

Dear Referee,

Thank you very much for your helpful comments and suggestions.

For better answering your main comments, we have rearranged them by issue.

The changes we propose in the text and figure captions, in response to all referees suggestions, are included in a revised version starting at page 4 of this document.

1. *The manuscript is based on the same model setup and boundary conditions as the earlier study by Sima et al. (2009), but focuses on a slightly different region. Therefore some of the conclusions in this ms. (e.g. importance of seasonality and vegetation) can already be found in the earlier study. Taking into account that the manuscript is a regional extension of the earlier study I am surprised by the somewhat lengthy presentation. Apart from the regional extension (which is logical given the loess sequence) the manuscript adds only little to the insights provided already in the earlier study. [...] I would strongly recommend to condense the manuscript substantially (specifically, introduction, results and conclusions).*

When we started this manuscript, we did consider the (easier) possibility of giving minimal explanations and referring the reader whenever possible to our previous modeling study on Western Europe. Taking into account the high spatial resolution of the numerical experiments (as imperfect as they may be with respect to the numerical setting), we opted for a more self-contained manuscript, which would better serve readers interested not only in the general mechanisms, but also in a detailed analysis of the results for Eastern Europe and comparisons with the West.

In this view, we have partially rewritten the introduction and conclusions to better show similarities and differences between the previous study and this manuscript, but have chosen not to modify the presentation of the results themselves. In the introduction we have also developed the presentation of the source areas, as suggested by Referee 2.

2. *I am also concerned with regard to the boundary conditions of the model, which are a mixture of glacial boundary conditions (partly outdated: ice sheet configuration), including somewhat arbitrary SST anomalies in the North Atlantic to represent a Heinrich stadial. As the authors mention there are several simulations with coupled ocean atmosphere models available for MIS 3 and it is surprising that the current study does not make use of this information. The discussion reads to some extent as an excuse for the use of a mixture of boundary conditions. However I do not find the line of argument very convincing with regard to having a setup that is representative of MIS 3. To me it would be more logical to present the study as a sensitivity experiment for some mixture of glacial boundary conditions (that were available at the time their model experiment was been conducted).*

In both Sima et al. (2009) and this manuscript (Section 4, Discussion, end of the 2nd paragraph) we have indeed written that we saw our simulations as sensitivity

experiments to changes in the North-Atlantic surface conditions as those suggested by data for DO and H events. To better stress this information, we added it in the description of the numerical simulations (Section 2, end of the 2nd paragraph) and in the conclusions (Section 5, 1st paragraph).

We have also slightly modified the abstract and the title (which was indeed misleading). The title is now:

“Modeling dust emission response to North-Atlantic millennial climate variations from the perspective of East European MIS 3 loess deposits”

As we wrote in the discussion, using the output of a coupled model for MIS3 instead of an LGM reconstruction for the sea-surface conditions would of course be an alternative worth exploring. Such outputs are more coherent with the rest of the numerical setup (inspired by the climate sequence around the Heinrich event 4), but also strongly model-dependent, as shown by Brandefelt et al. (2011). Here, we wanted to keep the set-up simple so that the perturbations can be ascribed to the SST anomalies over the North Atlantic and not to SST/sea-ice differences elsewhere. It is also true that these outputs were not available when we ran the experiments, and that it is not possible for us to run new experiments with exactly the same model because the model version we used does not run on the new super-computer.

We have made small changes to better present our choices in the Discussion, intended as a detailed explanation of the numerical set-up and its consequences (not as an excuse).

Minor comments:

p. 145, l. 9: Testing a correlation should be rephrased since it implies a statistical test.

The introduction was rewritten, this expression does not appear anymore.

p. 145, l. 14: Consider replacing H-events by H-stadials throughout the text because it is the cold phase and not the IRD input that determines the dust deposition anomalies.

We added in Section 2, end of the 2nd paragraph :

“In the following, we will use “H-stadial” when specifically referring to the cold climate interval associated with an H event (which is defined as an episode of massive iceberg release, recorded in marine sediments by layers rich in ice-rafted debris).”

and made the adequate changes in the text. No change was made where mentioning other studies that refer themselves to “H events”. Also, we kept the corresponding experiment name as “HE” for consistency with the previous study.

p. 149, l. 18: F is not defined.

Corrected.

p. 150, l. 19: Based on Sima et al. (2009) it is surprising that annual mean values are analyzed in the first place given that the seasonality was found to be very important in that study.

The precise “dusty season” is only determined later in the manuscript (Section 3.2), so that in order to determine the relative position of the sources we have examined both annual and monthly averages. The monthly averages are only partially shown, later, in figure 6, so we have omitted to mention them here. We do it now:

“More important, considering the strong seasonality of emissions for Western Europe (Sima et al. 2009): the strong westerly component is also found in monthly averages (not shown).”

p. 151, l. 22: It is interesting that the margin of the Fennoscandian ice sheet is discussed in this context. Is this also reflected in the ICE-4G reconstruction?

Yes (and it is practically the same in the ICE-5G reconstruction).

We added ($\sim 57^\circ\text{N}$ in our experimental setup, cf. the ICE-4G reconstruction).

p. 154, l. 22: Avoid “a bit lower”.

Replaced by “slightly lower”.

p. 168, l. 17: Replace hypotheses by assumptions.

Removed (conclusions partly rewritten).

Fig. 1: Mention in caption that blue/gray colors indicate elevation.

Corrected. Please note that the figure and the caption are slightly changed, following the publication of the Antoine et al. (2013) study.

This and most other figures: axis titles for latitude and longitude are missing.

We added “°N” and “°E” where appropriate on the concerned figures, except for figure 6 (18 panels, not enough space available), for which we included the explanation in the figure caption.

Fig. 6: Panels are much too small.

As actually formatted by CPD, yes (we are not happy about it, and hope the zoom on the pdf file helps).

In final form, each of the figures 6 and 7 is intended to occupy an A4 page.

Fig. 7/8: Indicate statistical significance of differences.

Fig. 7: White areas indicate no significant differences at 95% confidence level according to Student’s test - added in figure legend.

Fig. 8: White areas indicate masked regions as explained in the text; explanation added in the figure caption.

1 **Modeling dust emission response to North-Atlantic**
2 **millennial climate variations from the perspective of**
3 **East European MIS 3 loess deposits**

4
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18

18 **Abstract**

19 European loess sequences of the Marine Isotope Stage 3 (~60 - 25 kyr BP), show periods
20 of strong dust accumulation alternating with episodes of reduced sedimentation,
21 favoring soil development. In the western part of the loess belt centered around 50°N,
22 these variations appear to have been caused by the North Atlantic rapid climate
23 changes: the Dansgaard-Oeschger (DO) and Heinrich (H) events. It has been recently
24 suggested that the North-Atlantic climate signal can be detected further east, in loess
25 deposits from Stayky (50°05.65'N, 30°53.92'E), Ukraine. Here we use climate and dust
26 emission modeling to investigate this data interpretation. We focus on the areas north
27 and northeast of the Carpathians, where loess deposits can be found, and the
28 corresponding main dust sources must have been located as well. The simulations were
29 performed with the LMDZ atmospheric general circulation model and the ORCHIDEE
30 land-surface model. They represent a reference “Greenland stadial” state and two
31 perturbations, seen as sensitivity tests with respect to changes in the North-Atlantic
32 surface conditions between 30° and 63°N: a “DO interstadial” and an “H event”. The
33 main source for the loess deposits in the studied area is identified as a dust deflation
34 band, with two very active spots located west-northwest from our reference site.
35 Emissions only occur between February and June. Differences from one deflation spot
36 to another, and from one climate state to another, are explained by analyzing the
37 relevant meteorological and surface variables. Over most of the source region, the
38 annual emission fluxes in the “interstadial” experiment are 30 to 50% lower than the
39 “stadial” values; they would only be about 20% lower if the inhibition of dust uplift by
40 the vegetation were not taken into account. Assuming that lower emissions result in
41 reduced dust deposition leads us to the conclusion that the loess-paleosol stratigraphic
42 succession in the Stayky area reflects indeed North-Atlantic millennial variations. In the
43 main deflation areas of Western Europe, the vegetation effect alone determined most of
44 the (-50% on average) stadial-interstadial flux differences. Even if its impact in Eastern
45 Europe is less pronounced, this effect remains a key factor in modulating aeolian
46 emissions at millennial timescale. Conditions favorable to initiating particularly strong
47 dust storms within a few hundred kilometers upwind from our reference site, simulated
48 in the month of April of the “H event” experiment, support the correlation of H events
49 with peaks of grain-size index in some very detailed loess profiles, indicating increased
50 coarse sedimentation.

51

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Supprimé: Placed in Marine Isotope Stage 3 (~60 - 25 kyr BP) conditions, they only differ by the surface conditions imposed in the North Atlantic between 30° and 63°N.

