Modeled Downward Transport of a Passive Tracer over Western North America during an Asian Dust Event in April 1998

JOSHUA P. HACKER, IAN G. MCKENDRY, AND ROLAND B. STULL

Atmospheric Science Program, Department of Geography and Department of Earth and Ocean Sciences, The University of British Columbia, Vancouver, British Columbia, Canada

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ABSTRACT

An intense Gobi Desert dust storm in April 1998 loaded the midtroposphere with dust that was transported across the Pacific to western North America. The Mesoscale Compressible Community (MC2) model was used to investigate mechanisms causing downward transport of the midtropospheric dust and to explain the high concentrations of particulate matter of less than 10-μm diameter measured in the coastal urban areas of Washington and southern British Columbia. The MC2 was initialized with a thin, horizontally homogeneous layer of passive tracer centered at 650 hPa for a simulation from 0000 UTC 26 April to 0000 UTC 30 April 1998. Model results were in qualitative agreement with observed spatial and temporal patterns of particulate matter, indicating that it captured the important meteorological processes responsible for the horizontal and vertical transport over the last few days of the dust event. A second simulation was performed without topography to isolate the effects of topography on downward transport.

Results show that the dust was advected well east of the North American coast in southwesterly midtropospheric flow, with negligible dust concentration reaching the surface initially. Vertically propagating mountain waves formed during this stage, and differences between downward and upward velocities in these waves could account for a rapid descent of dust to terrain height, where the dust was entrained into the turbulent planetary boundary layer. A deepening outflow (easterly) layer near the surface transported the tracer westward and created a zonal-shear layer that further controlled the tracer advection. Later, the shear layer lifted, leading to a downward hydraulic acceleration along the western slopes, as waves generated in the easterly flow amplified below the shear layer that was just above mountain-crest height. Examination of 10 yr of National Centers for Environmental Prediction–National Center for Atmospheric Research reanalyses suggests that such events are rare.

1. Introduction

Air pollution in western North America has generally been considered to be of regional scope and local origin. However, recent observations show that 1) anthropogenic pollutants from rapidly industrializing east Asia reach western North America (Jaffe et al. 1999; Berntsen et al. 1999) and 2) mineral dust from semiarid regions of Asia, long known to influence the north Pacific including Hawaii (Duce et al. 1980; Merrill et al. 1989), can also reach North America (Husar et al. 2001, hereinafter HUS; McKendry et al. 2001, hereinafter MCK).

The potential impact of long-range transport of Asian pollutants to North America was highlighted during late April 1998, when mineral dust from a severe dust storm in the Gobi Desert of western China was rapidly transported across the north Pacific to create high surface concentrations of particulate matter of less than 10-μm diameter (PM_{10}) that led to reduced visibility across western North America for several days (HUS). Jaffe et al. (1999) document anthropogenic pollutants from Asia reaching the Pacific Northwest, but this was the first documented case of mineral dust reaching North America from Asia (although there is ice-core evidence of Asian dust reaching Greenland in the past; Biscaye et al. 1997). This spectacular and unusual event raised questions about mechanisms of free-tropospheric–planetary boundary layer (FT–PBL) exchange that would permit mineral aerosol transported in the middle and lower troposphere to be subsequently mixed to the surface in high concentrations (MCK). If mineral aerosol could be entrained into the PBL over western North America, then burgeoning east Asian anthropogenic emissions could suffer the same fate.

Although processes by which pollutants from the PBL find their way into the FT are well known (McKendry and Lundgren 2000), processes by which pollutants are transported downward to where they can be mixed back into the PBL after long-range transport are less well described. In this case study, the Asian dust event of April 1998 provides the motivation to examine mechanisms of FT–PBL exchange over mountainous western
North America. A numerical model with atmospheric tracer capability is utilized to identify important factors contributing to this event. Distinct spatial and temporal patterns of surface PM$_{10}$ concentrations give evidence that the model is capturing the important dynamics associated with the downward transport of the dust. The next section is a summary of the Asian dust event, and section 3 describes the experimental approach. Results are presented in section 4, including a brief comparison of modeled surface tracer distribution with the observed dust and an analysis of the transport mechanisms. A summary is presented in section 5.

