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Lembrechts, Jonas J., van den Hoogen, Johan, Aalto, Juha ORCID: https://orcid.org/0000-0001-6819-4911, Ashcroft, Michael B., De Frenne, Pieter, Kemppinen, Julia, Kopecký, Martin, Luoto, Miska, Maclean, Ilya M. D., Crowther, Thomas W., Bailey, Joseph ORCID: https://orcid.org/0000-0002-9526-7095, Haesen, Stef, Klinges, David H., Niittynen, Pekka, Scheffers, Brett R., Van Meerbeek, Koenraad, Aartsma, Peter, Abdalaze, Otar, Abedi, Mehdi, Aerts, Rien, Ahmadian, Negar, Ahrends, Antje, Alatalo, Juha M., Alexander, Jake M., Nina Allonsius, Camille, Altman, Jan, Ammann, Christof, Andres, Christian, Andrews, Christopher, Ardö, Jonas, Arriga, Nicola, Arzac, Alberto, Aschero, Valeria, Assis, Rafael L., Johann Assmann, Jakob, Bader, Maaike Y., Bahalkeh, Khadijeh, Barančok, Peter, Barrio, Isabel C., Barros, Agustina, Barthel, Matti, Basham, Edmund W., Bauters, Marijn, Bazzichetto, Manuele, Belelli Marchesini, Luca, Bell, Michael C., Benavides, Juan C., Luis Benito Alonso, José, Berauer, Bernd J., Bjerke, Jarle W., Björk, Robert G., Björkman, Mats P., Björnsdóttir, Katrin, Blonder, Benjamin, Boeckx, Pascal, Boike, Julia, Bokhorst, Stef, Brum, Bárbara N. S., Brůna, Josef, Buchmann, Nina, Buysse, Pauline, Luís Camargo, José, Campoe, Otávio C., Candan, Onur, Canessa, Rafaella, Cannone, Nicoletta, Carbognani, Michele, Carnicer, Jofre, Casanova-Katny, Angélica, Cesarz, Simone, Chojnicki, Bogdan, Choler, Philippe, Chown, Steven L., Cifuentes, Edgar F., Ciliak, Marek, Contador, Tamara, Convey, Peter, Cooper, Elisabeth J., Cremonese, Edoardo, Curasi, Salvatore R., Curtis, Robin, Cutini, Maurizio, Johan Dahlberg, C., Daskalova, Gergana N., Angel de Pablo, Miguel, Della Chiesa, Stefano, Dengler, Jürgen, Deronde, Bart, Descombes, Patrice, Di Cecco, Valter, Di Musciano, Michele, Dick, Jan, Dimarco, Romina D., Dolezal, Jiri, Dorrepaal, Ellen, Dušek, Jiří, Eisenhauer, Nico, Eklundh, Lars, Erickson, Todd E., Erschbamer, Brigitta, Eugster, Werner, Ewers, Robert M., Exton, Dan A., Fanin, Nicolas, Fazlioglu, Fatih, Feigenwinter, Iris, Fenu, Giuseppe, Ferlian, Olga, Rosa Fernández Calzado, M., Fernández-Pascual, Eduardo, Finckh,

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# **1** Global maps of soil temperature