51 1 Introduction

52 In Europe, a west-east eolian corridor was formed in glacial times between the British
53 and Fennoscandian ice sheet to the north and the relatively high mid-latitude European
54 relief (including the Alpine glacier) to the south (Fig. 1). Vast areas along this corridor
55 are generally flat (below 200m altitude), with the geological substratum mostly
56 represented by relatively easily erodible Tertiary or Cretaceous rocks (Asch et al.,
57 2005), and have periodically been subject to strong dust deflation under glacial climate
58 conditions. Deflatable material with a large range of grain sizes was made available by a
59 variety of mechanisms acting at local or regional scales, at timescales from seasonal to
60 millennial and orbital: the exposure of the continental shelf due to sea-level lowering,
61 grinding of rocks by ice sheets and glaciers, frost weathering, fluvial erosion by
62 periglacial rivers, eolian erosion by strong glacial winds, accentuated by a reduced
63 vegetation cover in a much colder and dryer climate than today. Particularly rich in
64 easily deflatable sand and silts were the exposed continental shelves and the periglacial
65 outwash plains, as well as the periglacial river valleys, mostly dried-out outside the
66 snowmelt period.

67 Part of the material deflated in these source areas has accumulated in the south of the
68 eolian corridor, forming a loess belt at about 50°N latitude. Some of the deposition
69 areas, located in a relief context allowing dust remobilization, could have been
70 “secondary dust sources”. Loess sedimentation rates have strongly varied at millennial
71 timescale, especially during Marine Isotope Stage 3 (MIS3, ~58,900-24,100 yr BP;
72 Martinson et al., 1987). High-resolution studies on sequences from Nussloch, Germany
73 (Rousseau et al., 2007; Antoine et al., 2009) have suggested that the sedimentation
74 variations in the Western Europe were correlated with the abrupt climate changes
75 known as Dansgaard-Oeschger (DO) events (Dansgaard et al., 1993) and Heinrich (H)
76 events (Heinrich, 1988; Broecker et al., 1992). The North-Atlantic cold episodes
77 identified in ice or marine cores, i.e. Greenland stadials (North Greenland Ice Core
78 Project, 2004; Rousseau et al., 2006) and H events, appear to correspond to periods of
79 loess accumulation, indicating a very active dust cycle caused by dry and windy
80 conditions. The warmer Greenland interstadials were associated to moister and less
81 windy conditions on the continent, with a less active dust cycle, favoring soil formation.
82 Alternating loess-paleosol units are recognizable especially after 40 kyr BP, when
83 the main loess sedimentation interval in Europe begins.

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Supprimé: Dust emission changes induced over Western Europe by the North-Atlantic millennial climate variation

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Supprimé: have been studied using an atmospheric general circulation model (Sima et al., 2009). The main aim was to test the correlation proposed by Rousseau et al. (2007) and

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Supprimé: aeolian sequences from the west of

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Supprimé: European loess belt centered around the 50°N latitude, and

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84 Rapid environmental changes have also been identified in loess sequences further east
85 (Haesaerts et al., 2003; Rousseau et al., 2001, 2007, 2011; Gerasimenko and Rousseau,
86 2008; Antoine et al., 2009). They are expressed in the loess-paleosol stratigraphic
87 succession, and in the variations of different indices: grain-size index, magnetic
88 properties, carbon isotope ratios, and, where available, in the pollen record. Following
89 investigations by Kukla (1977), a link between Central and Eastern Europe was
90 established on the basis of sequences from Dolni Vestonice, in the Czech Republic (e.g.,
91 Fuchs et al., 2012, Antoine et al., 2013), and Vyazivok, in Ukraine (e.g., Rousseau et
92 al., 2001). Recently, using high-resolution data from Nussloch, Germany, and another
93 Ukrainian site, Stayky, Rousseau et al. (2011) suggested that the North Atlantic climate
94 signal has been recorded throughout the European loess band, at least as far as 30°E.

95 In a previous study, we have used an atmospheric general circulation model (AGCM)
96 and offline dust emission calculations to investigate the impact of North-Atlantic
97 millennial climate changes on dust emission variations in Western Europe (Sima et al.,
98 2009). Three numerical simulations were run, designed as sensitivity experiments with
99 respect to SST variations in the North Atlantic as those associated with DO and H
100 events. We have analyzed the main western European deflation areas, with focus on the
101 exposed continental shelf in the English Channel and the North Sea (Juvigné, 1976;
102 Auffret, 1980; Auffret et al., 1982; Lauthridou et al., 1985; Antoine et al., 2003a). The
103 main results consisted in: a) a strong seasonality of emissions, which occurred overall
104 between February and June (with differences from one climate state to another), when a
105 compromise was achieved between snow melting, soil drying, and vegetation
106 development, b) considerably lower emission fluxes in the “Greenland interstadial”
107 experiment than in the “Greenland stadial” and “H event” simulations, supporting the
108 interpretation of loess sedimentation variations as being produced by the North-Atlantic
109 millennial variability. It was also shown that the vegetation, which inhibits eolian
110 erosion, has played a key role in determining the seasonal cycle of emissions and the
111 differences of dustiness between the relatively warm versus cold North-Atlantic phases.

112 Following the data study by Rousseau et al. (2011), which proposed a correlation
113 between Greenland, West and East European dust records, here we focus on Eastern
114 Europe. We use the same AGCM simulations and dust emission calculations as in Sima
115 et al. (2009), combined with information from the loess site of Stayky, in Ukraine
116 (briefly described in Section 2).

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Supprimé: The European loess belt continues eastward, along the west-east eolian corridor delimited, in glacial times, by the Fennoscandian ice sheet to the north and the relatively high mid-latitude European relief (including the Alpine glacier) to the south (Fig. 1). Where the Carpathians curve southward, the loess band widens, covering a large part of the East European plain. Loess sequences from this part of the continent also reveal rapid environmental changes

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Supprimé: Based upon these indices, a correlation was recently established between the loess sedimentation variations in Eastern and Western Europe

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Supprimé: High-resolution data have been used from two key loess sequences: Nussloch, in Germany

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Supprimé: These sequences are particularly detailed over the main loess sedimentation interval in Europe, ~40–15 kyr BP. Following investigations by Kukla (1977), a link between Central and Eastern Europe had already been shown on the basis of sequences from Dolni Vestonice, in the Czech Republic (e.g., Fuchs et al., 2012), and from another Ukrainian site, Vyazivok (Rousseau et al., 2001). Hence, it appears

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Supprimé: This data interpretation is the first aspect that we investigate here, by comparing simulated dust emissions north and northeast of the Carpathians (where the potential sources for the ~50°N aeolian deposits were most likely located), in the cold versus the warm North-Atlantic episodes.

117 After identifying the potential sources for the dust deposited around this site (Section
118 3.1) we investigate the impact of North-Atlantic SST changes on dust emission in these
119 areas. The “dusty season” is determined (Section 3.2), and the relevant climate variables
120 and surface conditions are analyzed on average over this period of the year (Section
121 3.3) with special attention to the role of vegetation. Furthermore, we examine in detail
122 the hypothesis that H events could be identified in European loess sequences as peaks of
123 grain size (Rousseau et al., 2007, 2011). We discuss the results (Section 4), draw the
124 conclusions and indicate some perspectives (Section 5).

126 2 Reference loess site, numerical simulations, dust emission 127 calculations

128 The reference loess site for this study is Stayky (50°05.65'N, 30°53.92'E, 194m asl), in
129 Ukraine, located by the Dnieper River, about 50 km south of Kiev. This outcrop was
130 chosen for its detailed record of the last climate cycle, during a preliminary
131 investigation of the numerous outcrops of the loess series studied in the area
132 (Gerasimenko and Rousseau, 2008). It is situated on a cliff ending the plateau on the
133 right bank of the river; the Dnieper river floodplain lies on the left bank. The sequence
134 corresponding to the last climatic cycle has been studied at high resolution by defining a
135 precise stratigraphy, sampling continuously for grain-size analysis, and taking sediment
136 for optically stimulated luminescence (OSL) dating (Rousseau et al., 2011). For the
137 interval 38 to 18 kyr BP, alternating loess and embryonic soils similar to the loess-
138 paleosol doublets observed at Nussloch (Germany) have been identified, as well as a
139 similar pattern of the grain-size index variations. A correlation was proposed between
140 the loess-embryonic soil doublets and the Greenland stadial-interstadial climate cycles.
141 Also, it was suggested that two particular peaks of the grain-size index might
142 correspond to H events 3 and 2.