2. The 1998 Asian dust event

A detailed examination of the 1998 dust event is provided in HUS, MCK, and Tratt et al. (2001, hereinafter TFW). In summary, during April 1998, several unusually intense dust storms were generated over the Gobi. These events are common during spring and produce a mineral aerosol rich in Fe, Si, Ca, and Al (Braaten and Cahill 1986). An unusually intense storm on 19 April produced a dust cloud that was rapidly transported across the Pacific. The dust cloud was tracked in satellite imagery and observed by LIDAR on both sides of the Pacific (TFW; Murayama et al. 2001, hereinafter MUR). Ground-level concentrations of PM$_{10}$ were observed to peak on 28–29 April from British Columbia to California, with 24-h average concentrations exceeding 100 $\mu$g m$^{-3}$ in parts of Washington on 29 April. The dust chemical composition over North America bore the signature of Asian dust (McKendry et al. 2000), was uniform, and had a volume mean diameter of 2–3 $\mu$m. The combination of an unusual dust storm in Asia and meteorological conditions conducive to the rapid transport across the Pacific is rare, as will be shown.

Observations from sparsely distributed routine monitoring stations provide the only information on the spatial and temporal patterns associated with the arrival of Asian dust in the PBL over western North America. In Fig. 1, the distribution of daily PM$_{10}$ concentrations is shown for 29 April 1998, the peak of the event. Contours in Fig. 1 were generated using an objective analysis scheme with weights determined by inverse squared distance. A zone of high concentrations (50–100 $\mu$g m$^{-3}$) was observed in northern Oregon, central Washington, and southern British Columbia. It is important to recognize that these PM$_{10}$ observations represent a combination of particulate matter of local origin and Asian dust. MCK estimate that up to 50% of PM$_{10}$ in the Lower Fraser Valley of southwestern British Columbia during the event was of Asian origin, while HUS estimate that Asian dust contributed about 40 $\mu$g m$^{-3}$ greater than the local regional PM$_{10}$ average of 10–25 $\mu$g m$^{-3}$ for April–May 1998.

Temporal variations during the event are not presented here because most stations do not monitor PM$_{10}$ on a daily basis (approximately 30 of 230 western U.S. stations sample on a daily basis; the remainder sample every sixth day). However, in southern British Columbia, MCK note that observed concentrations peaked in the southern central interior of the province first (28 April) and later shifted westward. On Vancouver Island, concentrations were highest on 30 April, and in the Lower Fraser Valley, they reached a maximum on 29 April. A similar pattern was observed in Washington (HUS). A second feature of the event was a strong diurnal variation in concentrations characterized by highest surface concentrations during the afternoon and lowest concentrations at night. This feature was observed as far south as California (HUS).

3. Modeling environment

a. MC2 characteristics and initial conditions

Meteorological transport processes associated with the April 1998 event were examined with the Canadian Mesoscale Compressible Community (MC2) atmospheric model (Benoit et al. 1997). In the MC2, advec-
tion is semi-Lagrangian with a semi-implicit time step, facilitating straightforward tracking of atmospheric tracers. The Recherche en Prévision Numérique full physics package is used, which is similar to that used by the Canadian Meteorological Centre (CMC) in their suite of operational models. The force–restore method (Dorff 1978) describes surface exchanges, and a 1.5-order turbulence closure (retaining a turbulent kinetic energy predictive equation, Benoit et al. 1989) parameterizes vertical turbulent diffusion. Surface values are specified at shelter height (2 m) and anemometer height (10 m) using a surface-layer model based on similarity theory. The MC2 uses a one-way (cascade) nesting strategy, in which coarser grid simulations provide initial and boundary conditions for finer grid simulations, and there is no upscale feedback. A $\Delta x = 90 \text{ km}$ grid simulation covering a large domain is initialized from 0000 UTC analyses. With this coarse mesh output, a $\Delta x = 30 \text{ km}$ grid initialized at 0600 UTC, which in turn drives a $\Delta x = 10 \text{ km}$ grid initialized at 1200 UTC for the smallest domain. Model domains are shown in Fig. 2. The 6-h offset between grid initialization is chosen to allow inertial gravity waves to disperse from the coarser grid before subsequent initialization of the finer grids.