2 Running head: Global maps of soil temperature

3

4 Jonas J. Lembrechts<sup>1,\*,x,</sup> Johan van den Hoogen<sup>2,\*,</sup> Juha Aalto<sup>3,4</sup>, Michael B. Ashcroft<sup>5,6</sup>, Pieter De Frenne<sup>7</sup>, Julia Kemppinen<sup>8</sup>, Martin Kopecký<sup>9,10</sup>, Miska Luoto<sup>11</sup>, Ilya M. D. Maclean<sup>12</sup>, Thomas W. Crowther<sup>2</sup>, Joseph J. Bailey<sup>13</sup>, 5 Stef Haesen<sup>14</sup>, David H. Klinges<sup>15,16</sup>, Pekka Niittynen<sup>11</sup>, Brett R. Scheffers<sup>17</sup>, Koenraad Van Meerbeek<sup>14</sup>, Peter 6 7 Aartsma<sup>18</sup>, Otar Abdalaze<sup>19</sup>, Mehdi Abedi<sup>20</sup>, Rien Aerts<sup>21</sup>, Negar Ahmadian<sup>20</sup>, Antje Ahrends<sup>22</sup>, Juha M. Alatalo<sup>23</sup>, 8 Jake M. Alexander<sup>24</sup>, Camille Nina Allonsius<sup>25</sup>, Jan Altman<sup>9</sup>, Christof Ammann<sup>26</sup>, Christian Andres<sup>27</sup>, Christopher 9 Andrews<sup>28</sup>, Jonas Ardö<sup>29</sup>, Nicola Arriga<sup>30</sup>, Alberto Arzac<sup>31</sup>, Valeria Aschero<sup>32,33</sup>, Rafael L. Assis<sup>34</sup>, Jakob Johann Assmann<sup>35,36</sup>, Maaike Y. Bader<sup>37</sup>, Khadijeh Bahalkeh<sup>20</sup>, Peter Barančok<sup>38</sup>, Isabel C. Barrio<sup>39</sup>, Agustina Barros<sup>40</sup>, 10 Matti Barthel<sup>27</sup>, Edmund W. Basham<sup>15</sup>, Marijn Bauters<sup>41</sup>, Manuele Bazzichetto<sup>42</sup>, Luca Belelli Marchesini<sup>43</sup>, 11 Michael C. Bell<sup>44</sup>, Juan C. Benavides<sup>45</sup>, José Luis Benito Alonso<sup>46</sup>, Bernd J. Berauer<sup>47,48</sup>, Jarle W. Bjerke<sup>49</sup>, Robert 12 G. Björk<sup>50,51</sup>, Mats P. 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Johan Dahlberg<sup>83,84</sup>, Gergana N. Daskalova<sup>85</sup>, Miguel Angel de Pablo<sup>86</sup>, Stefano Della Chiesa<sup>87</sup>, Jürgen Dengler<sup>88,89,68</sup>, Bart Deronde<sup>90</sup>, Patrice 19 20 Descombes<sup>91</sup>, Valter Di Cecco<sup>92</sup>, Michele Di Musciano<sup>93</sup>, Jan Dick<sup>28</sup>, Romina D. Dimarco<sup>94,95</sup>, Jiri Dolezal<sup>9,96</sup>, Ellen 21 Dorrepaal<sup>97</sup>, Jiří Dušek<sup>98</sup>, Nico Eisenhauer<sup>68,69</sup>, Lars Eklundh<sup>29</sup>, Todd E. Erickson<sup>99,100</sup>, Brigitta Erschbamer<sup>101</sup>, 22 Werner Eugster<sup>27</sup>, Robert M. Ewers<sup>102</sup>, Dan A. Exton<sup>103</sup>, Nicolas Fanin<sup>104</sup>, Fatih Fazlioglu<sup>60</sup>, Iris Feigenwinter<sup>27</sup>, 23 Giuseppe Fenu<sup>105</sup>, Olga Ferlian<sup>68,69</sup>, M. Rosa Fernández Calzado<sup>106</sup>, Eduardo Fernández-Pascual<sup>107</sup>, Manfred 24 Finckh<sup>108</sup>, Rebecca Finger Higgens<sup>109</sup>, T'ai G. W. Forte<sup>64</sup>, Erika C. Freeman<sup>110</sup>, Esther R. Frei<sup>111,112,113</sup>, Eduardo Fuentes-Lillo<sup>114,1,115</sup>, Rafael A. García<sup>114,116</sup>, María B. García<sup>117</sup>, Charly Géron<sup>118</sup>, Mana Gharun<sup>27</sup>, Dany Ghosn<sup>119</sup>, 25 Khatuna Gigauri<sup>120</sup>, Anne Gobin<sup>121,122</sup>, Ignacio Goded<sup>30</sup>, Mathias Goeckede<sup>123</sup>, Felix Gottschall<sup>68,69</sup>, Keith 26 27 Goulding<sup>124</sup>, Sanne Govaert<sup>7</sup>, Bente Jessen Graae<sup>125</sup>, Sarah Greenwood<sup>126</sup>, Caroline Greiser<sup>83</sup>, Achim Grelle<sup>127</sup>, Benoit Guénard<sup>128</sup>, Mauro Guglielmin<sup>129</sup>, Joannès Guillemot<sup>130,131</sup>, Peter Haase<sup>132,133</sup>, Sylvia Haider<sup>134,68</sup>, Aud H. 28 29 Halbritter<sup>135</sup>, Maroof Hamid<sup>136</sup>, Albin Hammerle<sup>137</sup>, Arndt Hampe<sup>138</sup>, Siri V. 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Verheijen<sup>300</sup>, Luis Villar<sup>301</sup>, Luca Vitale<sup>302</sup>, Pascal Vittoz<sup>303</sup>, Maria Vives-Ingla<sup>66</sup>, Jonathan von Oppen<sup>35,36</sup>, Josefine Walz<sup>304</sup>, Runxi 74 Wang<sup>128</sup>, Yifeng Wang<sup>287</sup>, Robert G. Way<sup>287</sup>, Ronja E. M. Wedegärtner<sup>267</sup>, Robert Weigel<sup>149</sup>, Jan Wild<sup>9,162</sup>, Matthew 75 76 Wilkinson<sup>44</sup>, Martin Wilmking<sup>219</sup>, Lisa Wingate<sup>104</sup>, Manuela Winkler<sup>168</sup>, Sonja Wipf<sup>247,111</sup>, Georg Wohlfahrt<sup>137</sup>, Georgios Xenakis<sup>305</sup>, Yan Yang<sup>306</sup>, Zicheng Yu<sup>307,308</sup>, Kailiang Yu<sup>309</sup>, Florian Zellweger<sup>113</sup>, Jian Zhang<sup>310</sup>, Zhaochen 77 Zhang<sup>310</sup>, Peng Zhao<sup>155</sup>, Klaudia Ziemblińska<sup>215</sup>, Reiner Zimmermann<sup>207,311</sup>, Shengwei Zong<sup>312</sup>, Viacheslav I. 78 79 Zyryanov<sup>220</sup>, Ivan Nijs<sup>1,+,</sup> Jonathan Lenoir<sup>276,+,x</sup>

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- 81 \*Jonas J. Lembrechts and Johan Van den Hoogen should be considered joint first author
- 82 <sup>+</sup> Ivan Nijs and Jonathan Lenoir should be considered joint senior author
- 83 \* Corresponding authors
- 84 \*\* See end of manuscript for affiliations and OrcIDs

85

- 86 **Corresponding authors**
- 87 Jonas Lembrechts (Jonas.lembrechts@uantwerpen.be, https://orcid.org/0000-0002-1933-0750, +).
- 88 Jonathan Lenoir (jonathan.lenoir@u-picardie.fr, https://orcid.org/0000-0003-0638-9582).

89

## 90 Abstract

91 Research in global change ecology relies heavily on global climatic grids derived from estimates of air temperature in open areas at around 2 m above the ground. These climatic 92 93 grids do not reflect conditions below vegetation canopies and near the ground surface, where critical ecosystem functions occur and most terrestrial species reside. Here, we provide global 94 maps of soil temperature and bioclimatic variables at a 1-km<sup>2</sup> resolution for 0–5 and 5–15 cm 95 96 soil depth. These maps were created by calculating the difference (i.e., offset) between in-97 situ soil temperature measurements, based on time series from over 1200 1-km<sup>2</sup> pixels (summarized from 8500 unique temperature sensors) across all the world's major terrestrial 98 99 biomes, and coarse-grained air temperature estimates from ERA5-Land (an atmospheric 100 reanalysis by the European Centre for Medium-Range Weather Forecasts). We show that 101 mean annual soil temperature differs markedly from the corresponding gridded air 102 temperature, by up to  $10^{\circ}$ C (mean =  $3.0 \pm 2.1^{\circ}$ C), with substantial variation across biomes and 103 seasons. Over the year, soils in cold and/or dry biomes are substantially warmer  $(+3.6 \pm 2.3^{\circ}C)$ than gridded air temperature, whereas soils in warm and humid environments are on average 104 slightly cooler (-0.7 ± 2.3°C). The observed substantial and biome-specific offsets emphasize 105 that the projected impacts of climate and climate change on near-surface biodiversity and 106 107 ecosystem functioning are inaccurately assessed when air rather than soil temperature is 108 used, especially in cold environments. The global soil-related bioclimatic variables provided 109 here are an important step forward for any application in ecology and related disciplines. Nevertheless, we highlight the need to fill remaining geographic gaps by collecting more in-110 situ measurements of microclimate conditions to further enhance the spatiotemporal 111 resolution of global soil temperature products for ecological applications. 112

113

- 114 Keywords: bioclimatic variables, global maps, microclimate, near-surface temperatures, soil-
- dwelling organisms, soil temperature, temperature offset, weather stations

