\kappa} \ln \frac{z_l}{z_{\mathrm{DEL}}}. \tag{12}$$

The function B reflects the fact that, while evaporating, the spray droplets are maintained at the evaporating temperature T_{ev}, which is the wet-bulb temperature corrected for salinity and droplet curvature effects, rather than at the air temperature (Andreas 1995):

$$B(T_{z_l}) = \left(1 + \frac{\epsilon L_v^2}{R_a c_{pa} T_{z_l}^2} q_s(T_{z_l})\right)^{-1}. \tag{13}$$

Here $\epsilon = 0.622$ and R_a is the gas constant for dry air.

The evaporation of spray modifies the vertical profiles of temperature and humidity in the near-surface layer below the lowest model level. Typically, this layer will be cooled and moistened. This leads to a negative feedback on the spray-mediated fluxes, and also alters the direct, interfacial fluxes, compared to (2) and (3). Moreover, the stability is affected, which influences turbulent diffusion and, hence, the fluxes. The net effect of such feedbacks will be to reduce the spray-induced changes in the heat fluxes realized at the lowest model level. Fairall et al. (1994) introduced a factor, α, to account for this and suggested a fixed value $\alpha = 0.5$. The same value is employed in the present study. Bao et al. (2000) and Kepert et al. (1999) noted that α should decrease when spray evaporation increases: higher evaporation rates enhance the negative feedback. Although this is indeed plausible, we feel that the fixed value $\alpha = 0.5$ is appropriate as long as the wind speeds are not extremely high (say $U_{10} < 32$ m s^{-1}), which is the case for the storms we study.

To get an impression of the magnitude of the spray-mediated fluxes, the estimates (10) and (11) are plotted as a function of wind speed in Fig. 1. The direct turbulent fluxes from (2) and (3) are also shown. The conditions are more or less typical for the midlatitudes. A number of observations can be made from this figure. First, the latent heat flux over sea is normally much larger than the sensible heat flux. Second, the spray-mediated heat fluxes increase much more rapidly with wind speed than the direct fluxes. Whereas, at a wind speed of 15 m s^{-1}, Q_l is only 10% of H_l, these fluxes are equally large for $U_{10} = 26$ m s^{-1}. Third, Q_s is small compared to H_s. Even at a wind speed of 30 m s^{-1}, and for an arbitrary choice of temperature and humidity (not shown in Fig. 1), Q_s is at most 10% of the direct flux.

Because the heat required for evaporation of the spray droplets is extracted from the air, the spray-mediated latent heat flux shows up as a sink in the boundary condition for sensible heat. Hence, the heat boundary

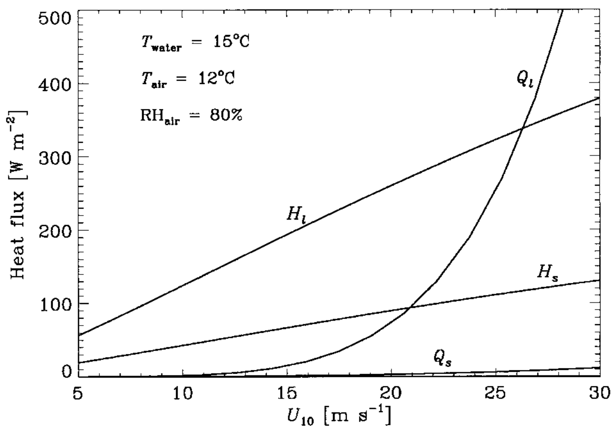

FIG. 1. Typical magnitude of the direct and spray-mediated sensible and latent heat flux as a function of wind speed. The fluxes have been calculated for a water temperature of 15°C, an air temperature (at 30-m height) of 12°C, and an RH of 80%.

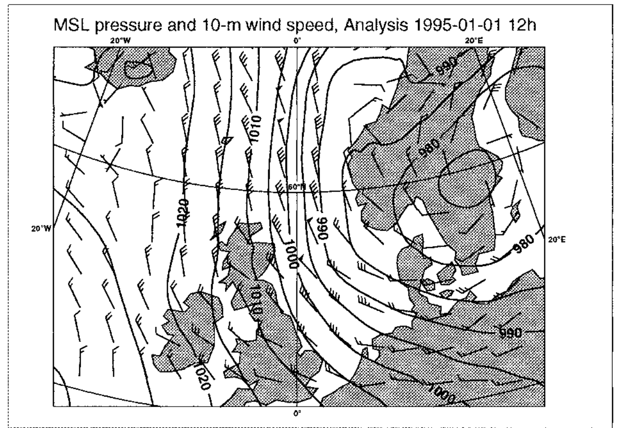

FIG. 2. Mean sea level isobars and 10-m wind speed flags from the HiRLAM analysis (reference run) at 1200 UTC 01 Jan 1995. Meaning of the wind speed flags: half barbs correspond to 5 kt, full barbs to 10 kt, and closed "triangles" to 50 kt (1 kt ≈ 0.514 m s^{-1}); barbs/triangles are located on the side of the wind direction.

conditions at the air–sea interface in the presence of spray become (see Fairall et al. 1994)

$$H_{s,\mathrm{tot}} = H_s + Q_s - Q_l, \tag{14}$$

$$H_{l,\mathrm{tot}} = H_l + Q_l. \tag{15}$$

These modified boundary conditions have been implemented in HiRLAM. Their impact is analyzed in the next section for two case studies of storm situations. Runs with original fluxes (H_s and H_l) will be referred to as reference runs, while those with modified fluxes will be termed spray runs.

Sea spray thus has the effect of redistributing energy from sensible to latent heat. On adding (14) and (15), we see that the total energy exchange between air and sea is, in principle, only changed by Q_s, which is small. However, the energy transport can be influenced indirectly by modifications of H_s and H_l due to, for example, changes in wind speed, temperature, and humidity at the lowest model level.

