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Changes in atmospheric shortwave absorption as important driver of dimming and brightening

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The amount of solar (shortwave) radiation reaching the Earth's surface underwent substantial variations over recent decades. Since the 1950s, surface shortwave radiation gradually decreased at widespread locations. In Europe, this so-called surface dimming continued until the late 1980s when surface brightening set in and surface shortwave radiation increased again. In China, the dimming leveled off in the 1980s but did not turn into brightening until 2005. Changes in clouds and aerosol are the prime potential causes for

19 the phenomenon but the scientific community has not yet reached consen-
 20 sus about the relative role of the different potential forcing agents. Here
 21 we bring together colocated long-term observational data from surface and
 22 space to study decadal changes of the shortwave energy balance in Europe
 23 and China from 1985-2015. Within this observation-based framework, we
 24 show that increasing net shortwave radiation at the Top-of-the-Atmosphere
 25 and decreasing atmospheric shortwave absorption each contribute roughly
 26 half to the observed brightening trends in Europe. For China, we find that
 27 the continued dimming until 2005 and the subsequent brightening occurred
 28 despite opposing trends in the Top-of-the-Atmosphere net shortwave radi-
 29 ation. This shows that changes in atmospheric shortwave absorption are a
 30 major driver of European brightening and the dominant cause for the Chi-
 31 nese surface trends. Although the observed variations can not be attributed
 32 unambiguously, we discuss potential causes for the observed changes.

33 It is well documented that the amount of shortwave radiation reaching the Earth's
 34 surface (I_{sfc}^\downarrow) underwent substantial decadal variations since measurements became avail-
 35 able in the first half of the 20th century [1]. Since the 1950s, I_{sfc}^\downarrow gradually decreased
 36 until the late 1980s at widespread locations [2]. In Europe and other parts of the world,
 37 this "dimming" phase was followed by a "brightening" period in which I_{sfc}^\downarrow started to
 38 increase again [3]. In China, the dimming leveled off in the 1980s but did not turn into
 39 brightening until 2005 [4].

40 Here we look at this phenomenon from a shortwave energy balance perspective. The
 41 part of the incoming shortwave radiation at the Top-of-the-Atmosphere (TOA) which is
 42 not reflected back to space is absorbed within the climate system ($I_{toa}^{net} = I_{toa}^\downarrow - I_{toa}^\uparrow$).
 43 This net incoming shortwave radiation at TOA itself is either absorbed in the atmosphere
 44 (A_{atm}) or absorbed at the surface (A_{sfc}): $I_{toa}^{net} = A_{atm} + A_{sfc}$. The surface absorption
 45 (A_{sfc}) is determined by I_{sfc}^\downarrow and the surface albedo (α) via: $A_{sfc} = I_{sfc}^\downarrow \cdot (1 - \alpha)$ [5, 6].
 46 Due to the principle of energy conservation, changes in the shortwave energy balance
 47 must be counterbalanced at all times. Thus, the changes in I_{sfc}^\downarrow during dimming and

48 brightening imply that other components of the shortwave energy balance must have
49 changed accordingly.

50 Evidence was reported that the dimming and brightening affected the climate sys-
51 tem by weakening and strengthening the surface temperature increase from greenhouse
52 warming [7, 8], by changing the global hydrological cycle [9] and local precipitation sys-
53 tems like the Asian monsoon [10], as well as by altering the global carbon cycle [11]. It
54 was also recognized that the impacts of dimming and brightening on the climate system
55 might be substantially different or even opposite, depending on which components of
56 the shortwave energy balance determine the surface changes [12, 8, 13].

57 The fact that surface temperatures effectively respond to changes in I_{toa}^{net} is well known
58 and even considered in developing geoengineering techniques (i.e., solar radiation modi-
59 fication) to reduce the impact from global greenhouse warming [14]. The surface temper-
60 ature response to changes in A_{atm} is more versatile. It not only depends on the change
61 in the surface energy balance but also on the change in atmospheric temperature, atmo-
62 spheric stability, and the coupling between surface and higher layers [15, 16, 17, 8, 18].

63 The global precipitation response to dimming and brightening might also be different
64 depending on whether it is governed by I_{toa}^{net} or A_{atm} . Evaporation – which, in a global
65 perspective equals precipitation – is determined by the surface radiation balance and is,
66 therefore, sensitive to changes in the downward solar radiation [9]. Since evaporation
67 also depends on surface temperatures and since surface temperature responds differently
68 to changes in I_{toa}^{net} and A_{atm} , the total precipitation response to dimming and
69 brightening might even have an opposite sign when A_{atm} instead of I_{toa}^{net} forces the surface
70 changes [19]. This also applies for changes in regional precipitation systems like the East
71 Asian summer monsoon [13].

72 To better understand the causes and impacts of declining surface solar radiation (dim-
73 ming) and subsequent recovery (brightening) we compiled the best available observa-
74 tional data sets from surface and space to be able to contrast top-of-the-atmosphere, in
75 atmosphere, and surface energy fluxes

76 Flux changes inferred from colocated observations.

77 We use I_{sfc}^\downarrow data from the Baseline Surface Radiation Network (BSRN [20]), the Global
78 Energy Balance Archive (GEBA [21]), and the China Meteorological Administration
79 (CMA [4]) and combine them with colocated surface albedo estimates from the Global
80 Land Surface Satellites (GLASS [22]), and colocated I_{toa}^{net} estimates from the DEEP-C
81 reconstruction [23, 24]. The combination of point and gridded data is justified as we
82 only select stations which are representative of their larger surroundings according to
83 objective criteria [25]. For the period where all data is available (i.e., 1985 to 2015), we
84 compute regionally averaged annual mean anomaly time series of all components of the
85 shortwave energy balance based on colocated homogeneous records [26] of 71 stations in
86 Europe and 61 stations in China. The methods section provides descriptions of the data
87 sets, the data processing, as well as details on the uncertainty propagation strategy.