143 The simulations have been carried out with the LMDZ.3.3 atmospheric general
144 circulation model (Jost et al., 2005) including the ORCHIDEE land-surface model
145 (Ducoudre et al., 1993; Krinner et al., 2005). Inspired by the GS9-H4-GIS8 sequence
146 around the H4 event (approx. 39 kyr BP; Bard et al., 2004), they represent a reference
147 glacial state (“Greenland stadial”, GS), a cold (“H event”, HE) and a warm
148 (“Dansgaard-Oeschger”, or “Greenland interstadial”, GIS) perturbation. Thus, the

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Supprimé: So, here we take as a reference the Stayky loess site (in Ukraine), where millennial-timescale variations are particularly well recorded, and identify

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Supprimé: Taking into account what we have learned on the strong seasonality of emissions in the Sima et al. (2009) s

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Supprimé: tudy, we first determine t

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Supprimé: simulated

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Supprimé: , which inhibits aeolian erosion

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Supprimé: Stadial-interstadial vegetation changes were identified in our previous work as the main factor by which the North-Atlantic millennial variations modulated dust emission in the western European deflation areas.

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Supprimé: , and designed to resemble the GS9-H4-GIS8 sequence around the H4 event (approx. 39 kyr BP; Bard et al., 2004)

149 orbital parameters (Berger, 1978; Berger and Loutre, 1991) were set to 39-kyr BP
150 values, and the CO₂ concentration to 209 ppmv (Petit et al., 1999). The ice-sheet
151 configuration at 14 kyr BP was selected from the ICE_4G reconstruction (Peltier, 1994),
152 as corresponding to a sea level similar to that at 39-kyr BP, approximately 60m lower
153 than today (Siddall et al., 2008). The land-sea mask of the LMDZ and SECHIBA
154 models was adapted to this sea level. In the absence of reconstructions or climate model
155 results for the MIS3 sea-surface temperatures (SSTs) and sea ice at the time when we
156 run the simulations, the GLAMAP2000 reconstruction (Sarnthein et al., 2003) for the
157 Last Glacial Maximum (LGM, approximately between 23 and 18 kyr BP) was used in
158 the reference glacial climate simulation GS. The cold and warm perturbations were
159 obtained by only altering the North Atlantic surface conditions in the latitudinal band
160 between 30°N and 63°N. All-year-long zonal SST anomalies of up to ±2°C (Cortijo et
161 al., 1997) were applied in this band, and sea ice was imposed where the SST was lower
162 than -1.8°C. The simulations are thus sensitivity experiments with respect to variations
163 in the North-Atlantic surface conditions as those associated with DO and H events. In
164 the following, we will use “H-stadial” when specifically referring to the cold climate
165 interval associated with an H event (defined as an episode of massive iceberg release
166 recorded in marine sediments by layers rich in ice-rafted debris).

167 In the ORCHIDEE model version we have used here (Krinner et al., 2005), the
168 computed leaf area index (LAI) varies between minimum and maximum values fixed
169 for each plant functional type (PFT) to standard values based on averaged observations,
170 and is only modulated by the AGCM-derived temperature. The maximum grid-cell
171 fraction that can be occupied by each PFT is also prescribed. In our paleoclimate
172 experiments we kept the present-day values, as recommended by the Paleoclimate
173 Modelling Intercomparison Project (e.g., Braconnot, 2004) for the LGM simulations.
174 The actual grid-cell fraction covered by a PFT depends on the imposed maximum
175 vegetation fraction and the computed LAI. In each experiment, the LMDZ-ORCHIDEE
176 model was run for a spin-up period of one year, followed by 20 years that were
177 analyzed.

178 The Sima et al. (2009) study has shown the importance of vegetation, as an inhibitor of
179 aeolian erosion, in modulating dust emission at millennial timescale in the western
180 european deflation areas. Therefore, here we calculate again separately the emitted ‘dry’

181 dust flux Fd , taking into account all factors but the vegetation effect and the F flux
182 including the vegetation effect. These fluxes are given by the following formulas:

183
$$Fd = C' * fd * w_{10m}^2 * (w_{10m} - w_{th}) \text{ for } w_{10m} > w_{th} \text{ (} Fd = 0 \text{ otherwise)}$$

184 and

185
$$F = Fd * fv = C' * E * w_{10m}^2 * (w_{10m} - w_{th}) \text{ for } w_{10m} > w_{th} \text{ (} F = 0 \text{ otherwise)}$$

186 where :

- 187 • C' is a constant for every grid cell that only depends on intrinsic characteristics
188 as the surface roughness (vegetation excluded), grain-size distribution and
189 texture of the bare soil. Here we take $C' = 5 \times 10^{-7} \text{ g m}^{-5} \text{ s}^2$ everywhere in our
190 domain of study, an intermediate value in the range of those determined by
191 Balkanski et al. (2004) for the present-day arid and semi-arid regions;
- 192 • fd , which we call “dry soil fraction”, quantifies the soil water effect on dust
193 emission. It equals the snow-free fraction of the grid cell if the soil is dry over
194 more than 5mm depth, and is 0 otherwise;
- 195 • fv , the vegetation factor, quantifies the vegetation effect of inhibiting wind
196 erosion. It is calculated as a function of the vegetated soil fraction $fveg$,
197 following the equation (6) of Fryrear (1985), corrected at low (<10%) and high
198 (>60%) vegetation cover:

199
$$fv = \min(1, 1.81 * \exp(-7.2 * fveg)) \text{ if } fveg < 0.6, \text{ and } fv = 0 \text{ otherwise.}$$

- 200 • $E = fd * fv$ is the “erodible fraction”, and represents the grid-cell fraction where
201 dust emission is allowed at any given moment by both soil humidity and
202 vegetation effects;
- 203 • w_{10m} is the 6-hourly averaged 10m-wind computed by the atmospheric model;
- 204 • w_{th} is the threshold wind speed for erosion, determined for each grid cell, same
205 as C' , by the intrinsic (bare) soil characteristics. As in Sima et al. (2009), a
206 constant value is used everywhere: 7 m s^{-1} , close to the lowest values for the
207 present-day deserts, either measured (Wang et al., 2003) or derived as a function
208 of soil characteristics (Marticorena and Bergametti, 1996; Laurent et al., 2005).

209 3 Results

210 3.1 Potential dust sources

211 In order to determine where the main source areas must have been located with respect
212 to the Stayky loess site, we analyze the wind direction at the surface and in altitude. We
213 take the 850hPa level (corresponding on average to an altitude of about 1500m asl) as
214 relevant for the medium-to-long distance dust transport. The mean annual wind
215 direction at this level has a strong westerly component in the reference state (Fig. 2), as
216 well as in the two perturbations (Rousseau et al., 2011, Fig 5 therein). More important,
217 considering the strong seasonality of emissions for Western Europe (Sima et al. 2009):
218 the strong westerly component is also found in monthly averages (not shown). To
219 identify the most probable position of the local source areas with respect to the site, we
220 examine the wind roses derived from 6-hourly 10m-winds for the 20 years analyzed for
221 each simulation (Fig. 3). Again, in all three states, westerly wind occurrences greatly
222 exceed the easterly ones. This explains why, despite the large amount of sand available
223 in the Dnieper river floodplain, east of Stayky, very little sand is found in the loess
224 deposit (Rousseau et al., 2011). Also, the loess site is located approximately 150 meters
225 higher than the valley, so the sand in the loess profile must have been transported during
226 rare strong easterly wind events.

227 The 10m-wind speed values are up to 20 m s^{-1} for the GS state, up to 21 m s^{-1} for GIS,
228 and about 22 m s^{-1} for HE, but the frequency of strong winds, exceeding 14 m s^{-1} , is not
229 high enough to see it in the plots. According to Sima et al. (2009), the yearly averaged
230 dust fluxes are not controlled by the strongest winds, but rather by the much more
231 frequent medium wind-speed category (from 9 to 14 m s^{-1} in the case of the western
232 European main sources). For HE, the strongest 10m-wind events, exceeding 20 m s^{-1} ,
233 occur in April and December (not shown). We will discuss this result in Section 3.2,
234 where we look at dust emission seasonality and the relationship with the identification
235 of H events in loess sediments as peaks of grain-size index.

236 Finally, considering the low end of the grain-size range in the Stayky profile (the clay
237 fraction, with diameters up to a few microns), most of the constituting material has
238 probably originated from sources not more than thousand km far from the site
239 (Rousseau et al., 2011). All these taken into account, we consider that the main potential
240 dust sources for Stayky must have been located between 15° and 35°E . This is the

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241 longitudinal range for which we will perform the dust emission calculations. The
242 latitudinal range of interest spans over a 10°-wide band centered on Stayky: from 45°N,
243 the latitude of the southern Carpathians, to 55°N, in the Baltic Sea, and close to the
244 Fennoscandian ice-sheet southern limit on the continent around 40 kyr BP (~57°N in
245 our experimental setup [cf. the ICE-4G reconstruction](#)). The resulting domain is shown
246 in Fig 2. When representing dust fluxes or surface conditions, we exclude the
247 Carpathians (by masking the areas with altitudes exceeding 500m), where no relevant
248 emission may occur. We also exclude the lowlands inside the mountain arch, as they are
249 unlikely to have contributed to dust deposition in the ~50°N band examined in this
250 study.

