

1 **Analysis of Saharan dust intrusions into the Carpathian Basin (Central Europe) over**
2 **the period of 1979–2011**

3

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18

19 **Abstract**

20

21 Aeolian dust particles and dust storms play substantial role in climatic and other

22 environmental processes of the Earth system. The largest and most important dust source

23 areas are situated in the Sahara, from where several hundred thousand tons of mineral dust is

24 emitted each year and transported towards the European continent. Here we show that 130

25 Saharan dust events (SDEs) reached the atmosphere of the Carpathian Basin from 1979 to

26 2011 by using the NASA's daily TOMS Aerosol Index data, satellite images and backward
27 trajectory calculations of NOAA HYSPLIT model. Monthly trends of dust events demonstrate
28 that the main period of dust transportation is in the spring, with a secondary maximum in the
29 summer (in July and August). This seasonal distribution match well the seasonality of
30 Saharan dust emissions. However synoptic meteorological conditions govern primarily the
31 occurrence of long-range dust transport towards Central Europe. Based on their different
32 meteorological backgrounds (geopotential field, wind vector and meridional flow), SDEs
33 were classified into three main types. By using composite mean maps of synoptic situations
34 and backward trajectories, the possible source areas have also been identified for the different
35 types of events. Finally, we provide a short discussion on how the African mineral dust could
36 contribute to the local aeolian sedimentation of the Carpathian Basin during the Plio-
37 Pleistocene.

38 39 **Highlights:**

40
41 130 Saharan dust events (SDEs) were identified in the Carpathian Basin (CB) atmosphere
42 from 1979 to 2011
43 Spring and summer are the typical seasons of SDEs
44 SDEs can be classified in three main types based on synoptic meteorological conditions
45 Saharan dust could have played a role in loess and red clay sedimentation of the CB during
46 the Plio-Pleistocene

47
48 **Key words:** mineral dust; dust storm; meteorology; Sahara; Carpathian Basin;

49 50 **1. Introduction**

51

52 Dust storms and related atmospheric mineral particles have been in the focus of
53 environmental and climatic studies for the last two decades (Stout et al., 2009). Previous
54 investigations confirmed that windblown dust is an active component of the climate system,
55 and can modify its elements via both direct and indirect effects (Harrison et al., 2001; Kohfeld
56 and Tegen, 2007; Maher, 2010; Pósfai and Buseck, 2010). Dust particles affect the Earth's
57 energy balance directly through absorption, scattering and reflection of incoming shortwave
58 and outgoing longwave radiation or by changing the albedo of (bright) surfaces (e.g. Arimoto,
59 2001). Indirectly, by acting as cloud condensation nuclei, mineral particles have also an effect
60 on atmospheric moisture balance (Rosenfeld et al., 2001; Sassen et al., 2003). Particles rich in
61 Fe have major impact on iron-limited oceanic ecosystems, and thus, dust can influence the
62 primary phytoplankton production and the carbon cycle through biogeochemical interactions
63 (Ridgwell, 2002).

64 The global annual input of mineral dust deflated from arid-semiarid areas can be set in the
65 range between 1 and 3 billion of tons (Tegen et al., 1996; Mahowald et al., 1999, 2006;
66 Ginoux et al., 2001). Most important sources are situated in Saharan and Sahel regions, which
67 are responsible for 50–70% of the global emission (Ginoux et al., 2001; Miller et al., 2004).
68 Four main pathways of Saharan dust transport can be distinguished: (1) southward to Gulf of
69 Guinea; (2) westward over the North Atlantic Ocean; (3) eastward to Middle East; and (4)
70 northward to Europe (for more details, see Engelstaedter et al., 2006; Goudie and Middleton,
71 2006).

72 The several hundred thousand tons of dust derived from Saharan sources influence numerous
73 constituents of European environmental systems (D'Almeida, 1986; Prospero, 1996). During
74 heavy dust-outbreaks, atmospheric dust concentration often exceed PM_{10} standards of the
75 European Union in Spain (Rodríguez et al., 2001), in Italy (Matassoni et al., 2011) and in

76 Greece (Gerasopoulos et al., 2006), thereby affecting human health (Griffin et al., 2001). The
77 strongly alkaline dust particles increase the pH of precipitation, thus reduce the frequency of
78 acid rains (Roda et al., 1993; Rogora et al., 2004; Špoler Čanić et al., 2009). As proposed by
79 Psenner (1999), permanent Saharan dust contributions to low-alkalinity European lakes
80 prevented them to become acidic during the late twentieth century. The accumulated dust
81 particles are even capable of modifying soil properties of a given region (Yaloon, 1997). As
82 such, terra rossa soils in Portugal (Jahn et al., 1991), in Spain (Muhs et al., 2010), in Italy
83 (Jackson et al., 1982), in Croatia (Durn et al., 1999), in Greece (MacLeod, 1980) and in
84 Turkey (Atalay, 1997) have been shown to be an alteration product of local material and far-
85 travelled African mineral dust.

86 Fine-grained particles lift to higher levels of the atmosphere and have a long atmospheric
87 residence time up to a few weeks (Pye, 1987). Aeolian dust from North Africa can often be
88 detected in Europe's high-latitude areas e.g. in British Isles (Wheeler, 1986), in Germany
89 (Klein et al., 2010), in Scandinavia (Franzén, 1994; Barkan and Alpert, 2010) and even in our
90 study area, the Carpathian Basin (CB) (Central Europe) (Borbély-Kiss et al., 2004; Koltay et
91 al., 2006; Szoboszlai et al., 2009). Nowadays the CB is generally not regarded as a dusty
92 place, except for episodic dust storms related to cold fronts invading the region at the
93 beginning of the vegetation period in the early spring. Still, Central Europe is lying in the D1b
94 zone of the "Saharan dust-fall map" of Stuut et al. (2009), implying that recent Saharan dust
95 material can be incorporated into the soil system and may increase its fine silt content (Stuut
96 et al., 2009). However, dust activity of the region was much more significant during the Plio-
97 Pleistocene periods, as it is shown by thick aeolian dust deposits covering more than half of
98 the area (e.g. Pécsi and Schweitzer, 1993; Kovács et al., 2008, 2011; Újvári et al., 2010;
99 Varga, 2011). It has been recognized that mineral dust particles of these aeolian sediments
100 originate mainly from local sources (e.g. alluvial plains), and only the clay and fine-silt

101 fractions may be linked to Saharan sources (Rózycki, 1991; Rousseau et al., 2007; Újvári et
102 al., 2012), similarly to Italian loess deposits (Cremaschi, 1990a, 1990b).