88 Figure 1 displays the station distribution in Europe and in China and shows the
89 estimated long term (2000-2015) annual mean A_{atm} for all stations expressed as a fraction
90 of I_{toa}^\downarrow . Globally averaged, roughly $23 \pm 2\%$ of I_{toa}^\downarrow is absorbed in the atmosphere [27,
91 28, 6, 29]. For Europe, the long term mean A_{atm} is close to the global value of 23% (see
92 Figure 1a) [30]. For China, however, we find that A_{atm} even exceeds 31% of I_{toa}^\downarrow at some
93 sites in highly developed regions in the southern and eastern parts of China.

94 Figure 2 shows the regionally averaged annual anomaly time series for I_{sfc}^\downarrow , α , A_{sfc} ,
95 A_{atm} , and I_{toa}^{net} as well as associated uncertainties, which we obtained by propagating
96 the measurement uncertainties using a bootstrapping approach (see methods section for
97 details). For the period where data of all components of the shortwave energy balance
98 are available (i.e., 1985-2015) we compute linear trends as well as their uncertainties and
99 significance for different sub-periods as shown in Figure 3 and in the respective tables
100 in the supplemental material.

101 For Europe, the well-documented decrease in I_{sfc}^\downarrow until roughly 1980 [3] and the fol-
102 lowing gradual increase of I_{sfc}^\downarrow [1, 31] is clearly visible in the lowest panel of Figure 2c.
103 The surface albedo only shows little variability and no trends. Consequently, the surface
104 absorbed flux (A_{sfc}) is closely following the variability of I_{sfc}^\downarrow . Over the whole 31-year

105 period A_{sfc} increases by $+1.7 \pm 0.1 W m^{-2} decade^{-1}$ with high statistical significance.

106 In Europe, I_{toa}^{net} tends to gradually increase throughout the whole period. However, the
107 observed trends in I_{toa}^{net} are considerably smaller and less significant than those observed
108 at the surface (see Fig. 3a & 3c), uncovering trends in A_{atm} which are comparable but of
109 opposite sign to those at the TOA. For the whole 31-year period we find trends in I_{toa}^{net} and
110 A_{atm} of $+1.0 \pm 0.1 W m^{-2} decade^{-1}$ and $-0.7 \pm 0.1 W m^{-2} decade^{-1}$, respectively. The
111 positive trend in the surface absorption during the European brightening period is thus
112 determined roughly equally from more incoming radiation at TOA and less radiation
113 absorbed within the atmosphere.

114 For some sub-periods, the trend in A_{atm} even outweighs the one of I_{toa}^{net} . The largest
115 trend in A_{atm} in Europe appears in the 15-year period centered around mid 1998. It
116 shows a decrease in A_{atm} of $-2.1 \pm 0.4 W m^{-2} decade^{-1}$ while in that period A_{sfc} and
117 I_{toa}^{net} show trends of $+3.6 \pm 0.3 W m^{-2} decade^{-1}$ and $+1.6 \pm 0.2 W m^{-2} decade^{-1}$.

118 In China, I_{sfc}^{\downarrow} shows a strong decrease prior to 1980 (see Fig. 2f). This "dimming"
119 leveled off in China during the 1980s and 1990s and I_{sfc}^{\downarrow} only started to recover to some
120 degree during the end of the observational period [32, 4]. This trend reversal is likely
121 related to the implementation of more rigorous air quality measures in China after 2000,
122 and thus to a reduction in air pollution [33, 34, 35, 36]. Also in China, the surface albedo
123 shows relatively little variability and no trend such that A_{sfc} again closely follows the
124 temporal evolution of I_{sfc}^{\downarrow} . For the whole 31-year period, the fluxes in China do not show
125 any significant trends. However, for shorter sub-periods large and significant trends are
126 apparent.

127 The trends in A_{sfc} in the first part of the total record are typically around $-1.0 \pm$
128 $0.2 W m^{-2} decade^{-1}$ with limited statistical significance. At the same time, more radia-
129 tion enters the climate system at the TOA. The negative trend at the surface and the
130 positive trend at the TOA can only occur at the same time when large positive trends in
131 A_{atm} are prevalent. These positive trends in A_{atm} range up to $+3.7 \pm 0.4 W m^{-2} decade^{-1}$
132 for some periods. For the 23-year period from 1985 to 2009 (centered mid 1996),
133 in which dimming in China occurs, a significant positive trend in A_{atm} of $+1.5 \pm$

134 $0.2 W/m^2 decade^{-1}$ coincides with trends in A_{sfc} and I_{toa}^{net} of $-0.7 \pm 0.2 W m^{-2} decade^{-1}$
135 and $+0.8 \pm 0.2 W m^{-2} decade^{-1}$, respectively.

136 For the period from 2001 to 2015 (centered in mid 2008), when pronounced brightening
137 is evident in China, trends of $-1.8 \pm 0.3 W m^{-2} decade^{-1}$, $-3.2 \pm 0.4 W m^{-2} decade^{-1}$,
138 and $+1.4 \pm 0.2 W m^{-2} decade^{-1}$ are observed for I_{toa}^{net} , A_{atm} , and A_{sfc} , respectively.

