

1 Simulations of a cold-air pool associated with elevated 2 wintertime ozone in the Uintah Basin, Utah

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8 9 10 **Abstract**

11 Numerical simulations are used to investigate the meteorological characteristics of the 31
12 January–6 February 2013 cold-air pool in the Uintah Basin, Utah, and the resulting high
13 ozone concentrations. Flow features affecting cold-air pools and air quality in the Uintah
14 Basin are studied, including: penetration of clean air into the basin from across the
15 surrounding mountains, elevated easterlies within the inversion layer, and thermally-driven
16 slope and valley flows. The sensitivity of the boundary layer structure to snow cover
17 variations and cloud microphysics are also examined. Snow cover increases boundary layer
18 stability by enhancing the surface albedo, reducing the absorbed solar insolation at the
19 surface, and lowering near-surface air temperatures. Snow cover also increases ozone levels
20 by enhancing solar radiation available for photochemical reactions. Ice-dominant clouds
21 enhance cold-air pool strength compared to liquid-dominant clouds by increasing nocturnal
22 cooling and decreasing longwave cloud forcing.

23 24 **1 Introduction**

25 High concentrations of near-surface ozone have an adverse impact on human health, including
26 respiratory irritation and inflammation, reduced lung function, aggravated asthma, and long-
27 term lung damage (Lippmann, 1993; Bell et al., 2004). Ozone is formed through
28 photochemical reactions of precursor pollutants, typically nitrogen oxides (NO_x) and volatile
29 organic compounds (VOCs), emitted from industrial sources and vehicles (Pollack et al.,

1 2013). Once thought to primarily be an urban, summer-time problem (due to the high
2 insolation required for photochemical reactions), high ozone levels and associated precursor
3 pollutants have recently been detected during the wintertime in snow-covered rural basins
4 with significant industrial fossil fuel extraction activities (Schnell et al., 2009; Helmig et al.,
5 2014; Lee et al, 2014; Warneke et al. 2014a,b; Li et al. 2014). Snow cover increases the
6 surface albedo and near-surface actinic flux (quantity of light available to molecules) leading
7 to photolysis rates notably larger (~ 50 %) than those observed in summer (Schnell et al.,
8 2009). In addition, the shallow and highly stable boundary layer often observed during the
9 wintertime in snow-covered rural basins further exacerbates the problem by trapping the high
10 ozone concentrations in the lowest several hundred meters of the atmosphere (Fig. 1).

11 High levels of ozone were first detected in Northeast Utah's Uintah Basin in 2009,
12 when 8 h average concentrations were over 100 ppb (Lyman et al., 2014). This value was well
13 above the US Environmental Protection Agency's (EPA) National Ambient Air Quality
14 Standard (NAAQS) of 75 ppb (EPA, 2014), and far above the background levels of ozone
15 near the earth's surface that typically range between 20–45 ppb (EPA, 2006). Fossil fuel
16 production has increased in the Uintah Basin over the last several years and will likely
17 continue to increase. Currently, there are over 11 200 producing wells in the basin (Helmig et
18 al., 2014) and over 3800 additional permit applications since the beginning of 2012. Extensive
19 scientific research has been conducted in the Uintah Basin to better understand the wintertime
20 rural ozone problem during the past several winters (Edwards et al., 2013; Lyman and
21 Shorthill, 2013; Stoeckenius and McNally, 2014; Helmig et al., 2014; Warneke et al. 2014a,
22 b). Considerable variations in late winter snow cover, which modulates the occurrence of high
23 ozone events in the Uintah Basin, are evident from year to year. Snow cover was largely
24 absent from the basin during February 2009, 2012, and 2014 and ozone levels remained low
25 during those months, while February 2010, 2011, and 2013 saw extensive snow cover and
26 several high ozone episodes.

27 The Uinta Mountains to the north, Wasatch Range to the west, and Tavaputs Plateau to
28 the south often confine cold air during winter within the topographic depression of the Uintah
29 Basin (Fig. 2). Such cold air pools (CAPs) form when synoptic and mesoscale processes lead
30 to persistent stable stratification in the boundary layer resulting from a combination of
31 warming aloft and cooling near the surface (Lareau et al., 2013). The high terrain
32 encompassing the basin and its large horizontal extent leave its central core less affected by

1 weak synoptic-scale weather systems, which results in longer-lived CAPs than those observed
2 in other locales (Zangl, 2005b; Lareau et al., 2013; Lareau and Horel, 2014; Sheridan et al.,
3 2014). CAPs are often associated with low clouds, fog, freezing precipitation, hazardous
4 ground and air travel, and elevated levels of particulate air pollution in valleys and basins
5 (Whiteman et al., 2001; Malek et al., 2006; Silcox et al., 2012; Lareau et al., 2013; Lareau,
6 2014; Lareau and Horel, 2014).

7 Numerical studies have examined the lifecycle of CAPs for a variety of idealized
8 (Zangl, 2005a; Katurji and Zhong, 2012; Lareau, 2014) and actual topographic basins
9 (Whiteman et al., 2001; Clements et al., 2003; Zangl, 2005b; Billings et al., 2006; Reeves and
10 Stensrud, 2009; Reeves et al., 2011; Lareau et al., 2013; Lareau and Horel, 2014; Lu and
11 Zhong, 2014). However, relatively few studies have examined the impact of snow cover,
12 clouds, and cloud microphysics on CAP formation and evolution. Zangl (2005a) found that
13 the limited heat conductivity of fresh snow was important for efficient cooling of the air near
14 the surface. Comparing simulations with a snow-covered and grass-covered sinkhole floor
15 suggested that the larger surface heat capacity of the grass floor resulted in more gradual
16 cooling, smaller afternoon-morning temperature difference, weaker static stability, and no
17 cloud cover. Billings et al. (2006) studied the impact of snow cover on a CAP in the Yampa
18 Valley, CO and found that snow-free simulations were incapable of producing the CAP.
19 Zangl (2005a) indirectly examined the effect of cloud particle phase on the formation of
20 CAPs in the Gstettneralm sink-hole, Austria. He found that an efficient drying mechanism to
21 remove fog was required, such as the nucleation and sedimentation of cloud ice, otherwise the
22 enhanced cloud longwave radiation inhibits the low-level cooling necessary for a strong CAP.
23 Numerical models often struggle to accurately simulate ice fogs that occur in some CAPs,
24 largely because the underlying ice fog microphysics are not well-understood (Gultepe et al.,
25 2014).

26 While the influence of snow and cloud cover, inter-basin flows, and terrain-flow
27 interactions on the evolution of the shallow, stable boundary layers associated with
28 wintertime high ozone episodes in the Uintah Basin has been recognized, those impacts have
29 only been partially explored (Lyman and Shorthill, 2013; Stoeckenius and McNally, 2014). In
30 this study, a high ozone episode from 31 January–6 February 2013 during the Uintah Basin
31 Winter Ozone Study (UBWOS) is examined. The Weather Research and Forecasting (WRF)
32 model is used to examine the sensitivity of CAP thermodynamic structure and wind flow

1 regimes to variations in snow cover, specification of snow albedo, and cloud microphysics,
2 while the Community Multi-Scale Air Quality (CMAQ) model is used to investigate the
3 impact of snow cover on ozone concentrations. Section 2 briefly describes the numerical
4 simulations and selected validating observations followed in Sect. 3 by an overview of the 31
5 January–6 February case study and modelling results. Section 4 illustrates the sensitivity of
6 simulated ozone concentrations during this period to snow cover. Discussion of the results
7 follows in Sect. 5. For further information, see also Neemann (2014).

8

9 **2 Data and methods**

10 **2.1 WRF and CMAQ models and observations**

11 Table 1 summarizes the WRF version 3.5 model setup used in this study. The WRF model is
12 nonhydrostatic, with a pressure-based, terrain-following vertical coordinate system.
13 Simulations herein used 41 vertical levels with the lowest 20 levels within approximately 1
14 km of the terrain surface. Three telescoping, one-way nested domains were employed to place
15 the highest-resolution nest over the Uintah Basin, with grid spacing of 12, 4, and 1.33 km,
16 respectively (Fig. 2a). Operational North American Mesoscale Model (NAM) analyses were
17 used to initialize atmospheric and land surface variables (except for snow variables, see the
18 following subsection) as well as provide the lateral boundary conditions for the outer domain
19 at 6 h intervals. We evaluate the core period (17:00 MST 31 January 2013 to 17:00 MST 6
20 February 2013) of the CAP in the Uintah Basin that lasted from 30 January to 10 February
21 2013.

22 WRF output from the 4 km domain was imported into the Utah Division of Air
23 Quality's (UDAQ) CMAQ model (version 5.0). The CMAQ model couples the
24 meteorological data from WRF with an emission inventory from the Uintah Basin developed
25 by UDAQ and chemistry-transport and photochemical subsystems to simulate concentrations
26 for a variety of chemical compounds and pollutants (Byun and Schere, 2006). The emission
27 inventory is for 2011 based on growth of oil & gas activities since 2006 (Barickman, 2014).
28 VOC oil & gas emission speciation profiles are provided by EPA's SPECIATE 4.3 database
29 (EPA, 2012). Default (i.e., not Uintah Basin specific) speciation profiles are used, which
30 assign generic natural gas flaring and crude oil storage tank profiles to the oil & gas
31 emissions. Since the UDAQ inventory and CMAQ model are available at a resolution of 4
32 km, that model was forced with WRF data from the 4 km nest (Fig. 2a).

1 Selected meteorological and surface ozone observations obtained during the UBWOS
2 were used to describe the overall evolution of the CAP episode and to compare to the model
3 results. A subset of six representative meteorological stations in the basin and archived in
4 MesoWest (Horel et al., 2002) was selected to validate simulated 2 m temperature (see Fig.
5 2). Vertical profiles of temperature, dew point temperature, and wind from rawinsondes
6 released at midday (11:00 MST) near Roosevelt on 1–6 February 2013 were used to evaluate
7 the model’s ability to reproduce the vertical structure of the boundary layer. Additional
8 profiles of wind, temperature, and chemical species in the boundary-layer available in the
9 east-central portion of the basin as part of UBWOS are being examined by other researchers
10 (e.g., Oltmans et al. 2014).

11 A Vaisala CL-31 laser ceilometer located at Roosevelt provided aerosol backscatter,
12 the presence of low clouds, and an estimate of the depth of the aerosol layer. Finally, snow-
13 cloud and nighttime microphysics RGB imagery from the NASA Short-term Prediction
14 Research and Transition Center (SPoRT) was used to determine the spatial extent of ice fog
15 within the basin.

16 **2.2 Prescribing initial WRF snow cover in Uintah Basin**

17 While NAM analyses represented the spatial coverage of snow during the 31 January–6
18 February 2013 period fairly well, they overestimated snow depth and snow water equivalent
19 (SWE) within the basin and underestimated them at higher elevations. In order to better
20 represent the actual snow surface conditions, an “idealized” layer of snow and SWE was
21 specified in the WRF initialization fields based on elevation in a manner similar to Alcott and
22 Steenburgh (2013). This prescribed snow cover was determined using: Snowpack Telemetry
23 observations; National Operational Hydrologic Remote Sensing Center analyses; Moderate
24 Resolution Imaging Spectroradiometer imagery; and manual and automated observations
25 from the Community Collaborative Rain, Hail, and Snow Network, and those collected during
26 the UBWOS campaign. As shown in Fig. 3c, the prescribed snow cover was applied within all
27 model domains with no snow cover outside of the Uintah Basin below an elevation of 2000 m
28 and a 17 cm snow depth from the basin floor up to an elevation of 2000 m. (Unless referenced
29 specifically as a.g.l., all elevations are given in m.s.l.). Above 2000 m, the snow depth was
30 elevation-dependent, increasing to 100 cm for elevations at 2900 m or higher.

31 In addition to poor representation of snow depth and SWE, the NAM analyses
32 underestimated snow albedo relative to observed shortwave radiation measurements at

1 Horsepool and Roosevelt (HOR and ROO, respectively in Fig. 2b). The surface albedo
2 averaged from 1 January–2 March 2013 at Horsepool was 0.82 (Roberts et al., 2014), which
3 is roughly 0.17 higher than the NAM analyses during the 31 January–6 February period. Very
4 low temperatures combined with repeated ice fog deposition onto the snow surface during
5 many nights apparently maintained the highly reflective surface. Hence, the snow albedo
6 variable in WRF was initialized to be 0.82 inside the basin. Furthermore, based on visual
7 observations of the snow covering nearly all of the sparse vegetation in the basin during the
8 31 January–6 February period, changes were made to the WRF vegetation parameter table for
9 the two dominant vegetation/land use types: “shrubland” and “cropland/grassland mosaic”.
10 For these vegetation types, 20 kg m^{-2} of SWE was allowed to fully cover the vegetation in the
11 Noah land surface model. The combination of increasing the snow albedo and modifying the
12 vegetation parameter table enabled the model surface to attain the high surface albedo
13 observed during the field campaign (compare Fig. 3a to b).

14 **2.3 Numerical sensitivity studies**

15 Sensitivity tests were conducted with the WRF model to evaluate the impact of variations in
16 cloud type and snow cover on CAPs in the Uintah Basin (Table 2). In order to test the
17 sensitivity of the Uintah Basin CAP to ice vs. liquid phase cloud particles, the default
18 Thompson microphysics scheme used in the BASE simulation was modified in the FULL
19 simulation to enhance the production of ice fog and low clouds by turning off cloud ice
20 sedimentation and the autoconversion of cloud ice to snow in the lowest 15 model layers (\sim
21 500 m a.g.l.). These changes allowed low-level cloud ice to remain suspended and thrive
22 through vapour deposition due to the lower vapour pressure over ice compared to water.
23 Recent research has shown that small ice particles suspended in ice fog have a much slower
24 rate of gravitational settling than the ice particles found in cirrus clouds (for which the settling
25 rates in the default WRF Thompson microphysics scheme were designed). Fall speeds are
26 often less than 1 cm s^{-1} for small ($< 20 \mu\text{m}$) ice fog particles (Heymsfield et al., 2013; Schmitt
27 et al., 2013; Kim et al., 2014), and can be more than 9 times slower than speeds calculated in
28 the original Thompson scheme for particles smaller than $15 \mu\text{m}$. Further, ice-dominant clouds
29 have reduced radiative effects compared to liquid-dominant clouds (Shupe and Intrieri, 2004),
30 allowing for stronger CAP formation, shallower PBLs, and lower near-surface temperatures.

31 The BASE and FULL simulations use the prescribed snow cover as shown in Fig. 3c.
32 As discussed in the Introduction, large snow cover variations are observed from one February

1 to another in the Uintah Basin. To examine the sensitivity of the conditions in the basin to
2 snow cover, the NONE simulation uses the same model configuration as the FULL simulation
3 for the 31 January–6 February period but snow is removed for elevations below 2000 m in the
4 basin (Fig. 3d), which is similar to what was observed during February 2012 and late
5 February 2014.

6

7 **3 Results**

8 **3.1 Overview of the 1–6 February 2013 CAP**

9 A deep upper-level trough and associated midlatitude cyclone moved across Utah from 28–30
10 January 2013, bringing very cold air aloft (700 hPa temperatures ~ -20 °C) and 1–5 cm of
11 light snowfall on top of a ~ 10 –20 cm base to the Uintah Basin. Following the upper-level
12 trough passage, 1–6 February was dominated by upper level ridging over the western United
13 States with large-scale subsidence and mid-level warming over the Uintah Basin. During the
14 CAP, warm air aloft (700 hPa temperatures between ~ -7 and 0 °C) was observed
15 overtopping very cold low-level air (diurnally ranging between ~ -18 and -5 °C), which
16 resulted in a strong capping inversion within the basin (Fig 4a-b). In addition, the presence of
17 fresh snow cover, quiescent surface weather conditions, and sufficient incoming solar
18 insolation to drive photochemistry set the stage for a high ozone episode.