3. Results and discussion

In this section, two case studies with HiRLAM are presented, in which the sensitivity of the atmosphere to sea spray–induced changes in the surface heat fluxes is studied. For both case studies, we have run the model for about four days, with a 6-h data assimilation cycle, producing forecasts up to 48 h.

a. Case 1

The first case concerns a storm on 1 January 1995. A low over southern Sweden and a high west of France cause strong winds over the North Sea from the north–northwest direction. The HiRLAM analysis at 1200 UTC is given in Fig. 2. The wind has been blowing with a more or less constant direction for almost 24 h, reaching speeds of over 25 m s^{-1}. The SST of the North Sea varies between 6° and 10°C. The advection of polar air makes the situation unstable: the air at 2-m height is on average 4°–7°C colder than the water. The air is also dry: the relative humidity varies mostly between 55% and 75%.

Figure 3 shows forecasts of wind speed and surface heat fluxes during the storm. The effect of spray on the latent heat flux is visible in Fig. 3c. Clearly, the major impact is found in the high wind speed regions (cf. Fig. 3a). The latent heat flux is enhanced with up to 140 W m^{-2}, which is a 40% increase compared to the reference run (Fig. 3b). Notice that, although the spray-mediated flux Q_l is always positive, the difference between spray and reference run can be negative. This is, for example, the case in the area west of Norway, where the wind speed is very low (Fig. 3a). In this area, the air has been moistened at previous times when the wind speed was high. This has resulted in a decreased direct flux at the time displayed in Fig. 3, while the spray-mediated flux is negligible.

Model predictions of several near-surface quantities for a particular location in the northern North Sea are presented in Fig. 4. The wind speed at this location reaches 25 m s^{-1}. During the whole period, the stratification is unstable, that is, upward heat fluxes. Due to spray, the latent heat flux is enhanced by more than 100 W m^{-2} (Fig. 4b) in the storm period. Likewise, the sensible heat flux decreases. These modifications lead to a cooler and moister lowest model level. In the model runs, the SST is kept constant during forecasts and is the same in the spray and reference runs. Therefore, Fig. 4c actually shows variations in lowest model level temperature. Note that in Fig. 4d, the relative humidity is plotted, which increases not only due to the higher specific humidity but also due to the lower saturation specific humidity caused by cooling.

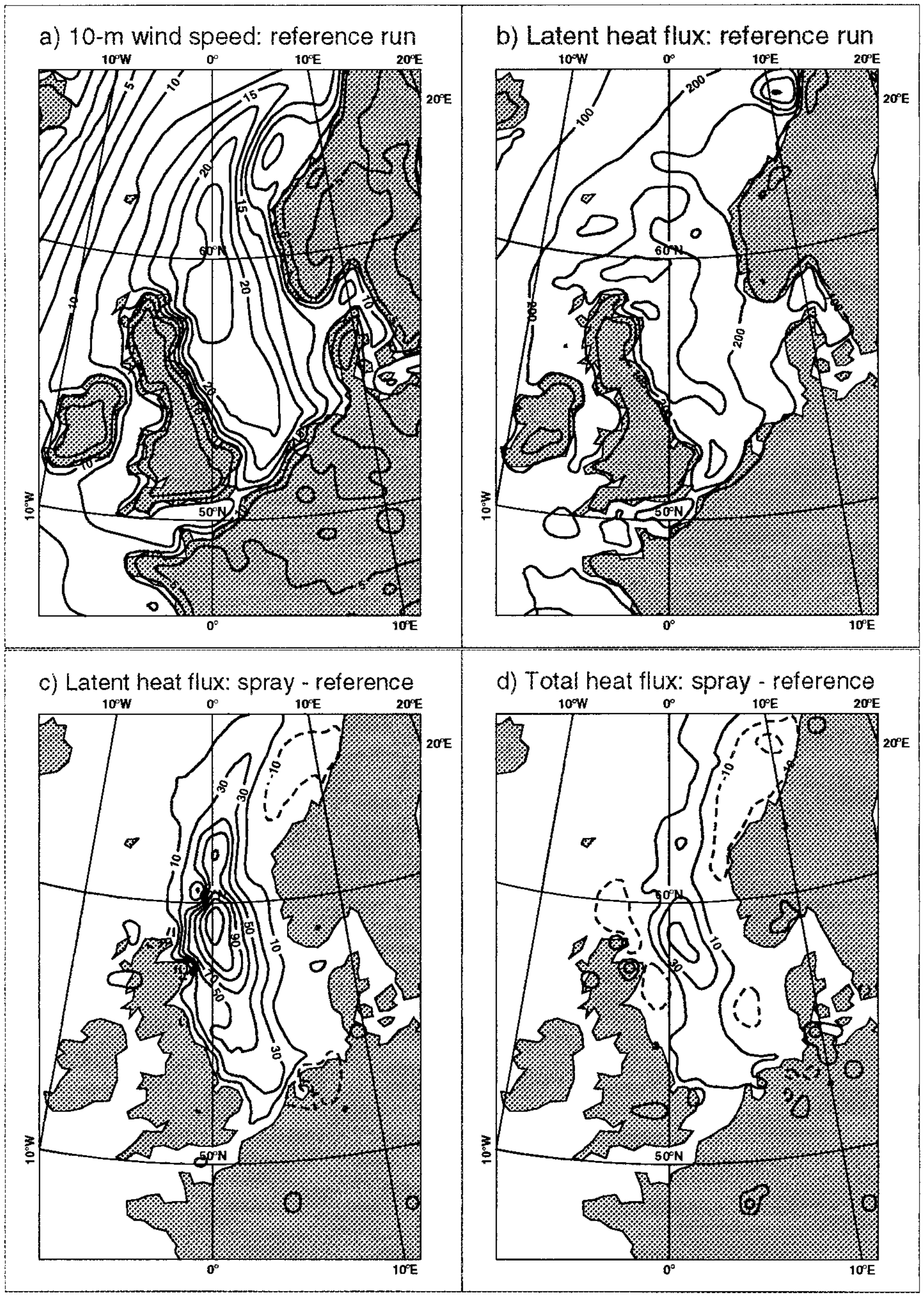

FIG. 3. Contour plots of wind speed and heat fluxes from the 12-h forecast valid at 1800 UTC 01 Jan 1995: (a) 10-m wind speed from reference run, (b) surface latent heat flux from reference run, (c) latent heat flux difference (spray − reference), and (d) total (latent plus sensible) heat flux difference. Contour intervals are (a) 2.5 m s^{-1}, (b) 100 W m^{-2}, (c) 20 W m^{-2}, and (d) 20 W m^{-2}.