251 For each simulated climate state, we compute yearly averaged dust emission fluxes over
252 the domain of interest (Fig. 4). In all climate states, emission mainly occurs in a NW –
253 SE band, located north and northeast of the Carpathians (Fig. 5). Two spots appear as
254 most active with respect to climate-related conditions and are placed west-northwest of
255 Stayky, constituting potential source areas for this reference site. The one closest to
256 Stayky, hereafter referred to as “Spot 1”, is in Ukraine, centered at about 51°N - 26°E
257 (S1 in Fig. 4a). It partly covers areas where loess deposits are located (Fig. 5), which
258 means that here dust remobilization might have been important. The second most active
259 region, “Spot 2”, is in Poland, centered at about 53°N – 19°E (S2 in Fig. 4a).

260 The extent of the potentially most active sources does not change significantly from the
261 GS to the HE climate state (Fig. 4 a,b), but annual mean dust fluxes are smaller for HE
262 than for GS, especially over Spot 2. A shrinking of the potential emission area can be
263 seen for the GIS compared to GS, as well as a decrease, stronger than in the HE case, of
264 the annual mean dust emission fluxes (Fig. 4c).

265 To explain the spatial distribution of the potential deflation areas and the differences of
266 dustiness between the simulated climate states (Fig.4), we need to examine the
267 variations of the relevant climate variables: wind, precipitation, temperature, as well as
268 the surface conditions determined by these variables: soil humidity, snow and
269 vegetation covers. The annual or seasonal means of these quantities are not quite
270 relevant for this matter (see Sima et al., 2009), so we first determine the period of the
271 year when dust emission occurs over our area of study, and then analyze the variables of
272 interest as averages on this period.

273 3.2 Seasonality of emissions

274 Sima et al. (2009) have shown the strongly seasonal nature of dust emission occurrence
275 over the large deflation areas formed by sea-level lowering in the English Channel and
276 the south of the North Sea. Here we remain in the same latitude range, and the annual
277 cycle of the main variables impacting dust emission resembles that for the west of
278 Europe (Fig 5 a,b in Sima et al., 2009). Winter is characterized by strong winds and
279 scarce vegetation, but snow cover and the high soil humidity prevent dust from being
280 uplifted. Conversely, in summer the wind weakens and, as the soil dries up, the
281 development of vegetation becomes the main surface process blocking dust
282 mobilization. These different conditions constraining dust emission determine the
283 potential deflation areas, and their seasonality. Thus, in our domain of interest, the main
284 emission band located north and northeast of the Carpathian Mountains is most active in
285 springtime, when a compromise is reached between soil humidity, wind and vegetation
286 conditions (Fig. 6). As in the western European source areas, the seasonal evolution of
287 dust emission intensity differs from a climate state to another. Furthermore, for each
288 climate state, the two most active spots show noticeable differences in their seasonality.
289 Spot 1 is the first to start emitting dust: in February for GS and GIS, and in March for
290 HE. In all three states, the most active period is April. The conditions become
291 unfavorable to dust emission in May for GIS, and in June for the other two states.

292 Spot 2 has the same general evolution, but with one month of delay with respect to Spot
293 1. It starts to significantly emit in March for GS and GIS, and in April for HE. For GS
294 and HE it is most active in May, and stops emitting in June, whereas for GIS the
295 emissions cease one month earlier.

296 If we consider the two most active areas together, the dusty season in our region of
297 interest lasts from February to June in the stadial state, from March to June in the HE
298 state, and from February to May in the interstadial state. For all months and climate
299 states the average 850hPa winds are from west or west-northwest (Fig. 6), so that the
300 deflation band we have identified may feed the European aeolian deposits located
301 farther eastward (Fig. 1). Considering the distance to our reference site (~300 km for
302 Spot 1, ~800 km for Spot 2), and the monthly means of 850hPa-wind direction over the
303 emission season, Spot 1 is the best candidate as a dust source for the loess deposits in
304 the Stayky area. Spot 2 certainly contributes as well, even though (again, considering

305 the monthly means of 850hPa-wind direction in Fig. 6) much of the dust emitted here is
306 probably transported on a more northern path.

307

308 **3.3 Climate variables, surface conditions and dust emission**

309 To explain the spatial distribution of the potential deflation areas, the differences of
310 dustiness and seasonality between the two most active spots, and between the simulated
311 climate states (Figs. 4 and 6), we need to examine the relevant climate variables and
312 surface conditions. As shown in Sect. 3.2, for all simulated climate states, the annual
313 amount of dust is only produced over a period between February and June. Therefore, in
314 the following, we analyze the variables and anomalies of interest as averages over this
315 “dusty season”.

316 The climate variables we address are (Fig. 7): (i) 2m-temperature, which impacts soil
317 humidity (through evaporation), snow cover extent and duration, and vegetation
318 development; (ii) precipitation, which in our study only impacts soil humidity and snow
319 cover, not vegetation (cf. Section 2.1), and (iii) 10m-wind, on which dust emission
320 fluxes strongly depend (cf. Section 2). For the surface conditions, we examine (Fig. 8):
321 the dry fraction fd , the vegetation factor fv and the resulting erodible fraction $E = fd*fv$.

322

323 **3.3.1 The reference GS state**

324 We focus on the domain for which we performed the dust calculations: 45°-55°N, 15°-
325 35°E, and on the resulting dust emission band shown in Fig. 4. In the reference GS
326 state, the average temperature over the investigated domain follows a north-south
327 gradient, with values ranging approximately from -4° to 6°C (Fig. 7a). This leads to a
328 faster snow melting and an enhanced surface evaporation in the southeast (SE) part
329 compared to the northwestern (NW) part of the emissions band (not shown).

330 Precipitation averages are between 1 and 1.5 mm/day, slightly lower in the SE (Fig. 7d).

331 These combined factors give better conditions for emission with respect to soil humidity
332 in the SE of the band. Thus, the calculated surface dry fraction fd is between 50-70% in
333 this region, and decreases to only 20-40% in the NW part (Fig. 8a).

334 In our simulations, vegetation development is only determined by temperature. Hence
335 the onset of the growth season starts later in the NW of the emission band. Thus, on

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336 average over the February to June interval, the vegetation inhibiting effect is less
337 effective in the NW of the emission band (mean vegetation factor $f_v > 0.7$) than in the
338 SE ($f_v < 0.5$) (Fig. 8d). The two spots clearly appearing in Fig. 4a as preferential
339 emission areas have high f_v values: 0.6 – 0.7 for Spot 1, and more than 0.7 for Spot 2.
340 When calculating the erodible fraction E (Fig. 8g), combination of f_d and f_v , the
341 gradient in the dry fraction of the surface, f_d , prevails. Thus, on average over the dusty
342 season, E is lower in the NW of the deflation band (10-15% in Spot 2) than in the SE
343 (15-20% in Spot 1 and more than 25% south of Stayky).

344 Dust emission fluxes depend on the erodible fraction and the cube of 10m-wind speed,
345 combined at fine timescale (6h in our case). The average 10m-wind speed increases
346 from less than 5 m s^{-1} in the SE to more than 6 m s^{-1} in the NW (Fig. 7g). This increase
347 prevails in the flux calculation over the decrease of E , resulting in stronger dust
348 emission in the NW than in SE of the deflation band. Thus, as shown in Fig. 4, more
349 dust is emitted in Spot 1 than in Spot 2 on average over the year (or over the dusty
350 season; the total amount is practically the same). Both spots can be identified in Figs. 7d
351 and 7g as areas of relatively high 10m-wind speed and low precipitation in our
352 investigated domain. The region of relatively high erodible fraction (25-30%) south of
353 Stayky does not correspond to high emission, because the wind is not strong enough.

354 The differences of seasonality between Spots 1 and 2 (Fig. 6) can also be explained by
355 considering the spatial distribution of temperature and precipitation averaged over the
356 dusty season (Fig. 7a,d), and the general evolution in the investigated area of the wind
357 speed, soil humidity (both decreasing from winter to summer) and vegetation cover
358 (better developed in summer than in winter). In all states, it is colder in Spot 2 than in
359 Spot 1, located more to the south, while the average precipitation amount is quite
360 similar. Considering the temperature impact on soil humidity and vegetation, this
361 explains why the emission period is delayed in Spot 2 compared to Spot 1 (by 1 month;
362 Fig. 6). It also explains why in the cold GS and HE states Spot 2 is most active a month
363 later than Spot 1, in May, in spite of the gradual decrease of the average 10m-wind from
364 winter to summer. The wind weakening is compensated for by a combination of drier
365 surface and vegetation developing later than in Spot 1 (where the maximum emission is
366 in April).