139 We can summarize the results as follows: In Europe, the increase in the amount of
140 net incoming radiation at the TOA and the decrease in atmospheric absorption each
141 contributed roughly half to the increase in surface absorption and therefore to the Euro-
142 pean brightening. In China, a transition from dimming to brightening is found around
143 the year 2005. Before that, increasing atmospheric absorption led to dimming at the
144 surface (and corresponding decrease in surface absorption) despite increasing TOA net
145 radiation. After 2005, all trends reversed in China and decreasing atmospheric absorp-
146 tion led to surface brightening (and corresponding increase in surface absorption) despite
147 decreasing TOA net radiation. This is to date the most direct observational evidence
148 that changes in atmospheric absorption are central to the brightening and dimming
149 phenomenon.

150 **Potential causes of the phenomenon.**

151 Because of the profound consequences of dimming and brightening for the climate sys-
152 tem, it is important to identify the governing processes behind the phenomenon. A
153 substantial body of literature exists, which investigated potential causes of the changes
154 in surface solar radiation. Scientific consensus about the relative role of different forc-
155 ing agents as well as the role of internal variability has, however, not yet been reached
156 [37, 1, 38]. This is also because most previous observational analyses studied dimming
157 and brightening from a surface perspective only, as long-term high-quality satellite ob-
158 servations of TOA fluxes and surface albedo were not yet available.

159 Our approach goes beyond this often used surface-only perspective and allows a si-
160 multaneous observation-based quantification of changes in the partitioning of I_{toa}^{net} , A_{sfc} ,

161 and A_{atm} during brightening and dimming, which in turn facilitates a more detailed in-
162 sight into the governing processes and their impacts on climate. However, a challenge in
163 all observation-based approaches is that various different forcing agents act simultane-
164 ously. It is thus difficult to attribute changes to individual forcing agents. Nevertheless,
165 our results offer additional constraints to judge which of the primary forcing agents –
166 notably clouds, surface albedo, water vapor, and scattering and absorbing aerosols –
167 may be able to explain the observed changes in shortwave fluxes (note that changes the
168 radiatively active gaseous compounds O_3 , N_2O , CH_4 , and CO_2 only induce very small
169 changes [39]).

170 Our main finding – namely that changes in atmospheric shortwave absorption are a
171 major driver for the changes in the shortwave energy balance – is especially evident in
172 China. There, a decrease in total cloud cover until 2005 was reported [40], while at
173 the same time emissions of scattering aerosols sharply increased [41]. After 2005, an
174 increase in clouds [42] and a decrease in emission of scattering aerosols were reported
175 [41]. Aerosol emissions possibly also forced some of the observed cloud changes through
176 aerosol cloud interactions [43]. In general, more radiation is reflected back to space when
177 more clouds and scattering aerosols are present leading to a smaller I_{toa}^{net} flux. The initial
178 increase and subsequent decrease in I_{toa}^{net} in our results then suggests that the changes
179 in cloud cover must have counterbalanced and outweighed the changes in scattering
180 aerosols. From changes in clouds and scattering aerosols alone one would expect that
181 the TOA and surface changes are of the same sign and magnitude, but the observations
182 actually show the opposite. Also, the radiative effect of clouds and scattering aerosols
183 is much larger at the TOA than on A_{atm} [12, 29, 44]. Thus, it is unlikely that changes
184 in clouds and scattering aerosols alone lead to observed changes in the shortwave fluxes
185 in China.

186 Observed changes in total water vapor path (WP) of -0.118 mm/year in China from
187 2000 to 2015 [45] fit the observed change in A_{atm} qualitatively. However, using an
188 empirical relation which links water vapor abundance (WP) to A_{atm} (in a pristine at-
189 mosphere) [29] via $A_{atm} = \alpha(2.1 + 0.86WP) + [15.7 + 3.3\ln(WP)]$ with climatological

190 $WP = 2.25 \text{ cm}$ [45] and albedo $\alpha = 0.1$ implies a change in A_{atm} due to changes in WP
191 of only 0.3 W m^{-2} from 2000 to 2015, which is much less than what is actually observed
192 (see also reference [46]).

193 Considering that α did not change drastically in the observational period (see Fig-
194 ure 2) this altogether points to absorbing aerosols – of which the most prominent species
195 is black carbon – as the prime forcing agent for the observed change in the shortwave
196 fluxes in China. Reconsidering changes in aerosol optical depth from observations [47],
197 reanalysis [48], and emission inventories [41] show consistent temporal changes in (ab-
198 sorbing) aerosols, but efforts to estimate the aerosol forcing from such data are subject
199 to substantial uncertainties [49]. The observed changes in A_{atm} might also have con-
200 tributed to the observed cloud changes via the semi-direct aerosol effect from absorbing
201 aerosols [50, 1, 51]. Our observational results also fit modeling efforts which utilize
202 radiative transfer models in combination with observations of aerosol optical proper-
203 ties which suggest that increasing atmospheric absorption due to increasing absorbing
204 aerosols prevail [52, 16, 53].

205 It is plausible that absorbing aerosols are also the main cause for the Chinese dimming
206 in the pre-satellite era and that the high A_{atm} (Fig. 1b) is a remnant of this strong
207 dimming. However, since direct continuous observations are not available, this can not
208 be demonstrated conclusively.

209 In China, observational evidence for changes in ambient surface temperatures, clouds,
210 wind speed, fog, precipitation, and changes of the Asian summer monsoon was reported
211 (see ref [43] and references therein). It is plausible that these changes are to some degree
212 caused by the changing A_{atm} as main radiative response to changing absorbing aerosols.