The wind speed is slightly reduced due to sea spray (see Fig. 4a, in particular on 1200 UTC 01 Jan 1995), which is explained as follows. In the spray run, the region close to the surface is more unstable than in the no-spray run. Hence, the wind feels the surface more strongly. Higher aloft, the vertical temperature profile is stabilized. This means that less momentum is exchanged with higher levels. Those effects lead to a decrease in wind speed at the lowest model level and also at 10-m height.

The difference in the total heat flux between the spray and reference runs at the same location is analyzed in Fig. 5. Part of this difference is caused by the spray-mediated sensible heat flux Q_s, plotted with long dashes. It is observed that Q_s is generally small, reaching a maximum of around 10 W m^{-2}. The other curves in

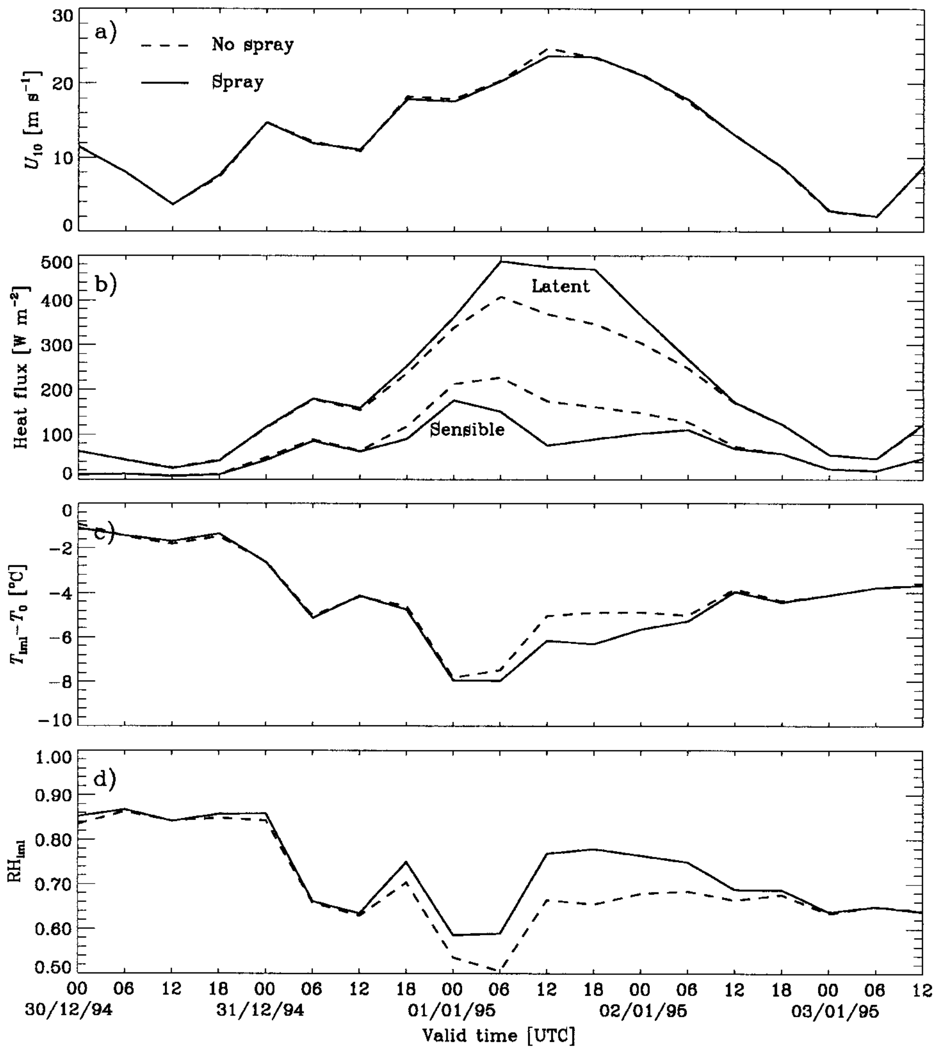

FIG. 4. Model predictions of near-surface quantities at location (59°N, 1°E) for reference (dashed line) and spray (solid line) runs: (a) 10-m wind speed, (b) surface heat fluxes, (c) difference between lowest model level (lml) temperature and SST, and (d) relative humidity at lml. The model output is taken from 12-h forecasts with valid time (time at which the forecast is valid) from 0000 UTC 30 Dec 1994 to 1200 UTC 3 Jan 1995.

Fig. 5 show the contributions of the various parameters in Eqs. (2) and (3) to modifying $H_s + H_l$. The partitioning into flux contributions was achieved by calculating the linear increments to the total flux due to changes in every single parameter. Figure 5 shows that the cooling due to spray leads to a higher direct heat flux, whereas the moistening reduces it (dashed and dash–dotted lines, respectively). Clearly, these effects do not cancel, which implies that the equivalent potential temperature at the lowest model level has changed. The inclusion of spray leads to a decrease in wind speed, as was noted before. This in turn causes a lower direct heat flux (dash–dot–dot–dotted line). Finally, the dotted line shows the joint effect of changes in C_H and C_E. With spray, the near-surface layer is more unstable, so that the exchange coefficients are larger. This implies larger direct heat fluxes, but the effect is only small (dotted line).