367

368 | **3.3.2 Changes of climate and surface variables in the “H-stadial” cold**
369 **perturbation, and consequences on dust emission**

370 In the HE experiment, the lower North-Atlantic SSTs imposed in the latitudinal band
371 30°-63°N result in an average cooling over the dusty season of 0.5 to 2°C in our
372 investigated domain (Fig. 7b), the anomaly being strongest in its W-NW part.
373 Precipitation only locally decreases, and by a small amount compared to the reference
374 GS state (Fig. 7e). The combination of these two factors increase the contrast in soil
375 humidity between the NW and the SE of the emission band, compared to GS (Fig. 8a,b).
376 Thus, the dry fraction fd decreases by up to 8% in the NW, but increases in the SW by
377 up to 6% (Fig.8b).

378 In our experiments, a delay in vegetation development and lower average vegetation
379 cover than for GS are straightforward consequences of the lower HE temperatures.
380 Thus, the vegetation factor f_v (anti-correlated with the vegetated soil fraction, as defined
381 in Sect. 2) is everywhere slightly higher than for GS (Fig. 8e). The resulting erodible
382 fraction anomaly is positive almost everywhere (Fig. 8h). The surface conditions are
383 thus better for deflation than in the GS state, but the average wind slightly decreases
384 compared to GS over most of the deflation band (Fig. 7h). The combined effect (at fine
385 timescale, here 6 hours) of these opposing variations on the dust emission change
386 between HE and GS is contrasted along the deflation band: from a strong decrease in
387 the NW to a slight increase in the SE (Fig. 9e). The HE fluxes are 50-80% of the GS
388 ones in Spot 2, and 70-100% in Spot 1. Both spots are still well identified as the most
389 active areas in the deflation band, with yearly average dust fluxes of up to $120 \text{ g m}^{-2} \text{ yr}^{-1}$
390 (Fig. 4b). The relative increase of emission fluxes south and east of Stayky is due to the
391 increase of the erodible fraction by more than 4%, in a zone where E was already high
392 for GS (20-25%). Nevertheless, the average winds are relatively weak, implying a low
393 frequency of significant emission events, so the average fluxes remain low ($< 60 \text{ g m}^{-2}$
394 yr^{-1}).

395

396 **3.3.3 “Stadial-Interstadial” changes of climate and surface variables, and**
397 **impact on dust emission**

398 We now analyze the effect of a North-Atlantic SST increase similar to that associated
399 with a Dansgaard-Oeschger warming event. The imposed SST perturbation results in an

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400 average temperature increase from 1.5° in the SE of our investigated domain to more
401 than 3°C in the NW (Fig. 7c). As in the case of the cold perturbation, there is little
402 change in precipitation (Fig. 7f). The resulting fd anomaly is positive almost
403 everywhere in the emission band (Fig. 8c), and higher in the NW (more than 8%) than
404 in the SE (up to about 4%). This anomaly distribution reduces the SE-NW contrast of fd
405 compared to the GS state.

406 The warmer climate favors vegetation development, so that the vegetation factor fv
407 decreases everywhere in the domain, by 0.1 to 0.2 in the emission band (Fig. 8f).

408 The resulting E anomaly is also negative everywhere (Fig. 8i). The average 10m-wind
409 speed decreases as well, more in the NW than in the SE of the domain, which attenuates
410 the NW-SE wind-speed gradient along the emission band compared to the GS state
411 (Fig. 7i). All these lead to a general decrease of the emission fluxes, which are now
412 mostly between 40-100 g m⁻² yr⁻¹ in the main spots, about half of the GS values (80-160
413 g m⁻² yr⁻¹). The decrease is stronger than in the HE experiment in the eastern half of the
414 band, including Spot 1 (Fig. 4c).

415

416 **3.3.4 The contribution of vegetation in modulating dust emission during** 417 **the North-Atlantic abrupt changes**

418 The Sima et al. (2009) study has shown that, for the main deflation areas of Western
419 Europe, stadial-interstadial changes in wind, precipitation, soil moisture and snow cover
420 did not produce changes in dust emission as important as indicated by the sedimentation
421 changes seen in the loess profiles. It was mainly the vegetation, by its effect of
422 inhibiting the aeolian erosion, which modulated the dust emissions in response to
423 climate variations. The inhibition was considerably more effective in the relatively
424 warmer GSI state (due to a better developed vegetation) than in the cold GS and HE
425 states. In order to assess the importance of this mechanism in the area investigated here,
426 further away from the North Atlantic region, in which the abrupt climate changes
427 originate, we analyze annual mean emission flux ratios HE/GS and GIS/GS in the
428 absence (Fd , Fig 9a-c) and in the presence of the vegetation effect (F , Fig 9d-f).

429 When only taking into account the effects of surface wind and precipitation (including
430 soil humidity and snow cover), dust emission occurs almost everywhere in our domain
431 (Fig. 9a). Annual mean dust fluxes (Fd) in GS locally exceed 220 g m⁻² yr⁻¹ in the two

432 most active spots. The HE/GS and GIS/GS flux ratios are quite similar in our band of
433 interest, mostly between 80 – 100% (Fig. 9b,c), meaning there is little difference
434 between the perturbed and reference states. In the GIS case, these values are too high to
435 be reconciled with the strong stadial-interstadial deposition differences indicated by the
436 loess record. Locally, they are even higher than those for the cold HE perturbation.
437 When adding the vegetation effect in the dust flux computation, the GS annual mean
438 dust fluxes (F) strongly decrease compared to Fd (Fig. 9d). The values in our two main
439 spots are now generally between 80 - 160 g m⁻² yr⁻¹. The band north and northeast of the
440 Carpathians clearly appears as the main emission area. Here, the HE/GS flux ratio does
441 not change much: an increase of about 10% can be seen especially in the eastern part of
442 the domain (Fig. 9e). On the contrary, in the GIS case, a shrinking of the deflation area
443 and a significant reduction of fluxes can be seen (Fig. 9d). The flux reduction is
444 strongest in the most active spots, where GIS fluxes are now 50 – 70% of the GS ones,
445 in better (qualitative) agreement with the loess data.

446 The considerable difference between annual mean emission fluxes without (Fd , Fig. 9a)
447 and with vegetation effect (F , Fig. 9d) is mainly due to the shortening of the emission
448 season, as shown by the Fd and F annual cycle averaged over each of the main Spots
449 (Fig. 10). Without vegetation, emission would occur all the year round, whereas taking
450 the vegetation effect into account restrains the emission to late winter and springtime.
451 The same was true for the main deflation areas of Western Europe: the English Channel
452 and the south of the North Sea (Sima et al., 2009, Fig. 5c therein). There are also some
453 differences. In the western European areas, in all three simulated states, the monthly
454 mean Fd was highest in May, month during which the attenuation of emission by the
455 developing vegetation was also strong. Taking this effect into account resulted in a
456 maximum emission flux F in April for GIS and GS. In Spot 1, Fd has similar values
457 over the dusty season for the three states, and reaches its maximum in April, one month
458 earlier than at the western sources. The vegetation effect in this month is considerably
459 weaker here than in the western sources (so that the maximum of emission flux F
460 remains in April), but is strong enough to differentiate the warm perturbation from the
461 cold states. Spot 2 is in an intermediary situation: both Fd and F reach their maximum
462 in May for GS and HE, and in April for GIS. Fd is higher for GIS than for HE, and both
463 are smaller than for GS. It is the vegetation effect that makes the GIS fluxes become
464 smaller than the HE ones.

465 4 Discussion

466 Our climate simulations and dust calculations bear some limitations and are idealized in
467 a number of aspects. In the few years since we have run them, new efforts have been
468 made towards better understanding various aspects of the abrupt climate changes, for
469 example, the sub-millennial structure of DO events (e.g., Capron et al., 2010), the
470 mechanism of stadial-interstadial oscillations (e.g., Arzel et al., 2012) or the Heinrich
471 event scenario (Alvarez-Solas and Ramstein, 2011). However, to date, a complete set of
472 sea-surface conditions does not exist for a sequence GS-H-GIS around an H event
473 occurring between the beginning of the main loess sedimentation period in Europe (~40
474 kyr BP) and the end of MIS 3 (~25 kyr BP); neither reconstructed, nor from coupled
475 model simulations.