213 Mitigation of aerosol pollution in China and associated surface brightening then may
214 not necessarily be associated with large additional temperature forcing as suggested
215 by modeling studies [54]. Our study demonstrates that the SW TOA forcing, thus the
216 shortwave energy input into the climate system, has been positive in the dimming period
217 from 1985 to 2005 but negative thereafter.

218 In Europe, A_{atm} and I_{toa}^{net} each contributed roughly half to the surface brightening.

219 Several studies highlighted that changes in clouds play a vital role for dimming and
220 brightening in Europe [55, 56, 57]. At the same time, other studies [58, 59] suggest
221 declining aerosols emissions in Europe [60] as main forcing for the European surface
222 trends. Although these decreasing emissions could also have led to declining aerosol-
223 cloud interactions, evidence was reported that these aerosol indirect effects contribute
224 little to the European brightening [61, 62, 55]. We demonstrate for the first time that also
225 in Europe changing A_{atm} significantly contributes to the surface changes. This would fit
226 to decreasing black carbon emissions in Europe obtained from emission inventories [60].
227 Our findings, therefore, point to a combination of clouds and aerosols as main forcing
228 agents in Europe, as also suggested previously [63, 55].

229 This study demonstrates that using combined colocated surface and TOA observations
230 allows valuable insight into the physical processes which govern the changes in the energy
231 balance. The observation based, quantitative estimates of decadal scale changes of the
232 shortwave energy balance components presented may also provide useful for the further
233 analysis and improvement of shortcomings of global climate models [64, 65].

234 A specific conclusion from our results, if combined with published estimates of changes
235 in different atmospheric constituents, is that dimming and brightening may not be at-
236 tributable to a single forcing agent but that it is a result of a complex interplay between
237 changes of different forcing agents where the role of absorbing aerosols has previously
238 been underestimated.

239 **Methods**

240 **Surface shortwave radiation.**

241 For I_{sfc}^\downarrow we utilize data from the Baseline Surface Radiation Network (BSRN [20]), the
242 Global Energy Balance Archive (GEBA [21]), and data from the China Meteorological
243 Administration (CMA) [4].

244 The GEBA and CMA data is available as monthly mean time series. For BSRN sta-
245 tions, monthly mean times series from the raw two- and one-minute data are calculated
246 according to the recommended procedure of ref. [66].

247 Only stations which are representative for a larger surrounding are included in the
248 analysis, to match with the scale of the TOA data. We consider three aspects of repre-
249 sentativeness [67, 25]: (1) the spatial distance up to which the temporal variability of a
250 I_{sfc}^\downarrow time series measured at a site can be considered representative (expressed in terms of
251 decorrelation length, δ), (2) spatial sampling biases (β), and (3) spatial sampling errors
252 (ϵ) which arise due to imperfect representativeness of the stations. In the terminology
253 of Schwarz et al. (2018) [25], stations which do not adequately represent the temporal
254 variations of I_{sfc}^\downarrow (i.e. $\delta < 2^\circ$) or have large (monthly) spatial sampling errors with
255 respect to the DEEP-C grid ($\epsilon > 16W/m^2$, 95% confidence level) are excluded from the
256 analysis. Spatial sampling biases are corrected for each station.

257 For the BSRN data, uncertainties of $\pm 8 Wm^{-2}$ and $\pm 5 Wm^{-2}$ (95% level) for monthly
258 and annual means were reported, respectively [68]. For the GEBA data, uncertainties
259 for monthly and annual means of $\pm 5\%$ and $\pm 2\%$ (root mean square error) were reported
260 [69]. The error of the CMA data likely does not exceed $\pm 5\%$ and $\pm 3\%$ for monthly and
261 annual data [70].

262 The long-term stability of the I_{sfc}^\downarrow observations is achieved by a regular calibration of
263 the sensors against a reference with known stability, which is traceable to a known refer-
264 ence. BSRN measurements are calibrated annually and tracable to the world radiometric
265 reference (WRR) from the World Radiation Center at the Physikalisch-Meteorologisches
266 Observatorium in Davos, Switzerland (PMOD/WRC) [71]. The CMA data is calibrated

267 using a multistep approach which is traceable to a national reference, which itself is
268 calibrated to the WRR every five years [72].

269 Within 38 years the WRR had a suggested total drift of less than 0.02% [68]. A quanti-
270 tative assessment of how that translates into the stability of long term I_{sfc}^\downarrow measurements
271 is lacking. However, for well calibrated instruments, we expect sufficient stability which
272 does not substantially influence the observed trends.

273 To avoid step-changes in the surface data which might occur due to changes in the in-
274 strumentation and/or relocation of the station, we assess the homogeneity of the monthly
275 I_{sfc}^\downarrow time series. The GEBA time series are tested regarding their homogeneity using
276 four different homogeneity tests as described in ref. [26]. The CMA data has been
277 analyzed and homogenized using sunshine duration data as described in ref. [4]. BSRN
278 data is expected to be homogeneous because of the rigorous measurement standards of
279 the BSRN [71].

280 After excluding stations with insufficient spatial representativeness, stations with lack-
281 ing homogeneity, or stations with less than 15 years of data for all shortwave energy
282 balance components, we use in total six BSRN and 65 GEBA stations for Europe and
283 one BSRN and 61 CMA stations for China.