To examine meteorological mechanisms by which mineral aerosol was transported to the PBL in the Lower Fraser Valley (Vancouver area), the MC2 was modified to advect and diffuse a passive tracer. A $\Delta x = 90 \text{ km}$ grid simulation was initialized 0000 UTC 26 April 1998, and 30-, and 10-km grids were run beginning 6 and 12 h later, respectively. The 96-h simulation ending at 0000 UTC 30 April used grids from the National Centers for Environmental Prediction (NCEP) Eta Model analyses for initial and boundary conditions. Vertically, the tracer was initially distributed as a 100-hPa deep ($\sim 1000 \text{ m}$) single layer, with a concentration maximum at 650 hPa. Specification of tracer depth was based on 1) serendipitous aircraft observations near Seattle on 27 April showing a distinct dust layer at about 2–3 km altitude and virtually no dust below (HUS) and 2) a short sensitivity experiment showing that a realistic surface-layer tracer distribution is modeled with a tracer initialized at that altitude (section 3b). To ensure that surface concentrations result from dominant downward-transport mechanisms, the layer used in this simulation is higher than the aircraft observed. Lidar backscatter data taken over Pasadena on 27 April (TFW) showed additional layers at higher altitudes. Multiple layers were excluded from the simulation because of the absence of complete tropospheric aerosol profile data over Washington/British Columbia with which to confirm the presence of midtropospheric dust and the sensitivity analysis demonstrating higher layers would contribute less to observed surface concentrations. Last, the simulated dust was distributed homogeneously horizontally and was continuously supplied at the boundaries of the model domain. To isolate the impact of the terrain on downward transport of tracer, simulations with both realistic terrain and flat terrain were conducted.

b. Meteorological and tracer distribution uncertainty

To evaluate the mechanisms responsible for high tracer concentration at ground level, a model simulation must capture the important features of both the meteorology and the surface tracer distribution. In a limited-area domain, the solution is strongly controlled by the boundary conditions (e.g., Warner et al. 1997). Given that the boundary conditions were analyses rather than forecasts, this simulation never deviates substantially from the analysis. This is no doubt aided by the fact that advective speeds were large for this case, rendering internal model physics relatively less important.

The simulation is a tool to understand some phenomena responsible for the observed surface dust distribution, and it is unnecessary to reproduce the event in detail. If modeled tracer resembles surface observations, the model is likely reproducing dynamics similar to what actually occurred. Because of scant dust plume observations, some arbitrariness is unavoidable in choosing the initial three-dimensional tracer distribution. Observations show that the dust plume generally tracked over the Pacific Northwest at upper levels just before and while concentrations peaked at the surface (HUS). Earlier experiments (MCK) used an initial horizontal distribution mimicking satellite imagery that showed the plume over the Pacific to be located in the midtropospheric flow maximum. Simulated surface tracer distributions were reasonable only in localized regions, and a tracer that was not perfectly collocated with the proper meteorological conditions did not reach the surface or was dispersed aloft. Because the goal here is not to perfectly reproduce surface tracer concentrations but rather to examine what is responsible for its arrival.
at the surface, the attempt at simulating a realistic horizontal distribution aloft was abandoned. Initializing with a horizontally homogeneous tracer distribution removes any assumptions about the horizontal distribution aloft and restricts our attention to vertical and low-level horizontal transport that occurs when the tracer is in the PBL. By maintaining the same profile at the boundaries throughout the simulation, the restriction maintains its integrity.

An extensive body of work examining haze layers over southern Africa (Garstang et al. 1996; Tyson et al. 1997; Piketh et al. 1999) links the vertical distribution of atmospheric aerosols to the thermal structure of the lower troposphere. Namely, aerosols concentrate in absolutely stable layers, and extensive vertical mixing generally occurs with deep convection that penetrates and destroys the stable layers. This phenomenon could account for the multiple peaks structure observed and reported in HUS, even though this case is in a synoptically active regime and the southern African work primarily examines persistent anticyclonic conditions. If stable layers develop in the MC2 simulation, the modeled tracer should also concentrate in these layers, and it is unnecessary to initialize with a tracer profile consistent with the thermal structure. The lead time of 2 or 3 days between model initialization and tracer arrival at the surface is sufficient to allow the model physics to vertically distribute the tracer according to the stability.