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476 We use LGM SSTs and sea-ice extent for the reference “stadial” state, which otherwise
477 is designed to correspond to 39 kyr BP. The sea-surface conditions follow a seasonal
478 cycle, but which does not change from one year to another. This lack of interannual
479 variability in the boundary conditions could affect the representation of extreme wind
480 events. As in most studies, no change of ice-sheet size and extent (and consequent
481 adjusting of sea level) associated with the DO and H events are represented. The SST
482 anomalies we apply in the North Atlantic in order to obtain the DO- and H event-like
483 perturbations are highly idealized and only depend on latitude. Nevertheless, as
484 thoroughly discussed in Sima et al. (2009), our experiment design allows us to test the
485 impact on dust emission of changes in the North-Atlantic sea-surface conditions as
486 those suggested by data for DO and H events. With this simple set-up, the perturbations
487 can be ascribed to the SST anomalies over the North Atlantic only, and not to SST or
488 sea-ice differences elsewhere.

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489 The relatively small differences of average wind and precipitation between the
490 simulated climate states are a consequence of the imposed zonal SST anomalies of only
491 up to 2°C. While the maximum anomaly of 2°C is set according to data, a more realistic
492 distribution of SST anomalies and of the resulting sea ice might increase these
493 differences. However, they would probably still not reach those obtained in other
494 numerical experiments employing very contrasted boundary conditions between
495 stadials, interstadials and H-stadials (e.g., Hostetler et al., 1999; Renssen and Bogaart,
496 2003).

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497 For forthcoming AGCM studies, an alternative to using reconstructed SSTs and
498 prescribed perturbations would be to employ the output of a coupled global climate
499 model (atmosphere – ocean - sea ice – land), after regridding at the finer resolution
500 generally required for the AGCM. This alternative, which would solve the interannual
501 variability issue, is certainly worth exploring, especially since coupled atmosphere-
502 ocean-sea-ice general circulation model experiments have started to address the MIS3
503 period (Merkel et al., 2010; Brandefelt et al., 2011). Such simulated MIS3 sea-surface
504 conditions would be more coherent with the rest of the numerical setup, and provide a
505 less idealized distribution of SST anomalies. One should keep in mind however that
506 they come with the model biases, and, cf. Brandefelt et al. (2011), are quite different
507 from one model to another.

508 An important limitation of our simulations concerns the vegetation treatment. In the
509 main deflation areas of Western Europe we have imposed a glacial-type vegetation
510 consistent with available paleodata (e.g., (Woillard, 1978; de Beaulieu and Reille, 1984,
511 1992; Rousseau et al., 1990; Hatté et al., 1998; Peyron et al., 1998; Müller et al., 2003;
512 Hatté and Guiot, 2005; Moine et al., 2008), only composed of boreal evergreen
513 needleleaf trees (up to 1% of a grid cell) and C3 grass (up to 80%). In the Eastern
514 Europe, the maximum fractional cover and the LAI limits for each PFT are prescribed
515 to present-day values, as for the LGM PMIP experiments. As mentioned by Woillez et
516 al. (2011), the present-day European vegetation includes considerable areas of
517 agricultural grass, therefore the landscape is not so different from the glacial one,
518 mainly represented by steppe or steppe-tundra. In our simulations, trees occupy less
519 than 10% of any given grid cell of the main emission band (Fig. 9d). Grass takes on
520 average on the dusty season 20-35% of each grid cell in the GS state, 25-50% in the
521 GIS state and 15-25% in the HE state, the rest of the cell being left to bare soil. Such
522 vegetation composition seems reasonable for the time slice we approach, at about 40
523 kyr BP, compared to the steppe or steppe-tundra predominating in Europe at the LGM.

524 In the NE of the domain, outside of the main band, some dust emission would occur as
525 well if vegetation were not accounted for (Fig. 9a). Here, grid cells are occupied all-
526 year-long by up to 30% trees. For the cold Greenland episodes, this might be an
527 overestimation, but we think it has no significant impact on our results: the differences
528 between the dust fluxes calculated without vs. with vegetation effect (Fig. 9) are the
529 direct consequence of the fact that each grid cell is partly covered by vegetation, no

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530 matter whether trees or grass (an effect expressed by the vegetation factor f_v). Also,
531 even for the warmest simulated state, GIS, the total vegetation fraction averaged over
532 the dusty season does not exceed 50% of a grid cell in most of the domain investigated
533 here, which is still coherent with a steppe-tundra environment.

534 The adjustment of vegetation to the climate conditions is only determined by
535 temperature in the configuration of ORCHIDEE used in this study. The glacial climates
536 we investigate were not only colder, but also drier than today in our area of interest.
537 More realistic simulations should also include the precipitation impact on vegetation, as
538 well as the effect of a lower atmospheric CO₂ concentration in glacial times than today.
539 However, it is difficult to validate simulated vegetation over our area of interest for the
540 main loess sedimentation period, due to the scarcity of pollen records compared to other
541 parts of Europe or glacial time slices. In the frame of the Stage 3 project (Barron and
542 Pollard, 2002), palynological data compiled from the four sites falling in our
543 investigated domain suggest tundra and temperate grassland around 50°N-20°E for the
544 interstadials, but give no information for the stadials (Huntley et al., 2003, et references
545 therein). For the Stayky area (approx. 50°N-30°E), Gerasimenko and Rousseau (2008)
546 indicate a transition from a forest-steppe environment before ~40 kyr BP to steppe
547 during the main loess sedimentation period, with arboreal pollen varying between ~10%
548 in the loess units and ~40% in the paleosols. The few simulations of the MIS3
549 vegetation, which could be used for comparison, either address the earlier part of MIS3,
550 with little loess sedimentation (e.g., GS12, at ~44 kyr BP, for Kjellström et al., 2010, or
551 GS15-GIS14, at ~55 kyr BP, for Van Meerbeeck et al., 2011), or give results in
552 discrepancy with the data on our area of interest (in particular for the tundra extent in
553 central Europe; Alfano et al., 2003, Huntley et al., 2003).

554 In the dust emission calculations, by choosing the erosion wind threshold close to the
555 lowest possible values (observed or derived as a function of soil characteristics), we aim
556 to include all possibly important dust sources in our domain of study. However, using
557 constant values for the threshold wind and the erosion potential implies homogeneous
558 soil characteristics (obstacles and mineralogy), which is not very realistic. In the general
559 case, the intensity and possibly even the location of the most active emission spots
560 would be affected by taking into account the surface inhomogeneity, which implies
561 variations of erosion threshold and potential across an investigated region. In our case,
562 the main emission band determined by the climate-related conditions does correspond to

563 surfaces favorable to deflation. Moreover, the erosion potential in this band decreases
564 from NW towards SW, so, if taken into account, it would accentuate the emission flux
565 gradient in Fig. 4a. Thus, Spot 2 falls in a roughly flat zone of Tertiary sediment, with
566 high erosion potential. Spot 1 lies in a more complex area with Cretaceous sedimentary
567 rocks in the western part, while in the eastern part, Neogene and less erodible
568 Precambrian rocks are mixed (Asch, 2005).

569 We note that the thickest European deposits are generally located along major river
570 valleys (of the Seine, the Rhine, the Danube, or the Dnieper). In glacial times, these
571 valleys used to be almost dried-out most of the year. Rich in sands and silts transported
572 by the rivers during the snow-melting period, they constituted important deflation areas.
573 Where the relief context favored the retention of the coarse deflated material, thick
574 deposits have formed within a short distance downwind (e.g., Antoine et al., 2001;
575 Smalley et al., 2009). This explains, the exceptional thickness (for Europe) of the loess
576 deposits at Nussloch (~13.5m for the 40-15 kyr BP interval in the P4 sequence; Antoine
577 et al., 2009), on the eastern bank of the Rhine valley, in the context of prevailing
578 westerly winds. In general, even though periglacial braided rivers used to be important
579 local sources for the coarse material in some of the European loess deposits, such details
580 cannot be captured at the resolution of an AGCM.

581 For the Stayky area, the prevailing winds are from west-northwest (Fig. 3; see also
582 Rousseau et al., 2007). Due to the relief configuration east of the Dnieper (a plain well
583 exposed to wind erosion), no loess deposit has formed downwind in the close vicinity of
584 the valley. The nearest loess deposits are located on the west bank of the river, and
585 contain little of the easily deflatable coarse material from the valley, brought by rare
586 strong easterly winds. In our reference sequence, the stratigraphic units corresponding
587 to the 40-15 kyr BP interval only add up to ~6.5m thickness. Thus, while both Nussloch
588 and Stayky sites have recorded millennial climate variations, their sensitivity to the
589 climate signal depended on the local relief context. At Stayky, without a strong local
590 source upwind, the relative contribution of more remote sources as those we identify
591 here must have been higher.