284 **TOA shortwave radiation.**

285 For the TOA net irradiance, we use the DEEP-C version 3 reconstruction, which is
286 available with a spatial resolution of 0.7° and monthly temporal resolution [23, 24]. The
287 reconstruction merges satellite SW TOA irradiances from the Clouds and Earth's Radi-
288 ant Energy System (CERES) Energy Balanced and Filled Ed2.8 data set (available since
289 02/2000; [73, 74]) and the Wide Field of View (WFOV) Ed.3 Rev1 data set from the
290 non-scanning instrument onboard of the Earth Radiation Budget Experiment (ERBE;
291 available 1985-1999; [75, 76]) satellites. For the reconstruction, the ERA-Interim at-
292 mospheric reanalysis [77] and a 25km resolution global atmospheric model (HadGEM3-
293 A-GA3) with five ensemble simulations from the UPSCALE project [78] were used to
294 homogenize the satellite data sets [23, 24]. For the period after March 2000, CERES

295 data is used. Before that, the reconstruction is based on an annual cycle calculated
296 from the first five complete years of the CERES data (2001-2005) to which ERA-I de-
297 seasonalized radiative flux anomalies are added. The data is then adjusted such that
298 the hemispheric ($60^{\circ}S - 0^{\circ}$ and $0^{\circ} - 60^{\circ}N$) mean deseasonalized anomalies match the
299 corresponding hemispheric ERBE time series. With this approach, the DEEP-C data
300 "combines the quality of the CERES data, stability of the ERBE data, and the realistic
301 circulation changes depicted by ERA-I" [23].

302 We provide a comparison between the DEEP-C reconstruction and the directly mea-
303 sured data in Figure 4 for the areas under investigation in this study. There, all regional
304 averages of the raw data for Europe and China from CERES EBAF v.4 (03/2000 to
305 present; [79]), the ERBE Scanner data (1985-1989) [75], and the ERBE WFOV Non-
306 scanner data Ed. 3 Rev.1 [76] and ERBE WFOV Nonscanner data Ed. 4 [80] (both
307 WFOV Nonscanner data available from 1985 to 1999) are shown.

308 The figure shows, that the DEEP-C data follows the same temporal evolution as the
309 directly measured data and that no spurious inhomogeneities or trends are visible in the
310 data record.

311 The uncertainty of regional I_{toa}^{net} from the DEEP-C reconstruction was estimated to
312 be $\pm 5.7 W m^{-2}$ for monthly and $\pm 2.1 W m^{-2}$ for annual means (one standard-deviation
313 confidence level) [24]. The stability of the the CERES instrument is on the order of
314 $\pm 0.2 W m^{-2} decade^{-1}$ [24] while the stability of the ERBE WFOV Nonscanner instru-
315 ment is on the order of $\pm 0.35 W m^{-2}$ during the period 1985-1999 [76].

316 **Albedo.**

317 We use the Global Land Surface Satellite (GLASS) white sky albedo as the main albedo
318 dataset. It is based on advanced very high resolution radiometer (GLASS-AVHRR;
319 1982 - 2015) and Moderate-resolution Imaging Spectrometer (GLASS-MODIS; 2000-
320 2015) observations [22]. The dataset provides a high-quality gap-free, long-term, self-
321 consistent albedo record since 1982 with similar quality as the Moderate Resolution
322 Imaging Spectroradiometer (MODIS) albedo data [81, 22]. The data is available in

323 8-day temporal resolution with 0.05° . We aggregate the observations temporally to
324 monthly means and spatially to the DEEP-C 0.7° grid.

325 For the period after 2000, we also use the MODIS white sky surface albedo from [82].
326 The MODIS data currently provides one of the most reliable albedo estimates and is,
327 therefore, used as reference data. The MODIS albedo accuracy has been proven to be
328 well within $\pm 5\%$ [83].

329 **Data processing.**

330 To calculate the surface absorbed flux, we multiply the monthly I_{sfc}^\downarrow with monthly mean
331 α estimates from GLASS-AVHRR. A step inhomogeneity in the GLASS-AVHRR data
332 around the year 2000 was corrected by subtracting the differences between the GLASS-
333 AVHRR long-term mean before and after the year 2000. Since the MODIS albedo is
334 currently the most reliable albedo estimate, we bias correct the GLASS-AVHRR data
335 by subtracting the differences between the MODIS and GLASS-AVHRR data.

336 To test the albedo's influence on A_{sfc} and A_{atm} we computed all fluxes by assuming a
337 constant α which we calculated from the MODIS data. The comparison of the shortwave
338 energy balance fluxes as calculated using GLASS-AVHRR and constant albedo reveals
339 that the albedo variability only has a minor influence on the A_{sfc} and A_{atm} (not shown).

340 Finally, we calculate A_{atm} by subtracting A_{sfc} from I_{toa}^{met} from co-located DEEP-C data.
341 These time series are then deseasonalized before we calculate annual mean anomalies for
342 each station (if at least nine of twelve months are available per year). We only consider
343 time series which have at least 15 annual mean values during the period 1985-2015.
344 Finally, we average the annual anomaly time series of all stations in Europe and China.

345 **Sensitivity to station selection**

346 To estimate the uncertainties in the computation of the regional mean time series with
347 respect to the station selection, we apply a bootstrapping approach (N=100) where we
348 randomly select sub-samples of all available stations. For the sub-sample of stations we

349 calculate the regional average time series as described above and compare it to the time
350 series where all stations are used. We found that the resulting regional average time series
351 and the corresponding trend estimates are rather insensitive to the station sampling. For
352 example, when two thirds of all available stations are used for the bootstrapping, the
353 mean standard deviation between the individual realizations for the regional mean is 0.2
354 / 0.4/ 0.5 W/m^2 for I_{toa}^{net} , A_{atm} and A_{sfc} , respectively.