103 The present paper is aimed at providing information on the frequency and seasonality of
104 recent Saharan dust intrusions that can reach the CB atmosphere. Besides, our goal is to
105 define the mean synoptic situations, typical transport pathways and source areas of this
106 airborne mineral dust material.

107

108 **2. Methods**

109

110 *2.1 Establishment of spatial and temporal changes of atmospheric dust using the TOMS*

111 *Aerosol Index*

112

113 Satellites represent the only data source with truly global coverage on most important dust
114 source areas and their emissions. For this study the Total Ozone Mapping Spectrometer's
115 (TOMS) aerosol data were employed to estimate atmospheric dust amount. The TOMS
116 aerosol index, as defined by the NASA/GSFC Ozone Processing Team, is a measure of how
117 much the wavelength dependence of backscattered UV radiation from an atmosphere
118 containing aerosols (Mie scattering, Rayleigh scattering, and absorption) differs from that of a
119 pure molecular atmosphere (pure Rayleigh scattering). Quantitatively, the aerosol index AI is
120 defined as

$$121 \quad AI = 100 \log_{10} \left(\frac{I_{360}^{meas}}{I_{360}^{calc}} \right), \quad (1)$$

122 where I_{360}^{meas} is the measured 360 nm TOMS radiance, and I_{360}^{calc} is the calculated 360 nm TOMS
123 radiance for a Rayleigh atmosphere (Herman et al., 1997). The TOMS sensors (on board of
124 different sun-synchronous NASA satellites) have the longest available global record (since

125 1978 November) with appropriate spatial (1×1.25 degree) and temporal (daily) resolution
 126 (Herman et al., 1997; Torres et al., 1998).
 127 Analyses of daily data-matrices were performed in MathWorks' MATLAB (R2007b)
 128 environment, while kriging of maps was processed in Golden Software SURFER 8 (Fig.1).
 129 The fractional data of 1993 and 1996 (caused by satellite failure), the periods with calibration
 130 problems of 2001–2004 (Kiss et al., 2007) and 2010–2011, and the four-yearly leap days (due
 131 to the matrix-operations) were excluded from the long-term mean mapping analyses (Table
 132 1). Some few dates have not been either available within other periods, so these were replaced
 133 by mean values of previous days.

134

135 Figure 1.

136

137 Table 1.

138

Data used	Satellite	Time-series	
01/01/1979–06/05/1993	Nimbus-7	$14 \times 365 + 126$	5236 days
06/05/1993–25/07/1996		No data	
25/07/1996–31/12/2000	EarthProbe	$4 \times 365 + 160$	1620 days
01/01/2001–31/12/2004		Calibration problems	
01/01/2005–31/12/2009	Aura/OMI	5×365	1825 days
01/01/2010–31/12/2011		Calibration problems	
01/01/1979–31/12/2011		$23 \times 365 + 286$	8681 days

139

140 *2.2 Identification of Saharan dust events (SDE) over the Carpathian Basin*

141

142 The daily TOMS AI values of the investigation area (45°–48.5°N, 16°–23°E) were
143 standardized following the work of Barkan et al. (2005):

$$144 \quad AI_{st} = \frac{(AI - AI_{mean})}{\sigma_{AI}}, \quad (2)$$

145 where AI_{st} is the daily standardized TOMS AI value, AI_{mean} is the yearly regional mean TOMS
146 AI and σ_{AI} is the standard deviation. Negative AI_{st} indicates below average values, whereas
147 positive values represent possible dusty episodes. During the SDE identification work
148 fractional data of 1993 and 1996, which have been excluded from mean mapping analyses,
149 were also employed. The periods of 2001–2004 and 2010–2011 could not be analysed in the
150 same way, due to calibration problems; for these intervals, the possible SDEs were identified
151 directly from daily TOMS AI maps.

152 A given supposed SDE is only accepted after being confirmed by satellite images of NOAA
153 AVHRR (Advanced Very High Resolution Radiometer – source:
154 <http://www.sat.dundee.ac.uk>), ESA Meteosat SEVIRI (Spinning Enhanced Visible and
155 Infrared Imager – source: <http://www.sat.dundee.ac.uk>) and Terra or Aqua MODIS (Moderate
156 Resolution Imaging Spectroradiometer – source: <http://modis.gsfc.nasa.gov>), and by
157 backward trajectory calculations of NOAA HYSPLIT (HYbrid Single-Particle Lagrangian
158 Integrated Trajectory) model (Draxler and Rolph, 2012; Rolph, 2012). The meteorological
159 input for the trajectory model was the NCEP/NCAR (National Centers for Environmental
160 Protection/National Center for Atmospheric Research) Reanalysis Project dataset (Kalnay et
161 al., 1996).

162

163 *2.3 Mean synoptic patterns and typical pathways of SDEs*

164

165 Daily geopotential height (at 700mb), wind vector and meridional flow data of the identified
166 SDEs were obtained from the NCEP/NCAR Reanalysis project (Kalnay et al., 1996). The

167 selection of 700mb level was based on previous analyses of long-range Saharan dust transport
168 episodes (Hamonou et al., 1999; Barkan et al., 2005; Dayan et al., 2007). These
169 meteorological data were classified according to their synoptic patterns, and composite mean
170 maps of each type were compiled using the Daily Mean Composite application of NOAA
171 Earth System Research Laboratory (<http://www.esrl.noaa.gov/psd/>). In order to identify the
172 typical dust transportation routes, the daily backward trajectories of all type SDEs were
173 plotted on multiple trajectory maps, using the NOAA HYSPLIT model (Draxler and Hess,
174 1997).