The initial tracer should be placed high enough so that it will only reach the surface where the modeled vertical transport is large but low enough so that the amount of tracer reaching the surface can be qualitatively compared with observations. Results from four simulations, each initialized with different profiles (Fig. 3), demonstrate how the surface tracer distributions vary with the initial height of the tracer layer (Fig. 4). As may be expected, a higher initialized layer results in lower simulated surface concentrations. The location of

![Tracer Concentration](image)

**Fig. 3.** Initial tracer profiles for the four runs of the sensitivity study.

![Surface tracer concentration](image)

**Fig. 4.** Surface tracer concentration for the four runs corresponding to the initial profiles shown in Fig. 3, valid 1200 UTC 29 Apr 1998. Units are relative.
the maxima over the Vancouver area and south of Puget Sound does not change between runs 2 and 4. The quantities of tracer reaching the surface in run 4 were deemed substantial for evaluation in more detail. Besides satisfying the criteria, Run 4 is also initialized with a layer that is higher than any of the lowest observed dust layers (HUS; TFW). Thus, surface concentrations are conservative in the sense that the height of origin is not as low as the observations allow.

4. Simulation results

a. Synoptic background

Synoptic conditions, as simulated by the MC2, are shown in Fig. 5. A 500-hPa ridge and a surface high pressure center persisted throughout the period. Geopotential heights (500 hPa) over the coast increased as the ridge amplified, and the strongest flow aloft moved northward. Simultaneously, the ridge axis moved slightly westward. During the period of the simulation, a surface anticyclone propagated eastward and strengthened. When surface isobars are broadly parallel to the coast over Washington and southern British Columbia (Fig. 5b), continental air generally flows through gaps in the mountain ranges and spills down western slopes. Such events are known as “outflow” in British Columbia and usually occur during winter months (Jackson 1996). They can cause strong gap-flow events (Jackson and Steyn 1994) and can be responsible for cold-air outbreaks along the coast and significant airborne transport of local crustal material out of the Fraser Valley in southern British Columbia (McKendry 2000). The outflow during this April 1999 event was weak as compared with the much more intense winter events; nonetheless, it was important for dust transport, as will be shown.

b. Surface tracer distribution

Simulated surface-layer tracer concentrations and surface-layer winds at 30-h intervals are shown in Fig. 6. Initial maximum concentrations in the layer at 650 hPa were scaled to have a dimensionless value of 1000. Thus, the surface values in Fig. 6 are relative to this maximum tracer value.

Tracer at ground level was first apparent in the southeastern portion of the simulation domain on the morning of 27 April (Fig. 6a). By midday on 28 April, the highest ground-level concentrations were along the Washington/British Columbia interior border and to the south of Puget Sound in southwestern Washington. These two zones correspond with the highest average elevations in the Cascade Range. By this time, outflow had strengthened considerably at the surface and continued to do so throughout the simulation. By late afternoon on 29 April (0000 UTC 30 April; Figs. 6e,f) surface tracer concentrations had increased in magnitude and spread horizontally across a broad zone including southern Vancouver Island. A second zone of high surface concentrations also formed on the lee (east) side of the Rocky Mountains, bordering British Columbia and Alberta. For tracer mixed downward in the vicinity of the Cascades, simulated outflow winds were responsible for advection of the near-surface tracer westward and ultimately offshore.

Sudden arrival of the tracer at the surface in a zone extending from immediately north of the Washington/British Columbia border southward through western central Washington and near-surface westward transport of the tracer are roughly consistent with observations described by MCK and shown in Fig. 1. These results, combined with the fact that the simulated flow largely reproduces the analyses, suggest that the simulation captures the salient meteorological mechanisms responsible for bringing the dust downward and advecting it toward the coastal urban areas.

Comparing the modeled surface tracer concentration with the observations simply provides confidence that the simulation is capturing the gross features evident in the observations and therefore some of the important dynamics of the event. The modeled tracer concentra-
tions are instantaneous values, and the observations in Fig. 1 are a 24-h average, making a direct, quantitative comparison impossible. It is true that the simulated maximum concentration, just west of southern Vancouver Island, appears to be farther west than the observed peak in Fig. 1. This may result from the tracer being advected too quickly or a lack of any tracer deposition in the MC2. A maximum also could have actually occurred offshore, but it was not observed. Even in the worst case that the tracer advected too quickly, the analysis of the vertical and horizontal transport of the tracer is still valid.

c. Vertical transport processes

In Fig. 7, a west–east transect through the Lower Fraser Valley illustrates processes contributing to the downward transport of the tracer. In this cross section, the Pacific Ocean is west of 126°W, Vancouver Island is at 125°W, and Vancouver is at approximately 123.25°W longitude. Georgia Strait lies in the flat area between 123° and 124°W.