355 **Uncertainty estimation**

356 To propagate the measurement uncertainties to the regional average time series and
357 to the trend estimated for the different periods we apply a bootstrapping approach
358 (N=1000) where we artificially add Gaussian random noise to the measured time se-
359 ries. The standard deviation of the noise is chosen such that it corresponds to the (1σ)
360 measurement uncertainties (including the spatial sampling error) of the different fluxes.
361 The station time series with the random noise are then processed as outlined above to
362 calculate the regional means. For each realization we compute regional time series and
363 the trend estimates for all periods. The differences in the trend estimates between the
364 different realizations can be interpreted as propagated measurement uncertainties. We
365 show the statistics of the different realizations of this bootstrapping approach in Figure 3
366 and in the corresponding tables in the supplemental material in terms of standard devi-
367 ation of the slope estimate and percent of realizations which show statistical significant
368 (Wald Test with t-distribution of the test statistic) trend estimates on the 95%-level for
369 a given period.

370 **Code availability**

371 All code used in this study to perform the analyses and to create the figures can be
372 made available upon request from the corresponding author.

373 **Data availability**

374 The DEEP-C data is available via <http://dx.doi.org/10.17864/1947.111>. The GLASS
375 data is available via <http://glcf.umd.edu/data/abd/>. The BSRN data is available via
376 <https://bsrn.awi.de/>. The GEBA data is available via <http://www.geba.ethz.ch/>.
377 The CMA data can be accessed from the China Meteorological Administration [http:](http://www.cma.gov.cn/)
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605 **Addendum**

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Competing Interests The authors declare that they have no competing financial interests.

Author contributions M.S., D.F., and M.W. designed the study. Y.S. processed the in-situ data over China. R.A.
618 provided the DEEP-C data and helped interpreting it. M.S. did the coding and data analysis with
619 help of all coauthors. M.S., D.F., M.W. wrote the paper with contributions from all coauthors.

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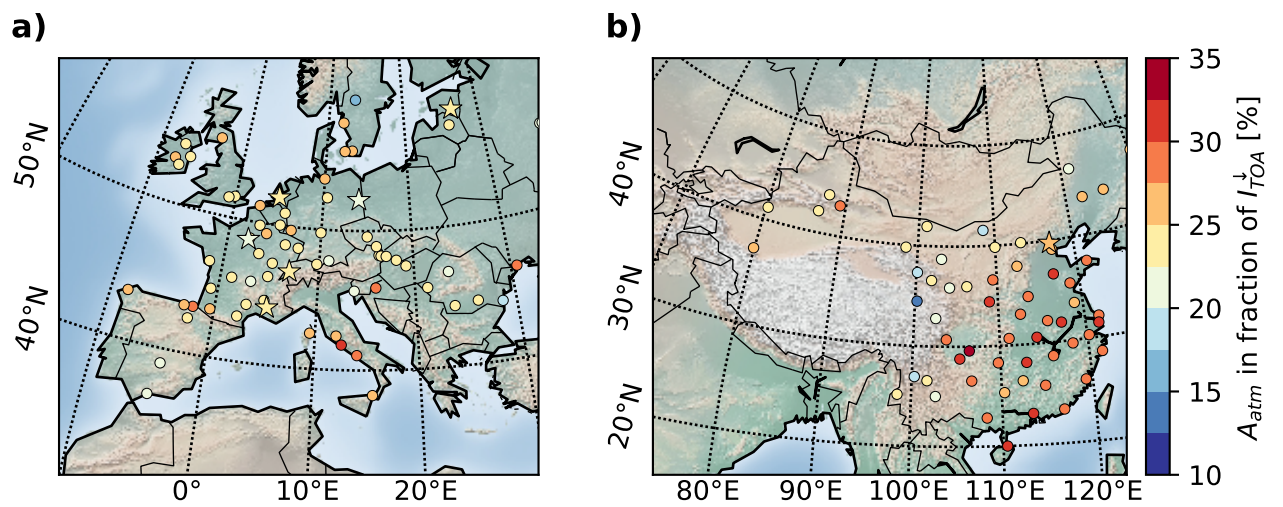


Figure 1: Long term mean (2000-2015) fractional atmospheric shortwave absorption (A_{atm}) for Europe (a) and China (b). Values are given as a fraction of TOA incoming radiation (I_{toa}^{\downarrow}). Points show GEBA and CMA stations. Stars show BSRN stations.