# 116 Introduction

With the rapidly increasing availability of big data on species distributions, functional traits 117 and ecosystem functioning (Bond-Lamberty & Thomson, 2018, Bruelheide et al., 2018, 118 Kissling et al., 2018, Kattge et al., 2019, Lenoir et al., 2020), we can now study biodiversity 119 and ecosystem responses to global changes in unprecedented detail (Senior et al., 2019, 120 Steidinger et al., 2019, Van Den Hoogen et al., 2019, Antão et al., 2020). However, despite 121 this increasing availability of ecological data, most spatially-explicit studies of ecological, 122 biophysical and biogeochemical processes still have to rely on the same global gridded 123 temperature data (Soudzilovskaia et al., 2015, Van Den Hoogen et al., 2019, Du et al., 2020). 124 125 Thus far, these global gridded products are based on measurements from standard 126 meteorological stations that record free-air temperature inside well-ventilated protective shields placed up to 2 m above-ground in open, shade-free habitats, where abiotic conditions 127 may differ substantially from those actually experienced by most organisms (World 128 Meteorological Organization, 2008, Lembrechts et al., 2020). 129

130 Ecological patterns and processes often relate more directly to below-canopy soil temperature rather than to well-ventilated air temperature inside a weather station. Near-131 surface, rather than air, temperature better predicts ecosystem functions like biogeochemical 132 cycling (e.g., organic matter decomposition, soil respiration and other aspects of the global 133 134 carbon balance) (Schimel et al., 2004, Pleim & Gilliam, 2009, Portillo-Estrada et al., 2016, Hursh et al., 2017, Gottschall et al., 2019, Davis et al., 2020, Perera-Castro et al., 2020, Jian et 135 al., 2021). Similarly, the use of soil temperature in correlative analyses or predictive models 136 may improve predictions of climate impacts on organismal physiology and behaviour, as well 137 as on population and community dynamics and species distributions (Körner & Paulsen, 2004, 138 Schimel et al., 2004, Ashcroft et al., 2008, Kearney et al., 2009, Scherrer et al., 2011, Opedal 139 et al., 2015, Berner et al., 2020, Zellweger et al., 2020). Given the key role of soil-related 140 141 processes for both aboveground and belowground parts of the ecosystem and their feedbacks to the atmosphere (Crowther et al., 2016), adequate soil temperature data are 142 critical for a broad range of fields of study, such as ecology, biogeography, biogeochemistry, 143 agronomy, soil science and climate system dynamics. Nevertheless, existing global soil 144 temperature products such as those from ERA5-Land (Copernicus Climate Change Service 145

(C3S), 2019), with a resolution of 0.08 × 0.08 degrees (≈ 9 × 9 km at the equator), remain too
coarse for most ecological applications.

The direction and magnitude of the difference or offset between in-situ soil temperature and 148 149 coarse-gridded air temperature products result from a combination of two factors: (i) the (vertical) microclimatic difference between air and soil temperature, and (ii) the (horizontal) 150 151 mesoclimatic difference between air temperature in flat, cleared areas (i.e., where 152 meteorological stations are located) and air temperature within different vegetation types 153 (e.g., below a dense canopy of trees) or topographies (e.g., within a ravine or on a ridge) 154 (Lembrechts et al., 2020, De Frenne et al., 2021). In essence, the offset is thus the combination 155 of both the vertical and horizontal differences that result from factors affecting the energy 156 budget at the Earth's surface, principally radiative energy: the ground absorbs radiative 157 energy, which is transferred to the air by convective heat exchange, evaporation and spatial variation in net radiation, and lower convective conductance near the Earth's surface results 158 159 in horizontal and vertical variation in temperature (Richardson, 1922, Geiger, 1950). Both these vertical and horizontal differences in temperature vary significantly across the globe 160 and in time as a result of environmental conditions affecting the radiation budget (e.g., as a 161 result of topographic orientation, canopy cover or surface albedo), convective heat exchange 162 163 and evaporation (e.g., foliage density, variation in the degree of wind shear caused by surface 164 friction) and the capacity for the soil to store and conduct heat (e.g., water content and soil 165 structure and texture) (Geiger, 1950, Zhang et al., 2008, Way & Lewkowicz, 2018, De Frenne et al., 2019). 166

167 While the physics of soil temperatures have long been well-understood (Richardson, 1922, 168 Geiger, 1950), the creation of high-resolution global gridded soil temperature products has not been feasible before, partially due to the absence of detailed global in-situ soil 169 170 temperature measurements (Lembrechts & Lenoir, 2019, Lembrechts et al., 2020). Recently, 171 however, the call for microclimate temperature data representative of *in-situ* conditions (i.e., 172 microhabitat) as experienced by organisms living close to the ground surface or in the soil has become more urgent (Bramer et al., 2018). In this paper, we address this issue by generating 173 174 global gridded maps of below-canopy and near-surface soil temperature at 1-km<sup>2</sup> resolution 175 (in line with most existing global air temperature products). These maps are more 176 representative of the habitat conditions as experienced by organisms living under vegetation

canopies, in the topsoil or near the soil surface. They were created using the abovementioned 177 offset between gridded air temperature data and *in-situ* soil temperature measurements. We 178 179 expect these soil temperature maps to be substantially more representative of actual 180 microclimatic conditions than existing products as they capture relevant near-surface and below-ground abiotic conditions where ecosystem functions and processes operate (Daly, 181 182 2006, Bramer et al., 2018, Körner & Hiltbrunner, 2018). Indeed, the offset between free-air (macroclimate) and soil (microclimate) temperature, and between cleared areas and other 183 habitats, can easily reach up to ±10°C annually, even at the 1-km<sup>2</sup> spatial resolution used here 184 185 (Zhang et al., 2018, Lembrechts et al., 2019, Wild et al., 2019).

186 To create the global gridded soil temperature maps introduced above, we used over 8500 187 time series of soil temperature measured *in-situ* across the world's major terrestrial biomes, 188 which are compiled and stored in the SoilTemp database (Lembrechts et al., 2020) (Fig. 1a, Supplementary Material Fig. S1) and averaged into 1200 (or 1000 for the second soil layer) 189 190 unique 1-km<sup>2</sup> pixels. First, to illustrate the magnitude of the studied effect, we visualized the global and biome-specific patterns in the mean annual offset between in-situ soil temperature 191 (0–5 cm and 5–15 cm depth) and coarse-scale interpolated air temperature from ERA5-Land 192 193 using the average within  $1 \times 1$  km grid cells. Hereafter, we refer to this difference between 194 soil temperature and air temperature as the *temperature offset* (or offset), sensu (De Frenne 195 et al., 2021); elsewhere called the surface offset (Smith & Riseborough, 1996, Smith & 196 Riseborough, 2002)). Secondly, we used a machine learning approach with 31 environmental predictor variables (including macroclimate, soil, topography, reflectance, vegetation and 197 anthropogenic variables) to model the spatial variation in monthly temperature offsets at a 1 198 199 × 1 km resolution for all continents except Antarctica (as not covered by many of the used 200 predictor variable layers). Using these offsets, we then calculated relevant soil-related bioclimatic variables (SBIO), mirroring the existing global bioclimatic variables for air 201 202 temperature. Finally, we compare our new global soil temperature product with a similar one 203 calculated using coarser-resolution soil temperature data (0.1 × 0.1 degrees) from ERA5-Land (Copernicus Climate Change Service (C3S), 2019). 204

# 205 Methods

## 206 Data acquisition

6

Analyses are based on SoilTemp, a global database of microclimate time series (Lembrechts *et al.*, 2020). We compiled soil temperature measurements from 9362 unique sensors (mean duration 2.9 years, median duration 1.0 year, ranging from 1 month to 41 years) from 60 countries, using both published and unpublished data sources (Fig. 1, Supplementary Material Fig. S1). Each sensor corresponds to one independent time series.