19 Ice fog and low stratus were commonly observed during the 31 January–6 February
20 2013 CAP in the lowest reaches of the basin, typically breaking up during the late morning
21 and afternoon hours into hazy skies. A mid-day satellite image on 2 February 2013 indicates
22 that the lower elevations of the Uintah Basin were snow covered with fog and stratus confined
23 to the lowest elevations of the basin (Fig. 5a). A Visible Infrared Imaging Radiometer Suite
24 (VIIRS) nighttime microphysics RGB image during the previous night (Fig. 5b) helps
25 confirm that the low clouds in the basin are ice-phase stratus and fog. This product
26 discriminates particle phase by combining data from the 3.9, 10.8, and 12.0 micron infrared
27 channels. Liquid-phase low stratus and fog are represented by aqua/green colours (e.g.,
28 southern ID and portions of western and central UT) while the yellow/orange colours evident
29 in the basin are typically associated with ice-phase stratus and fog. The elevation dependence
30 of the fog/stratus is evident in Fig. 5b by the cloud tendrils extending up the river valleys
31 within the basin. The clouds observed within the basin generally occurred within the lowest

1 300 m of the boundary-layer within a shallow mixed layer with high relative humidity with
2 respect to ice (Fig. 4a and c). The strong stability observed on 4 February extends upwards to
3 2750 m with decreasing moisture aloft. Weak easterly winds of $2\text{--}3\text{ m s}^{-1}$ at the base of the
4 stable layer near 2000 m give way to westerly winds of $\sim 10\text{--}12\text{ m s}^{-1}$ near the top of the
5 inversion layer (Fig. 4e and g). The observed mixed layer was shallower the following day as
6 a result of thinner surface ice fog (Fig. 4b) with $\sim 10\%$ lower relative humidity within the
7 upper portions of the CAP. Weak easterly winds are present again near 2000 m with westerly
8 winds of $7\text{--}9\text{ m s}^{-1}$ in the upper reaches of the capping stable layer.

9 The observed ozone concentrations in the basin exceeded the EPA standard of 75 ppb
10 beginning during the afternoon of 1 February at Horsepool and Ouray (HOR and OUR in Fig.
11 2b) and continued to increase through 6 February (Fig. 6a). A weak weather system moved
12 across the basin after 17:00 MST 6 February that lowered the ozone concentrations. However,
13 elevated ozone levels continued until 9 February, after which a stronger weather system with
14 sufficient cold-air advection aloft to destabilize the column moved through the region (not
15 shown). Ozone concentrations near the small cities of Roosevelt and Vernal reach lower
16 afternoon peaks and decrease to background levels at night as a result of NO_x titration
17 (Edwards et al., 2013; Lyman et al., 2014).

18 Figure 6b presents the time evolution of aerosol backscatter, low clouds, and an
19 estimate of the depth of the aerosol layer from the Roosevelt laser ceilometer during the 31
20 January–6 February period. Aerosol backscatter profiles collected at 16 s intervals are
21 averaged into hourly profiles. Fewer aerosols were observed on 1 February followed early the
22 next morning by the development of ice fog evident as well in Fig. 5b. Then, a semi-regular
23 pattern developed over the next several days with shallow nighttime fog and low clouds
24 thinning by mid-day and followed by a deeper layer of aerosols in the afternoon that quickly
25 collapsed at sunset. The ceilometer backscatter data also corroborates other observations that
26 the fog and low cloud occurrence in the basin peaked during 3–4 February. During that time,
27 significant hoar frost was observed on trees and other surfaces after sunrise with light
28 accumulations of snow crystals falling out of the ice clouds in Roosevelt later in the morning.
29 The high levels of aerosol backscatter on 5–6 February diminished near 17:00 MST 6
30 February as a result of the weak weather system mentioned earlier (not shown).

31

32 **3.2 BASE simulation**

1 WRF model simulations were conducted to improve our understanding of the spatiotemporal
2 characteristics of temperature, wind, and moisture throughout the basin during the 31
3 January–6 February 2013 CAP and to investigate the role of snow cover and low clouds on
4 the CAP’s evolution (Table 2). Evaluation of the BASE model simulation on the 1.33 km
5 inner grid relative to the available observations confirms that the WRF model captures the
6 salient temperature, moisture and wind features of the CAP episode throughout the 31
7 January-6 February period above the boundary layer, i.e., above ~ 2100 m. For example, the
8 simulated potential temperature profiles from the BASE simulation at elevations above 2100
9 m agree well with the observed mid-day profiles at Roosevelt (Fig. 4a and b). In addition, the
10 weak easterly flow of $\sim 2 \text{ m s}^{-1}$ observed between 1750 and 2050 m near the base of the
11 stable layer aloft is also evident in the BASE simulation (Fig. 4e–h).

12 However, the simulated surface-based mixed layers at mid-day in the BASE
13 simulation are unrealistically warm and deep compared to observations (Fig. 4a and b). The 2
14 m temperature bias (model – observed) for the BASE simulation averaged over the 6
15 representative surface stations in the centre of the basin is $1.65 \text{ }^\circ\text{C}$ (Table 3). These biases are
16 partially related to thick layers of liquid fog and stratus in the BASE simulation as
17 demonstrated in the next subsection.

18

19 **3.3 Sensitivity to cloud type**

20 Straightforward modifications to the Thompson microphysics scheme (Sect. 2.3) employed in
21 the FULL model run make it possible to examine the sensitivity of the CAP simulations to
22 cloud type. As detailed in Table 2, the FULL simulation has cloud ice sedimentation and
23 cloud ice autoconversion to snow turned off in the lowest 15 model levels. These
24 modifications force the WRF model to produce and maintain clouds dominated by ice-phase
25 particles, and effectively act to achieve similar results as decreased gravitational settling rates
26 introduced for ice fog by Kim et al. (2014). Returning to Fig. 4a and b, the mid-day potential
27 temperature profiles at Roosevelt from the FULL simulation exhibit lower temperatures near
28 the surface and a thinner CAP compared to the BASE simulation. Further, the bias and mean
29 errors relative to the observations of the FULL simulation compared to the BASE simulation
30 are reduced by $\sim 1 \text{ }^\circ\text{C}$ (Table 3). When averaged over the entire simulation period, the
31 reduced 2 m temperatures throughout the basin demonstrate the colder CAP in the FULL

1 simulation relative to the BASE simulation (Fig. 7). Figure 8a indicates the ~ 1.5 °C
2 difference between those two fields in the interior of the Basin.

3 Comparing the temporal evolution of the potential temperature, cloud water and ice
4 profiles at Horsepool between the BASE and FULL simulations further illustrates the impact
5 of cloud type on CAP thermodynamics (Fig. 9). The 1–3 °C colder surface temperatures
6 noted in the FULL compared to the BASE simulation are associated with extensive ice fog
7 that occurred in the FULL simulation between the surface and the bottom of the capping
8 inversion (Fig. 9a and b). The base of the capping inversion (approximately represented by
9 the ~ 280 K potential temperature isotherm in Fig. 9a-b and the ~ 290 K isotherm in Fig 9c)
10 associated with the top of the stratus clouds in BASE (Fig. 9a) averages 100–200 m higher
11 than the top of the ice fog simulated in FULL (Fig. 9b). The ice and liquid clouds simulated in
12 FULL and BASE also have a diurnal cycle, with higher liquid and ice cloud amounts during
13 the night than during the day, but the simulated cloud occurrences are overestimated in both
14 simulations compared to ceilometer observations, particularly on the 5th and 6th of February
15 (Figs 6b and 9b). However, the surface-based depth of the ice fog in FULL is more realistic
16 than the deeper and elevated stratus cloud seen in BASE and the simulated vertical
17 temperature profile in FULL also more closely matches available observations (e.g., Fig. 4a
18 and b).

19 The improved vertical temperature and cloud profiles and 2 m temperatures in the
20 FULL simulation are related to the compositional change of the fog and stratus clouds in the
21 CAP, i.e., cloud water in the BASE simulation compared to cloud ice in the FULL simulation.
22 Snapshots of the cloud characteristics at 23:00 MST 4 February (Fig. 10) reflect similar total
23 cloud amounts and coverage. The BASE run is dominated by liquid-phase particles (Fig. 10c)
24 while the FULL run is dominated by ice-phase particles (Fig. 10d). The preferential tendency
25 for stratus clouds in the BASE simulation due to its deeper CAP leads to cloud cover
26 extending outward farther away from the lowest elevations of the basin compared to the
27 shallower surface-based fogs typically produced during the FULL run. Although the
28 elimination of cloud ice sedimentation leads to greater cloud mass in that run relative to the
29 BASE simulation (compare Fig. 10c to d), the cloud water in the BASE run results in 70–80
30 W m^{-2} of downwelling longwave radiation in the core of the basin while the cloud ice in the
31 FULL run produces only 40–70 W m^{-2} over the same region (compare Fig. 10e to f).
32 Averaged over the entire 6 day period, downwelling longwave radiation from the cloud water

1 is $10\text{--}20 \text{ W m}^{-2}$ more than from the cloud ice (Fig. 8b), which is consistent with the elevated
2 temperatures over the entire period as well (Fig. 8a). The greatest difference in 2 m
3 temperature is at the low elevations in centre of the basin, while the greatest difference in
4 longwave radiation is mid-way up the basin slope where cloud water is present in the BASE
5 run and cloud ice is not found in the FULL simulation.

6

7 **3.4 Sensitivity to snow cover**

8 The simulation with no snow cover in the basin (NONE) for the 31 January–6 February 2013
9 period is now compared to the FULL simulation. The lack of snow in the basin increases the
10 average CAP temperatures by as much as $8 \text{ }^\circ\text{C}$ (Figs. 7 and 9), which is unrealistic relative to
11 those observed (Table 3). While the CAP depth in the NONE simulation is also unrealistically
12 deep (because this particular CAP was snow-covered), the lack of snow has negligible effects
13 aloft (Figs. 4a and b, 9). The modelled afternoon mixed layer depths of 400–800 m in the
14 NONE simulation were somewhat less than those observed in the basin during snow-free
15 conditions the previous winter (Lyman and Shorthill 2013), and clouds formed within the
16 NONE simulations whereas clouds were only infrequently observed in the Uintah Basin in
17 winter 2011–2012 when snow cover was absent. However, a number of factors contribute to
18 mixed layer depth and cloud occurrence, including the strength of the synoptic-scale capping
19 inversion and boundary-layer relative humidity, so evaluating overall model simulation
20 performance during snow-free conditions within the basin would require simulating periods
21 when snow was not present, such as during the 2011–2012 winter. Several interrelated
22 processes contribute to the high low-level temperatures and deep afternoon CAP in the NONE
23 simulation relative to the FULL simulation. First, when the snow is removed from the basin
24 floor, the thermal conductivity of the land surface increases, and the decrease in surface
25 albedo results in greater absorption of solar radiation. Second, the sensitivity of the CAP to
26 ice-phase microphysics is minimized in the NONE simulation since the boundary layer over
27 the bare ground/vegetation is too warm (i.e., higher than $-12 \text{ }^\circ\text{C}$) to nucleate cloud ice. The
28 resulting liquid-phase stratus in the NONE simulation leads to increased longwave radiation
29 at the surface. Finally, the cloud thickness and occurrence in the warmer NONE boundary-
30 layer is reduced compared to FULL, resulting in greater incoming short-wave radiation (Fig.
31 9c).

1

2 **3.5 Flow features**

3 While the observations collected during the UBWOS field campaigns are the most extensive
4 available to date for studying the thermodynamic and dynamic conditions in the Uintah Basin
5 (Lyman and Shorthill, 2013; Stoeckenius and McNally, 2014), the majority of them consist of
6 enhanced surface observations throughout the basin combined with vertical profiles at only a
7 few locations (e.g., Horsepool, Ouray, and Roosevelt). The FULL simulation is used here to
8 examine the four-dimensional fields of temperature, wind, and moisture to help identify
9 relevant physical processes. We focus on several flow features evident in the FULL
10 simulation that could be validated using the available data and which likely play an important
11 role to transport pollutants within the CAP.

12 **3.5.1 Clean-air intrusions into the basin**

13 CAP structure varies extensively, both temporally and spatially, over the course of the FULL
14 simulation. Time height potential temperature profiles at Horsepool suggest that the CAP is
15 initially confined to elevations below 1700 m before it deepens to a base near 1850 m early on
16 3 February (Fig. 9b). By midday on 4 February, the inversion base retreats to 1800 m, and
17 eventually lowers to ~ 1700 m from early on 6 February through the end of the simulation.
18 The CAP is continually modulated by synoptically-driven mid-level flow atop the CAP,
19 forcing it to “slosh” back and forth within the basin. Ridging aloft can lead to flow
20 surmounting the surrounding terrain from nearly every direction from the southwest to the
21 north. Downsloping flows mixing higher potential temperature and cleaner air downward into
22 the basin are common and their impact depends on the stability and strength of the flow
23 across the upwind barriers. For example, when the cross-barrier flow had a northerly
24 component across the high Uintah Mountains during the 2013 winter, a notable strengthening
25 of the inversion top due to subsidence warming of flow descending in the lee of the mountains
26 was evident in the Uintah Basin (not shown).

27 The CAP may become displaced or tilted through hydrostatic and dynamic processes,
28 which can then be disrupted by changes in wind speed above the CAP (Lareau and Horel,
29 2014). These disruptions produce gravity current features as the CAP rebounds, causing
30 relatively large changes in depth (a few hundred meters) within just a few hours. Figure 11
31 shows an example of this type of behaviour. Strong westerly flow crossing the mountain
32 barrier to the west of the basin at 23:00 MST 3 February is highlighted by a narrow band of

1 increased westerly to northwesterly flow at 2.3 km over the western portion of the basin (Fig.
2 11a). The cross section of potential temperature from west to east through the centre of the
3 basin at the same time is shown in Fig. 11b. The westerly downslope winds have eroded and
4 depressed the CAP ~200 m along the western basin slope. The FULL simulation suggests that
5 weakening westerly winds over the next several hours lead to the CAP rebounding westward
6 with the inversion base quickly rising to ~ 1900 m, roughly level with the rest of the basin
7 (Fig. 11b). Lower ozone concentrations were also advected down the western slope
8 of the basin by the westerly winds while the primary region of high ozone
9 concentrations simulated in the eastern sector of the basin remained undisturbed
10 (not shown).

11 **3.5.2 East-west cross basin transport**

12 Easterly flow immediately above the shallow mixed layer is evident in the mid-day soundings
13 at Roosevelt on a number of days (Fig. 4). The ceilometer data at Roosevelt (Fig. 6b) as well
14 as ozone tethered observations at Ouray (Schnell et al., 2014) suggest that aerosols, ozone
15 precursors, and ozone extend upward into this layer of easterly flow likely as a result of weak
16 turbulence and entrainment (Cai and Luhar, 2002; Salmond, 2005). The ozone precursors
17 from eastern basin source regions that are able to leak into the easterly flow layer may then be
18 transported westward to portions of the basin that have more limited precursor sources,
19 allowing ozone production to take place more widely (Karion et al., 2014; Oltmans et al.
20 2014).

21 Figure 12 shows the time-averaged zonal wind component from the FULL simulation
22 along the cross section shown in Fig. 2b, split into daytime and nighttime periods. Synoptic
23 westerly flow dominates above 2200 m with easterly flow present a few hundred meters
24 above the basin floor. The core of the easterly flow coincides with the strongest stability (see
25 Fig. 9b) in the basin and lies between 1800–2000 m. Although this feature is relatively weak
26 ($\sim 1 \text{ m s}^{-1}$ during the day, 0.5 m s^{-1} at night), it is persistent enough to appear as a coherent
27 spatial pattern when averaged over the 6 day period. During the day, the core of the easterly
28 flow is more intense aloft, and the west-east spatial extent is greater (compare Fig. 12a to b).
29 At night, the easterly flow exhibits a weaker and more regional core shifted to the eastern
30 portion of the basin and extending down to the surface (Fig. 12b).

31 **3.5.3 Thermally-driven valley and slope flows**

1 Thermally-driven daytime upvalley/upslope and nighttime downvalley/downslope flows were
2 observed within the basin by Lyman et al. (2013), while cross-basin elevated easterly flows
3 100-300 m a.g.l. (possibly associated with basin-scale thermal gradients) were observed in
4 rawinsonde soundings between 1-6 February 2013 at Roosevelt (Fig. 4). Within the model
5 simulations, it appears that both additive and destructive interactions between the cross-basin
6 elevated easterly flows and near-surface daytime upvalley/upslope and nighttime downvalley/
7 downslope flows are occurring (Fig. 12). While basin-scale thermal gradients likely drive the
8 elevated easterly flow, those gradients are at times in concert with and at other times
9 interfering with more localized thermal gradients within drainages and along slopes. We
10 hypothesize that the basin-scale thermal gradients are associated with either (1) elevated
11 heating on the western slope of the basin, or (2) interactions between the westerly downslope
12 flow, the cold air pool, and the mountain slope.