Thus, while the main effect of spray is a redistribution of the energy flux from sensible to latent, the total flux can change due to indirect effects. This is also seen in Fig. 3d, where the total heat flux is significantly larger in the spray run. At the location of Fig. 5, the change in total heat flux is negative in the first part and positive in the second part of the storm. On average, there is no change, except for the Q_s contribution.

The time series shown in Fig. 4 are typical for locations in the North Sea. The changes due to spray can be considerable. To check whether these effects are also present in observations, we have compared model predictions of 2-m temperature and humidity against all available observations over the sea in the area (50°–

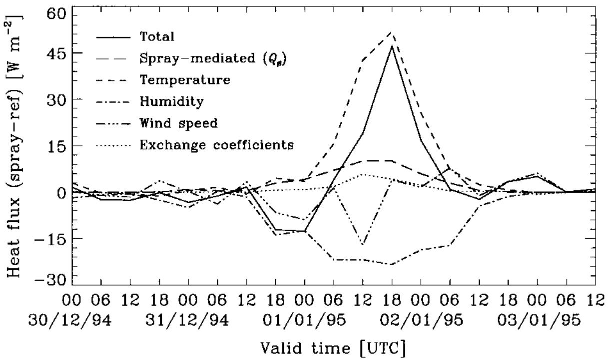

FIG. 5. Difference in heat flux between spray and reference runs at location (59°N, 1°E): the total heat flux difference consists of the spray-mediated sensible heat flux Q_s, and differences in the direct fluxes due to changes in temperature, specific humidity, wind speed, and exchange coefficients.

70°N, 10°W–10°E). The inclusion of spray reduces the temperature bias from 0.78° to 0.49°C, while the bias in humidity increases from 0.27 to 0.41 g kg^{-1}, averaged over the four days of the run. This change in bias is, of course, due to the storm period. In contrast, the standard deviation remains at the same level of 1.5°C and 0.67 g kg^{-1}, respectively. Moreover, if we select only those observations taken at high wind speeds, then the standard deviation also remains unchanged. Thus, no improvement of the skill of the model is noticed in this sense. It is probably difficult anyway to reach an improvement of the standard deviation because of the coarse resolution of the SST field that is used in HiRLAM. The 2-m temperature and humidity are tightly coupled to the SST. Hence, they also suffer from lack of spatial variation.

Figure 6 shows cross sections of the difference in temperature and specific humidity between the spray and reference runs. Below 850 hPa, the atmosphere is cooled by up to 1.5°C due to spray evaporation. At the lower levels, the air is moistened, which is consistent with the increased latent heat flux. However, around 800 hPa, drying is observed. The reason for this is that spray stabilizes the boundary layer, thus reducing its height, and the height at which the lowest clouds are formed. Hence, due to condensation, the humidity can become lower than in the reference run. At higher levels, where the clouds are also present in the reference run, the spray run is again moister. The additional water vapor, added to the atmosphere by spray evaporation, condensates, at least partly, between 750 and 500 hPa. The associated latent heat release leads to a general slight warming at these midlevels.

Such features are still visible when the air column has been advected over land. Figure 7 compares vertical profiles in De Bilt at the time when the largest differences between the spray and reference runs are present. In the spray run, the boundary layer is more humid as a consequence of the extra moisture fed into the air above the sea. The lower atmosphere is cooled, whereas at higher levels warming is observed, as noted above. Unfortunately, we cannot discriminate between the two runs on the basis of the observations. The discrepancy between model forecast and observations is much larger than the difference between both model runs, and the effect is too indirect to see whether the impact of taking into account spray evaporation is positive or negative.

The increased condensation at higher levels is expected to lead to enhanced rainfall. Figure 8a shows the total precipitation from a 24-h forecast. In the region with high wind speeds, precipitation rates are around 9 mm day^{-1} with a maximum of 14 mm. The spray run gives rise to an increase in rainfall with 1 to 2 mm over a large area, while the maximum increase is around 7 mm (Fig. 8b). Since 1 mm of water in 24 h corresponds to a latent heat flux of around 30 W m^{-2}, these numbers indicate that the increase in precipitation is of the same order as the increase in latent heat flux from the sea (cf. Fig. 3c). Thus, for this case, most of the extra moisture added to the atmosphere returns as rain, which is expected when the stratification is unstable up to a sufficient height.

b. Case 2

The second case, in January 1999, concerns a depression moving from south to east of Iceland. Figure 9 shows its development. According to the analyses, the minimum central pressure is reached on 1800 UTC 15 January, after a very fast deepening of 35 hPa in 18 h. The accompanying wind speeds reach 30 m s^{-1}. The SST varies over a wide range, from near 0°C in the north to 10°C in the south.

Figure 10 shows the forecasts of wind speed and heat fluxes at the time when the depression has reached its maximum intensity. The highest latent heat fluxes are present in the southwestern part of the depicted area (Fig. 10b), since there the SST is high and the air is dry. The spray acts to increase the latent heat flux, with a maximum difference of over 200 W m^{-2} (Fig. 10c). As discussed before, the latent heat flux can locally be lower in the spray run than in the reference run, when a period with low winds follows a period with high winds. This is the case in the eye of the depression. From Fig. 10d, it again appears that the modifications of the latent and sensible heat flux do not necessarily cancel. Due to feedback effects, the total heat flux increases in some regions, whereas it decreases in others.