We used time series spanning a minimum of one month, with a temporal resolution of four 212 hours or less. Sensors of any type were included (Supplementary Material Table S1), as long 213 as they measured in situ. Sensors in experimentally manipulated plots, i.e., plots in which 214 215 microclimate has been manipulated, such as in open top chambers, were excluded. Most data (> 90%) came from low-cost rugged microclimate loggers such as iButtons (Maxim Integrated, 216 217 USA) or TMS4-sensors (Wild et al., 2019), with measurement errors of around 0.5–1°C (note 218 that we are using degree Celsius over Kelvin throughout, for ease of understanding), while in a minority of cases sensors with higher meteorological specifications such as industrial or 219 scientific grade thermocouples and thermistors (measurement errors of less than 0.5°C) were 220 used. Contributing datasets mostly consisted of short-term regional networks of microclimate 221 measurements, yet also included a set (< 5%) of soil temperature sensors from long-term 222 223 research networks equipped with weather stations (e.g., Pastorello et al., 2017). By combining 224 these two types of data, a much higher spatial density of sensors and broader distribution of 225 microhabitats could be obtained than by using weather station data only.

About 68% of sensors were deployed between 2010 and 2020 and 93% between 2000 and 2020; we thus focus on the latter period in our analyses. Additionally, given the relatively short time frame covered by most individual sensors and thus the lack of spatially unbiased long-term time series, we were not able to test for systematic differences in the temperature offset between old and recent data sets, and thus we did not correct for this in our models. We strongly urge future studies to assess such temporal dynamics in the offset once longterm microclimate data have become sufficient and more available.

For each of the individual 9362 time series, we calculated monthly mean, minimum (5% percentile of all monthly values) and maximum (95% percentile) temperature, after checking all time series for plausibility and erroneous data. These monthly values, while perhaps not fully intercomparable between the northern and southern hemisphere, are those that have

traditionally been used to calculate bioclimatic variables (Fick & Hijmans, 2017). Months with 237 more than one day of missing data, either at the beginning or end of the measurement period, 238 or due to logger malfunctioning during measurement, were excluded, resulting in a final 239 240 subset of 380,676 months of soil temperature time series that were used for further analyses. For each sensor with more than twelve months of data, we calculated moving averages of 241 242 annual mean temperature, using each consecutive month as a starting month and calculating the mean temperature including the next eleven months. We used these moving averages to 243 make maximal use of the full temporal extent covered by each sensor, because each time 244 245 series spanned a different time period, often including parts of calendar years only.

The selected dataset contained sensors installed strictly belowground, measuring temperature at depths between 0 and 200 cm below the ground surface. Sensors recording several measurements at the same site but located at different (vertical) depths were included separately (the 9362 unique sensors thus came from 7251 unique loggers).

Sensors were grouped in different soil depth categories (0–5, 5–15, 15–30, 30–60, 60–100, 100–200 cm, Supplementary Material Table S2) to incorporate the effects of soil temperature dampening associated with vertical stratification. We limited our analyses to the topsoil (0–5 cm) and the second soil layer (5–15 cm), as we currently lack sufficient global coverage to make accurate models at deeper soil depths (8519 time series, about 91%, came from the two upper depth layers). Due to uncertainty in identification of these soil depths between studies (e.g., due to litter layers), no finer categorisation is used.

We tested for potential bias in temporal resolution (i.e., measurement interval) by calculating mean, minimum and maximum temperature for a selection of 2000 months for data measured every 15 minutes, and the same data aggregated to 30, 60, 90, 120 and 240 minutes. Monthly mean, minimum and maximum temperature calculated with any of the aggregated datasets differed on average less than 0.2°C from the ones with the highest temporal resolution. We were thus confident that pooling data with different temporal resolutions of 4 hours or finer would not significantly affect our results.

264 **Temperature offset calculation** 

For each monthly value at each sensor location (see Supplementary Material Table S3 for 265 number of data points per month), we extracted the corresponding monthly means of the 2 266 m air temperature from the European Centre for Medium-Range Weather (ECMWF) 267 268 Forecast's 5<sup>th</sup> reanalysis (ERA5) (from 1979–1981) and ERA5-Land from 1981–2020 269 (Copernicus Climate Change Service (C3S), 2019), hereafter called ERA5L. The latter dataset 270 models the global climate with a spatial resolution of  $0.08 \times 0.08$  degrees ( $\approx 9 \times 9$  km at the 271 equator) with an hourly resolution, converted into monthly means using daily means for the whole month. Similarly, monthly minima and maxima were obtained from TerraClimate 272 273 (Abatzoglou *et al.*, 2018) for the period 2000 to 2020 at a 0.04  $\times$  0.04 degrees ( $\approx$  4  $\times$  4 km at 274 the equator) resolution. Monthly means for TerraClimate were not available, and we 275 therefore estimated them by averaging the monthly minima and maxima. Finally, we also 276 obtained monthly mean temperatures from CHELSA (Karger et al., 2017a, Karger et al., 2017b) 277 for the period 2000 to 2013 at a 30  $\times$  30 arc second ( $\approx$  1  $\times$  1 km at the equator) resolution. In 278 our modelling exercises (see section 'Integrative modelling' below), we opted to use the mean 279 temperature offsets as calculated based on ERA5L rather than on CHELSA. While CHELSA's higher spatial resolution is definitely an advantage, its time period (stopping in 2013) 280 281 insufficiently overlapped with the time period covered by our *in-situ* measurements (2000 to 282 2020), soil temperature offsets based on the CHELSA dataset were only used for comparative purposes. We used TerraClimate to model offsets in monthly minimum and maximum 283 temperature. 284

We calculated moving annual averages of the gridded air temperature data in the same way as for soil temperature. These were used to create annual temperature offset values following the same approach as above.