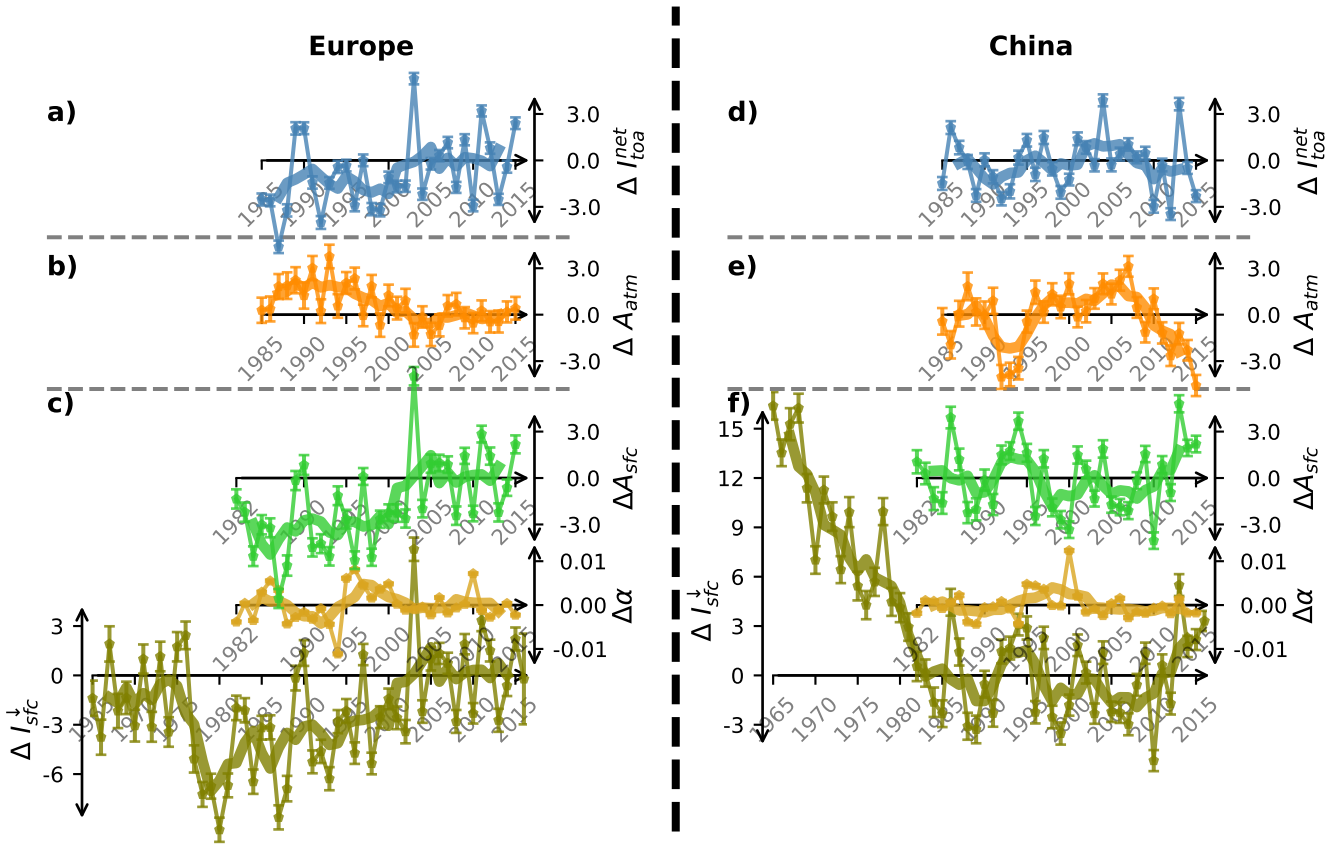


Figure 2: **Anomaly time series of shortwave energy balance quantities.** Shown are regional mean time series for Europe (a-c) and China (d-f) for TOA net shortwave flux (a,d), atmospheric shortwave absorption (b, e), as well as downward surface solar shortwave radiation, albedo, and surface shortwave absorption (c,f). Time series are shown as deviations (Δ) from long term means of the reference period 2000-2015 in W/m^2 for all fluxes and in albedo units for $\Delta\alpha$. Thin lines show station averaged annual means. Vertical bars indicate the 95% uncertainty range for propagated measurement uncertainties using a bootstrapping approach (as described in the methods section). Thick lines show centered five-year running means.