175

176 **3. Results and discussion**

177

178 *3.1 Frequency of Saharan dust intrusions into the Carpathian Basin*

179

180 During the investigation period of 1979–2011 130 Saharan dust episodes could be identified
181 in the CB atmosphere. The time series of annual number of SDEs are characterized by high
182 amplitude annual variations (Fig. 2). Outstanding values of 1984, 1985, 1992, 2000, 2001 and
183 2008 with 8 or more SDEs are in contrast with the years of 1981, 1991, 2003, 2004, 2006 and
184 2009, when the number of identified events was far below the average. The causal
185 relationships among the changing annual frequency of SDEs, Saharan and Sahel droughts,
186 climate teleconnections and atmospheric circulation patterns are not yet fully understood. In
187 some cases, the increased number of identified SDEs over the CB is coincident with Saharan
188 drought periods (e.g. in 1983–84), but in other cases this relationship become increasingly
189 uncertain (e.g. in 2008 and 2009).

190

191 Figure 2.

192

193 Monthly values of dust events demonstrate that the main period of dust transportation is in the
194 spring, with a secondary maximum in the summer (in July and August), and dust activities in
195 February and October are also fairly high (Fig. 3). This seasonality pattern of our observations
196 fairly agrees with reports of several previous studies on Saharan dust events (Moulin et al.,
197 1998; Middleton and Goudie, 2001; Barkan et al., 2005; Engelstaedter et al., 2006; Goudie
198 and Middleton, 2006 and references therein).

199 It is worth mentioning that, according to the proton-induced X-ray emission analyses and
200 backward air trajectories of Borbély-Kiss et al. (2004) and Koltay et al. (2006), the maxima of
201 SDEs in Hungary appear in March and November. However, some of the identified events of
202 these studies could be originated from the southernmost part of the Sahel or even from the
203 Sudanian savannas, where biomass burning is the most important source of aerosols (Koltay
204 et al., 2006).

205

206 Figure 3.

207

208 *3.2 Synoptic patterns of Saharan dust transport towards the Carpathian Basin*

209

210 In spring and summer, the thermal convective activity forces the injection of particles to
211 higher atmospheric levels. It was recognized by Israelevich et al. (2002) that during these
212 periods a permanent reservoir of dust exists in the atmosphere above the major source areas,
213 and the occurrence of long-range dust transport is primarily governed by meteorological
214 conditions. Our monthly mean TOMS AI maps also confirmed these findings for the period of
215 1979–2009 (Fig. 4).

216

217 Figure. 4.

218

219 However, the strong upper level flows and resulting dust outbreaks can be connected to
220 various meteorological conditions. Based on the daily geopotential height and wind maps, and
221 backward trajectories of the SDEs, three main types of synoptic patterns and dust transport
222 pathways are distinguishable.

223

224 3.2.1 Type-1

225

226 66.2% (n=86) of the total SDEs belong to Type-1, which is a similar synoptic situation to that
227 mentioned by Barkan et al. (2005) concerning Saharan dust transport towards Italy. In this
228 case, a deep, well-developed trough emanated from the Bay of Biscay to the Atlantic coast of
229 Africa, and the eastern cell of the divided subtropical high over North Africa causes the strong
230 southwestern flow. The composite mean 700mb geopotential height and wind flow maps of
231 the identified Type-1 SDEs represent this situation in detail (Fig. 5). The dust particles
232 penetrate from North Africa into the atmosphere of the CB by strong SW flow across the
233 western or central basin of the Mediterranean Sea.

234 Seasonality of the identified 86 Type-1 SDEs display similar bimodal pattern compared to the
235 total SDEs. Monthly frequencies of the discussed type are typically high from March through
236 August. This seasonal distribution is the consequence of the synoptic background, as it is the
237 period of the northward migration of the subtropical high-pressure belt (actually a series of
238 high-pressure centres). In June, due to intense heating of the mainland, a low-pressure region
239 develops in Central Europe. The thermal low creates a steady pressure gradient between the
240 land and the Atlantic Ocean (this month is the wettest in the CB), and the resulting persistent
241 NW-flow blocks the penetration of dusty Saharan air masses.

242

243 Figure 5.

244

245 3.2.2 Type-2

246

247 33 SDEs were identified as Type-2 events, which is 25.4% of the total observations. Dust
248 events related to Type-2 are generated by southerly winds associated with depressions centred
249 in the central Mediterranean. The dust transport is generated by warm sector winds on
250 foreshore of the eastward moving cyclones (Fig. 6) and the strongest meridional wind flows are
251 located above the Ionian Sea and the southern part of the Apennine Peninsula, suggesting
252 straight dust intrusions from the south into the CB. Mediterranean low-pressure centres are
253 developing and the Type-2 SDEs occur typically in spring, before the intensification of the
254 Azores High caused by northward migration of subtropical highs.

255

256 Figure 6.

257

258 3.2.3 Type-3

259

260 11 episodes (8.5% of the total) were classified as Type-3 SDE, at times when a high pressure
261 centre is established over NW Africa and SW Europe. The strong flows of the anticyclone
262 carry dust particles from western parts of the Sahara northward along the coastline or over the
263 eastern Atlantic, at the mid-latitudes the dust-laden air mass moves eastward due to the
264 westerlies (Fig. 7). This type of Saharan dust transport is responsible for the observed dusty
265 events in western Europe (Pye, 1987): in Ireland (e.g. Tullet, 1978; Vernon and Reville, 1983)
266 and in Britain (e.g. Mill and Lempfert, 1904).

267

268 Figure 7.

269

270 *3.3 Typical pathways of dust transport and possible North African sources of mineral dust*

271

272 Related to the main synoptic situations, three typical routes of Saharan dust transport can be
273 identified based on the calculated backward trajectories of SDEs (Fig. 8). Type-1 synoptic
274 pattern favours to northeastward dust transport from the NW Sahara across the western basin
275 of the Mediterranean Sea. During Type-2 events, the outbreaks of dust-laden air reach the CB
276 directly from the south, and rarely from the SE. The longest (ca. 6500 km) and least common
277 dust pathways are associated with the Type-3 SDEs from the western parts of Sahara across
278 the eastern Atlantic and western Europe.