Including sea spray can change the development of the depression. Figure 11a shows the pressure in the center of the depression as a function of time, as calculated in the reference run. The changes due to spray are presented in Fig. 11b. The analyses and 12-h forecasts both show that, during the deepening of the depression, the spray run has lower central pressures than the reference run. The effect is largest in the 12-h forecasts, where the maximum pressure difference is about

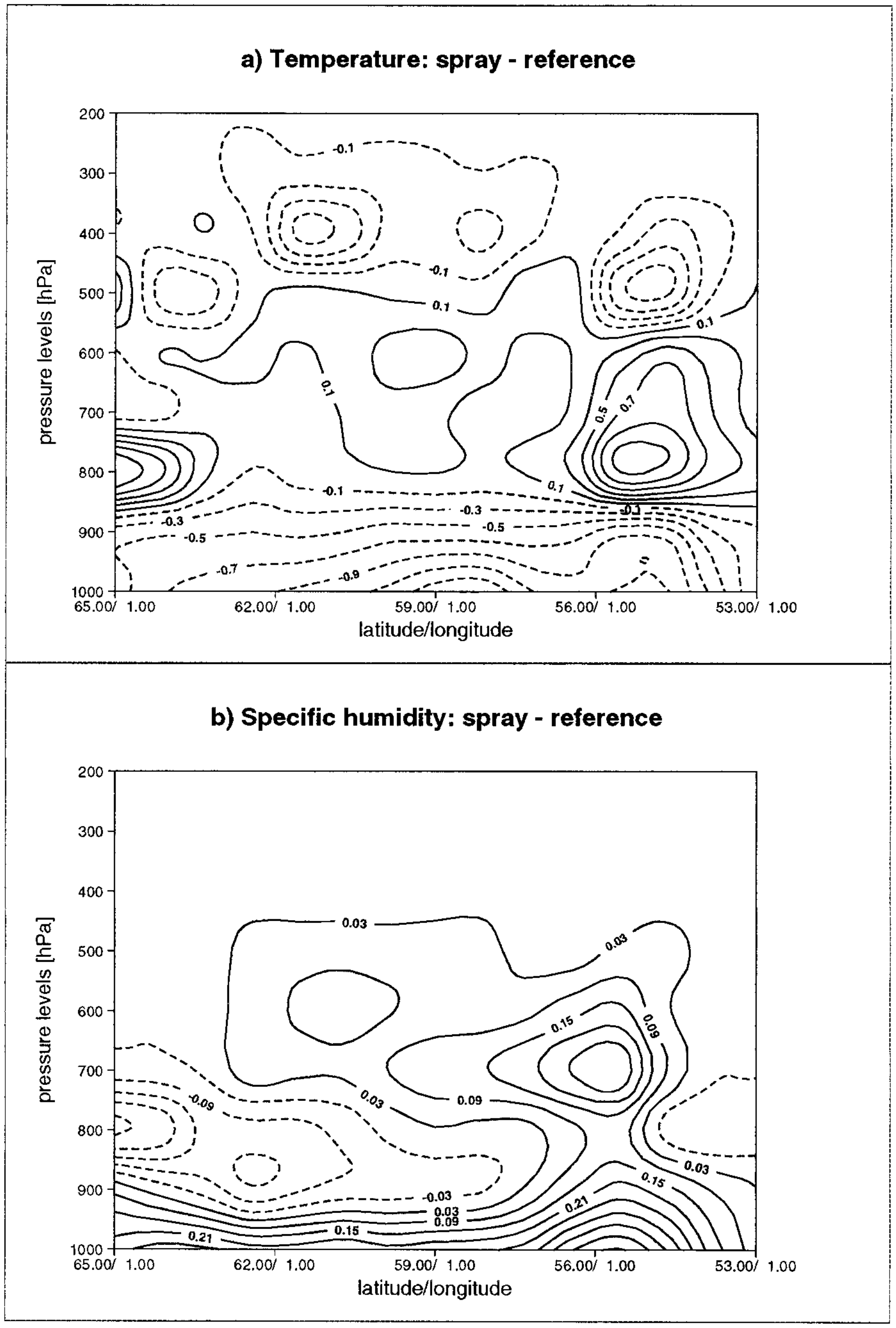

FIG. 6. Cross sections between (65°N, 1°E) and (53°N, 1°E) of (a) temperature difference and (b) specific humidity difference between spray and reference runs. The model output is taken from the 12-h forecast valid at 1800 UTC 01 Jan 1995. Contour intervals are (a) 0.2°C and (b) 0.06 g kg^{-1}.

1.3 hPa. The analyses are less affected by spray, which is expected, since they tend to observations every 6 h. After the minimum central pressure has been reached, the differences between spray and reference runs vanish. Similar results are obtained for longer forecast periods. We note that the location of the center of the depression is not significantly influenced by including spray.

In the forecast cycle, differences between forecasts are partly caused by differences in the analyses they start with. Thus, to make a cleaner comparison, we have also made two separate pairs of forecasts (see Fig. 11), where the spray and reference runs begin with the same analysis. The forecasts started at 0000 UTC 15 January (s1) show the same picture of lower central pressures in the spray run. The difference is around 1 hPa when the depression has reached its minimum central pres-