Flow aloft during the dust event was broadly perpendicular to the ridge lines of the Cascades, the British Columbia Coast Mountain Range, and the Rocky Mountains. At Δx = 10 km, the MC2 simulated strong mountain wave activity aloft (Figs. 7b,d,f) and a shear layer near crest height between the developing easterly outflow and predominately southwesterly flow aloft (Figs. 7a,c,e). Aloft, the slow descent of an eastward-moving tracer (shaded in left panels) offshore throughout the simulation is indicative of large-scale subsidence. Southwesterly winds remained constant in the upstream branch of the ridge over British Columbia at 500 hPa, while wind speeds decreased slightly as the ridge amplified and the jet core moved northward (Fig. 5). Near the surface, developing outflow conditions were evident as a deepening layer with an easterly wind component. As the outflow deepened, the directional shear layer between it and the southwesterlies aloft rose from approximately 850 to about 750 hPa at the end of the simulation. By 1800 UTC 28 April (Fig. 7c), the tracer over the mountains was subject to much stronger subsidence than over the ocean. Near the surface, the tracer was advected westward through the coastal valleys, reaching Vancouver a few hours later. A large proportion of the tracer was able to reach the surface.

Stable stratification and weak shear in the southwesterlies aloft allowed vertically propagating mountain waves to amplify up to about 500 hPa (Fig. 7b,c,f). Vertical velocities generated through a deep layer were on the order of 0.1 m s⁻¹, with the downward velocities as much as 0.1 m s⁻¹ greater than upward velocities. The zonal wavelength was approximately 45 km, and wind speeds at 500 hPa were about 10 m s⁻¹. Under these conditions, an air parcel would travel one wavelength in 1.25 h, during which time it would descend 450 m. Net downward transport by these waves rapidly enabled the eastward-travelling tracer to be intercepted by the PBL over higher topography and the inland valleys, despite the likelihood that the additional subsidence limited the depth of the PBL. Outflow in the PBL then transported it coastward. Later in the simulation (Fig. 7f), mountain wave activity aloft abated. In the easterly flow near the surface, isentropes deflected downward on the western slopes, a feature that grew more pronounced later in the simulation, suggesting stronger downward motion here as the simulation progressed.

A parallel simulation without terrain yields a first-order approximation of the effect of the western cordillera on downward transport (Fig. 7g,h). Though not shown, the long-wave pattern aloft was not greatly affected by the exclusion of terrain, and large-scale subsidence was similar for both simulations. However, by 1800 UTC 28 April in the flat terrain case, the absence of the tracer at low levels and isentropes that did not lower significantly (as in the real terrain case) indicate that total subsidence was less. Variations in vertical velocities across the transect and perturbations to the isentropes in the flat terrain case (Fig. 7g,h) can be attributed to differential heating over Vancouver Island (125°W) and the mainland coast (123°W).

Figure 8 shows average vertical velocities over the eastern portion of the model domain and provides further evidence of the importance of terrain-induced wave activity in transporting dust downward. With realistic terrain, vertical velocities at 600 hPa were negative throughout the simulation and displayed a significant diurnal fluctuation characterized by the strongest downward velocities during morning. With no terrain, vertical velocities were of smaller magnitude and ranged between net upward and downward motion. The same diurnal fluctuation is evident in this case and suggests that thermal forcing is responsible for the diurnal periodicity in the vertical motion field. MCK noted a diurnal variation in observed dust concentrations across the region, with maxima during the day. On the basis of these model results, early morning downward transport and then entrainment and advection within the boundary layer likely contributed to observed daytime peak ground-level concentrations. Nocturnal deposition processes may also have contributed to the observed diurnal modulation of surface concentrations (HUS).