279

280 Figure 8.

281

282 It is clearly visible on the mean aerosol maps that large proportion of emitted Saharan dust
283 can be associated with some distinct source areas situated in topographic lows or on the flanks
284 of topographic highs (Fig. 9; Middleton and Goudie, 2001; Prospero et al., 2002; Washington
285 et al., 2003; Engelstadter et al., 2006). Most of these sources were flooded during the
286 Pleistocene and Holocene pluvial periods, and nowadays can be characterized with ephemeral
287 rivers and streams, alluvial fans, playas and saline lakes favouring the accumulation of fine-
288 grained material. Large sand seas cannot be regarded as effective source areas due to the
289 shorter atmospheric residence time and shorter transport of sand particles. However, the
290 bombardment energy of sand grains can disrupt the hardened and compacted smooth surfaces
291 of salt lakebeds and playas, thereby enhancing the amount of emitted dust particles (Gillette,

292 1999). Many of these source areas are located in proximity to large sand seas (Prospero et al.,
293 2002).

294

295 Figure 9.

296

297 The mean aerosol and detailed topographic maps, as well as previous studies (e.g. Prospero et
298 al., 2002) and backward trajectory analyses allows for locating the most probable dust loading
299 regions over North Africa. Extensive areas with high dust activity can be found in the western
300 region of the Sahara. One of these major sources is the territory of a series of playas and
301 seasonal streams (marked with 1 at Fig. 9) at the slopes of Zemmour Massif and Adrar
302 Soutouf Highlands, running parallel to the Atlantic coast. These areas, together with the dust
303 hotspot (2) next to the Inner Niger Delta (Mali) could be the source of mineral dust of Type-1
304 and seldom Type-3 dusty events of the CB. As this region is often fully obscured by the
305 atmospheric dust, a closer identification of exact locations of different types is difficult.

306 The large alluvial fans and wadis at the western slopes of Ahaggar Moutains (3) can be
307 characterized by high mean aerosol index and by several trajectories of both Type-1 and
308 Type-2. Furthermore, the depression of Tidikelt at the northern part of this region, surrounded
309 by plateaus (the Tanezrouft to the south and Plateau du Tademait to the north), by mountains
310 (Ahaggar and Tassili-n-Ajjer to the east) and by the sand sea of Erg Chech to the west has an
311 extensive ephemeral drainage system including several wadis from elevated regions, seasonal
312 marshes and mud flats (Glaccum and Prospero, 1980).

313 Similarly to Tidikelt, the system of salt and dry lakes in the lowlands south of the Tell Atlas
314 (4) acts also as a major source of dust material of both the identified Type-1 and Type-2
315 SDEs. Dust activity seems to be the largest between and to the south of the two largest salt

316 lakes (Chott Melrhir and Chott Jerid), in the foreland of Grand Erg Oriental (Prospero, J.M. et
317 al., 2002).

318 The dusty area expanding from the northern hillslopes of Tibesti across Cyrenaica to the
319 Qattara Depression (5) is the easternmost region responsible for SDEs over the CB. These
320 sources are associated with alluvial fans and extensive wadi systems on flanks of topographic
321 highs, and ephemeral salt lakes in the low-lying areas. Dust entrainment from this region
322 towards Central Europe only occurs during the existence of Type-2 synoptic patterns.

323 It is notable that backward trajectories have not indicated any events originating from the
324 Bodele depression for the entire studied period of 1979–2011, despite the fact that it is
325 considered as the most intense dust source globally (McTainsh and Walker, 1982; Prospero,
326 2002; Engelstaedter et al., 2006).

327

328 *3.4 Sahara as a source of Plio-Pleistocene aeolian dust deposits (red clay and loess) in the*
329 *Carpathian Basin?*

330

331 As it has been mentioned earlier, in some periods of Earth's history the amount of
332 atmospheric dust and frequency of dust storms increased by several orders of magnitude,
333 compared to the present situation (Mahowald et al., 1999, 2006; Kohfeld and Harrison, 2001).

334 The Pleistocene glacials were such dusty periods that widely distributed, thick loess deposits
335 were formed all over the mid-latitudes (Muhs and Bettis, 2003). Wind-blown loess and loess-
336 like deposits (underlain by aeolian red clay) are covering almost half of the CB (Pécsi 1990).

337 Previous studies (Kovács, 2008; Kovács et al., 2008, 2011; Varga, 2011; Varga et al., IN
338 PRESS) revealed that the bimodal grain-size distribution curves of the aeolian dust deposits of
339 the CB are representing two main sediment populations and may be interpreted as a mixture
340 of local and far-travelled dust material. Recent observations of dust emissions (Pye, 1987,

341 1995) demonstrated that the coarse-grained (~10–60 μm) component of aeolian dust deposits
342 was transported by surface winds in short suspension episodes, during discontinuous local
343 dust storms. The fine-grained (~1–8 μm) component is generally transported by upper level
344 flows even from remote deflation areas (Pye, 1987; Tsoar and Pye, 1987; Sun et al., 2002,
345 2004; Stuut et al., 2009).

346 The more seasonal distribution of precipitation, stronger winds and intensifying cyclogenesis
347 triggered by more frequent incursions of cold Arctic air masses, and the long-term variations
348 of African summer monsoon (Larrasoña et al., 2003) led to increased dust emissions from
349 the Sahara during glacial phases. Analyses of Mediterranean marine sediments also
350 confirmed the enhanced wind-blown dust supply from the Sahara during cold periods
351 (Moreno et al., 2002; Hoogakker et al., 2004; Larrasoña et al., 2003, 2008). Accordingly, a
352 mineral dust contribution to loess sediments of the Carpathian Basin is supposed. Indeed, as it
353 is demonstrated by the Late Pleistocene background dust concentration (Varga et al., IN
354 PRESS) and Sr-Nd isotope database for the CB (Újvári et al., 2012), an admixture of Saharan
355 dust into loess cannot be dismissed, however significant contribution of North African
356 material seems to be unlikely based on grain size considerations.

357 The relationship between large-scale quasi-periodic climate patterns (El Niño Southern
358 Oscillation, North Atlantic Oscillation) and Saharan dust emission is controversial. As
359 proposed by Prospero and Lamb (2003), large dust outbreaks could be associated with major
360 El Niño events. During the Pliocene, a permanent El Niño-like state (El Padre) has influenced
361 the climate conditions (Ravelo et al., 2006; Shukla et al., 2009), thus the Saharan dust
362 outbreaks could be more dominant factors in Pliocene dust accumulations in the CB which
363 material subsequently preserved as red clay in the basin.