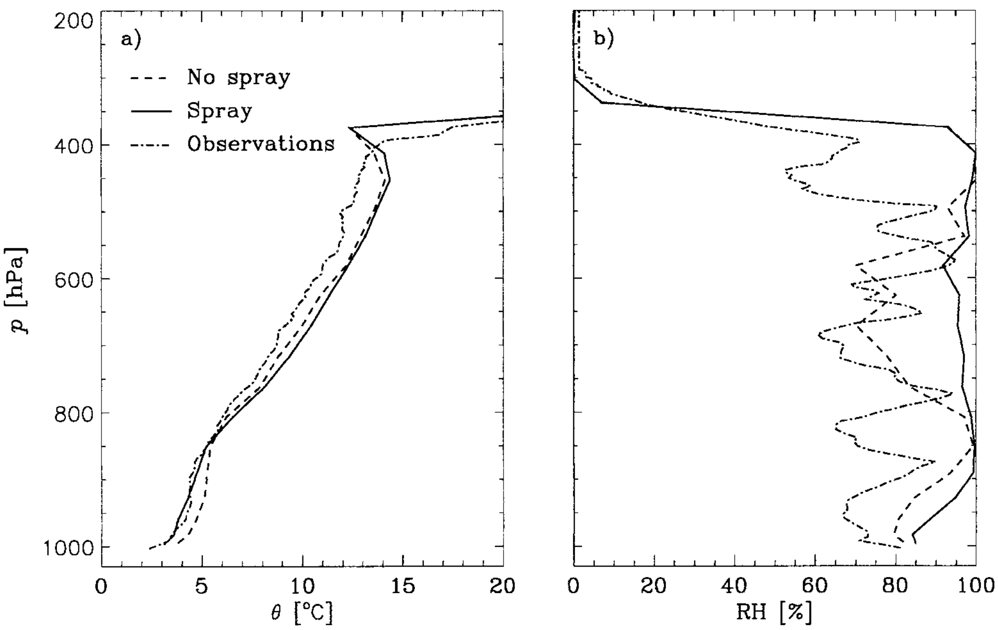

FIG. 7. Comparison of (a) vertical potential temperature and (b) relative humidity profiles from 24-h forecast with radiosonde observations in De Bilt, Netherlands (52.1°N, 5.2°E), at 0000 UTC 02 Jan 1995: (solid line) spray run, (dashed line) reference run, and (dash–dotted line) observations.

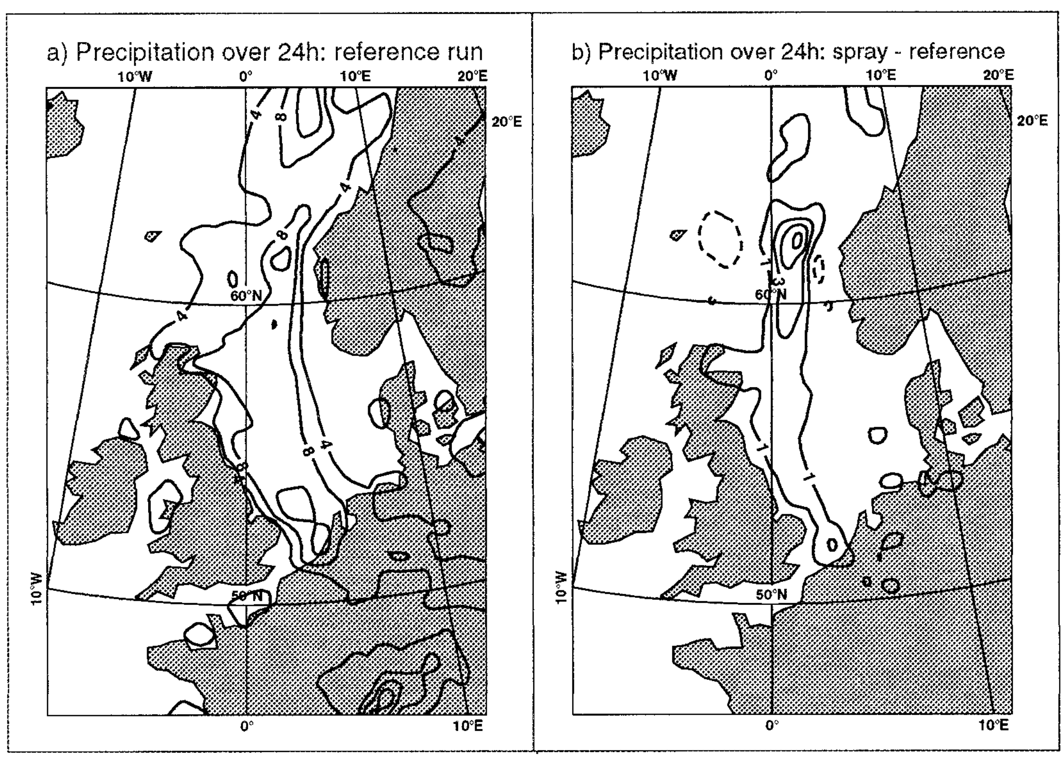

FIG. 8. Cumulative precipitation from 24-h forecast valid at 0600 UTC 02 Jan 1995: (a) reference run and (b) difference between spray and reference runs. Contour intervals are (a) 4 and (b) 2 mm.

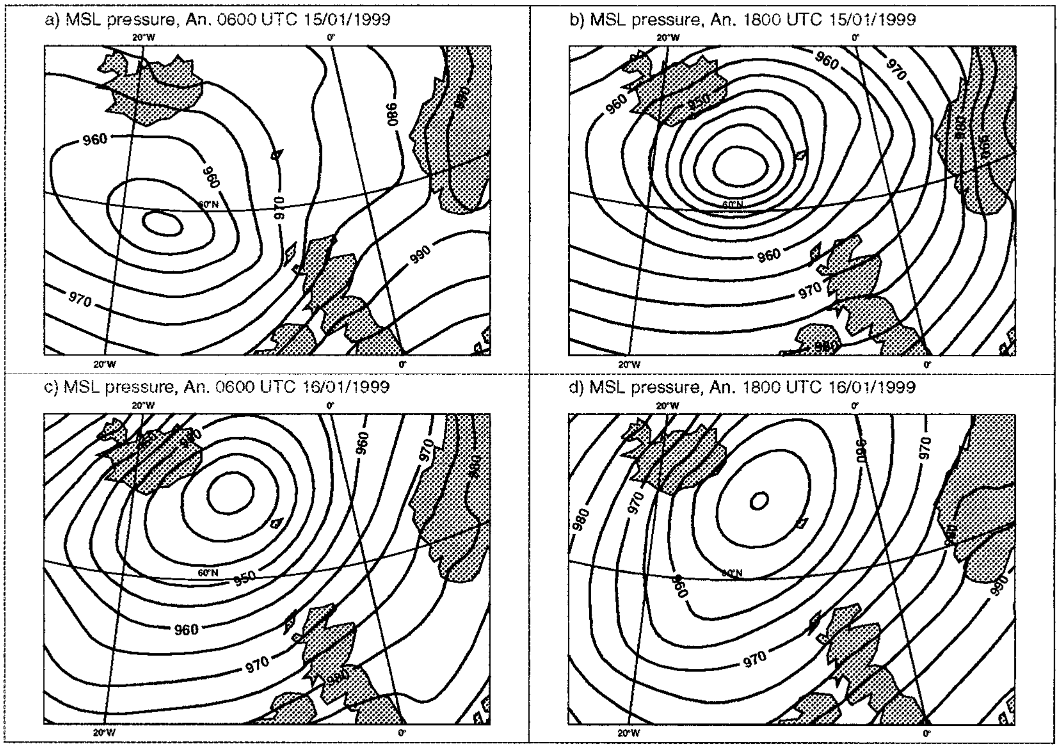

FIG. 9. Mean sea level isobars from the HiRLAM analyses (reference version) at (a) 0600 UTC 15 Jan 1999, (b) 1800 UTC 15 Jan 1999, (c) 0600 UTC 16 Jan 1999, and (d) 1800 UTC 16 Jan 1999.