d. Stability and shear effects

Bulk Richardson number (Riₜ) calculations along the same west–east transect as Fig. 7 illustrate the potential for turbulence in the PBL from shear and mountain effects (Fig. 9). Although turbulence occurs when the gradient Richardson number (Ri) is less than a critical value of 0.25, such is not the case when using Riₜ. The wind and temperature differences across thick layers are inadequate to resolve thin turbulent layers. Thus, there is a probability of turbulence even when Riₜ > 0.25 (Stull 1988). Based on Stull’s criteria, the shear envi-
The turbulent environment during the morning and afternoon of 29 April suggests that turbulence generated by the flow over the mountains was primarily responsible for the tracer entering the low-level outflow. The $R_i$ in the shear layer between the southwesterlies aloft and the outflow are between 0.25 and 5.0. Although the band of low $R_i$ loses its coherence over the mountains, the PBL is visible both in the early morning and in the afternoon. Absence of a significant diurnal periodicity in $R_i$ suggests that wind shear over the complex terrain was primarily responsible for generating turbulence rather than convection from surface heating. However, with the addition of daytime heating, $R_i$ de-
Fig. 7. Zonal cross section (49.27°N) for the Δx = 10 km simulation valid 1200 UTC 29 Apr 1998. The left column shows relative tracer concentration (shaded) and wind barbs (m s⁻¹), and the right column shows vertical velocity (m s⁻¹) (shaded) and potential temperature (K). The times correspond to Fig. 6. (g), (h) The same as (c) and (d) for the simulation done without topography.
creased in the afternoon, particularly over Vancouver Island (125°W) and the first ridge on the mainland (Fig. 9b). Because RiB values were calculated using centered finite differences, the lowest level is not included. Consequently, the top of the PBL appears during the day as a thin layer of values between 0.25 and 1.0 over the ocean (Fig. 9b).

Once mixed into the PBL, the tracer was advected coastward in the easterly flow. Along western slopes of Vancouver Island, the Coast Mountain Range, and the Cascade Range, the shear and stability environment caused further downward acceleration. At 1200 UTC 29 April (Fig. 9a), the modeled atmosphere was stably stratified through most of the troposphere, with a slight decrease in stability above 700 hPa. Stability was greatest in the lower troposphere, allowing gravity wave generation in the easterly outflow. The wind reversal (critical level) between the outflow and the southwesterlies aloft could be expected to absorb wave energy. This would lead to a hydraulic (Bernoulli) acceleration similar to that modeled in idealized studies (Clark and Peltier 1984) and observed along the mountains of the northwestern United States (Colle and Mass 1998). The RiB was lowest on the western slopes of Vancouver Island and the coastal mountains in the easterly flow (Figs. 7a,b) and was below 0.25 locally. This is consistent with gravity wave breaking in the hydraulic flow down the slopes (Clark and Peltier 1984). Persistent low RiB on the eastern slopes of Vancouver Island at 124°W was likely the result of the overwater fetch of the outflow. Cool air flowing over the Strait of Georgia between the British Columbian mainland coast and Vancouver Island is subject to heating from the warmer ocean below, leading to convection and low RiB values. Similarly, at the end of the simulation, outflow had reached well offshore, and RiB suggests deeper convection near the western edge of the domain (Fig. 9b).

The tracer caught in the easterly outflow was forced rapidly toward the coastal valley floors. Time series of zonal wind (not shown) are consistent with an increasing nocturnal cross-mountain pressure gradient, strengthening surface high pressure, and with downslope wind speeds reaching a maximum in the early morning hours. Throughout the simulation, low-level stratification and subsidence were sufficient to limit PBL heights to less than 1000 m but would not prevent entrainment of the near-surface tracer because PBL convection was strong. Consequently, simulation results indicate that downslope transport loaded the stable air within coastal valleys with the tracer prior to any convective PBL developing during daylight hours. This tracer was then further mixed to the surface by entrainment and downmixing associated with growth of the convective mixed layer.

A change from southwesterly to southerly winds at 700 hPa and deepening of the outflow layer combined to decrease gravity wave activity aloft over the last day of the simulation (Fig. 7). A southerly shift reduced the cross-range wind component on the last day of the simulation, but the northwest–southeast orientation of the coast mountains still allowed some wave generation in the stable environment. As described above, deepening outflow raised the critical shear level and captured the wave energy at low altitude.