592 The emission flux calculations use 6-hourly winds, but even this high time series
593 frequency does not capture the shorter episodes of strong wind, which mainly control
594 the total amount of emitted dust. A way to compensate for that would be to lower the
595 emission threshold. Changing this threshold from the 7 m s^{-1} value used here to 6 m s^{-1}

596 obviously increases the mean annual flux (not shown), but only slightly widens the
597 main emission areas, and does not affect the location of the most active spots or the
598 relative differences between the simulated climate states.

599 The simulated monthly mean 10m-winds in our investigated domain during the dusty
600 season (Fig. 6) are in agreement with the W-NW wind direction inferred from field
601 observations by Rozycki (1967) and Léger (1990). These studies describe ridge-like
602 features called gredas, elongated in the main wind direction, varying from NW-SE to N-
603 S around the Carpathians.

604 Considering the predominant wind direction, and the distance to our reference site, Spot
605 1 is particularly well placed as a source for the Stayky deposits. Dust calculations only
606 taking into account the wind and soil humidity conditions give emission fluxes by 10 to
607 30% lower in the warm GIS perturbation than in the GS reference state. Adding the
608 vegetation effect increases the difference by another 10 to 20%.

609 Spot 2 is the largest and most intense deflation area of the simulated emission band in
610 the reference GS experiment, without as well as with the vegetation effect. In the GIS
611 simulation, dust fluxes are only by up to 20% smaller than in the reference state before
612 applying the vegetation inhibition factor. The vegetation effect further reduces them by
613 20-30%.

614 The vegetation effect not only determines a strong decrease of the GIS emission fluxes
615 compared to the GS ones, particularly in the most active spots, but also decreases the
616 size of the band where significant emission occurs (Figs. 4, 9). Without a transport and
617 deposition model, the impact on the sedimentation rates cannot be calculated.
618 Nevertheless, as the simulated slightly lower precipitation and slightly stronger winds in
619 the GS and HE states favor the transport compared to the GIS state, we may reasonably
620 suppose that considerably more emitted dust would lead to considerably more
621 deposition during the cold North-Atlantic episodes than during the relatively warmer
622 ones, in agreement with the loess data. For example, at Nussloch (Germany), stadial
623 loess sedimentation rates are up to 5 times higher than the interstadial ones (Rousseau et
624 al., 2007). Thus, the key role of vegetation in modulating stadial-interstadial dust
625 emission variations is confirmed.

626 Also, in both main spots, the GIS surface winds are lower than the GS ones not only on
627 average over the dusty season (Fig. 7i), but also on average over each month of this

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628 season (not shown). This result is consistent with the grain-size variations in the Stayky
629 loess profile, indicating a coarser sedimentation in stadial than in interstadial episodes.

630 Concerning the H-stadials, our modeling experiments suggest a reduction of dust
631 emission with respect to a stadial state. When only taking into account the wind and soil
632 humidity effects, the simulated emission flux decrease is even stronger locally than for
633 the interstadial. Including the effect of vegetation, less developed in a colder climate,
634 attenuates the difference of emission fluxes between a stadial and an H-stadial (whereas
635 it amplifies the stadial-interstadial differences, as seen above). In our experiments, the
636 flux ratio HE/GS is up to 10% higher with than without the vegetation effect, but the
637 HE fluxes remain smaller than the GS ones. This is somehow counterintuitive, because
638 colder climates are associated with higher loess sedimentation rates, generally
639 interpreted as a result of stronger winds and dryer conditions, favoring both the
640 emission and the transport of dust. This is certainly true for “cold and dry” vs. “warm
641 and humid” climates, like glacial (loess sedimentation) vs. interglacial (no loess
642 sedimentation) or, at a finer timescale, stadial (high loess sedimentation rate) vs.
643 interstadial (reduced or no sedimentation).

644 The emission attenuation suggested by our experiments for an H-stadial compared to a
645 stadial state can be understood if we think of difference between stadial and H-stadial
646 as a change from “cold and dry” to “colder and drier”. Indeed, the lower dust emission
647 fluxes in our HE experiment than in the GS one are associated with lower precipitation
648 and weaker winds, the former favoring the dust transport, the later hindering it. Again, a
649 transport and deposition model would be needed to determine the net effect on the
650 sedimentation rates in the investigated domain, and more specifically at the reference
651 site. But even if we used such a model, loess stratigraphy offers no element to confront
652 the results, as there is practically no way to distinguish between dust layers deposited at
653 different rates in similarly dry conditions.

654 In the case of interstadials, the lower emission activity is associated with wetter soil
655 conditions, favorable to pedogenesis, and the resulting soils (well developed or in
656 embryonic form only), are distinguishable in the sediment (Rousseau et al., 2007;
657 2011). In contrast, only exceptionally it is possible to find in stratigraphic profiles
658 particular features susceptible to be associated with H events. It is the case of the
659 millimetric sandy laminations identified in particular loess units at the Nussloch loess
660 site, in Germany, resulted from a combination of strong wind events and coarser

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661 deposition (Lautridou et al., 1985; Derbyshire and Mellors, 1988). Otherwise, loess
662 studies suggest that H events only could be associated to peaks in some very detailed
663 grain-size index records (Porter and An, 1995; Antoine et al., 2001, 2009; Rousseau et
664 al., 2002, 2007). In theory, if such records had a fine enough resolution, and the
665 different variations could be dated with a reasonable precision, it would be possible to
666 distinguish the sedimentation rates corresponding to the different climate episodes. In
667 practice, to date, no loess profile allows such quantitative estimations. Qualitatively, as
668 the sandy laminations, the grain-size peaks (reflecting coarser deposition) are
669 interpreted as indicating episodes of particularly strong wind. Such very strong winds
670 are able to bring more medium-to-coarse material from the nearby sources to the
671 considered deposition site, while from the remote sources still only finer material can
672 travel the longer distance. Thus, the coarser deposition also reflects an increased relative
673 contribution of the nearby vs. remote emission areas to the sedimentation at a given site.

674 Looking at the numerical results from this perspective, we note that for Spot 1, close to
675 Stayky, the monthly mean emission fluxes are the highest in the month of April of the
676 HE state (Fig. 6). Also, the average wind in April for HE in Spot 1 is directed eastward
677 at the 850hPa level as well as at 10m (not shown). So, it is in the HE state, during this
678 particular month, that Spot 1 may have the highest contribution to dust deposition at
679 Stayky of all months and analyzed climate states. In addition, in the grid cell
680 corresponding to Stayky, the few strongest 10m-wind events over the year, exceeding
681 20 m s^{-1} on average over 6h, also occur in April (and in December, but this is outside
682 the dusty season). Even though in this cell the emission dust flux in HE's month of
683 April is lower than in the main emission spots, $20\text{-}25 \text{ g m}^{-2} \text{ month}^{-1}$ only, the proximity
684 to the deposition site makes it an important potential contributor to the Stayky
685 sediments. Thus, our modeling results support the identification of H events in loess
686 sequences as peaks of grain-size index.

687

688 **5 Conclusions**

689 Following the Sima et al. (2009) study on the impact of North-Atlantic abrupt climate
690 changes on dust emission in Western Europe, and the correlation proposed by Rousseau
691 et al. (2011) between Greenland, West and East European dust records, we have
692 focused here on the Eastern European dust sources. The same simulations have been

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693 used, including a reference “Greenland stadial” experiment GS, and two perturbations,
694 seen as sensitivity tests with respect to changes in the North-Atlantic surface conditions:
695 a “Greenland interstadial” GIS and a “H-stadial” HE (i.e., the cold climate episode
696 associated with a Heinrich event). We have combined results from these numerical
697 experiments and dust emission calculations, with information from the loess site of
698 Stayky, in Ukraine.

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699 A band stretching north and northeast of the Carpathians appears as an important
700 deflation area, potential source for the eastern European loess deposits located around
701 50°N latitude. Two spots are particularly active, one in Ukraine (Spot 1), the other in
702 Poland (Spot 2). Located west-northwest from Stayky, they are well placed to be the
703 main dust sources for our reference site.

704 The general conclusions of the previous study on the Western Europe (strong
705 seasonality of emissions, difference of dusty season from one climate state to another,
706 higher emission fluxes in the “stadial” than in the “interstadial” state, importance of the
707 vegetation) are found to apply to Eastern Europe as well.

708 In the deflation band identified here, taken as a whole, emissions mainly occur from
709 February to June in the GS experiment (compared to February-May in the West), from
710 March to June in the HE experiment (same in the West), and from February to May in
711 the “GIS” simulation (February-April in the West). Thus, the beginning of the dusty
712 season, constrained by soil humidity and snow conditions, is the same for East and
713 West, while the end, determined by vegetation development, is slightly later in the East
714 in the GIS and GS states. The resemblances are due to the fact that in our simulations
715 there are no strong differences of precipitation or temperature (the main variables
716 impacting the continental surface conditions) between West and East along the ~50°N
717 latitudinal band where the main deflation areas are located. The differences are mainly
718 due to a delay in vegetation development in Spot 2, still allowing some emissions later
719 than in the other investigated sources.