13 During the night (Fig. 12b), drainage flows are evident by light westerly winds in the
14 lowest 100 m a.g.l. on the west side of the basin in combination with light easterly winds on
15 the east side. This pattern reverses during the day (Fig. 12a); however, the cross-basin
16 easterlies appear to accentuate the upvalley/upslope winds at ~ 1800 m. As with any basin or
17 mountain range, the diurnal flow patterns within the Uintah Basin are complex. An
18 examination of mean wind direction during the day (not shown) highlights areas of upslope
19 easterly flow within the CAP in the western half of the basin. Outside of the CAP, however,
20 to the north and west of the basin, synoptic west-northwesterly flow dominates. This
21 demonstrates how the strong stability above the CAP is able to effectively shield the basin
22 interior from synoptic flows, allowing for weak thermally-driven circulations to become
23 important.

24 **3.5.4 Effects of snow cover on terrain-flow interactions**

25 The sensitivity of terrain-flow interactions to the presence or absence of snow cover in the
26 Uintah Basin is briefly examined here. Comparison of the cross sections of time-averaged
27 zonal winds from the FULL (Fig. 12a and b) and NONE (Fig. 12c and d) simulations are
28 consistent with earlier results: the removal of snow cover only affects the near-surface
29 atmosphere below the capping inversion. The weaker stability within the capping inversion in
30 the NONE simulation likely allows the synoptic-scale westerlies to extend further down
31 toward the basin floor. This extension appears to diminish the intensity of the easterly winds
32 within the lower reaches of the inversion layer that would be expected in NONE given the

1 lack of snow cover. Comparable differences are evident during the day (Fig. 12a and c) and
2 night (Fig. 12b and d) with weaker and lower elevation easterly flow aloft when snow cover is
3 removed. However, the intensity of the upvalley/upslope and downvalley/downslope flows
4 near the surface remains largely the same and is actually increased during the day on the
5 western side of the basin in the NONE simulation.

6

7 **4. Ozone**

8 **4.1 Overview**

9 The January–March 2013 period featured seven persistent CAPs with high ozone
10 concentrations in the Uintah Basin (Stoeckenius and McNally, 2014). The CAP that began on
11 31 January led to increasing ozone concentrations over the next week (Fig. 6). Ozone
12 concentrations started out relatively low on 31 January (~ 20 to 60 ppb) and gradually built to
13 a maximum of 154 ppb at Ouray on 6 February. Two key characteristics of ozone
14 concentrations in the Uintah Basin are the (1) maintenance of high ozone levels above
15 background levels over night in some areas of the basin, and (2) the pooling of the highest
16 ozone values in lower elevations and river valleys, particularly in the southeastern quadrant
17 near Horsepool and Ouray (Fig. 13). Data collected from ozonesondes and tethersondes
18 during February 2013 show that the vertical extent of maximum ozone concentrations was
19 typically limited to 1700 m and below, or in the lowest 200–300 m a.g.l. of the boundary-
20 layer (Schnell et al., 2014). A gradient in concentrations was noted above this level, with
21 ozone concentrations returning to background levels above 1900 m (Karion et al., 2014;
22 Oltmans et al. 2014).

23 **4.2 Sensitivity of ozone concentrations to snow cover**

24 While ozone concentrations in the Uintah Basin are recognized to be strongly controlled by
25 snow cover, the presence of snow has two complementary effects: (1) higher albedo
26 enhancing photochemistry and (2) reduced near-surface temperatures; shallower CAP; and
27 possibly enhanced east–west cross-basin transport a few hundred meters above the surface.
28 For example, crude estimates of the actinic flux from the WRF FULL and NONE simulations
29 provide an example of these complementary effects. The cloud ice typically present in the
30 colder CAP found in the FULL simulation allows greater penetration of solar radiation to the
31 surface than the cloud water observed between 2–4 February in the NONE simulation (Fig. 9

1 b and c). Hence, more downward solar radiation during that time was then available to be
2 reflected by the snow cover.

3 The objective of this phase of the study is to simply assess the sensitivity of WRF-
4 CMAQ simulated ozone concentrations to snow cover during a CAP. The potential
5 shortcomings of driving CMAQ from imperfect atmospheric information and emissions
6 inventories (Sect. 2.1) as well as the limitations of CMAQ are not addressed. The mean ozone
7 concentrations near the surface throughout the basin averaged over the 6 afternoons (11:00 to
8 17:00 MST) from 1–6 February 2013 are generally 15–30 % greater when the CMAQ model
9 is forced by the FULL simulation compared to the NONE simulation (compare Fig. 14a to b).
10 As expected, ozone concentrations simulated by the CMAQ model are highest in the
11 southeastern portion of the basin where the emission of ozone precursors (NO_x and VOCs) is
12 greatest (Barickman, 2014). The region where average surface concentrations are greater than
13 75 ppb is ~ 6 times larger in the FULL simulation than that in the NONE simulation. In
14 addition, the peak ozone concentration simulated in the FULL case is 16 ppb higher than that
15 from the NONE case (Table 4) and the timing and magnitude of the peak value on 6 February
16 in the FULL case is comparable to that observed (see Figs. 6 and 13). A comparison of east–
17 west vertical cross sections of ozone (averaged along a 24 km wide swath approximately 25
18 km south of the red line in Fig. 2b) demonstrates the vertical extent of the higher ozone
19 concentrations generated in the FULL vs. NONE simulations (Fig. 14c and d).

20 Ozone concentrations from the two CMAQ simulations are compared to those
21 observed at Roosevelt and Horsepool in Fig. 14e and f. CMAQ struggles to simulate the
22 ozone buildup at Roosevelt in the western portion of the basin whether snow is present or not
23 (Fig. 14f). Closer to the primary precursor emission sources in the southeastern section of the
24 basin, substantially higher ozone concentrations are evident at Horsepool in the FULL
25 simulation compared to when snow is removed (Fig. 14e).

26 A time-height plot of ozone concentration and potential temperature at Horsepool
27 from the FULL and NONE simulations helps to highlight some of the deficiencies of the
28 CMAQ simulations for this case (Fig. 9b-c). While the largest concentrations of ozone are
29 confined within the CAP, elevated concentrations in excess of 75 ppb extend higher than
30 observed at Horsepool (Karion et al., 2014). In addition, CMAQ fails to build ozone
31 concentrations from day-to-day through the event (Fig. 6). Instead, the highest concentrations
32 appear to be controlled by the simulated mixed layer depth, e.g., concentrations are high

1 during the late afternoon/early evening on 1 and 2 February, when the mixed layer is shallow,
2 then they decrease on the 3rd and 4th as the inversion base lifts to ~ 1800 m. As the inversion
3 base lowers again on 5 and 6 February, concentrations increase with a maximum during the
4 afternoon of the 6th. A similar evolution is noted in the NONE simulation, but the mixed
5 layer is much deeper, concentrations are lower, and the maximum occurs on the afternoon of
6 the 5th (Fig. 9c). However, the modelled ozone concentrations on some days with no snow
7 cover appear to be unrealistically high compared to observations during other snow-free
8 periods. While this inverse relationship between mixed layer depth and ozone concentrations
9 is understandable physically, i.e., when the inversion base lowers it effectively decreases CAP
10 volume, the observations during this case suggest other processes play a role as well.

11

12 **5 Summary and discussion**

13 The 31 January–6 February 2013 CAP in the Uintah Basin is examined and simulations are
14 used to evaluate its sensitivity to cloud microphysics and snow cover. Output from
15 meteorological simulations was input into the CMAQ model to relate ozone production to
16 snow cover. The key findings of this study can be summarized as follows:

- 17 • The WRF mesoscale meteorological model was able to simulate a strong wintertime
18 capping temperature inversion above the boundary-layer within the Uintah Basin
19 irrespective of surface or cloud characteristics.
- 20 • The CAP characteristics below ~ 2100 m (stable layer intensity, vertical structure, and
21 boundary-layer flows) are heavily influenced by the presence of snow cover and are
22 modulated to a lesser extent by the numerical treatment of cloud microphysics.
- 23 • The default settings in the Thompson microphysics scheme produce dense, liquid-
24 phase low clouds and fog that were not observed during this event, whereas restricting
25 cloud ice sedimentation and conversion to snow in the lowest model layers resulted in
26 more realistic vertical profiles of temperature and low clouds.
- 27 • Intrusions of clean air into the basin as a result of terrain-flow interactions, east-to-
28 west cross-basin advection above the surface, and shallow thermally-driven slope and
29 valley circulations are likely important factors for mixing pollutants throughout the
30 Uintah Basin.

- CMAQ model-derived estimates of ozone concentrations agree better with observations (1) during the daytime than during the nighttime and (2) near the highly dense precursor emission sources located in the southeast quadrant of the basin (Fig. 14).” The numerical simulations presented herein provide additional support to previous observational evidence that snow cover affects ozone concentrations by (1) cooling the near-surface layer and thereby strengthening the CAP and increasing stability further aloft, and (2) increasing the surface albedo and subsequent photolysis rates, contributing to rapid ozone production.

As in many model sensitivity studies focused on specific physical processes, there are a number of caveats to consider. First, the work presented here has been limited to a single CAP event. In order to obtain a more thorough understanding of how cloud microphysics and snow cover affect the evolution of CAPs, their wind flow patterns, and resulting impacts on air quality, further cases need to be examined. For example, if no clouds are allowed to form during the CAP lifecycle, this results in shallower mixed-layer heights and a colder CAP by several °C than that simulated with ice fog present (Neemann 2014). Second, the modelling capability for the highly stable CAP meteorological conditions in the Uintah Basin lags behind typical meteorological situations; improvements in the parameterization of stable boundary layers and ice fog processes in numerical models are needed in order to obtain improved CAP simulations (Holtslag et al., 2013; Gultepe et al., 2014). Third, the idealized prescription of snow depth and albedo to constant values throughout the basin are imperfect estimates. Improvements in the representation of snow variables in meteorological and air quality models and analysis initialization fields in regions with shallow, persistent snow cover such as the Uintah Basin are needed. Fourth, additional data (particularly with regard to chemical species) exists in the east-central part of the basin (e.g. at Horsepool and Ouray) that could be used to evaluate meteorological and chemical model performance. Finally, significant uncertainty exists regarding precursor emission estimates within the basin. We elaborate further on several of these points in the following paragraphs.

As discussed by Gultepe et al. (2014), additional research is needed to understand ice fog microphysics and how to parameterize these processes in numerical models. Future research to investigate the impact of employing the recent WRF ice-fog scheme of Kim et al. (2014) on cloud formation in the Uintah Basin is recommended. For this study, we neglected the fall speed of the ice fog particles to ensure that cloud ice was retained by the modified Thompson

1 microphysics scheme. In addition, the effects of the unusually high ozone and particulate
2 concentrations in the Uintah Basin on the ice nucleation processes are unknown, although
3 studies suggest ice fog can be enhanced by anthropogenic activities (Benson, 1965; Kumai
4 and O'Brien, 1965; Schmitt et al., 2013; Kim et al., 2014). While we did not find any
5 perceptible difference in CAP simulations by varying the cloud droplet concentrations in the
6 Thompson scheme from the default ($100 \times 10^6 \text{ m}^{-3}$) to those typically assumed for continental
7 ($300 \times 10^6 \text{ m}^{-3}$) or hypothetical polluted continental ($1000 \times 10^6 \text{ m}^{-3}$) situations, we
8 recommend further testing along these lines, including testing the newly available aerosol-
9 aware Thompson scheme (Thompson and Eidhammer, 2014).

10 Further work to improve parameterization schemes for modelling very stable boundary
11 layers and their impact on CAP simulations is also needed (Baklanov et al., 2011). PBL
12 schemes have difficulties handling low clouds, vertical temperature profiles, 2 m
13 temperatures, and mixing in stably stratified conditions (Reeves et al., 2011; Shin and Hong,
14 2011; Zhang et al., 2013). Many schemes typically allow too much turbulent mixing in stable
15 conditions, which results in boundary layers that are too deep (Holtslag et al., 2013). While
16 the MYJ PBL scheme was ultimately selected for this study, the Asymmetric Convective
17 Model, Grenier-Bretherton-McCaa, and Bretherton-Park PBL schemes were also tested in
18 addition to the Yonsei University (YSU) scheme with and without the Jimenez surface layer
19 formulation and updated stability functions (Jimenez et al., 2012). The MYJ was chosen
20 because in simulating this particular CAP it best represented the combination of moisture,
21 stability, and temperature characteristics that were observed in the Uintah Basin for the
22 simulated period. Further testing of PBL schemes performance in the presence of high wind
23 speeds above the stable PBL are also needed (e.g., Hu et al. 2013 found that modifications to
24 the YSU nighttime velocity scale improved the simulations).

25 Snow cover and albedo were shown to have a prominent impact on simulated CAP
26 evolution and ozone concentrations. However, in remote locations such as the Uintah Basin,
27 where snow cover is typically very thin ($\sim 5\text{--}10$ cm) and variable, accurately assessing snow
28 mass or water equivalent for input into numerical models can be difficult (Jeong et al., 2013).
29 The impacts of modifying the albedo from the default NAM initialization to that observed (an
30 increase of 0.17, see Section 2.2) on the CAP meteorology were relatively small (not shown).
31 However, larger impacts resulting from modest changes in snow albedo are likely to be
32 observed within photochemical models (e.g., Ahmadov et al. 2014). Because of the spatio-

1 temporal variability of snow depth and albedo within the Uintah Basin during winter seasons,
2 the need for more sophisticated representation of snow variables in meteorological and air
3 quality models in this region is apparent, and worthy of future research to better quantify the
4 impact of these improvements on ozone simulations. Proper treatment of both the spatial
5 extent of snow cover as well as the snow surface using a snow physics model driven by local
6 atmospheric and chemical properties (e.g., the three-layer snow model within Noah Multi-
7 Parameterization land surface model; Niu et al., 2011) may be needed to obtain a sufficiently
8 accurate evolution of the snowpack and surface albedo. Additional research is also needed to
9 understand the complex cycling of water over the thin snowpacks in the Uintah Basin and its
10 impact on surface albedo, i.e., the interplay of very small sublimation rates, formation of ice
11 fogs, and deposition of ice crystals back onto the snow surface.

12 Finally, as discussed in Sect. 2.1, the CMAQ emission inventory used in this study was
13 prepared to represent oil & gas activities in 2011 (Barickman, 2014). The emission inventory
14 and VOC speciation profiles for the Uintah Basin remain uncertain and are the subject of
15 ongoing research (e.g., Warneke et al. 2014a, b). The recent work by Ahmadov et al. (2014)
16 illustrates the effect of variations in emission scenarios on ozone levels within the Uintah Basin.
17 Data collected during the 2013 UBWOS will add to the fidelity of these profiles as
18 measurements are incorporated into future inventories. For example, a better understanding
19 regarding how formaldehyde becomes highly concentrated in the basin (through direct
20 emission or secondary chemical reactions) is needed.

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1 **References**

- 2 Ahmadov, R., McKeen, S., Trainer, M., Banta, R., Brewer, A., Brown, S., Edwards, P. M.,
3 de Gouw, J. A., Frost, G. J., Gilman, J., Helmig, D., Johnson, B., Karion, A., Koss, A.,
4 Langford, A., Lerner, B., Olson, J., Oltmans, S., Peischl, J., Pétron, G., Pichugina, Y.,
5 Roberts, J. M., Ryerson, T., Schnell, R., Senff, C., Sweeney, C., Thompson, C., Veres, P.,
6 Warneke, C., Wild, R., Williams, E. J., Yuan, B., and Zamora, R.: Understanding high
7 wintertime ozone pollution events in an oil and natural gas producing region of the western
8 US, *Atmos. Chem. Phys. Discuss.*, 14, 20295-20343, doi:10.5194/acpd-14-20295-2014, 2014.
- 9 Alcott, T. I. and Steenburgh, W. J.: Orographic influences on a Great Salt Lake-effect snow-
10 storm, *Mon. Weather Rev.*, 141, 2432–2450, doi:10.1175/MWR-D-12-00328.1, 2013.
- 11 Baklanov, A. A., Grisogono, B., Bornstein, R., Mahrt, L., Zilitinkevich, S. S., Taylor,
12 P., Larsen, S. E., Rotach, W. M., and Fernando, H. J. S.: The nature, theory, and modelling of
13 atmospheric planetary boundary layers, *B. Am. Meteorol. Soc.*, 92, 123–128,
14 doi:10.1175/2010BAMS2797.1, 2011.
- 15 Barickman, P.: Emission inventory development activities, in: Final report: 2013 Uinta Basin
16 winter ozone study, edited by: Stoeckenius, T. and McNally, D., ENVIRON International
17 Corporation, Novato, California, Chapter 9, 1–7, available at:
18 <http://www.deq.utah.gov/locations/uintahbasin/studies/UBOS-2013.htm> (last access: 13 June
19 2014), 2014.
- 20 Bell, M. L., McDermott, A., Zeger, S. L., Samet, J. M., and Dominici, F.: Ozone and short-
21 term mortality in 95 US urban communities, 1987–2000, *J. Am. Med. Assoc.*, 292, 2372–
22 2378, doi:10.1001/jama.292.19.2372, 2004.
- 23 Benson, C. S.: Ice fog: low temperature air pollution, Geophysical Institute of the University
24 of Alaska, Fairbanks, Alaska, 198 pp., available at:
25 www.dtic.mil/dtic/tr/fulltext/u2/631553.pdf (last access: 13 June 2014), 1965.
- 26 Billings, B. J., Grubišić, V., and Borys, R. D.: Maintenance of a mountain valley cold pool:
27 a numerical study, *Mon. Weather Rev.*, 134, 2266–2278, doi:10.1175/MWR3180.1, 2006.
- 28 Byun, D. and Schere, K. L.: Review of the governing equations, computational algorithms,
29 and other components of the Models-3 Community Multi-Scale Air Quality (CMAQ)
30 modelling system, *Appl. Mech. Rev.*, 59, 51–77, doi:10.1115/1.2128636, 2006.