A discrete Fourier analysis of the vertical velocities at 600 hPa over the eastern portion of the cross section
in Fig. 7 clearly shows this trend of reduced wave activity later in the period (Fig. 10). Peaks in the curves between wavelengths of λ = 40 and 50 km correspond to the mountain waves displayed in Fig. 7. The dual-peak structure is an artifact of the Fourier analysis, and a single peak would be evident if the analysis was entirely robust. The small sample, including only 64 grid points in the eastern part of the domain, leaves the analysis prone to errors related to discretization, such as aliasing and phase cancellation. Smaller peaks near 75 km correspond to land–sea differences, because waves of this length show up in the simulation that did not include topography (Figs. 7g,h).

e. Horizontal tracer distribution

Flow aloft exerted substantial control over the horizontal distribution of tracer in the simulation. A persistent, elongated region of divergence at 600 hPa, associated with the wind maximum aloft (Fig. 7a) led to a minimum in the horizontal tracer distribution in a swath cutting diagonally across British Columbia immediately to the north of Vancouver Island (Fig. 11a). Mountain waves generated in the Δx = 30 km grid appear in Fig. 11a as alternating centers of divergence and convergence through central British Columbia. A south–north transect through 120°W (Fig. 11b) shows the same split in the tracer distribution, with an absence in tracer aloft at 53°–56°N, where divergence in the upper level jet was strongest.

That the tracer reached the surface to the south of the swath of divergence, but not to the north, can be explained by the evolution of the mountain wave activity and outflow conditions. As the wind maximum in the upstream branch of the upper-level ridge moved northward during the simulation, so did the primary axis of vertically propagating mountain waves. Prior to 1800 UTC 28 April, vertical velocities were much stronger over Washington, and later in the simulation, the vertical velocities were strongest over central British Columbia. In addition, the outflow layer deepened and strengthened northward (Fig. 6b,d,f). Time evolution of surface-layer tracer concentrations reflects the northward propagation of processes that brought the tracer to the surface and later advected it westward (Fig. 6a,c,e).

f. Frequency of meteorological conditions

To estimate the probability of a similar flow pattern occurring and thereby transporting pollutants to populated coastal regions after initial interception of pollutants by mountain wave activity, daily average NCEP–NCAR (National Center for Atmospheric Research) reanalyses from 1990 to 1999, inclusive, were examined. In so doing, it is assumed that similar zonal wind shear has the potential to generate similar wave dynamics. Because outflow conditions often persist for several days, daily mean averages were deemed sufficient to estimate the probability of occurrence. During the April 1998 outflow event, the zonal wind aloft (500 hPa) was greater than 10 m s⁻¹ while the zonal wind near the surface (925 hPa) was less than 2.5 m s⁻¹. This combination was used as the criterion for comparison (i.e., zonal wind shear of 15 m s⁻¹).

Table 1 shows frequencies of occurrence of zonal
wind shear when both the westerly component aloft and the easterly component near the surface are in the range of this case study. Occurrences in the table’s second column, which is the strictest criterion, are strongly skewed toward winter months, when outflow conditions are more common (Jackson 1996). The third column evaluates a relaxed criterion, in which the magnitude of the wind shear is similar to this case, but either the westerly component aloft or the near-surface easterly component has a smaller magnitude. Probabilities are not cumulative; that is, occurrences in column three exclude those in column two. The total probability that the zonal wind shear is greater than 15 m s$^{-1}$ was $1.2 + 17.4 = 18.6\%$. In both cases here, $U_{925} < 0$ and $U_{500} > 0$, denoting a zonal wind reversal between these levels. In the 10 yr of reanalyses, no days were found that contained a zonal wind reversal with shear less than 15 m s$^{-1}$, suggesting that daily averaged analyses do not capture such events. Those are likely not synoptically driven, as was this case, but are products of local topographically driven circulation with a diurnal evolution.

5. Summary and conclusions

A mesoscale model was used to simulate the arrival of mineral aerosol from an intense dust storm in the Gobi over western North America during late April 1998. In the absence of detailed meteorological measurements during the event, the model was used to elucidate FT–PBL exchange mechanisms by which dust was mixed downward and subsequently affected coastal regions. A 4-day simulation, nesting grids to $\Delta x = 10$ km, provided a realistic representation of the surface dust distribution when it was initialized with a thin, horizontally homogeneous layer of the tracer with a maximum at 650 hPa. Modeled spatial and temporal patterns of tracer concentrations were qualitatively consistent with observations during the event. A parallel simulation was completed without topography, showing that topography was important in generating downward vertical velocities. Last, 10 yr of NCEP–NCAR reanalysis grids were examined to gain an estimate of the frequency of similar meteorological conditions.