364

365 **4. Conclusions**

366

367 A multi-year record of various data sources including TOMS Aerosol Index, satellite images,
368 HYSPLIT backward trajectory counting and synoptical meteorology were used to identify and
369 analyse Saharan dust events in the CB atmosphere over the period of 1979–2011. In this
370 period, 130 SDEs were detected altogether, primarily during spring and summer. Based on the
371 daily geopotential height, wind flow and meridional wind vector maps of dusty days, the
372 SDEs were classified into three main types. In the case of Type-1, a trough emanates from the
373 direction of Bay of Biscay to the Atlantic coast of Africa. The eastern cell of the divided
374 subtropical high and the cyclonic stream of a low-pressure system cause the strong SW flow.
375 During Type-2 events, dust transportation can be connected to the warm sector winds on the
376 foreshore of an eastern moving Mediterranean cyclone. The relatively seldom Type-3 events
377 are responsible for the longest dust transportation from the western parts of the Sahara along
378 the western fringe of an anticyclone and by the westerlies.

379 Our results nicely demonstrate that Saharan dust could often be detected in the CB
380 atmosphere. SDEs, during the dustier periods of the Pliocene and Pleistocene could have
381 served as significant source of clay and fine silt particles which then may have contributed to
382 the widely distributed aeolian dust deposits in the basin.

383

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385

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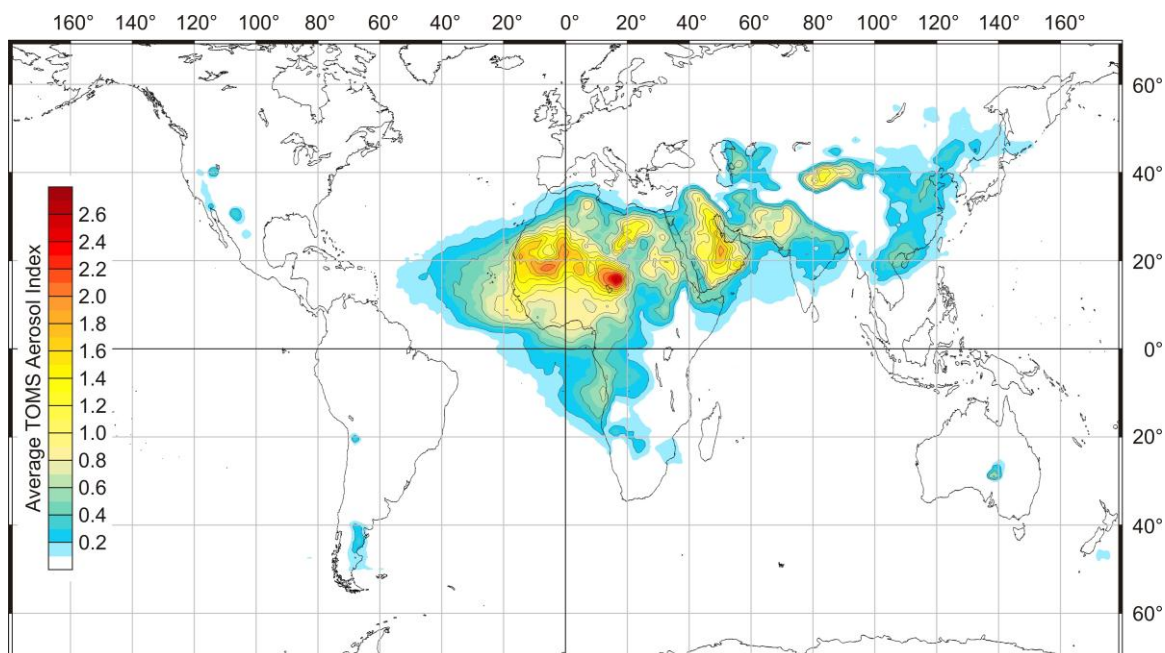
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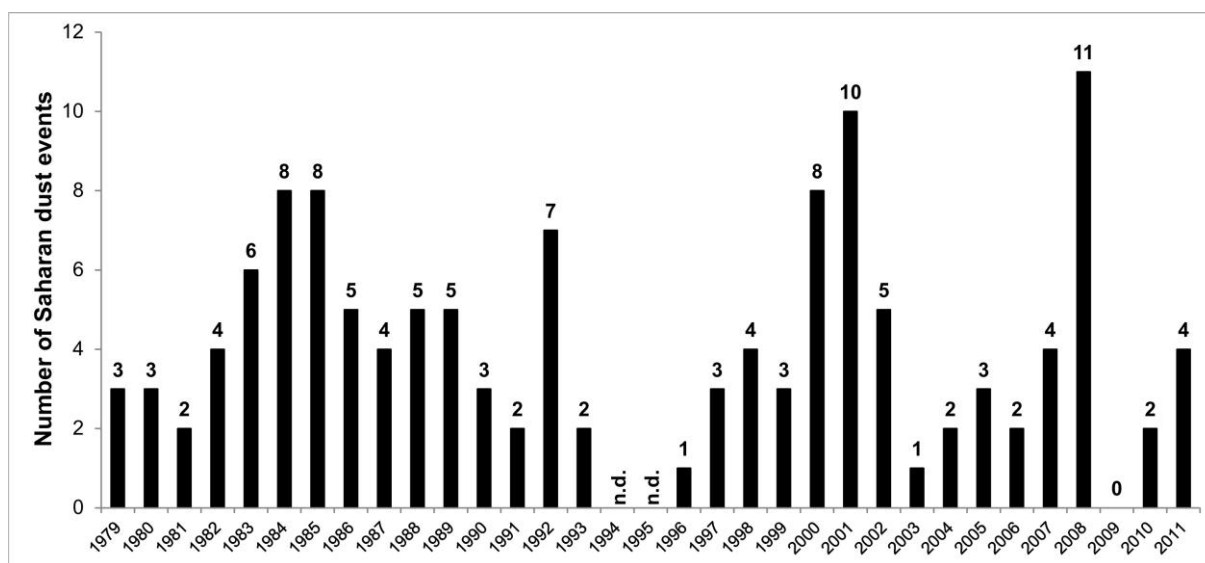
633 **Figures**



634

635 **Figure 1.** Mean global TOMS AI map of the investigated 23 full years from 1979 to 2009.