1 Cai, X.-M. and Luhar, A. K.: Fumigation of pollutants in and above the entrainment zone into
2 a growing convective boundary layer: a large-eddy simulation, *Atmos. Environ.*, 36, 2997–
3 3008, doi:10.1016/S1352-2310(02)00240-6, 2002.

4 Chen, R. and Dudhia, J.: Coupling an advanced land surface–hydrology model with the Penn
5 State–NCAR MM5 modelling system. Part I: model implementation and sensitivity, *Mon.*
6 *Weather Rev.*, 129, 569–585, 2001.

7 Clements, C. B., Whiteman, C. D., and Horel, J. D.: Cold-air-pool structure and evolution in
8 a mountain basin: Peter Sinks, Utah, *J. Appl. Meteorol.*, 42, 752–768, doi:10.1175/1520-
9 0450(2003)042<0752:CSAEIA>2.0.CO;2, 2003.

10 Edwards, P. M., Young, C. J., Aikin, K., deGouw, J., Dube, W. P., Geiger, F., Gilman, J.,
11 Helmig, D., Holloway, J. S., Kercher, J., Lerner, B., Martin, R., McLaren, R., Parrish, D. D.,
12 Peischl, J., Roberts, J. M., Ryerson, T. B., Thornton, J., Warneke, C., Williams, E. J., and
13 Brown, S. S.: Ozone photochemistry in an oil and natural gas extraction region during winter:
14 simulations of a snow-free season in the Uintah Basin, Utah, *Atmos. Chem. Phys.*, 13, 8955–
15 8971, doi:10.5194/acp-13-8955-2013, 2013.

16 Environmental Protection Agency, (EPA), Air quality criteria for ozone related
17 photochemical oxidants, EPA 600/R-05/004aF, US Environ. Prot. Agency, Research Triangle
18 Park, NC, 821, pp., 2006.

19 EPA, SPECIATE 4.3: addendum to SPECIATE 4.2, speciation database development
20 documentation, EPA/600/R-11/121, US Environ. Prot. Agency, Research Triangle Park, NC,
21 28 pp., available at:
22 <http://www.epa.gov/ttn/chief/software/speciate/speciate4/addendum4.2.pdf> (last access: 13
23 June 2014), 2011.

24 EPA: National ambient air quality standards (NAAQS), available at: [http://www.epa.gov/air/
25 criteria.html](http://www.epa.gov/air/criteria.html), (last access: 30 April), 2014.

26 Gultepe, I., Kuhn, T., Pavolonis, M., Calvert, C., Gurka, J., Heymsfield, A. J., Liu, P. S. K.,
27 Zhou, B., Ware, R., Ferrier, B., Milbrandt, J., and Bernstein, B.: Ice fog in arctic during
28 FRAM- ICE fog project: aviation and nowcasting applications, *B. Am. Meteorol. Soc.*, 95,
29 211–226, doi:10.1175/BAMS-D-11-00071.1, 2014.

1 Helmig, D., Thompson, C. R., Evans, J., Boylan, P., Hueber, J., and Park, J.-H.: Highly
2 elevated atmospheric levels of volatile organic compounds in the Uintah Basin, Utah.
3 *Environ. Sci. Technol.*, 48, 4707–4715, doi:10.1021/es405046r, 2014.

4 Heymsfield, A. J., Schmitt, C., and Bansemer, A.: Ice cloud particle size distributions and
5 pressure-dependent terminal velocities from in situ observations at temperatures from 0° to
6 –86 °C, *J. Atmos. Sci.*, 70, 4123–4154, doi:10.1175/JAS-D-12-0124.1, 2013.

7 Holtslag, A. A. M., Svensson, G., Baas, P., Basu, S., Beare, B., Beljaars, A. C. M., Bosveld,
8 F. C., Cuxart, J., Lindvall, J., Steenveld, G. J., Tjernstrom, M., and Van De Wiel, B. J. H.:
9 Stable atmospheric boundary layers and diurnal cycles: challenges for weather and climate
10 models, *B. Am. Meteorol. Soc.*, 94, 1691–1706, doi:10.1175/BAMS-D-11-00187.1, 2013.

11 Horel, J., Splitt, M., Dunn, L., Pechmann, J., White, B., Ciliberti, C., Lazarus, S., Slemmer, J.,
12 Zaff, D., and Burks, J.: Mesowest: cooperative mesonets in the western United States, *B. Am.*
13 *Meteorol. Soc.*, 83, 211–225, doi:10.1175/1520-0477(2002)083<0211:MCMITW>2.3.CO;2,
14 2002.

15 Hu, X. M., P. M. Klein, and M. Xue (2013), Evaluation of the updated YSU planetary
16 boundary layer scheme within WRF for wind resource and air quality assessments, *J Geophys*
17 *Res-Atmos*, 118(18), 10490-10505, Doi 10.1002/Jgrd.50823.

18 Iacono, M. J., Delamere, J. S., Mlawer, E. J., Shephard, M. W., Clough, S. A., and Collins,
19 W. D.: Radiative forcing by long-lived greenhouse gases: calculations with the AER radiative
20 transfer models, *J. Geophys. Res.*, 113, D13103, doi:10.1029/2008JD009944, 2008.

21 Janjic, Z. I.: The Step-mountain eta coordinate model: further developments of the
22 convection, viscous sublayer, and turbulence closure schemes, *Mon. Weather Rev.*, 122, 927–
23 945, doi:10.1175/1520-0493(1994)122<0927:TSMECM>2.0.CO;2, 1994.

24 Jeong, J.-H., Linderholm, H. W., Woo, S.-H., Folland, C., Kim, B.-M., Kim, S.-J., and Chen,
25 D.: Impacts of snow initialization on subseasonal forecasts of surface air temperature for the
26 cold season, *J. Climate*, 26, 1956–1972, doi:10.1175/JCLI-D-12-00159.1, 2013.

27 Jiménez, P. A., Dudhia, J., González-Rouco, J. F., Navarro, J., Montávez, J. P., and García-
28 Bustamante, E.: A revised scheme for the WRF surface layer formulation, *Mon. Weather*
29 *Rev.*, 140, 898–918, doi:10.1175/MWR-D-11-00056.1, 2012.

1 Kain, J. S.: The Kain–Fritsch convective parameterization: an update, *J. Appl. Meteorol.*, 43,
2 170–181, doi:10.1175/1520-0450(2004)043<0170:TKCPAU>2.0.CO;2, 2004.

3 Karion, A., Oltmans, S., Petron, G., Sweeney, C., and Schnell, R.: Analysis of aircraft
4 observations, in: Final report: 2013 Uinta Basin winter ozone study, edited by: Stoeckenius,
5 T. and McNally, D., ENVIRON International Corporation, Novato, California, chapter 4, 1–
6 26, available at: <http://www.deq.utah.gov/locations/uintahbasin/studies/UBOS-2013.htm>, last
7 access: 13 June 2014.

8 Katurji, M. and Zhong, S.: The influence of topography and ambient stability on the
9 characteristics of cold-air pools: a numerical investigation, *J. Appl. Meteorol. Climatol.*, 51,
10 1740–1749, doi:10.1175/JAMC-D-11-0169.1, 2012.

11 Kim, C. K., Stuefer, M., Schmitt, C. G., Heymsfield, A. J., and Thompson, G.: Numerical
12 modelling of ice fog in interior Alaska using the weather research and forecasting model, *Pure
13 Appl. Geophys.*, 1–20, doi:10.1007/s00024-013-0766-7, 2014.

14 Kumai, M. and O’Brien, H. W.: A study of ice fog and ice-fog nuclei at Fairbanks, Alaska,
15 part II. US Army Materiel Command Cold Regions Research and Engineering Laboratory,
16 Hanover, New Hampshire, 19 pp., available at: [http://www.dtic.mil/cgi-bin/GetTRDoc?
17 Location=U2&doc=GetTRDoc.pdf&AD=AD0676811](http://www.dtic.mil/cgi-bin/GetTRDoc?Location=U2&doc=GetTRDoc.pdf&AD=AD0676811) (last access: 13 June 2014), 1965.

18 Lareau, N. P.: The Dynamics of persistent cold-air pool breakup, Ph.D. thesis, University of
19 Utah, Salt Lake City, Utah, 138 pp., 2014.

20 Lareau, N. P. and Horel, J. D.: Dynamically induced displacements of a persistent cold-air
21 pool, *Bound.-Layer Meteorol.*, accepted, 2014.

22 Lareau, N. P., Crosman, E. T., Whiteman, C. D., Horel, J. D., Hoch, S. W., Brown, W. O. J.,
23 and Horst, T. W.: The persistent cold-air pool study, *B. Am. Meteorol. Soc.*, 94, 51–63,
24 doi:10.1175/BAMS-D-11-00255.1, 2013.

25 Lee, L., Wooldridge, P. J., Gilman, J. B., Warneke, C., de Gouw, J., and Cohen, R. C.: Low
26 temperatures enhance organic nitrate formation: evidence from observations in the 2012
27 Uintah Basin Winter Ozone Study, *Atmos. Chem. Phys. Discuss.*, 14, 17401–17438,
28 doi:10.5194/acpd-14-17401-2014, 2014.

29 Li, R., Warneke, C., Graus, M., Field, R., Geiger, F., Veres, P. R., Soltis, J., Li, S.-M.,
30 Murphy, S. M., Sweeney, C., Pétron, G., Roberts, J. M., and de Gouw, J.: Measurements of

1 hydrogen sulfide (H₂S) using PTR-MS: calibration, humidity dependence, inter-comparison
2 and results from field studies in an oil and gas production region, *Atmos. Meas. Tech.*
3 *Discuss.*, 7, 6205-6243, doi:10.5194/amtd-7-6205-2014, 2014.

4 Lippmann, M.: Health effects of tropospheric ozone: review of recent research findings and
5 their implications to ambient air quality standards, *J. Expo. Anal. Env. Epid.*, 3, 103–129,
6 1993.

7 Lu, W. and Zhong, S.: A numerical study of a persistent cold air pool episode in the Salt Lake
8 Valley, Utah, *J. Geophys. Res.-Atmos.*, 119, 1733–1752, doi:10.1002/2013JD020410, 2014.

9 Lyman, S. and Shorthill, H. (Eds.): Final report: 2012 Uintah Basin winter ozone and air
10 quality study, Commercialization and Regional Development, Utah State University, Vernal,
11 Utah, 285 pp., available at: http://rd.usu.edu/files/uploads/ubos_2011-12_final_report.pdf (last
12 access: 13 June 2014), 2013.

13 Lyman, S., Mansfield, M., Shorthill, H., Anderson, R., Mangum, C., Evans, J., and Shorthill,
14 T.: Distributed measurements of air quality and meteorology, in: Final report: 2013 Uinta
15 Basin winter ozone study, edited by: Stoeckenius, T. and McNally, D., ENVIRON
16 International Corporation, Novato, California, chapter 3, 1–35, available at:
17 <http://www.deq.utah.gov/locations/uintahbasin/studies/UBOS-2013.htm>, last access: 13 June
18 2014.

19 Malek, E., Davis, T., Martin, R. S., and Silva, P. J.: Meteorological and environmental aspects
20 of one of the worst national air pollution episodes (January 2004) in Logan, Cache Valley,
21 Utah, USA, *Atmos. Res.*, 79, 108–122, doi:10.1016/j.atmosres.2005.05.003, 2006.

22 Neemann, E. M.: Analysis and simulation of a cold-air pool and high wintertime ozone
23 episode in Utah's Uintah Basin, M.S. thesis, University of Utah, Salt Lake City, Utah, 94 pp.,
24 2014.

25 Niu, G.-Y., Yang, Z.-L., Mitchell, K. E., Chen, F., Ek, M. B., Barlage, M., Kumar, A.,
26 Manning, K., Niyogi, D., Rosero, E., Tewari, M., and Xia, Y.: The community Noah land
27 surface model with multiparameterization options (Noah-MP): 1. Model description and
28 evaluation with local-scale measurements, *J. Geophys. Res.*, 116, D12109,
29 doi:10.1029/2010JD015139, 2011.

30 Oltmans, S.J., A Karion, R.C. Schnell, G. Petron, C. Sweeney, S. Wolter, D.
31 Neff, S. A. Montzka, B. R. Miller, D. Helmig, B. J. Johnson, and J. Hueber: A high ozone

1 episode in winter 2013 in the Uinta Basin oil and gas region
2 characterized by aircraft measurements.
3 *Atmos. Chem. Phys. Discuss.*, 14, 20117-20157, 2014

4 Pollack, I. B., Ryerson, T. B., Trainer, M., Neuman, J. A., Roberts, J. M., and Parrish, D. D.:
5 Trends in ozone, its precursors, and related secondary oxidation products in Los Angeles,
6 California: a synthesis of measurements from 1960 to 2010, *J. Geophys. Res.-Atmos.*, 118, 1–
7 19, doi:10.1002/jgrd.50472, 2013.

8 Reeves, H. D. and Stensrud, D. J.: Synoptic-scale flow and valley cold pool evolution in the
9 western United States, *Weather Forecast.*, 24, 1625–1643,
10 doi:10.1175/2009WAF2222234.1, 2009.

11 Reeves, H. D., Elmore, K. L., Manikin, G. S., and Stensrud, D. J.: Assessment of forecasts
12 during persistent valley cold pools in the Bonneville Basin by the North American Mesoscale
13 Model, *Weather Forecast.*, 26, 447–467, doi:10.1175/WAF-D-10-05014.1, 2011.

14 Roberts, J. M., Veres, P. R., Yuan, B., Warneke, C., Geiger, F., Edwards, P. M., Wild, R.,
15 Dube, W., Petron, G., Kofler, J., Zahn, A., Brown, S. S., Graus, M., Gilman, J., Lerner, B.,
16 Peischl, J., de Gouw, J. A., Li, R., Bates, T., Quinn, P., Koss, A., Li, S.-M., Parrish, D. D.,
17 Senff, C. J., Langford, A. O., Banta, R., Martin, R., Zamora, R., Murphy, S., Soltis, J., and
18 Field, R.: Analysis of aircraft observations, in: Final report: 2013 Uinta Basin winter ozone
19 study, edited by: Stoeckenius, T. and McNally, D., ENVIRON International Corporation,
20 Novato, California, Chapter 5, 1–96, available at: [http://www.deq.utah.gov/locations/
21 uintahbasin/studies/UBOS-2013.htm](http://www.deq.utah.gov/locations/uintahbasin/studies/UBOS-2013.htm), last access 13 June 2014.

22 Salmond, J. A.: Wavelet analysis of intermittent turbulence in the very stable nocturnal
23 boundary layer: implications for the vertical mixing of ozone, *Bound.-Lay. Meteorol.*, 114,
24 463–488, doi:10.1007/s10546-004-2422-3, 2005.

25 Schmitt, C. G., Stuefer, M., Heymsfield, A. J., and Kim, C. K.: The microphysical properties
26 of ice fog measured in urban environments of interior Alaska, *J. Geophys. Res. Atmos.*, 118,
27 11136–11147, doi:10.1002/jgrd.50822, 2013.