288 The offset between the *in situ* measured soil temperature in the SoilTemp database and the 2 m free-air temperature obtained from the air-temperature grids (ERA5L, TerraClim and 289 290 CHELSA, hereafter called 'gridded air temperature') was calculated by subtracting the 291 monthly or annual mean air temperature from the monthly or annual mean soil temperature. 292 Positive offset values indicate a measured soil temperature higher than gridded air temperature, while negative offset values represent cooler soils. Similarly, monthly minimum 293 294 and maximum air temperature were subtracted from minimum and maximum soil temperature, respectively. Monthly minima and maxima of the soil temperature were 295

calculated as, respectively, the 5% lowest and highest instantaneous measurement in that
month, to correct for outliers, which can be especially pronounced at the soil surface (Speak *et al.*, 2020). As a result, patterns in minima and maxima are more conservative estimates
than if we had used the absolute lowest and highest values.

300 Importantly, the temperature offset calculated here is a result of three key groups of drivers: (1) height effects (2 m versus 0–15 cm below the soil surface); (2) environmental or habitat 301 effects (e.g., spatial variability in vegetation, snow or topography); and (3) spatial scale effects 302 (resolution of gridded air temperature) (Lembrechts et al., 2020). We investigated the 303 304 potential role of scale effects by comparing gridded air temperature data sources with different resolutions (ERA5L, TerraClimate and CHELSA, see below). Height effects and 305 306 environmental effects are however not disentangled here, as the offset we propose 307 incorporates both the difference between air and soil temperature (vertically), as well as the difference between free-air macroclimate and in situ microclimate (horizontally) in one 308 measure (Lembrechts et al., 2020). While it can be argued that it would be better to treat 309 310 both vertical and horizontal effects separately, this would require a similar database of coupled *in-situ* air and soil temperature measurements, which is not yet available. Using *in* 311 312 situ measured air temperature could also solve spatial mismatches (i.e., spatially averaged air 313 temperature represents the whole 1 to 81 km<sup>2</sup> pixel, depending on pixel size, not only the 314 exact location of the sensor). However, coupled air and soil temperature measurements are not only rare, but the air temperature measurements also have large measurement errors, 315 316 especially in open habitats (Maclean et al., 2021). These errors can be up to several degrees in open habitats when using non-standardized sensors, loggers and shielding (Holden et al., 317 318 2013, Terando et al., 2017, Maclean et al., 2021). Hence, using in situ measured air temperature without correcting for these measurement errors would be misleading. 319

#### 320 Global and biome-level analyses

For the purpose of visualization, annual offsets were first averaged in hexagons with a resolution of approximately 70,000 km<sup>2</sup>, using the dggridR-package (version 2.0.4) in R (Barnes *et al.*, 2017) (Fig. 1). Next, we plotted mean, minimum and maximum annual soil temperature as a function of corresponding gridded air temperature from ERA5, TerraClimate and CHELSA and used generalized additive models (GAMs, package mgcv 1.8-31; Wood, 2012) to visualise deviations from the 1:1-line (i.e., temperature offsets deviating from zero,Supplementary Figs. S4-5).

328 All annual and monthly values within each soil depth category and falling within the same 1-329 km<sup>2</sup> pixel were aggregated as a mean, resulting in a total of c. 1200 unique pixels at 0–5 cm, 330 and c. 1000 unique pixels at 5–15 cm each month, across the globe (Supplementary Material Table S3). This averaging includes summarizing the data over space, i.e., multiple sensors 331 within the same 1-km<sup>2</sup> pixel, and time, i.e., data from multi-year time series from a certain 332 sensor, to reduce spatial and temporal autocorrelation and sampling bias. We assigned these 333 334 1-km<sup>2</sup> averages to the corresponding Whittaker biome of their georeferenced location, using the package *plotbiomes* (version 0.0.0.9901) in R (Fig. 1 c, d, Supplementary Material Table 335 336 S4-5 (Stefan & Levin, 2018)). We ranked biomes based on their offset and compared this with 337 the mean annual precipitation in each biome (Fig. 1b). This was done separately for each air temperature data source (ERA5L, TerraClimate and CHELSA), soil depth (0–5 cm, 5–15 cm) 338 and timeframe (ERA5L 1979–2020, 2000–2020), as well as for the offset between monthly 339 340 minimum and maximum soil temperature and the minimum and maximum gridded air temperature from TerraClimate. Our analyses showed that patterns were robust to variation 341 342 in spatial resolution, sensor depth, climate interpolation method and temporal scale (Supplementary Material Figs. S2–5). 343

## 344 Acquisition of global predictor variables

345 To create spatial predictive models of the offset between *in-situ* soil temperature and gridded air temperature, we first sampled a stack of global map layers at each of the logger locations 346 347 within the dataset. These layers included long-term macroclimatic conditions, soil texture and 348 physiochemical information, vegetation, radiation, and topographic indices as well as 349 anthropogenic variables. Details of all layers, including descriptions, units, and source information, are described in Supplementary Data S1. In short, information about soil texture, 350 351 structure and physiochemical properties was obtained from SoilGrids (version 1 (Hengl et al., 2017)), limited to the upper soil layer (top 5 cm). Long-term averages of macroclimatic 352 353 conditions (i.e., monthly mean, maximum and minimum temperature, monthly precipitation) 354 was obtained from CHELSA (version 2017 (Karger et al., 2017a)), which includes climate data 355 averaged across 1979–2013, and from WorldClim (version 2 (Fick & Hijmans, 2017)). Monthly

snow probability is based on a pixel-wise frequency of snow occurrence (snow cover >10%) 356 in MODIS daily snow cover products (MOD10A1 & MYD10A1 (Hall et al., 2002)) in 2001–2019. 357 Spectral vegetation indices (i.e., averaged MODIS NDVI product MYD13Q1) and surface 358 359 reflectance data (i.e., MODIS MCD43A4) were obtained from the Google Earth Engine Data Catalog (developers.google.com/earth-engine/datasets) and averaged from 2015 to 2019. 360 Landcover and topographic information were obtained from EarthEnv (Amatulli et al., 2018). 361 Aridity index (AI) and potential evapotranspiration (PET) layers were obtained from CGIAR 362 (Zomer *et al.*, 2008). Anthropogenic information (population density) was obtained from the 363 364 EU JRC (ghsl.jrc.ec.europa.eu/ghs\_pop2019.php). Aboveground biomass data were obtained 365 from GlobBiomass (Santoro, 2018). RESOLVE ecoregion classifications were used to 366 categorize sampling locations into biomes (Dinerstein et al., 2017). With this set of predictor 367 variables, we included information on all different categories of drivers of soil temperature. 368 An important variable that had to be excluded was snow depth, due to the lack of a relevant 369 1-km<sup>2</sup> resolution global product. The final set of predictor variables included 24 'static' 370 variables and eight monthly layers (i.e., maximum, mean, and minimum temperature, precipitation, cloud cover, solar radiation, water vapour pressure, and snow cover). As cloud 371 372 cover estimates were not available for high-latitude regions in the Northern Hemisphere in 373 January and December due to a lack of daylight, we excluded cloud cover as an explanatory 374 variable for these months (i.e., 'EarthEnvCloudCover MODCF monthlymean XX', with XX representing the months in two-digit form Supplementary Data S1). 375

All variable map layers were reprojected and resampled to a unified pixel grid in EPSG:4326 (WGS84) at 30 arc-sec resolution ( $\approx 1 \times 1$  km at the equator). Areas covered by permanent snow or ice (e.g., the Greenland ice cap or glaciated mountain ranges, identified using SoilGrids) were excluded from the analyses. Antarctic sampling points were excluded from the modelling data set owing to the limited coverage of several covariate layers in the region.