sure, remains there for about 12 h, and then vanishes. The forecasts started at 1800 UTC (s2), which is near the time of maximum intensity, show no spray impact.

In principle, spray leads to stabilization of the boundary layer and destabilization of the surface layer, which results in lower wind speeds (see case 1). However, in this case, the deepening of the cyclone is accompanied by higher wind speeds. The maximum wind speed at the lowest model level increases by 2%, from 36.8 m s^{-1} without spray to 37.5 m s^{-1} with spray, according to the s1 forecasts. As in case 1, precipitation rates are higher in the spray runs, which is consistent with the enhanced surface latent heat flux.

Earlier studies on the effects of sea spray evaporation have mainly focused on tropical cyclones. It is interesting to relate the present results to those studies.

Bao et al. (2000) investigated the impact of sea spray evaporation on the development of a hurricane using a coupled atmosphere–ocean–wave model. They took spray into account in a similar way as in the present study, but with an additional evaporation partitioning parameter β in relation (14):

$$H_{s,\mathrm{tot}} = H_s + Q_s - \beta Q_l. \qquad (16)$$

They divided the spray-mediated latent heat flux into two parts: $Q_l = Q_{l1} + Q_{l2}$. For Q_{l1}, the heat required for evaporation is taken from the droplets; it corresponds to the cooling of the droplets from the air temperature to their evaporating temperature. For Q_{l2}, the necessary heat is provided by the air. The parameter β is then defined as $\beta = Q_{l2}/Q_l$. Bao et al. (2000) studied the cases of $\beta = 0$, 0.5, and 1. When $\beta = 1$, the impact of spray on the intensity of the cyclone is small, of the same order as in this study. In contrast, $\beta = 0$ leads to a much more intense cyclone. The reason for this is that the total heat flux increases enormously, thus fueling the cyclone. In our opinion, the parameter β should, however, be close to 1. With the assumption that all droplets reach the evaporating temperature T_{ev}, the part of the latent heat flux termed Q_{l1} can be written in a similar way as Eq. (10):

$$Q_{l1} = 6.4 \times 10^{-8} \alpha U_{10}^{3.4} \gamma \rho_a c_{pa} (T_{z_l} - T_{\mathrm{ev},z_l}). \qquad (17)$$

Hence, Q_{l1} is of the same order as Q_s and, consequently, much smaller than Q_l.

Using a high-resolution tropical cyclone model, Wang et al. (1999) found that sea spray can substantially reduce the intensification rate of a cyclone in the early stage of its development, but that the final intensity is hardly affected. However, repeating their experiments with different initial conditions, Wang et al. (2001) observed no effect on the early intensification rate but found a moderate decrease of the final central pressure. They obtained an 8% increase in maximum wind speed, which is larger than the 2% we find. However, midlatitude and tropical cyclones have quite different spatial scales and forcing mechanisms. In particular, the latter are more sensitive to the air–sea fluxes for their dynam-

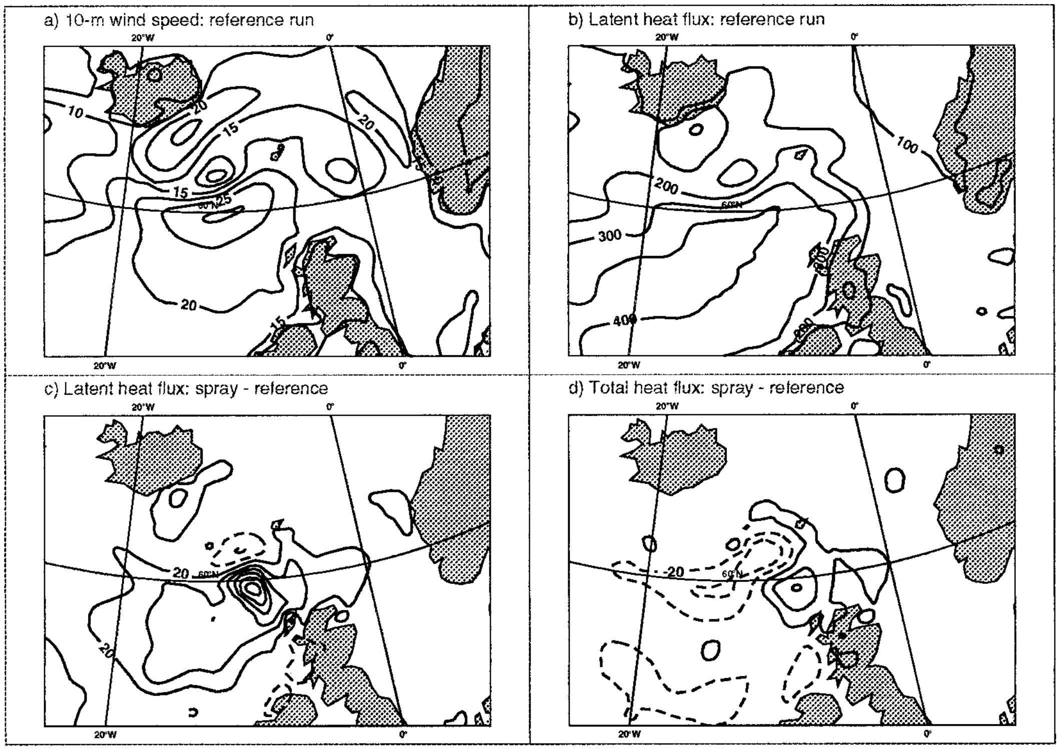

FIG. 10. Contour plots of wind speed and heat fluxes from the 12-h forecast valid at 1800 UTC 15 Jan 1999: (a) 10-m wind speed from reference run, (b) surface latent heat flux from reference run, (c) latent heat flux difference (spray − reference), and (d) total (latent plus sensible) heat flux difference. Contour intervals are (a) 5 m s^{-1}, (b) 100 W m^{-2}, (c) 40 W m^{-2}, and (d) 40 W m^{-2}.