28 Schnell, R. C., Oltmans, S. J., Neely, R. R., Endres, M. S., Molenaar, J. V., and White, A. B.:
29 Rapid photochemical production of ozone at high concentrations in a rural site during winter,
30 *Nat. Geosci.*, 2, 120–122, doi:10.1038/NGEO415, 2009.

1 Schnell, R., Johnson, B., Cullis, P., Sterling, C., Hall, E., Albee, R., Jordan, A., Wendell, J.,
2 Oltmans, S., Petron, G., and Sweeney, C.: Tethered ozonesonde and surface ozone
3 measurements in the Uintah Basin, winter 2013, in: Final report: 2013 Uinta Basin winter
4 ozone study, edited by: Stoeckenius, T. and McNally, D., ENVIRON International
5 Corporation, Novato, California, chapter 8, 1–48, 2014.

6 Shin, H. H. and Hong, S.-Y.: Intercomparison of planetary boundary-layer parameterizations
7 in the WRF model for a single day from CASES-99. *Bound.-Lay. Meteorol.*, 139, 261–281,
8 doi:10.1007/s10546-010-9583-z, 2011.

9 Shupe, M. D. and Intrieri, J. M.: Cloud radiative forcing of the arctic surface: the influence of
10 cloud properties, surface albedo, and solar zenith angle, *J. Climate*, 17, 616–628,
11 doi:10.1175/1520-0442(2004)017<0616:CRFOTA>2.0.CO;2, 2004.

12 Silcox, G. D., Kelly, K. E., Crosman, E. T., Whiteman, C. D., and Allen, B. L.: Wintertime
13 PM_{2.5} concentrations in Utah’s Salt Lake Valley during persistent multi-day cold-air pools,
14 *Atmos. Environ.*, 46, 17–24, doi:10.1016/j.atmosenv.2011.10.041, 2012.

15 Stoeckenius, T. and McNally, D. (Eds.): Final report: 2013 Uinta Basin winter ozone study,
16 ENVIRON International Corporation, Novato, California, 367 pp., available at: [http://www.
17 deq.utah.gov/locations/uintahbasin/studies/UBOS-2013.htm](http://www.deq.utah.gov/locations/uintahbasin/studies/UBOS-2013.htm), last access: 13 June 2014.

18 Sheridan, P. F., Vosper, S. B., and Brown, A. R.: Characteristics of cold pools observed in
19 narrow valleys and dependence on external conditions, *Q. J. Roy. Meteor. Soc.*, 140, 715–
20 728, doi:10.1002/qj.2159, 2014.

21 Thompson, G. and Eidhammer, T.: A study of aerosol impacts on clouds and precipitation
22 development in a large winter cyclone, *J. Atmos. Sci.*, accepted, 2014.

23 Thompson, G., Field, P. R., Rasmussen, R. M., and Hall, W. D.: Explicit forecasts of winter
24 precipitation using an improved bulk microphysics scheme, Part II: Implementation of a new
25 snow parameterization, *Mon. Weather Rev.*, 136, 5095–5115, doi:10.1175/2008MWR2387.1,
26 2008.

27 Warneke, C., Geiger, F., Edwards, P. M., Dube, W., Pétron, G., Kofler, J., Zahn, A., Brown,
28 S. S., Graus, M., Gilman, J., Lerner, B., Peischl, J., Ryerson, T. B., de Gouw, J. A., and
29 Roberts, J. M.: Volatile organic compound emissions from the oil and natural gas industry in
30 the Uinta Basin, Utah: point sources compared to ambient air composition, *Atmos. Chem.
31 Phys. Discuss.*, 14, 11895–11927, doi:10.5194/acpd-14-11895-2014, 2014.

1 Warneke, C., Veres, P. R., Murphy, S. M., Soltis, J., Field, R. A., Graus, M. G., Koss, A., Li,
2 S.-M., Li, R., Yuan, B., Roberts, J. M., and de Gouw, J. A.: PTR-QMS vs. PTR-TOF
3 comparison in a region with oil and natural gas extraction industry in the Uintah Basin in
4 2013, *Atmos. Meas. Tech. Discuss.*, 7, 6565-6593, doi:10.5194/amtd-7-6565-2014, 2014.

5 Whiteman, C. D., Zhong, S., Shaw, W. J., Hubbe, J. M., Bian, X., and Mittelstadt, J.:
6 Cold pools in the Columbia Basin, *Weather Forecast.*, 16, 432–447, doi:10.1175/1520-
7 0434(2001)016<0432:CPITCB>2.0.CO;2, 2001.

8 Zängl, G.: Formation of extreme cold-air pools in elevated sinkholes: an idealized numerical
9 process study, *Mon. Weather Rev.*, 133, 925–941, doi:10.1175/MWR2895.1, 2005a.

10 Zängl, G.: Wintertime cold-air pools in the Bavarian Danube Valley Basin: data analysis and
11 idealized numerical simulations, *J. Appl. Meteorol.*, 44, 1950–1971, doi:10.1175/JAM2321.1,
12 2005b.

13 Zhang, H. L., Z. X. Pu, and X. B. Zhang (2013), Examination of Errors in Near-Surface
14 Temperature and Wind from WRF Numerical Simulations in Regions of Complex Terrain,
15 *Weather Forecast*, 28(3), 893-914, doi: 10.1175/Waf-D-12-00109.1.

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1 Table 1. Summary of WRF setup and parameterizations.

Parameter	Chosen Setup	Reference
Initial/Boundary Conditions	NAM Analysis	
Vertical Levels	41	
Domains	3 one-way nests	
Resolution	12 km, 4 km, 1.33 km	
Time Step	45 s, 15 s, 5 s	
Microphysics	Thompson	Thompson et al. (2008)
Shortwave Radiation	RRTMG	Iacono et al. (2008)
Longwave Radiation	RRTMG	Iacono et al. (2008)
Boundary Layer	Mellor-Yamada-Janjic (MYJ)	Janjic (1994)
Surface Layer	Eta Similarity	
Land Surface	Noah	Chen and Dudhia (2001)
Cumulus	Kain-Fritsch (12 km domain only)	Kain (2004)
Diffusion	2nd order on coordinate surfaces	

1 Table 2. Overview of WRF simulations.

Simulation Name	Prescribed Snow Cover	Cloud Ice Sedimentation	Cloud Ice Auto-conversion to Snow
BASE	Full Snow in basin	ON	ON
FULL	Full Snow in basin	OFF	OFF
NONE	No Snow below 2000 m in basin	OFF	OFF

2

1 Table 3. 2-m temperature errors from WRF simulations. Mean errors calculated from the six
2 surface stations in Fig. 1.5b during the 1-6 February 2013 period.

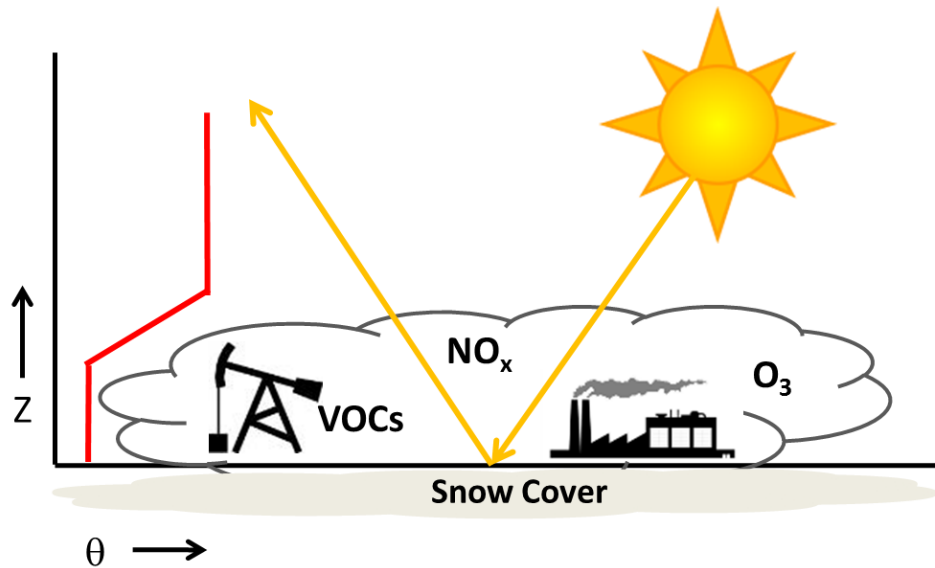
Simulation	Bias (C)	Mean Abs Error (C)	RMSE (C)
BASE	1.65	3.25	3.97
FULL	0.11	2.44	2.98
NONE	7.71	7.74	8.29

3

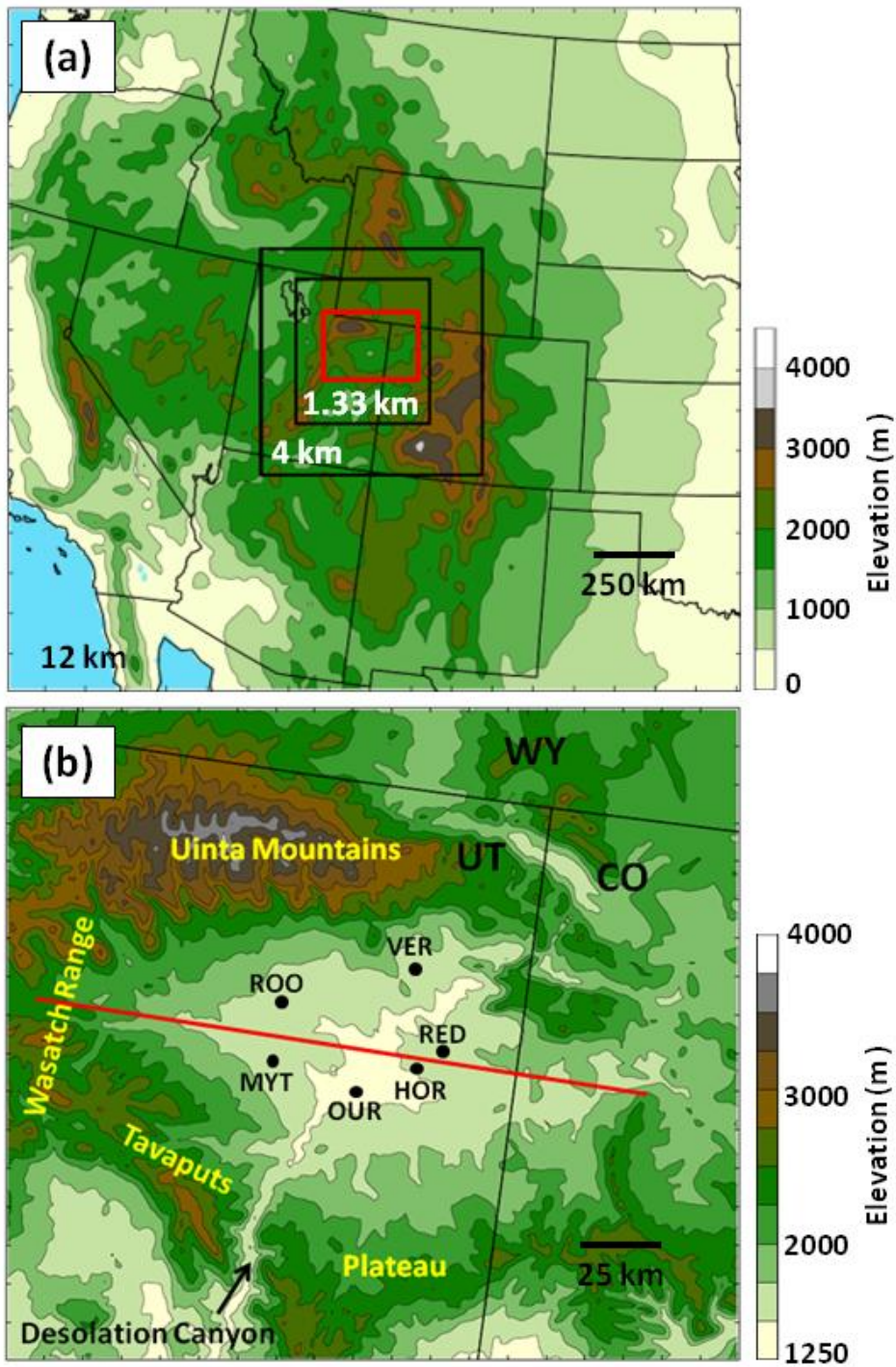
1 Table 4. Ozone concentration statistics from CMAQ model forced by FULL and NONE
2 simulations during the 1-6 February 2013 period.

	FULL	NONE
Highest mean O ₃ - Afternoon (ppb)	97.2	81.2
Highest mean O ₃ - Non afternoon (ppb)	61.9	51.0
Maximum Hourly O ₃ (ppb)	134.4	118.0
Area of mean afternoon O ₃ > 75 ppb (km ²)	896	144

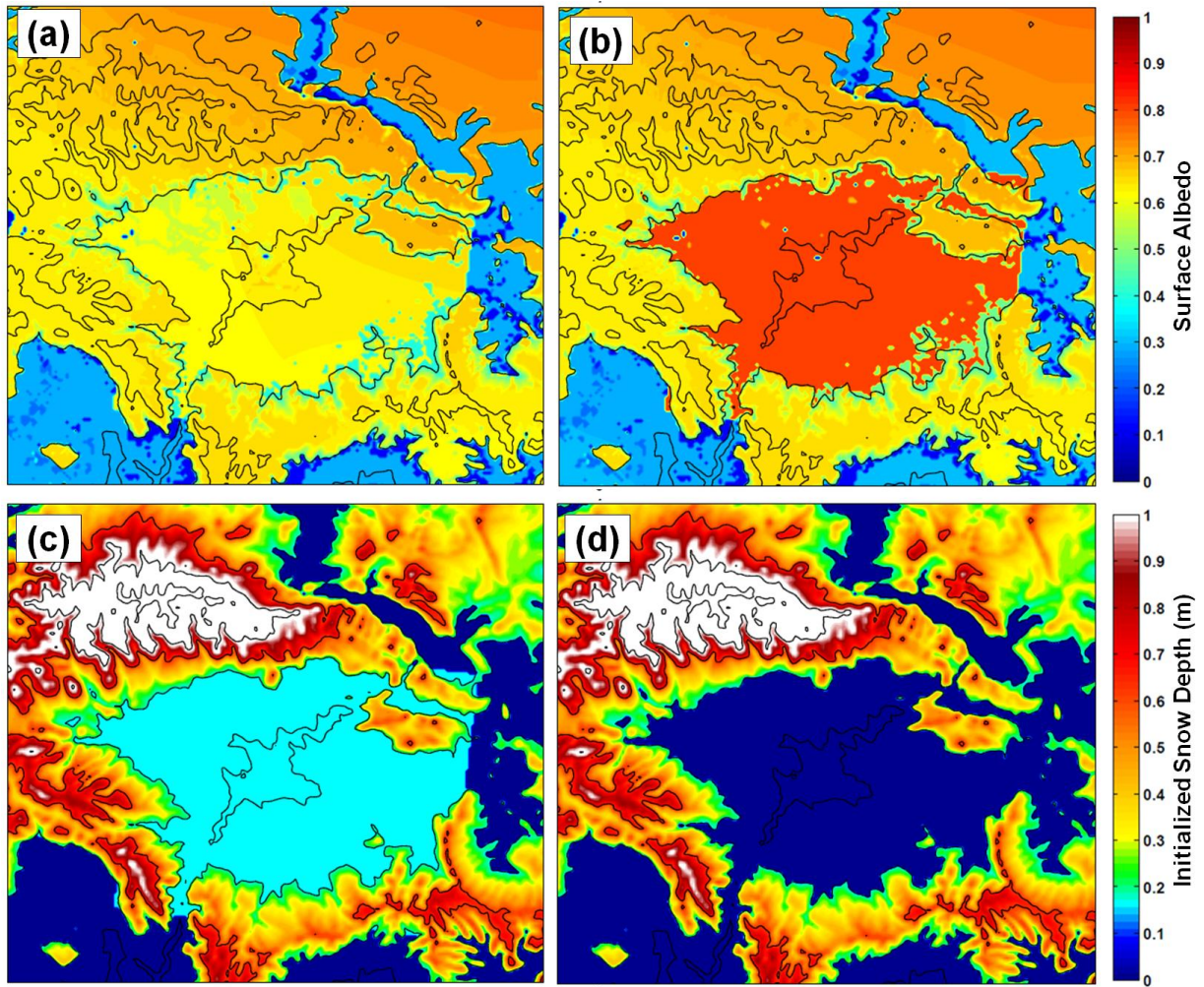
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- 1
- 2 Figure 1. Schematic of factors contributing to high ozone concentrations. Potential
- 3 temperature profile (red line) with stable layer trapping ozone precursors (NO_x and VOCs)
- 4 within the cold-air pool. Snow cover reflects solar radiation, increases photolysis rates, and
- 5 leads to enhanced ozone (O₃) concentrations near the surface. Ice fogs are common in the
- 6 cold-air pool.
- 7



1
 2 Figure 2. (a) WRF 12-, 4-, and 1.33-km domains with terrain contoured every 500 m. (b)
 3 Uintah Basin subdomain with terrain contoured every 250 m and major geographic features
 4 labelled. Black dots indicate locations of surface stations used for verification: Horsepool
 5 (HOR), Myton (MYT), Ouray (OUR), Red Wash (RED), Roosevelt (ROO), and
 6 Vernal(VER). Red box in (a) indicates close-up region shown in (b) and the red line in (b)
 7 indicates position of vertical cross sections shown later.