# 381 *Modelling*

To generate global maps of monthly temperature offsets (Fig. 2), we trained Random Forest (RF) models for each month, using the temperature offsets as the response variables and the global variable layers as predictors (Breiman, 2001, Hengl *et al.*, 2018). We used a geospatial RF modelling pipeline as developed by van den Hoogen *et al.* (2021). RF models are machine

learning models that combine many classification trees using randomized subsets of the data, 386 with each tree iteratively dividing data into groups of most closely related data points (Hengl 387 et al., 2018). They are particularly valuable here due to their capacity to uncover nonlinear 388 389 relationships (e.g., due to increased decoupling of soil from air temperature in colder and thus 390 snow-covered areas) and their ability to capture complex interactions among covariates (e.g., 391 between snow and vegetation cover) (Olden et al., 2008). Furthermore, they may currently 392 have advantages over mechanistic microclimate models for global modelling (Maclean & Klinges, 2021), as the latter require highly detailed physical input parameters for calibration, 393 394 and current computational barriers preclude global assessments at a 1 km<sup>2</sup> resolution and 395 over multiple decades. Nevertheless, we urge future endeavours to compare and potentially 396 improve our results with estimates based on such mechanistic models.

397 We performed a grid search procedure to tune the RF models across a range of 52 398 hyperparameter settings (variables per split: 2–14, minimum leaf population: 2–5, in all combinations adding up to 52 models, each time with 250 trees). During this procedure, we 399 400 assessed each of the 52 model's performance using k-fold cross-validation (k = 10; folds 401 assigned randomly, stratified per biome). The models' mean and standard deviation values were the basis for choosing the best of all evaluated models. This procedure was repeated for 402 403 each month separately for the two soil depth layers (0-5 cm, 5-15 cm), for offsets in mean, 404 minimum and maximum temperature. The importance of predictor variables was assessed using the variable importance and ordered by mean variable importance across all models. 405 406 This variable importance adds up the decreases in the impurity criterion (i.e., the measure on 407 which the local optimal condition is chosen) at each split of a node for each individual variable 408 over all trees in the forest (van den Hoogen *et al.*, 2021).

# 409 Soil bioclimatic variables

The resulting global maps of the annual and monthly offsets between mean, minimum and maximum soil and air temperature were used to calculate relevant bioclimatic variables following the definition used in CHELSA, BIOCLIM, ANUCLIM and WorldClim (Xu & Hutchinson, 2011, Booth *et al.*, 2014, Fick & Hijmans, 2017, Karger *et al.*, 2017a)(Fig. 3–4). First, we calculated monthly soil mean, maximum and minimum temperature by adding monthly temperature offsets to the respective CHELSA monthly mean, maximum and minimum temperature (Karger *et al.*, 2017a). Next, we used these soil temperature layers to compute
11 soil bioclimatic layers (SBIO, Table 1) (O'Donnell & Ignizio, 2012). Wettest and driest
quarters were identified for each pixel based on CHELSA's monthly values.

#### 419 *Model uncertainty*

420 To assess the uncertainty in the monthly models, we performed a stratified bootstrapping procedure, with total size of the bootstrap samples equal to the original training data (van 421 422 den Hoogen et al., 2021). Using biomes as a stratification category, we ensured the samples 423 included in each of the bootstrap training collections were proportionally representative of 424 each biome's total area. Next, we trained RF models (with the same hyperparameters as selected during the grid-search procedure) using each of 100 bootstrap iterations. Each of 425 426 these trained RF models was then used to classify the predictor layer stack, to generate perpixel 95% confidence intervals and standard deviation for the modelled monthly offsets (Fig. 427 5a, Supplementary Material Fig. S6a). The mean R<sup>2</sup> value of the RF models for the monthly 428 mean temperature offset was 0.70 (from 0.64 to 0.78) at 0-5 cm and 0.76 (0.63-0.85) at 5 to 429 15 cm across all twelve monthly models. Mean RMSE of the models was 2.20°C (1.94–2.51°C) 430 431 at 0–5 cm, and 2.06°C (1.67–2.35°C) at 5–15 cm.

Importantly, model uncertainty as reported in Fig. 5a and Supplementary Material Fig. S6a
comes on top of existing uncertainties in (1) *in-situ* soil temperature measurements and (2)
the ERA5L macroclimate models as used in our models. However, both of those are usually
under 1°C (Copernicus Climate Change Service (C3S), 2019, Wild *et al.*, 2019).

436 To assess the spatial extent of extrapolation, which is necessary due to the incomplete global 437 coverage of the training data, we first performed a Principal Component Analysis (PCA) on the full environmental space covered by the monthly training data, including all explanatory 438 variables as used in the models, and then transformed the composite image into the same PC 439 spaces as of the sampled data (Van Den Hoogen et al., 2019). Next, we created convex hulls 440 441 for each of the bivariate combinations from the first 10 to 12 PCs, covering at least 90% of the 442 sample space variation, with the number of PCs depending on the month. Using the coordinates of these convex hulls, we assessed whether each pixel fell within or outside each 443 444 of these convex hulls, and calculated the percentage of bivariate combinations for which this

was the case (Fig. 5b, Supplementary Material Fig. S6b). This process was repeated for eachmonth, and for each of the two soil depths separately.