Model results indicate that several processes combined to bring particles from aloft (650 hPa) to the surface. First, large-scale subsidence associated with a strengthening high pressure center inland and vertically propagating mountain waves combined to force the tracer to near the height of the topography over British Columbia and Washington. The tracer was incorporated into the low-level outflow by entrainment processes at the top of the PBL in the mountains and inland valleys and through turbulence generated in the shear layer residing near crest height. The tracer was then advected westward and forced into the coastal valleys by hydraulic (Bernoulli) acceleration along the western slopes, which reached a maximum in the early morning hours. A shallow PBL could then mix it down to the surface during the day. Outflow dynamics through the mountain gaps may also have played a role in the transport of the tracer from the east of the mountain peaks toward the coast.

As the ridge strengthened during the time period of the simulation, the region most affected by down-mixing processes and coastward advection moved northward. This included the vertically propagating mountain waves aloft, the eastward advection near the surface, and the hydraulic downslope acceleration over western slopes. The northern extent of the ground-level tracer was constrained by a broadly zonal swath of upper-level divergence that reduced the midtropospheric source of the tracer available to downward-transport processes.

Zonal wind shear was responsible for much of the free-atmosphere dynamics that brought the tracer downward in the simulation, including vertically propagating mountain waves in the westerly flow, potential mixing in the shear layer, eastward transport near the surface, and a hydraulic acceleration with possible wave breaking in the low-level easterlies. Daily average NCEP–NCAR reanalyses from 1990 to 1999 suggest that the probability of similar meteorological conditions and comparable zonal wind shear with comparable zonal velocities is 1.2%. This may be interpreted as an upper-bound on the probability because other factors, such as stability and cross-mountain pressure gradient, were not considered and because of the coarse vertical resolution of the analysis. The probability of a similar dust event would be difficult to estimate, but it is bound to be even smaller because the frequency of occurrence was largely skewed toward the winter months whereas Asian dust storms are skewed toward spring (Braaten and Cahill 1986). Dust must also be transported across the Pacific without dispersing, being scoured out, or settling before it reaches the coast of North America. The probability

<table>
<thead>
<tr>
<th>Month</th>
<th>$U_{925} &lt; -5$, $U_{500} &gt; 10$ m s$^{-1}$</th>
<th>$U_{500} - U_{925} &gt; 15$ m s$^{-1}$</th>
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<td>Jan</td>
<td>10</td>
<td>83</td>
</tr>
<tr>
<td>Feb</td>
<td>5</td>
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<tr>
<td>Jul</td>
<td>0</td>
<td>24</td>
</tr>
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<td>Aug</td>
<td>0</td>
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</tr>
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<td>Sep</td>
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<tr>
<td>Oct</td>
<td>2</td>
<td>87</td>
</tr>
<tr>
<td>Nov</td>
<td>7</td>
<td>87</td>
</tr>
<tr>
<td>Dec</td>
<td>13</td>
<td>78</td>
</tr>
<tr>
<td>Total</td>
<td>42</td>
<td>635</td>
</tr>
<tr>
<td>Probability</td>
<td>1.2%</td>
<td>17.4%</td>
</tr>
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</table>
that similar wind shear magnitude occurred while allowing one or both of the zonal wind velocities to be less than in this case was much higher, at 18.6%, but again this must be interpreted as an upper limit and does not guarantee similar dynamics.

Results described herein suggest that the meteorological conditions that permit interception of pollutants by the mountain ranges of western North America and then subsequent transport of pollutants coastward are rare. However, further research is required to investigate the role of the western cordillera in intercepting the full range of pollutants emanating from Eurasia. It is conceivable that pollutants (crustal or anthropogenic) may be intercepted by other processes (e.g., precipitation) or mixed downward but not coastward. Clearly, this has important implications for regional environmental quality, particularly in light of predictions of growing emissions from Eurasia and emerging evidence that trans-Pacific transport of anthropogenic pollutants is not unusual.

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REFERENCES