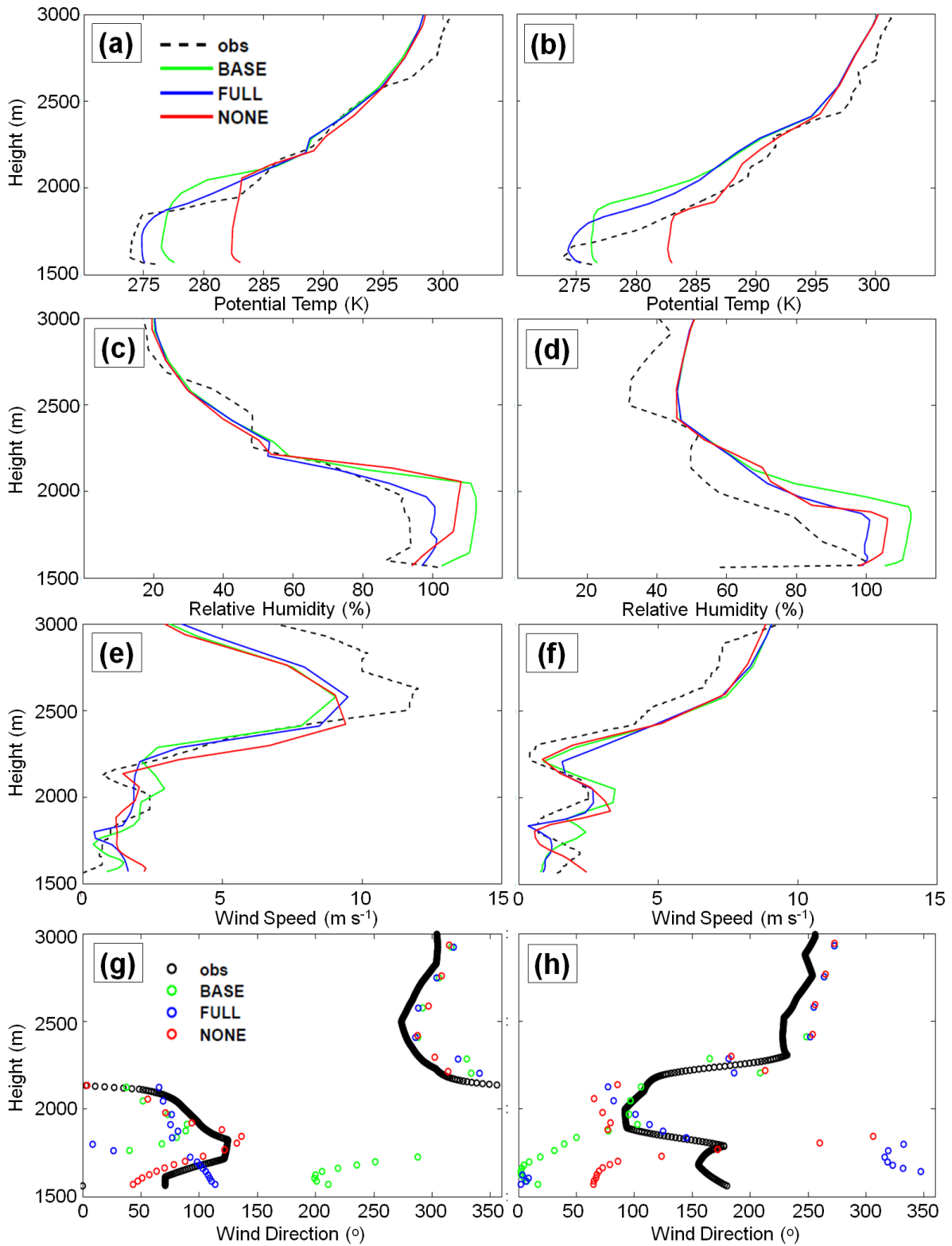


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2 Figure 3. WRF surface albedo (top) at 18:00MST 31 January 2013 for (a) before and (b) after
 3 modifications to WRF snow albedo and vegetation parameter table. Initialized snow depth
 4 (bottom, in m) at 17:00MST 31 January 2013 for (c) “Full Snow” cases (BASE/FULL) and
 5 (d) “No Snow” case (NONE). Terrain contoured every 500 m in black.

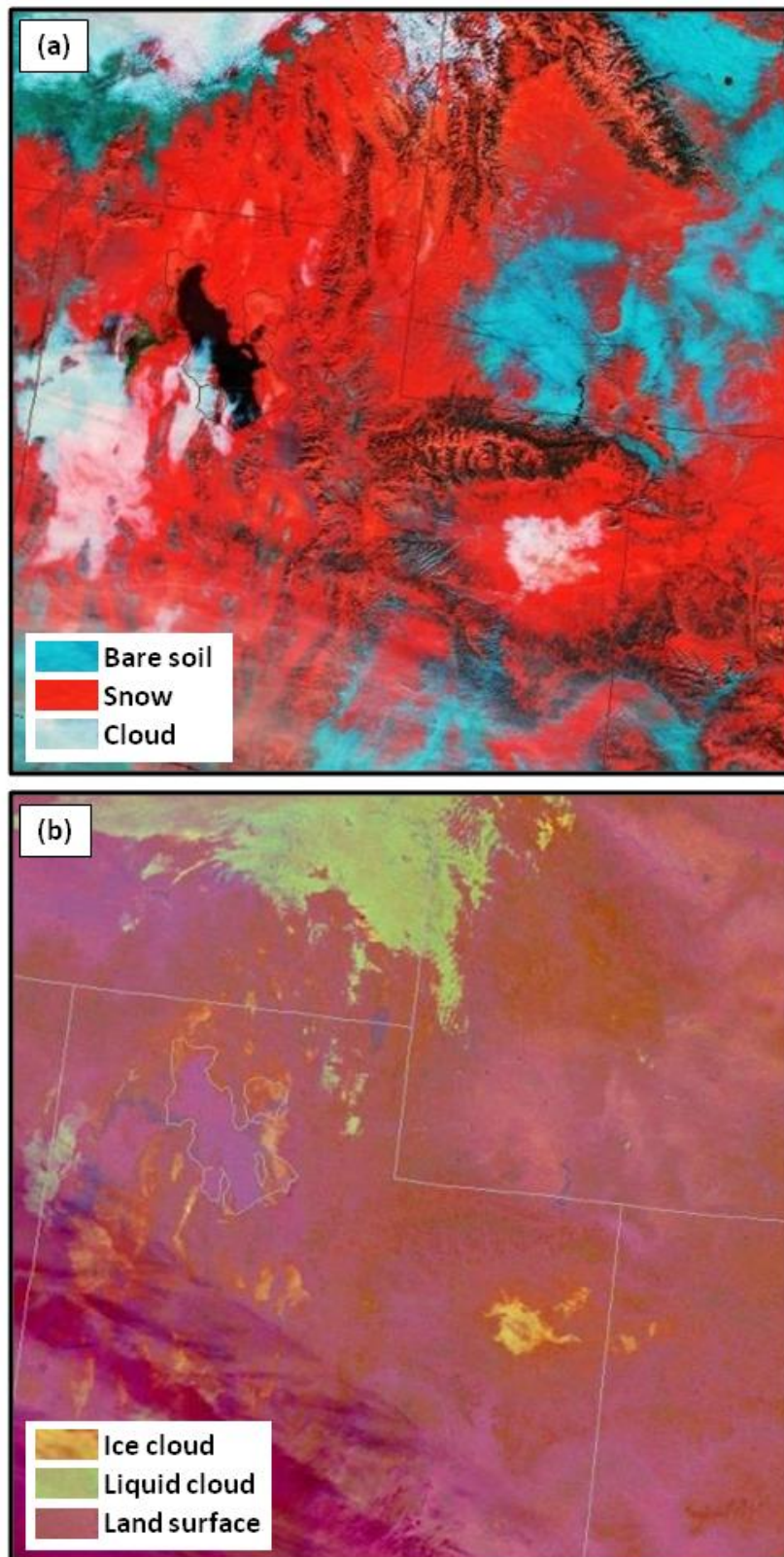
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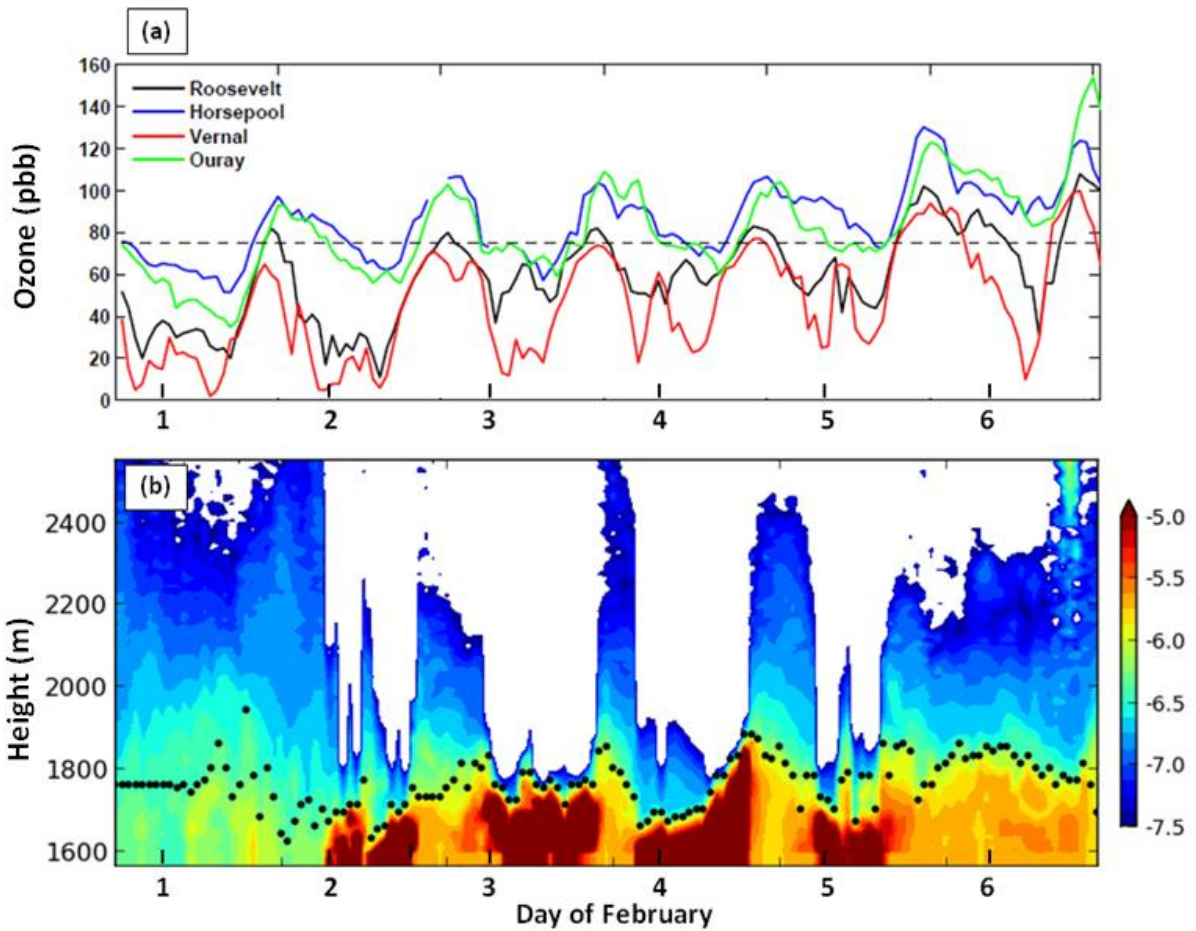
2 Figure 4. Observed and simulated vertical profiles at Roosevelt of (a, b) potential temperature,
 3 (c, d) relative humidity with respect to ice, (e, f) wind speed, and (g, h) wind direction for
 4 11:00 MST 4 February 2013 (left) and 11:00 MST 5 February 2013 (right).



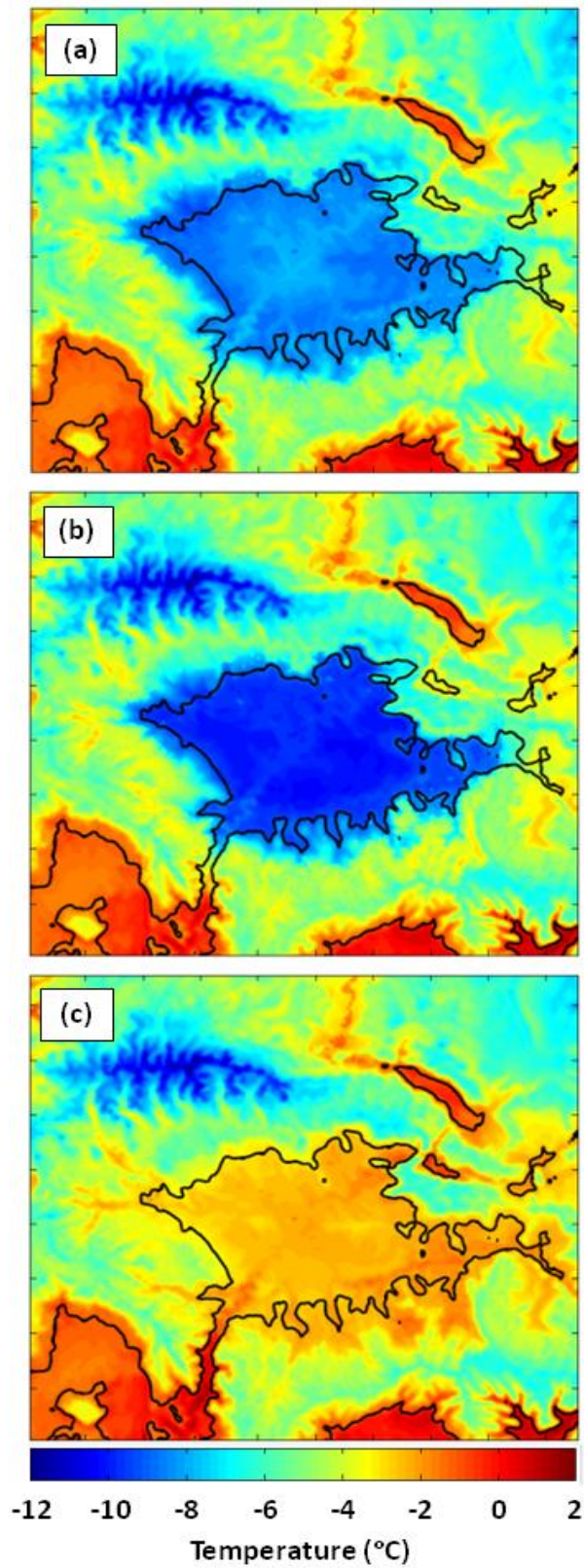
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2 Figure 5. SPoRT-derived VIIRS satellite images: (a) Snow-Cloud product at 1115 MST 2