447 These uncertainty maps are important because one should be careful with extrapolation 448 beyond the range of conditions covered by the environmental variables included in the 449 original calibration dataset, especially in the case of non-linear patterns such as modelled here. The maps are provided as spatial masks to remove or reduce the weighting of the pixels 450 for which predictions are beyond the range of values covered by the models during 451 calibration. To assess this further, we used a spatial leave-one-out cross-validation analysis to 452 test for spatial autocorrelation in the data set (Supplementary Material Fig. S7) (van den 453 Hoogen et al., 2021). This approach trains a model for each sample in the data set on all 454 455 remaining samples, excluding data points that fall within an increasingly large buffer around 456 that focal sample. Results show lowest confidence for May to September at 5–15 cm, likely driven by uneven global coverage of data points. 457

458 Finally, we compared the modelled mean annual temperature (SBIO1, topsoil layer) with a similar product based on monthly ERA5L topsoil (0-7 cm) temperature with a spatial 459 460 resolution of 0.1 × 0.1 degrees (Copernicus Climate Change Service (C3S), 2019). The corresponding SBIO1 based on ERA5L was calculated using the means of the monthly 461 462 averages for each month over the period 1981 to 2016, and averaging these 12 monthly values into one annual product. We then visualized spatial differences between SBIO1 and 463 464 ERA5, as well as differences across the macroclimatic gradient, to identify mismatches between both datasets. 465

All geospatial modelling was performed using the Python API in Google Earth Engine (Gorelick *et al.*, 2017). The R statistical software, version 4.0.2 (R Core Team, 2020), was used for data visualisations. All maps were plotted using the Mollweide projection, which preserves relative areas, to avoid large distortions at high latitudes.

# 470 Sources of uncertainty

The temporal mismatch between the period covered by CHELSA (1979-2013) and our *in-situ* measurements (2000-2020) prevented us from directly using CHELSA climate to calculate the temperature offsets used in our models. This temporal mismatch might affect the offsets

calculated here because the relationship between temperature offset and macroclimate will 474 change through time as the climate warms. Similarly, inter-annual differences in offsets due 475 476 to specific weather conditions cannot be implemented in the used approach. However, we 477 are confident that, at the relatively coarse spatial (1 km<sup>2</sup>) and temporal (monthly averages) 478 resolution we are working at, our results are sufficiently robust to withstand these temporal 479 issues, given that we found high consistency in offset patterns between the different 480 timeframes and air temperature datasets examined (Supplementary Material Figs. S2–5). 481 Nevertheless, we strongly urge future research to disentangle these potential temporal 482 dynamics, especially given the increasing rate at which the climate is warming (Xu et al., 2018, 483 GISTEMP Team, 2021).

484 Similarly, a potential bias could result from the mismatch in method and resolution between 485 ERA5L – used to calculate the temperature offsets – and CHELSA, which was used to create the bioclimatic variables. However, even though temperature offsets have slightly larger 486 variation when based on the coarser-grained ERA5L-data than on the finer-grained CHELSA-487 488 data, Supplementary Material Figs. S2–5 show that relationships between soil and air temperature are largely consistent in all biomes and across the whole global temperature 489 490 gradient. Therefore, the larger offsets created additional random scatter, yet no consistent 491 bias.

Finally, we acknowledge that the 1-km<sup>2</sup> resolution gridded products might not be 492 493 representative of conditions at the *in-situ* measurement locations within each pixel. This issue 494 could be particularly significant for different vegetation types (here proxied at the pixel level using total aboveground biomass (unit: tons/ha i.e., Mg/ha, for the year 2010; Santoro, 2018) 495 and NDVI (MODIS NDVI product MYD13Q1, averaged over 2015–2019)). To verify this, we 496 497 compared a pixel's estimated aboveground biomass with the dominant *in-situ* habitat (forest 498 versus open) surrounding the sensors in that pixel (Supplementary Table S6). Importantly, all 499 sensors installed in forests fell indeed in pixels with more than 1 ton/ha aboveground 500 biomass. Similarly, 75% or more of sensors in open terrain fell in pixels with biomass estimates 501 of less than 1 ton/ha. Only in the temperate woodland biome was the match between *in-situ* habitat estimates and pixel-level aboveground biomass lower, with less than 95% of sensors 502 503 in forested locations correctly placed in pixels with more than 1 ton/ha biomass, and less than 50% of open terrain sensors in pixels with less than 1 ton/ha biomass. While our predictions 504

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will thus not be accurate for locations within a pixel that largely deviate from average
conditions (e.g., open terrain in pixels identified as largely forested, or vice versa), they should
be largely representative for those pixel-level averages.

# 508 **Results**

# 509 Biome-wide patterns in the temperature offset

510 We found positive and negative temperature offsets of up to 10°C between in situ measured 511 mean annual topsoil temperature and gridded air temperature (mean =  $3.0 \pm 2.1$ °C standard deviation, Fig. 1, 0–5 cm depth; 5–15 cm is available in Supplementary Material Figs. S2, 5). 512 The magnitude and direction of these temperature offsets varied considerably within and 513 across biomes. Mean annual topsoil temperature was on average 3.6 ± 2.3°C higher than 514 515 gridded air temperature in cold and/or dry biomes, namely tundra, boreal forests, temperate grasslands, and subtropical deserts. In contrast, offsets were slightly negative in warm and 516 517 wet biomes (tropical savannas, temperate forests, and tropical rainforests) where soils were, 518 on average, 0.7 ± 2.7°C cooler than gridded air temperature (Fig. 1b, Supplementary Material Figs. S2 and 5; note, however, the lower spatial coverage in these biomes in Fig. 1a, c, d, 519 Supplementary Material Table S4). Temperature offsets in annual minimum and maximum 520 521 temperature amounted to c. 10°C maximum. While annual soil temperature minima were on 522 average higher than corresponding gridded air temperature minima in all biomes, temperature offsets of annual maxima followed largely the same biome-related trends as 523 524 seen for the annual means, albeit with the higher variability expected for temperature 525 extremes (Supplementary Material Figs. S2g, S2h, S4g, S4h). Using different air temperature data sources did not alter the annual temperature offset and biome-related patterns (see 526 527 Methods and Supplementary Material Figs. S2–5).

528 Soils in the temperate seasonal forest biome were on average 0.8°C (± 2.2°C) cooler than air 529 temperature within 1-km<sup>2</sup> grid cells of forested habitats, and 1.0°C (± 4.0°C) warmer than the 530 air within 1-km<sup>2</sup> grid cells of non-forested habitats, resulting in a biome-wide average of 0.5°C 531 (Supplementary Material Table S7). Similar patterns were observed in other biomes.

## 532 Temporal and spatial variation in temperature offsets

Our Random Forest outputs highlighted a strong seasonality in monthly temperature offsets, 533 especially towards higher latitudes (Fig. 2). High-latitude soils were found to be several 534 degrees warmer than the air (monthly offsets of up to 25°C) during their respective winter 535 536 months, and cooler (up to 10°C) in summer months, both at 0–5 cm (Fig. 2) and 5–15 cm 537 (Supplementary Material Fig. S8) soil depths. In the tropics and subtropics, soils in dry biomes 538 (e.g., in the Sahara Desert or southern Africa) were predicted to be warmer than air throughout most of the year, whilst soils in mesic biomes (e.g., tropical biomes in South 539 America, central Africa and Southeast Asia) were modelled to be consistently cooler, at both 540 541 soil depths. These global gridded products were then used to create temperature-based 542 global bioclimatic variables for soils (SBIO, Fig. 3, Supplementary Material Fig. S9).