ics. Hence, comparing these results is not straightforward.

In the present case study, including spray leads to a slight deepening of the depression. This effect seems to be systematic and was also observed in another case that is not presented here. Yet, a 2% increase in maximum wind speed is not much. Probably, the slightly increased intensity can be explained by the general small enhancement of the total surface heat flux due to the Q_s contribution by spray. It must, however, be noted that changes in surface pressure are rather indirect consequences, which must be viewed with caution.

4. Conclusions

In this study, the sensitivity of the atmosphere to sea spray evaporation is investigated using the limited-area NWP model HiRLAM. The effect of sea spray is taken into account via a modification of the surface heat fluxes, estimated with the bulk parameterization of Fairall et al. (1994). This modification mainly implies a redistribution of the energy from sensible to latent heat.

Two case studies of intense midlatitude storms are analyzed. The first case concerns a period with high wind speeds over the North Sea. The second case handles a rapidly deepening depression southeast of Iceland. In both cases, the surface heat fluxes are considerably affected by the presence of spray. The increase in latent and the decrease in sensible heat flux lead to a substantial cooling and moistening at the lowest model levels.

Indirect effects are reported as well. A large part of the additional moisture transported to the air returns as precipitation. The latent heat released with the formation of this precipitation leads to a slight warming at the higher levels. The depression in the second case study intensifies when sea spray is taken into account, but only marginally. The reason for this is probably that spray only has a small systematic effect on the total air–sea energy flux.

It must be noted that verification of the results is difficult. The impact in the NWP model is relatively small and localized in time and space, and observations are not conclusive as to whether the impact is positive or negative. Moreover, most effects (e.g., the increase in precipitation and the slight intensification of a depression) are rather indirect. The physics modules in an NWP model have compensating errors. Thus, even if taking into account sea spray evaporation leads to better surface fluxes, then the model performance does not necessarily improve.

This study has attempted to give an indication of the typical impact of sea spray evaporation above the midlatitude oceans in numerical weather prediction. The

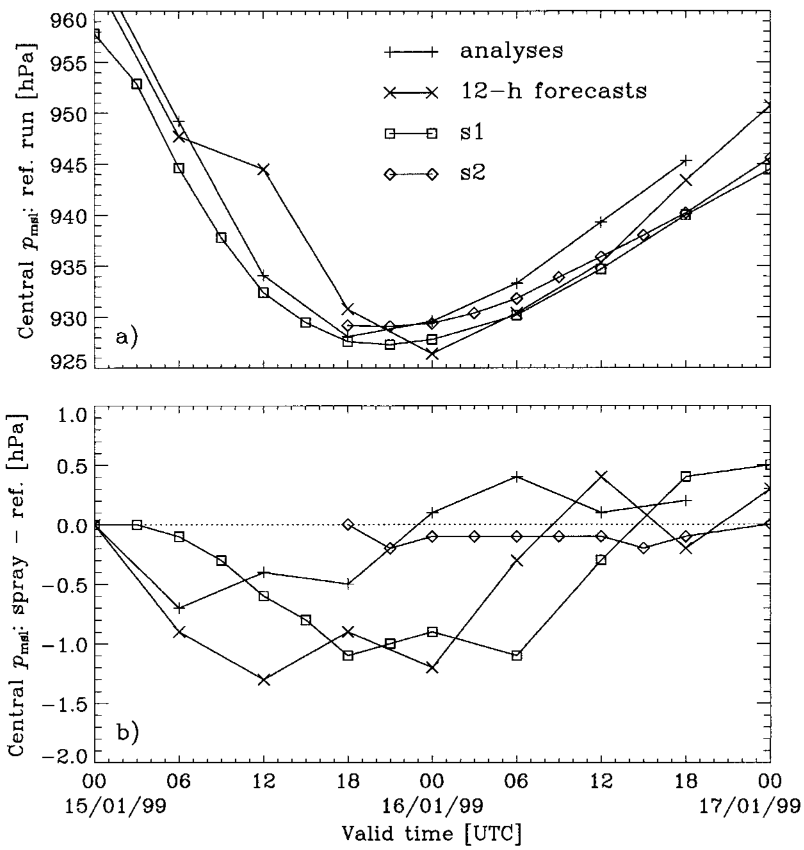

FIG. 11. Mean sea level pressure in the center of the depression: (a) reference run and (b) difference between spray and reference runs. The plus and cross symbols refer to the forecast cycle, which was started on 0000 UTC 14 Jan 1999: pluses are analyses and crosses are 12-h forecasts. The squares and diamonds refer to the separate forecasts: squares (s1) started on 0000 UTC 15 Jan 1999; diamonds (s2) started on 1800 UTC 15 Jan 1999.

results naturally depend on the parameterization of the contribution of spray to the surface heat fluxes. This in turn depends heavily on the amount of spray production, which is poorly known. Thus, the sensitivity of the atmospheric impact to the underlying spray production function needs further investigation. At the same time, measurements of air–sea heat fluxes and spray droplet distributions for wind speeds well above 20 m s^{-1} are needed to make more accurate estimates of the importance of spray evaporation.

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