3 February 2013 and (b) Nighttime Microphysics RGB product at 0331 MST 2 February 2013.

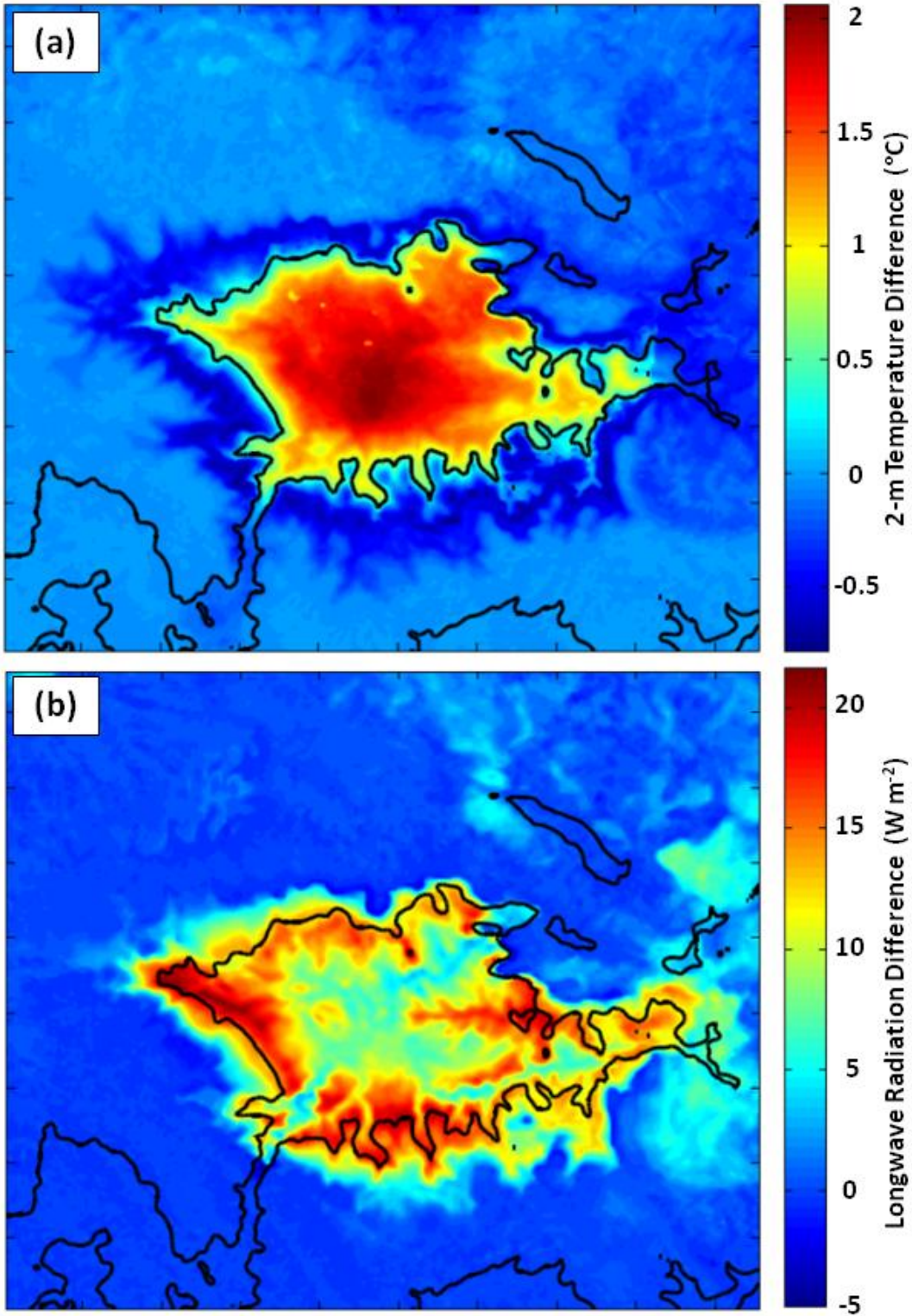


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 2 Figure 6. (a) Hourly ozone concentrations from 31 January-6 February 2013 (thick black tick
 3 marks denote 00:00 MST) for Roosevelt (black), Horsepool (blue), Vernal (red), and Ouray
 4 (green) with the 75 ppb (8-hour mean) NAAQS denoted by the dashed line. (b) Ceilometer
 5 backscatter (shaded) and estimated aerosol depth (black dots) as a function of height (m) at
 6 Roosevelt from 31 January-6 February 2013. Red, yellow, blue, and white shading denote
 7 fog and stratus clouds, high aerosol concentrations; low aerosol concentrations, and beam
 8 attenuation, respectively.

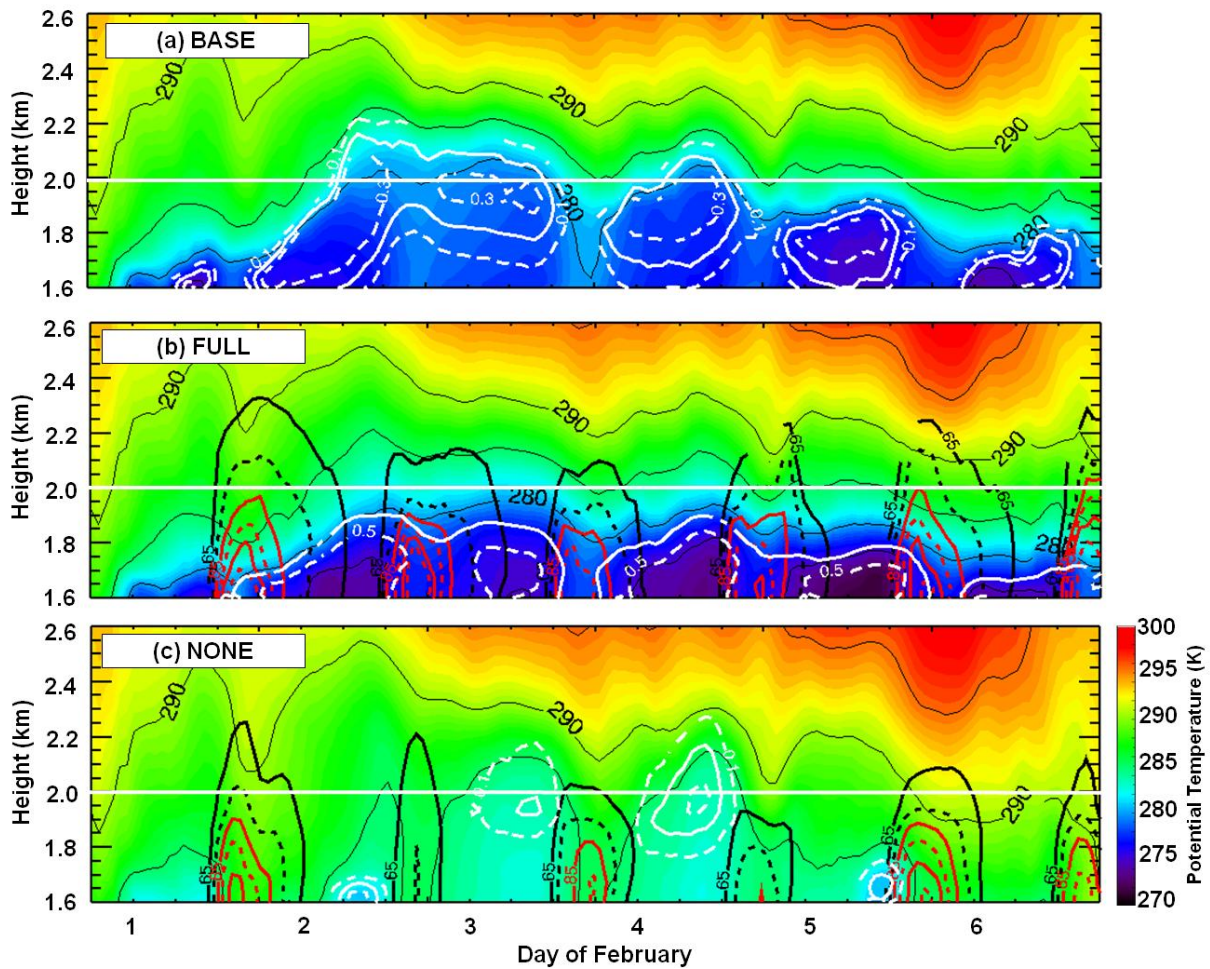


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2 Figure 7. Average 2-m temperature (in °C according to the scale below) between 17:00 MST
 3 31 January and 17:00 MST 6 February 2013 from (a) BASE, (b) FULL, and (c) NONE
 4 simulations.



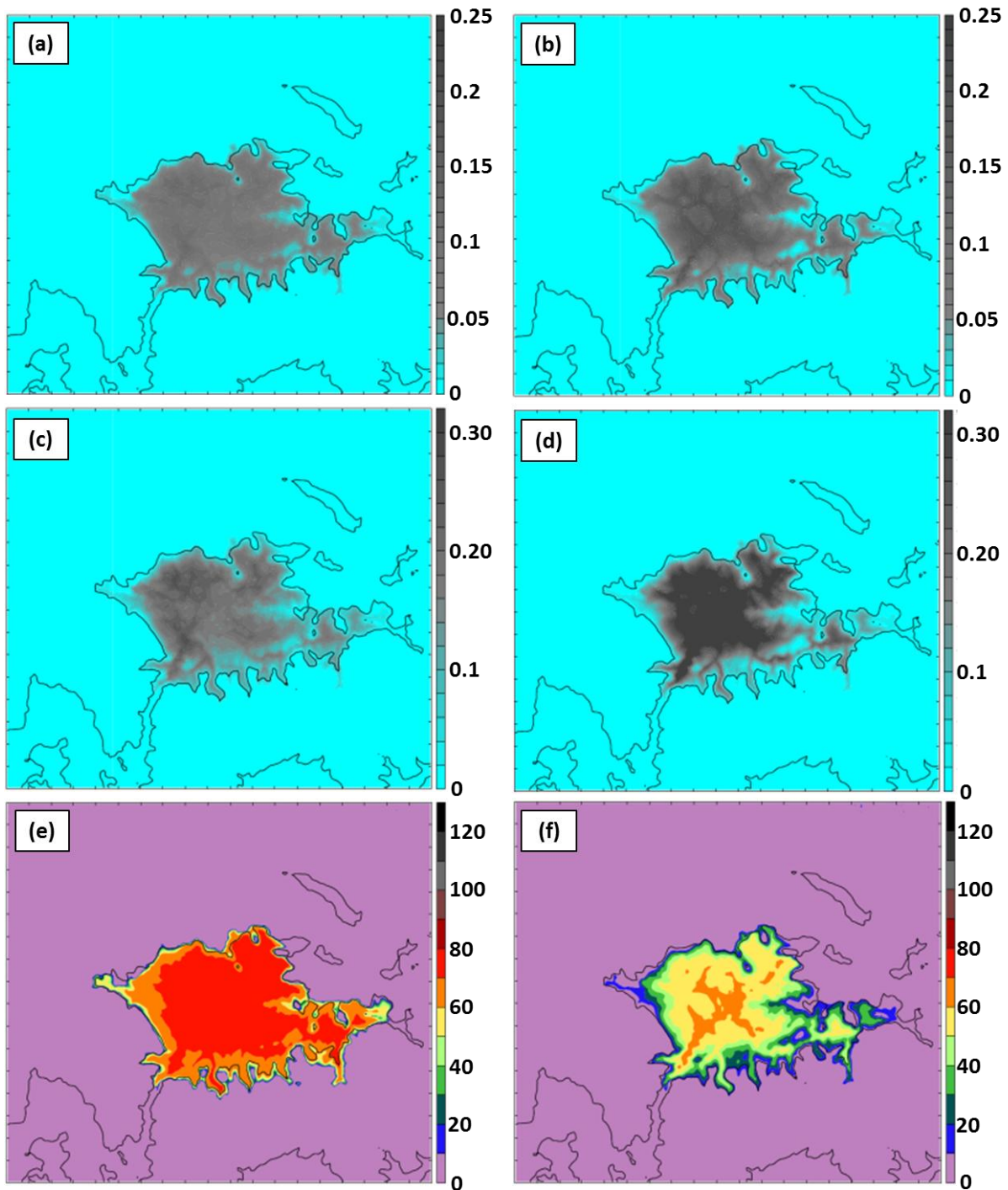
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2 Figure 8. Average difference (BASE – FULL) for period 17:00 MST 31 January to 17:00
3 MST 6 February 2013 in: (a) 2-m temperature (in $^{\circ}\text{C}$ according to the scale to the right) and
4 (b) downwelling longwave radiation (in W m^{-2} according to the scale on the right).



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 2 Figure 9. Time-height plot of potential temperature shaded in K according to the scale on the
 3 right at Horsepool from 17:00 MST 31 January to 17:00 MST 6 February 2013 (dates are
 4 labelled at 00:00 MST) from (a) BASE, (b) FULL, and (c) NONE simulations. Cloud water
 5 amounts in (a) and (c) of 0.1, 0.2, and 0.3 g kg^{-1} are denoted by dashed white, solid white,
 6 and dashed white lines, respectively. Cloud ice amounts in (b) of 0.3 and 0.5 g kg^{-1} are
 7 denoted by solid white and dashed white lines, respectively. Ozone concentrations are
 8 contoured every 10 ppb, starting at 65 ppb and alternate between solid and dashed every 10
 9 ppb. Ozone concentrations above (below) 85 ppb are contoured in red (black). Plotted ozone
 10 concentrations represent the maximum value for each hour in a 40 by 40 km region
 11 encompassing Ouray and Horsepool.