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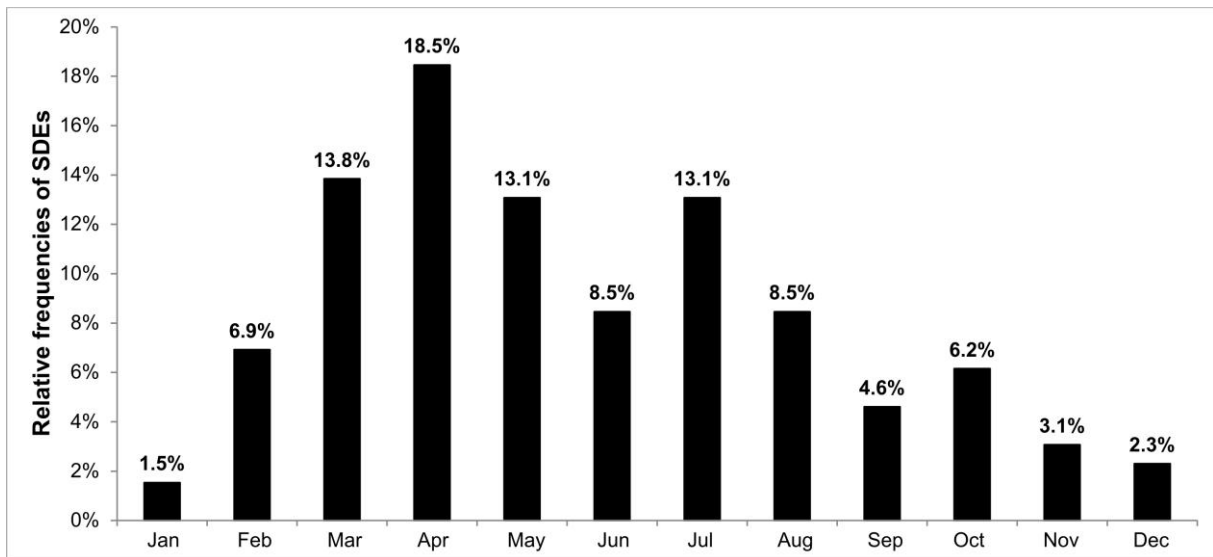


637

638 **Figure 2.** Annual number of identified Saharan dust intrusions in the atmosphere of the

639 Carpathian Basin. The data from years of 1993 and 1996 are fractional; n.d. means no data.

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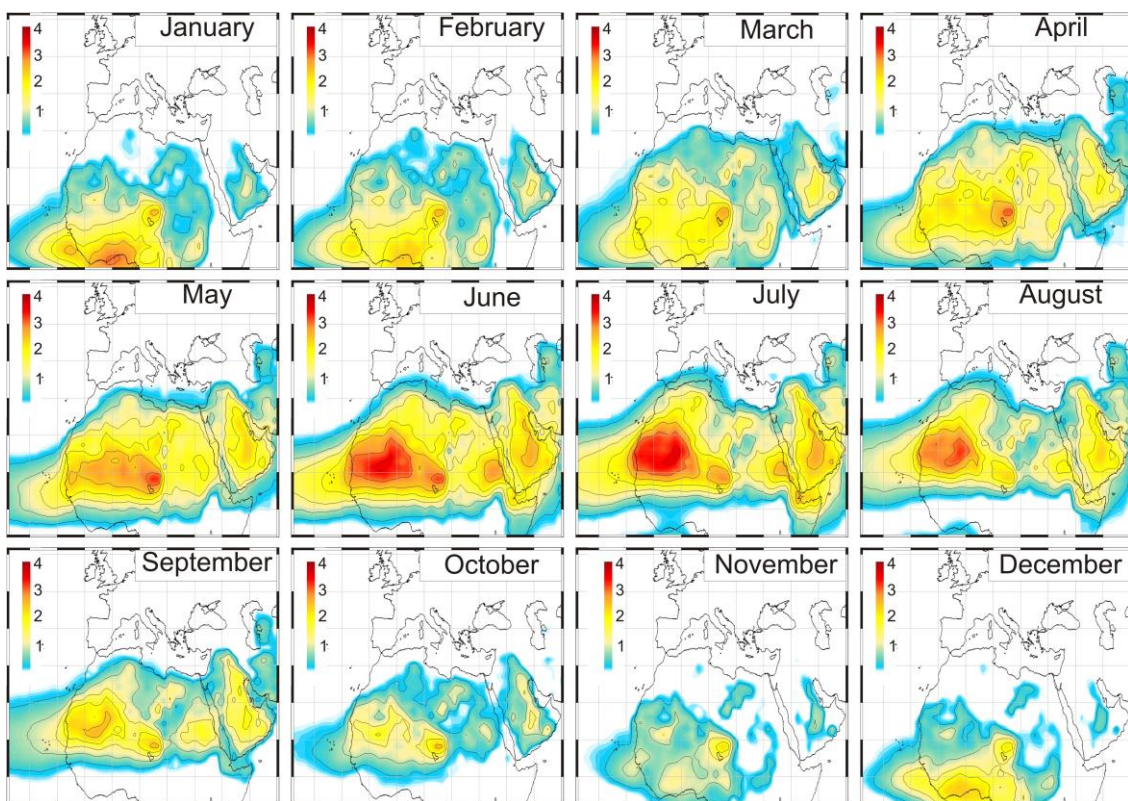


641

642 **Figure 3.** The seasonality of identified North African dust episodes over the Carpathian Basin

643 (1979–2009).