720 Indeed, a more detailed analysis for the area investigated here shows differences: (a)
721 between Spots 1 and 2 for each climate state, and (b) from one climate state to another
722 for each spot individually. In our simulations, they are caused by the differences of
723 temperature (indirectly, via the impact on soil humidity and vegetation): due to the

724 geographical position in the case (a), because Spot 2 is more to the north, closer to the
725 ice sheet, and due to the imposed North-Atlantic SST anomalies in the case (b).

726 Furthermore, in the main deflation band in Eastern Europe, emission fluxes in the GIS
727 experiment are 50-70% of the GS ones (the ratio was less than 10% for the English
728 Channel area, and 10 to 80% for the area south of the North Sea, including the exposed
729 continental shelf). The vegetation, better developed in the warmer climate, and thus
730 protecting the soil more efficiently from aeolian erosion, is responsible for about half of
731 the flux difference. Its contribution in modulating the response of dust emission
732 intensity to the North-Atlantic millennial variability is less important than in the main
733 western European sources, but still significant. The simulated weaker winds and slightly
734 higher precipitation in interstadial conditions suggest less favorable conditions for
735 transport than in a stadial. The modeling results are thus qualitatively consistent with
736 the stadial-interstadial sedimentation variations in the Stayky loess profile, and in the
737 European loess sequences in general.

738 In the HE experiment, emission fluxes are generally lower than the GS ones. The
739 simulated climate is slightly drier, but also slightly less windy over the region studied
740 here. A transport and deposition model would be needed to evaluate the resulting
741 change of average sedimentation rate at a loess site; the resolution and dating
742 uncertainties of the available loess profiles do not allow a comparison with such a result
743 anyway. A more detailed analysis than in the previous study allows nevertheless to
744 investigate the hypothesis suggested by some loess data studies, i.e., that H-stadials
745 could be identified in some of the most detailed loess profiles as peaks of the grain-size
746 index. Such peaks represent brief intervals of coarser sedimentation due to particularly
747 strong winds, increasing the relative contribution of the nearby vs. remote sources. Our
748 simulations support this interpretation, pointing to the month of April of the HE
749 experiment as the month with strongest winds in the immediate vicinity of Stayky,
750 where some dust mobilization occurs, and highest dust emission in the main deflation
751 spot 1, only a few hundred kilometers away, associated with dominant 850 hPa winds
752 directed towards the Stayky area, susceptible to transport more relatively coarser
753 material to our reference site.

754 This study proposes another way to put together loess data and climate simulations to
755 critically assess the modeling results, and test data interpretation. Investigating
756 mechanisms and regional details strongly benefits from the “zoom” capacity of the

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Supprimé: Considering the identified deflation band as a whole, dust emission mainly occurs from February to June in the “Greenland stadial” GS experiment, from March to June in the “Heinrich event” HE experiment, and from February to May in the “Greenland interstadial” GIS simulations. The beginning of the dusty season is constrained by soil humidity and snow conditions, while the end is determined by vegetation development.

In each simulated climate state, the dusty season in Spot 1 is one month earlier than for Spot 2. This happens mainly because Spot 1 is located more south, farther away from the ice sheet, so that air temperature is higher on average. Therefore, the soil dries earlier in the year, allowing dust emission to begin, but the vegetation also develops earlier and reaches more rapidly the critical threshold above which emission is completely inhibited.

The same mechanisms are responsible for the differences of dusty season between the different simulated climate states. The main cause is again the difference of temperature, which comes, in this case, from the imposed changes in the North Atlantic sea-surface conditions.

In the main deflation band, emission fluxes are by 30 to 50% lower in the GIS experiment than in the GS one. About half of the emission flux difference is due to the vegetation, which is better developed in the warmer climate, and thus protects more efficiently the soil from aeolian erosion. This confirms the key role of vegetation in modulating the response of dust emission intensity in Europe to the North-Atlantic millennial variability. Furthermore, the simulated weaker winds and slightly higher precipitation in interstadial conditions suggest less favorable conditions for transport than in a stadial. Our modeling results are thus qualitatively consistent with the stadial-interstadial sedimentation variations in the Stayky loess profile, and in the European loess sequences in general.

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Supprimé: As said above, the temporal resolution of European loess profiles is not very high; in the best cases it can get below a century for certain layers of particularly thick sediments. On the modeling side, quite a number of hypotheses on the boundary conditions must be made to simulate MIS3 climates. Also, the horizontal resolution of most AGCM experiments makes it difficult to correlate data from particular sites with numerical results.

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Supprimé: , using the same numerical means and methods as the one dedicated to Western Europe (Sima et al., 2009), and an approach adapted to Eastern Europe,

757 LMDZ AGCM, and from analyzing the results at timescales ranging from yearly
758 averages down to high frequency time series (6h in our case).

759 For our future simulations we will consider two main changes: forcing the AGCM with
760 sea-surface conditions issued by MIS3 simulations with a coupled ocean-atmosphere
761 model, instead of the GLAMAP dataset for the LGM, and imposing a vegetation
762 distribution consistent with the simulated glacial climates instead of the present-day
763 distribution. Also, the effect of precipitation will be taken into account along with that
764 of temperature in computing the vegetation changes. We also plan to simulate the entire
765 dust cycle (emission, transport and deposition), in view of a more quantitative
766 comparison to European loess data.

767

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771 performed using HPC resources of the Commissariat à l’Energie Atomique, France.

772

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Supprimé: Taking advantage of recent modeling advancements and results can certainly improve this work.

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1001 Figure captions

1002 **Figure 1.** Map of the thickest European loess deposits (in yellow), in the context of the
1003 Last Glacial Maximum (21 kyr BP) ice sheets (light blue) and sea level (modified from
1004 Antoine et al., 2013, based on data from compilations kindly provided by D. Haase
1005 from Haase et al., 2007, and J. Ehlers from Ehlers et al., 2011). Blue/gray colors
1006 indicate depth/elevation with respect to the actual sea level.

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1007 **Figure 2.** Mean annual 850 hPa wind speed and direction for the reference GS state
1008 (modified from Rousseau et al., 2011). The area investigated in this study stretches
1009 between 15°-35°E and 45°-55°N.

1010 **Figure 3.** 10m-wind roses derived from 6-hourly 10m winds at Stayky for 20 years of
1011 simulation for each of the three climate states. Relevant winds for dust emission are
1012 those exceeding the threshold erosion wind speed, 7 m s⁻¹ in this study. Winds below 14
1013 m s⁻¹ are much more frequent than those above 14 m s⁻¹ (which cannot even be seen on
1014 the plots); cf. Sima et al (2009), they determine most of the emitted amount of dust.

1015 **Figure 4.** Annual means of dust emission fluxes F (g m⁻² yr⁻¹) for the three simulated
1016 climate states: GS (left), HE (middle), and GIS (right). The flux calculation includes the
1017 vegetation effect. Emission mainly occurs in a NW-SE band, with two most active
1018 spots, S1 and S2.

1019 **Figure 5.** Annual mean of dust emission fluxes for GS (Fig. 4a) superimposed on the
1020 topographic map.

1021 **Figure 6.** Monthly means of dust emission fluxes outside the Carpathians in the three
1022 simulated climate states for January to June (for each panel, the x-axis represents
1023 longitude (°E), and the y-axis, latitude (°N)). Wherever the slightest emission occurs,
1024 the monthly average wind vectors at 850 hPa indicate the direction in which the dust is
1025 most likely transported. Little or no dust is emitted in this area in the rest of the year.

1026 **Figure 7.** February to June averages of 2m-temperature (a-c), precipitation (d-f) and
1027 10m-wind (g-i) for the GS state (left column), and anomalies HE-GS (center column)
1028 and GIS-GS (right column). In white, areas where the differences are not significant at
1029 the 95% confidence level (Student's t-test).

1030 **Figure 8.** Averages over the dust emission period (February to June) for dry soil
1031 fraction f_d , vegetation factor f_v and erodible soil fraction E in the GS state (left column),

1032 | and anomalies HE-GS (center) and GIS-GS (right). Masked in white, the Carpathians
1033 | (altitudes exceeding 500m) and the lowlands inside the mountain arch (cf. Section 3.1).

1034 | **Figure 9.** Mean annual dust fluxes in the reference climate state GS (left) and ratios of
1035 | dust fluxes HE/GS (center) and GIS/GS (right), without (a-c) and with (d-f) vegetation
1036 | effect.

1037 | **Figure 10.** Annual cycle of emitted dust flux averaged on each of the main deflation
1038 | spots in the three simulated climate states, without (F_d) and with vegetation effect (F)