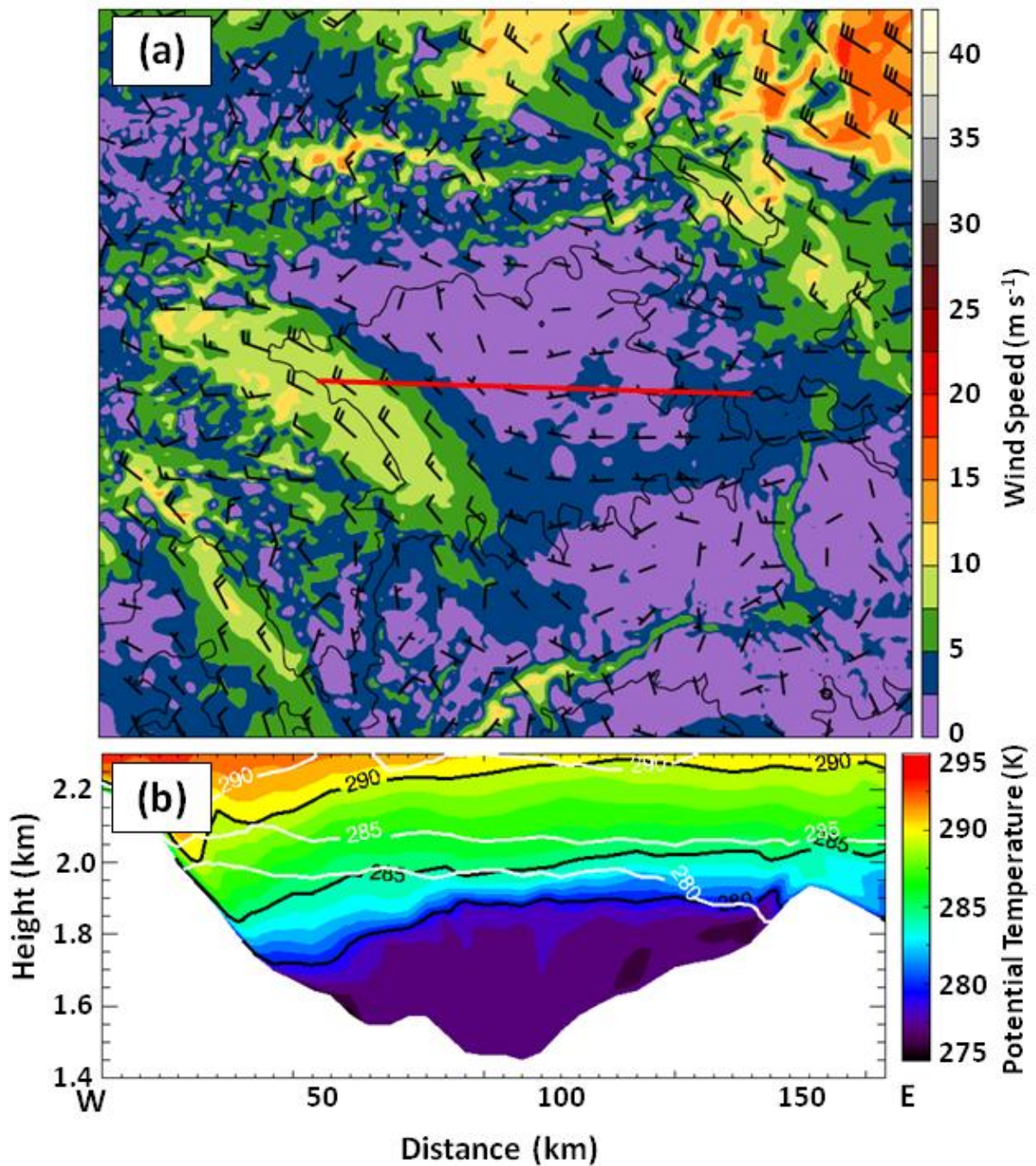
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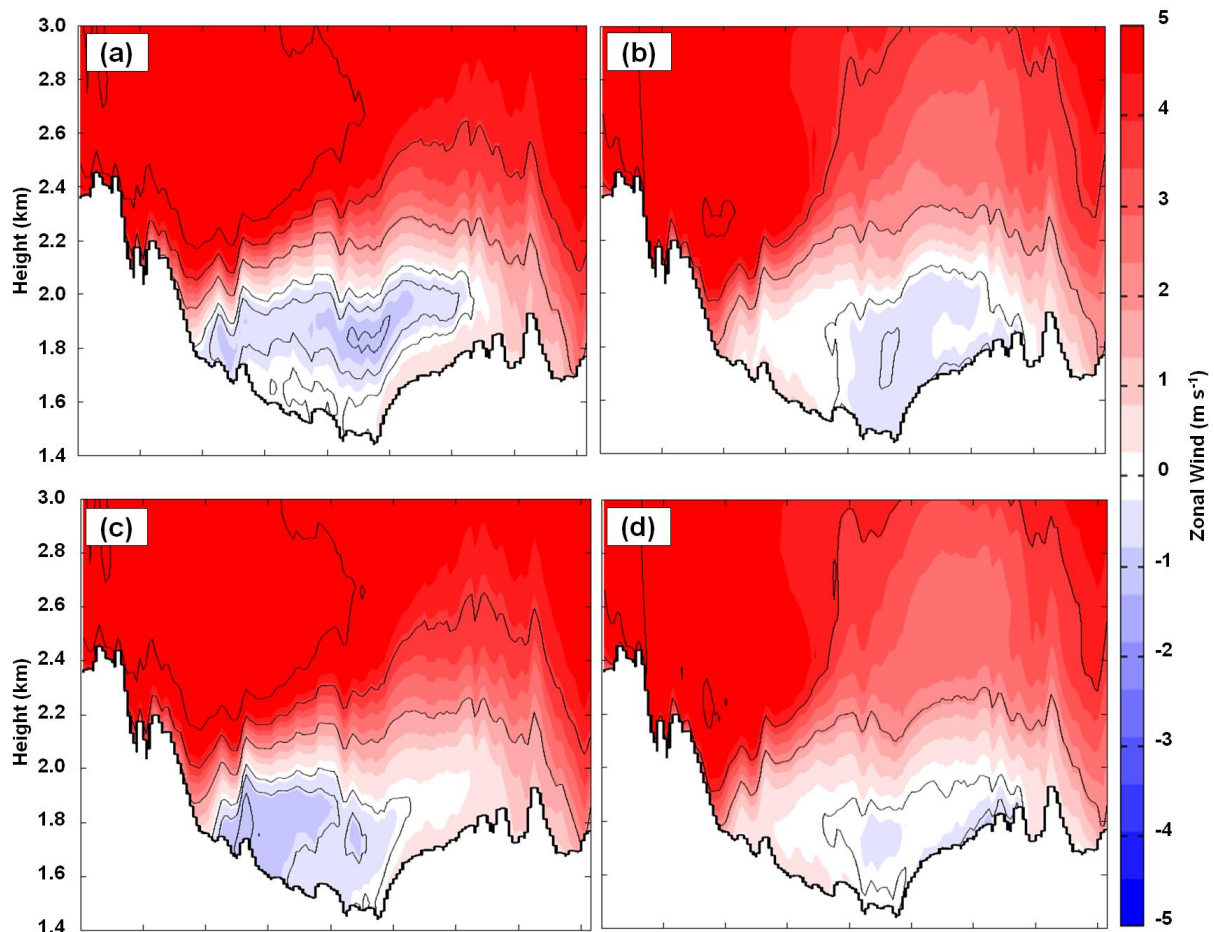


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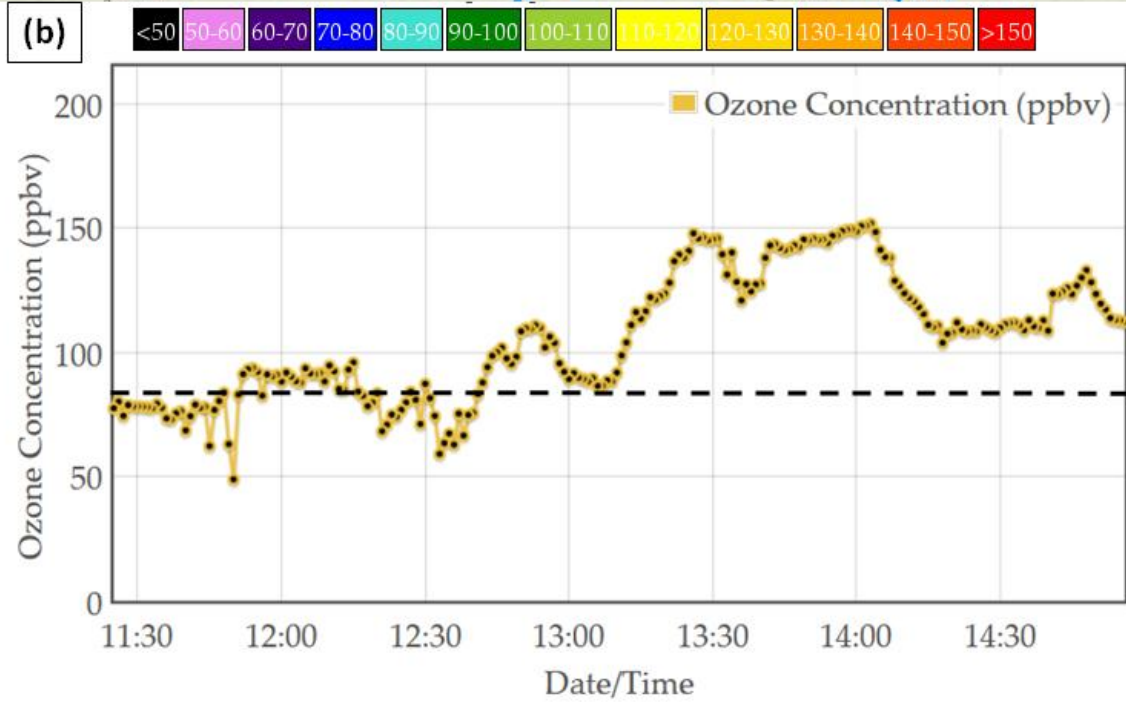
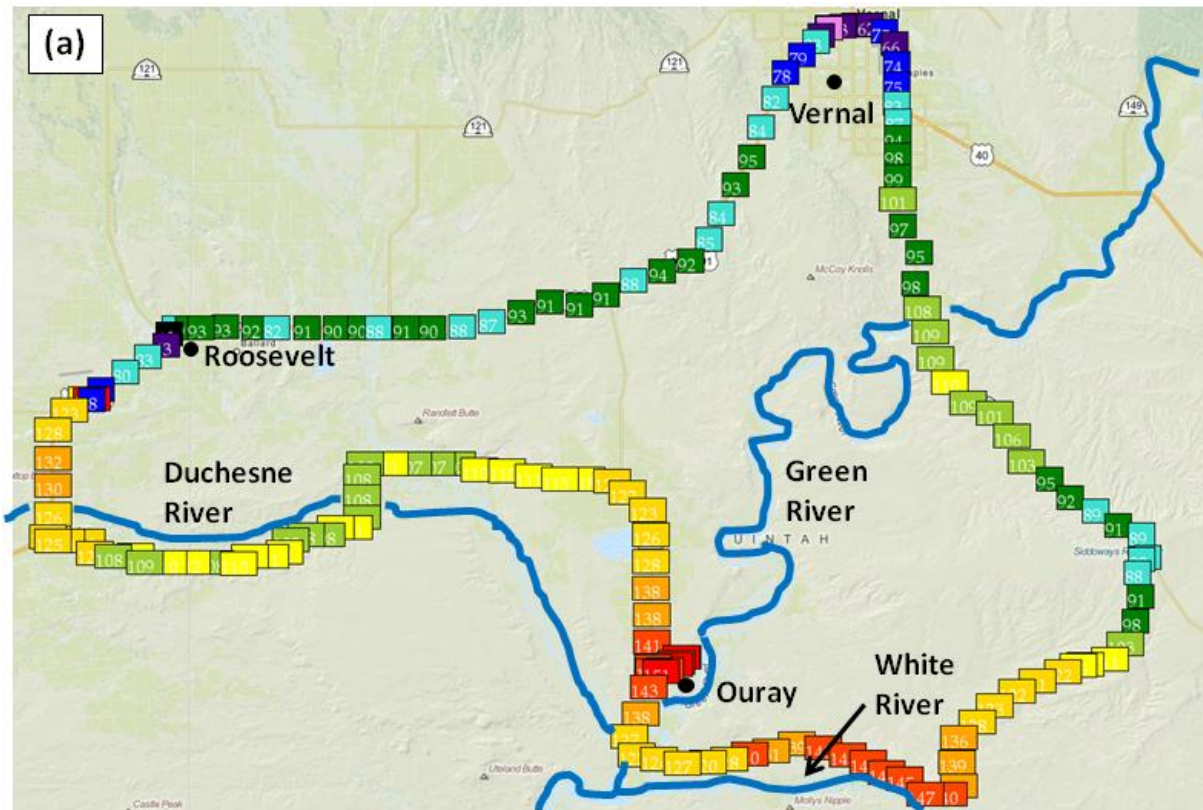
2 Figure 10. Cloud characteristics from BASE (a, c, e) and FULL (b, d, f) simulations at 23:00
 3 MST 4 February 2013. (a, b) Integrated cloud amount (in mm according to the scale on the
 4 right), (c) mean cloud water in bottom 15 model levels (in g kg⁻¹ according to the scale on the
 5 right), (d) mean cloud ice in bottom 15 model levels (in g kg⁻¹ according to the scale on the
 6 right), (e, f) net downwelling longwave radiation from clouds (in W m⁻² according to the scale
 7 on the right).



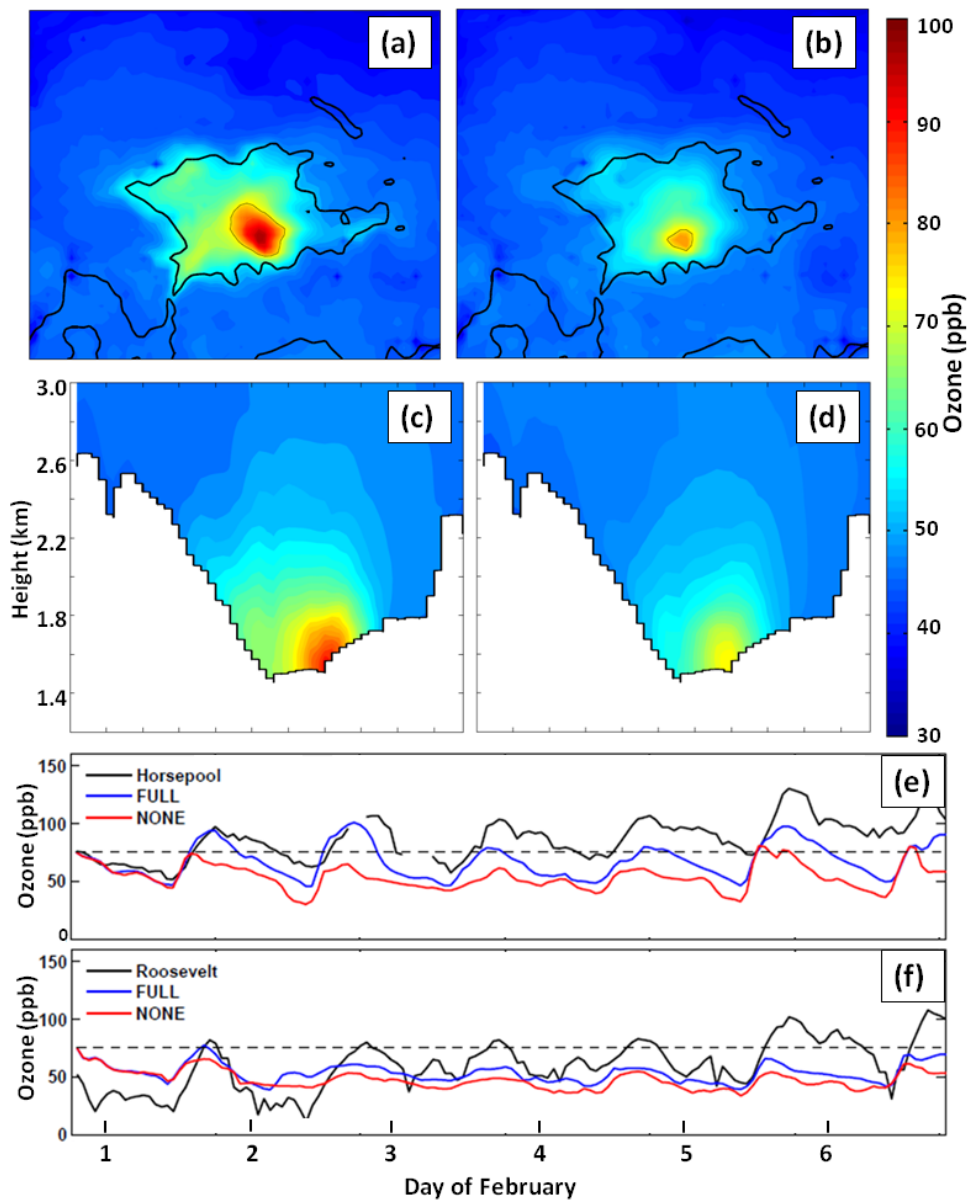
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 2 Figure 11. FULL simulation at 23:00 MST 3 February 2013 for (a) 2.3 km MSL wind speed
 3 (in m s^{-1} according to the scale on the right) and barbs (full barb 5 m s^{-1}). (b) Vertical cross
 4 section of potential temperature at 23:00 MST 3 February shaded in K according to the scale
 5 on the right and at 05:00 MST 4 February (in K according to labelled white contours) along
 6 red line in (a).



1
2 Figure 12. Average zonal wind in the vicinity of the cross-section in Fig. 2b for 17:00 MST
3 31 January to 17:00 MST 6 February 2013 period. The FULL simulation (top) and NONE
4 simulation (bottom) results for (a, c) daytime hours (0800 to 17:00 MST) and (b, d) nighttime
5 hours (1800 to 07:00 MST). Westerly (easterly) winds shaded in m s^{-1} according to the scale
6 on the right in red (blue) with westerly (easterly) winds contoured every 2 m s^{-1} (-0.5 , -1 , and
7 -2 m s^{-1} only). Values are averaged over a 26-km wide swath perpendicular to the cross
8 section.
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 2 Figure 13. University of Utah Mobile transect of ozone concentration from 1130 to 15:00
 3 MST 6 February 2013 as a function of (a) geographic location and (b) time. Dashed black
 4 line represents NAAQS for ozone (75 ppb). Ozone concentrations in (a) are according to
 5 color scale.



1
 2 Figure 14. (top) Average ozone concentration (in ppb according to scale on the right) during
 3 1100-17:00 MST 1–6 February 2013 on the lowest CMAQ model level (~17.5 m) from (a)
 4 FULL and (b) NONE simulations. The thin black line outlines regions where the ozone
 5 concentration exceeds 75 ppb while the reference terrain elevation of 1800 m is shown by the
 6 heavy black line. (bottom) Average ozone concentration during 1100-17:00 MST 1–6
 7 February 2013 from (c) FULL and (d) NONE simulations along cross section approximately
 8 25 km south of the red line in Fig. 2b. Values averaged over 24-km wide swath perpendicular
 9 to the cross section. Time Series of ozone concentrations from (e) Horsepool, and (f)
 10 Roosevelt. Observations, CMAQ output from FULL and NONE simulations in blue, red, and
 11 black respectively (thick black tick marks denote 00:00 MST). The NAAQS of 75 ppb is
 12 denoted by the thin black dashed line.