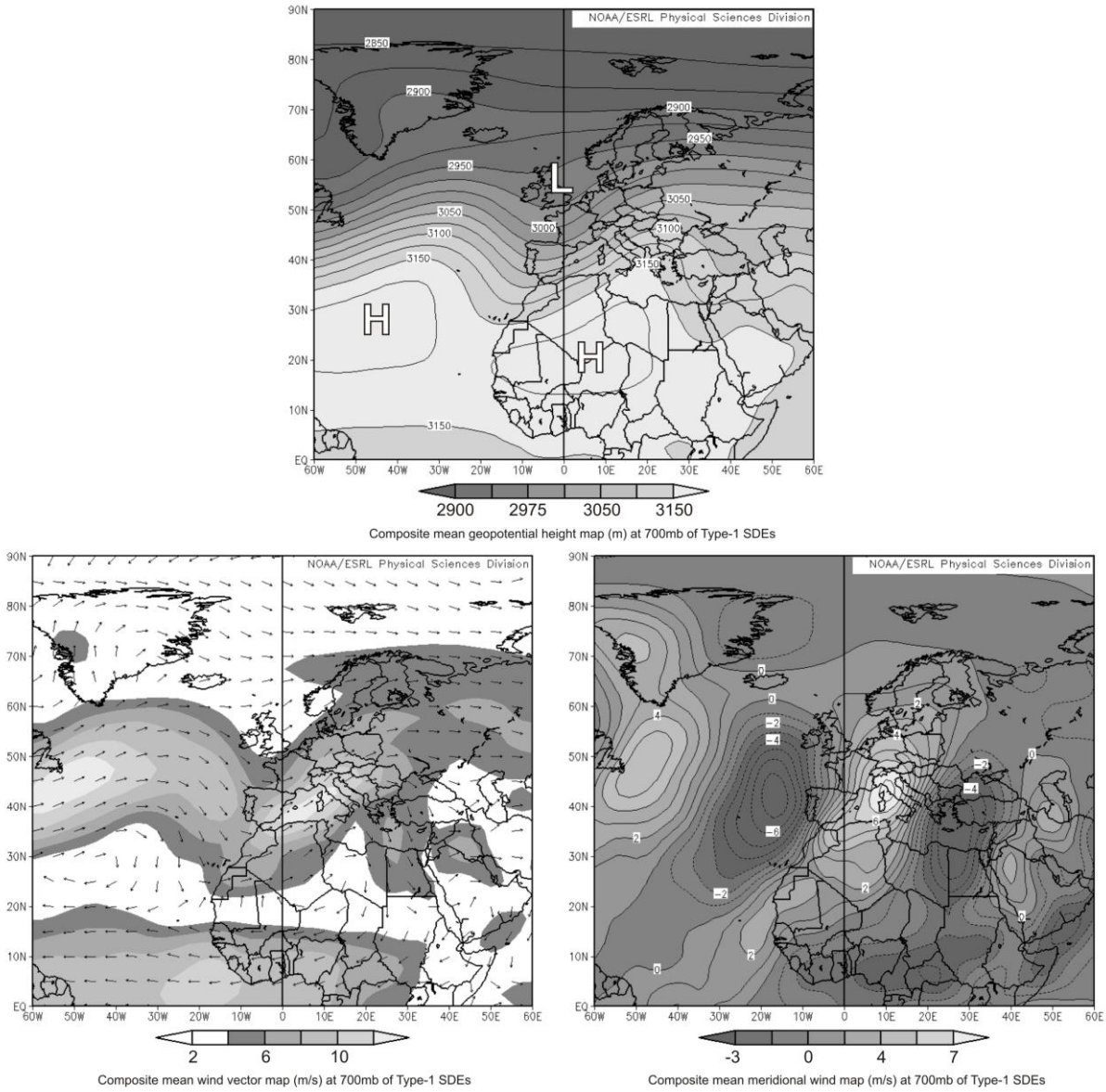
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646 **Figure 4.** Monthly mean TOMS AI maps of North Africa (1979–2009).

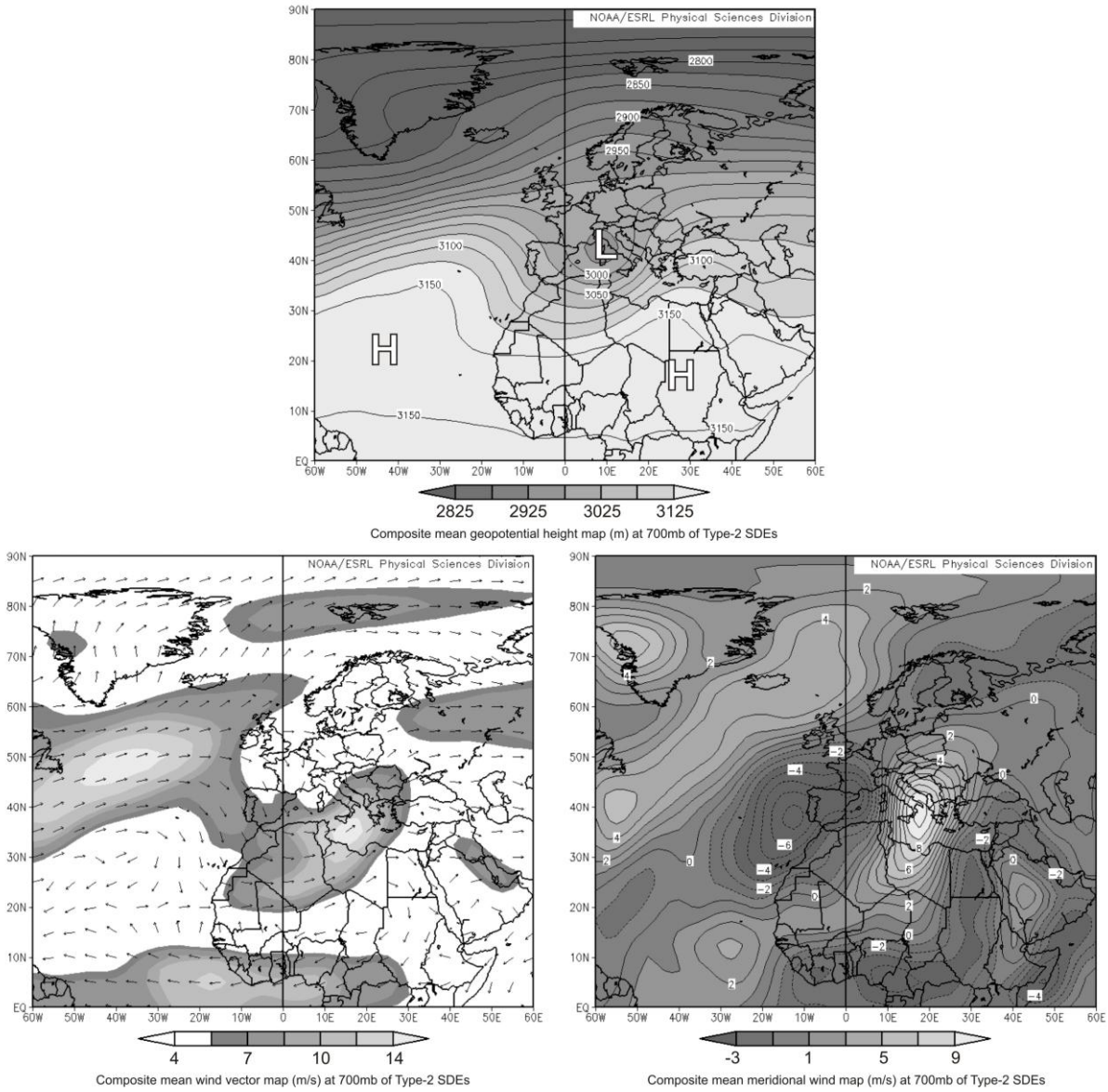
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649 **Figure 5.** Composite mean maps of the Type-1 synoptic situations (n=86).