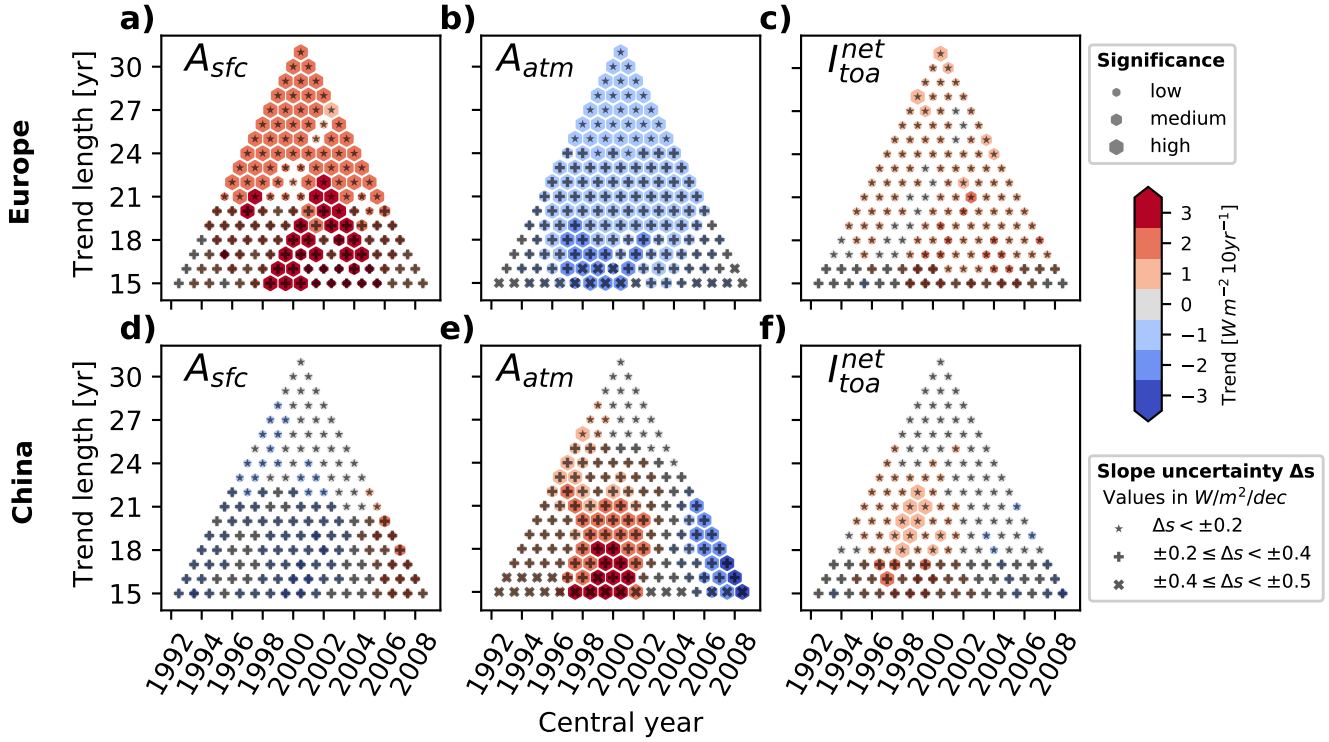


Figure 3: Trend matrices for surface shortwave absorption (a, d), atmospheric shortwave absorption (b, e), and TOA net shortwave radiation (c, f) for Europe (top row) and China (bottom row). Shown are linear trend estimates for different periods. The x-axis shows the central year of the trend window while the y-axis shows the length of the trend. The slopes, statistical significance, and uncertainties of the estimated trends are indicated by color, the size of the markers, and symbols, respectively. The uncertainties were derived by propagating the measurement uncertainties using a bootstrapping approach (see methods section for details).

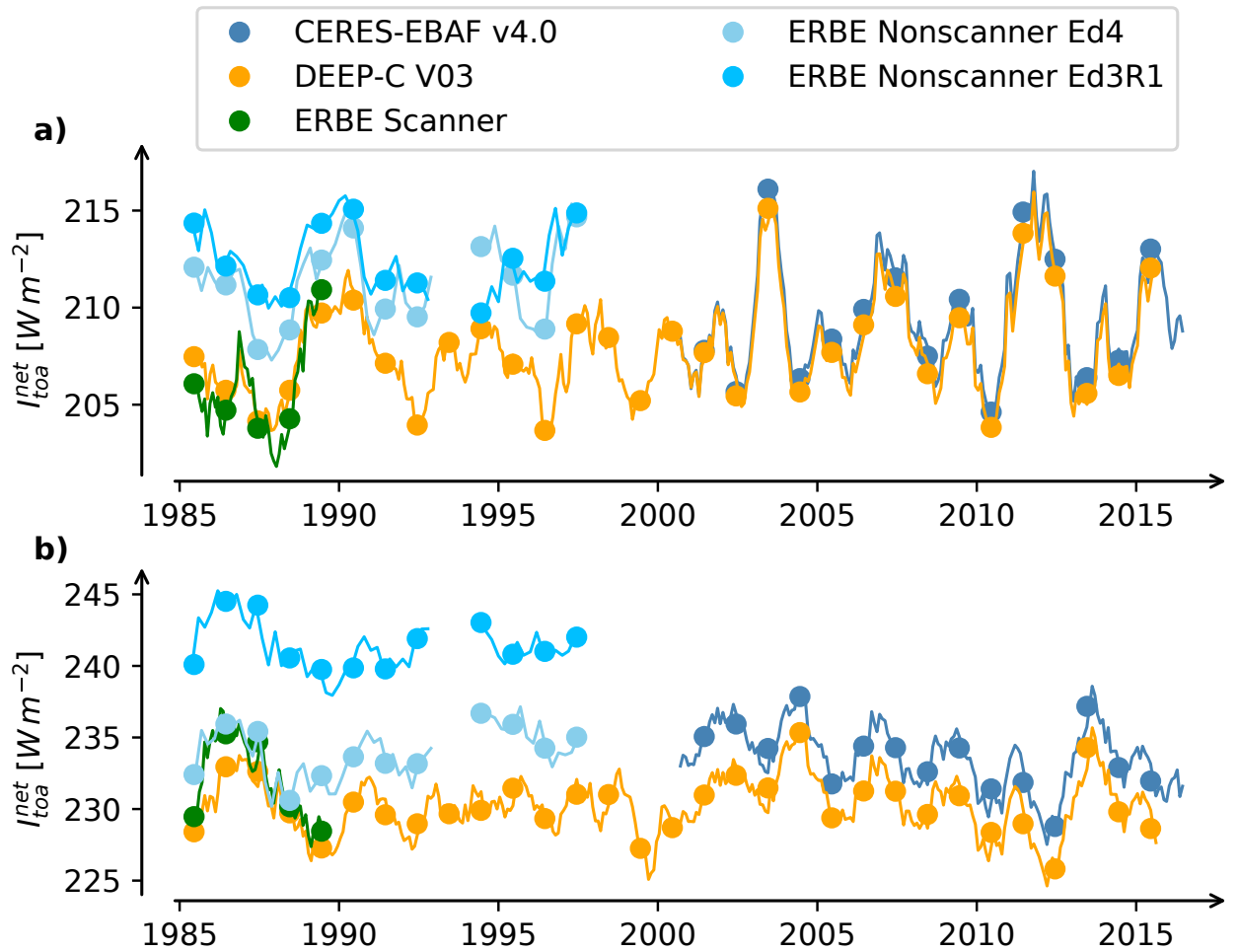


Figure 4: **Comparison of TOA net shortwave fluxes from different data sources.**

Shown are regional average TOA net shortwave flux time series of the CERES-EBAF v.4.0, DEEP-C V03, ERBE Scanner, and ERBE Nonscanner (Ed3R1 and Ed4) data sets for (a) Europe ($40^{\circ}N-50^{\circ}N$; $10^{\circ}W-20^{\circ}E$) and (b) China ($20^{\circ}N-40^{\circ}N$; $100^{\circ}E-120^{\circ}E$). Points show annual means while the thin lines show 12-month running means.