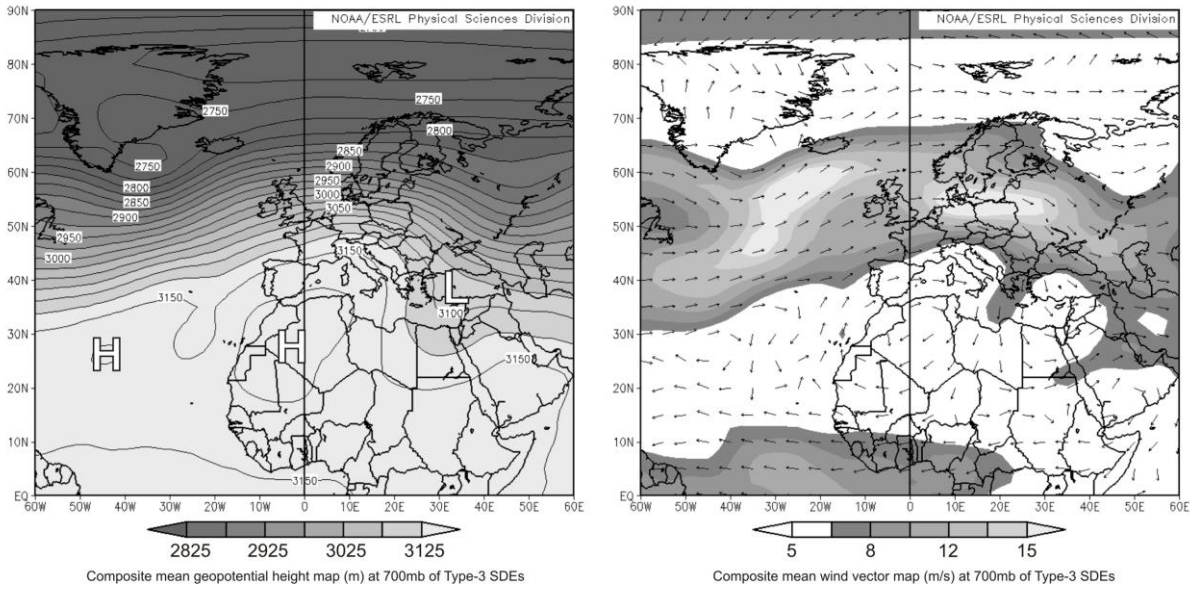
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652 **Figure 6.** Composite mean maps of the Type-2 synoptic situations (n=33).

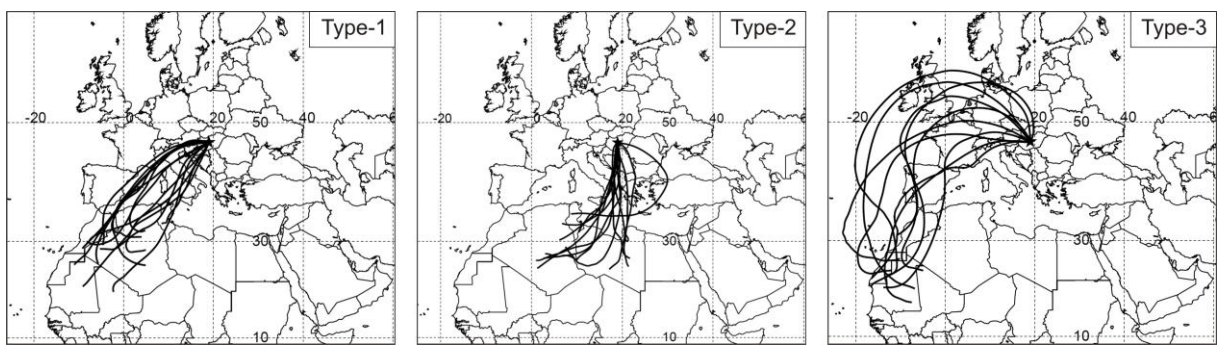
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655 **Figure 7.** Composite mean maps of the Type-3 synoptic situations (n=11).

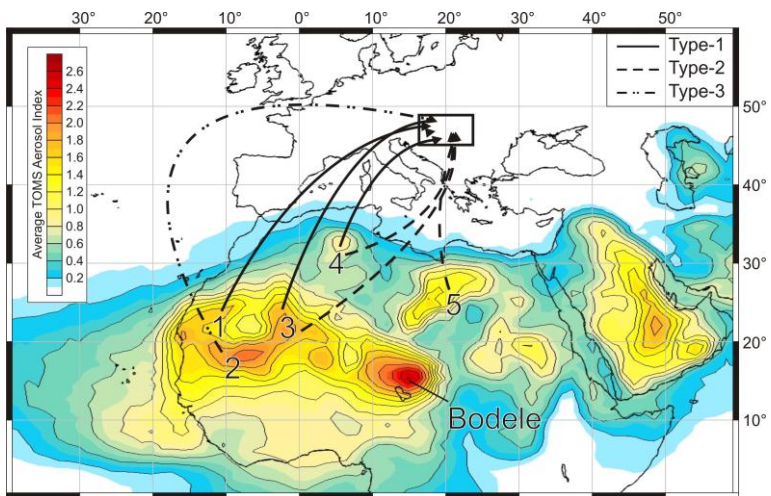
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658 **Figure 8.** Typical pathways of dust transport trajectories toward the Carpathian Basin.

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660

661 **Figure 9.** Major source areas and schematic transport routes of Saharan dust identified over
662 the Carpathian Basin (the numbers are explained in the text).