#### 543 *Global variation in soil temperature*

544 We observed 17% less spatial variation in mean annual soil temperature globally (expressed 545 by the standard deviation) than in air temperature, largely driven by the positive offset between soil and air temperature in cold environments (Fig. 4). Importantly, our machine 546 learning models slightly (up to 1°C, or around 10% of variation) underestimated temperature 547 offsets at both extremes of the temperature gradient at the 1-km<sup>2</sup> resolution (Supplementary 548 549 Material Fig. S10) and likely even more in comparison with finer-resolution products. Estimates of the reduction in variation across space are thus conservative, especially in the 550 551 coldest biomes. The reduction in spatial temperature variation was observed in all cold and 552 cool biomes, with tundra and boreal forests having both a significant positive mean temperature offset and a reduction of 20% and 22% in variation, respectively (Fig. 4c). In the 553 warmest biomes (e.g., tropical savanna and subtropical desert), however, we found an 554 555 increase in variation of, on average, 10%.

556 Our bootstrap approach to validate modelled monthly offsets indicated high consistency 557 among the outcomes of 100 bootstrapped models (Fig. 5, Supplementary Material Fig. S6a), 558 with standard deviations in most months and across most parts of the globe around or below 559  $\pm 1^{\circ}$ C. One exception to this was the temperature offset at high latitudes of the northern 560 hemisphere during winter months (standard deviation up to  $\pm 5^{\circ}$ C in the 0–5 cm layer). 561 Predictive performance was comparable across biomes, although with large variation in data 562 availability (Supplementary Material Fig. S11). The importance of predictor variables in the RF models was largely consistent across months. Macroclimatic variables such as incoming solar radiation as well as long-term averages in air temperature and precipitation were by far the most influential explanatory variables in the spatial models of the monthly temperature offset (Supplementary Material Figs. S12, 13).

567 We highlight that the current availability of *in-situ* soil temperature measurements is 568 significantly lower in the tropics (Supplementary Material Table S5), where our model had to 569 extrapolate temperatures beyond the range used to calibrate the model (Fig. 5b, 570 Supplementary Material Fig. S6b).

571 Finally, our comparison with a mean annual soil temperature product derived from the coarse-resolution ERA5L topsoil temperature showed that spatial variability, e.g., driven by 572 573 topographic heterogeneity, is much better captured here than in the coarser resolution of the 574 ERA5L-based product (Fig. 6c-e). Nevertheless, our predictions at the coarse scale showed to 575 be condensed within a 5°C range of values from the ERA5L-predictions, for more than 95% of pixels globally. Noteworthy, our predictions resulted in consistently cooler soil temperature 576 577 predictions than topsoil conditions provided by ERA5L across large areas, such as the boreal and tropical forest biomes (Fig. 6a, b). Additionally, our models predicted lower values for 578 579 SBIO1 than ERA5L in all regions with mean annual soil temperature below 0°C, except for a few locations around Greenland and Svalbard (Fig. 6a, b). 580

# 581 **Discussion**

## 582 Global patterns in soil temperature

583 We observed large spatiotemporal heterogeneity in the global offset between soil and air 584 temperature, often in the order of several degrees annually and up to more than 20°C during 585 winter months at high latitudes. These values are in line with empirical data from regional studies (Zhang et al., 2018, Lembrechts et al., 2019, Obu et al., 2019). Both annual and 586 monthly offsets showed clear discrepancies between cold and dry versus warm and wet 587 588 biomes. The modelled monthly offsets covaried strongly negatively with both long-term averages in free-air temperature and solar radiation, linking to the well-known decoupling of 589 soil from air temperature due to snow (for cold extremes in cold and cool biomes) (Grundstein 590 591 et al., 2005). However, the secondary importance of variables related to precipitation and soil

structure hints to the additional distinction between wet and dry biomes at the warm end of 592 the temperature gradient. There, buffering due to shading, evapotranspiration and the 593 594 specific heat of water (mostly against warm extremes in warm and wet biomes) results in 595 cooler soil temperature (Geiger, 1950, Grundstein et al., 2005, Hennon et al., 2010, Wang & 596 Dickinson, 2012, De Frenne et al., 2013, Grünberg et al., 2020), while such buffering is not as 597 strong in warm and dry biomes due to the lower water availability (Wang & Dickinson, 2012, Greiser et al., 2018, Zhou et al., 2021). As such, these results highlight strong macroclimatic 598 impacts on the soil microclimate across the globe (see also De Frenne et al., 2019), yet with 599 600 soil temperature importantly non-linearly related to air temperature at the global scale. This 601 confirms that the latter is not sufficient as a proxy for temperature conditions near or in the 602 soil. With our soil-specific global bioclimatic products, we have provided the means to correct 603 for these important region-specific, non-linear differences between soil and air temperature 604 at an unprecedented spatial resolution.

#### 605 Drivers of the temperature offset

606 Our empirical modelling approach enabled us to accurately map global patterns in soil 607 temperature. In doing so we did not aim to disentangle the mechanisms governing the 608 temperature offset: such an endeavour would require modelling the biophysics of energy 609 exchange at the soil surface across biomes (Kearney et al., 2019, Maclean et al., 2019, 610 Maclean & Klinges, 2021). Importantly, many of the predictor variables used in our study (e.g., 611 long-term averages in macroclimatic conditions or solar radiation) are unlikely to represent 612 direct causal relationships underlying the temperature offset, but may rather indirectly relate to many ensuing factors that affect the functioning of ecosystems at fine spatial scales which, 613 614 in turn, feedback on local temperature offsets, such as energy and water balances, snow cover, wind intensity and vegetation cover (De Frenne et al., 2021). For example, while 615 616 increased solar radiation itself would theoretically result in soils warming more than the air, 617 high solar radiation at the global scale often coincides with high vegetation cover blocking 618 radiation input to the soil, thus correlating with relatively cooler soils (De Frenne *et al.*, 2021). 619 Our results highlight, however, that the complex relationship between microclimatic soil 620 temperature and macroclimatic air temperature is predictable across large spatial extents 621 thanks to broad scale patterns, even if this is governed by a multitude of local-scale factors 622 involving fine spatiotemporal resolutions. Nevertheless, the predictive quality of our models

was lower in high latitude regions, where high variation in the *in situ* measured offsets – likely
driven by the interactions between snow, local topography and vegetation – reduced
predictive power of the models at the 1-km<sup>2</sup> resolution (Greiser *et al.*, 2018, Way &
Lewkowicz, 2018, Grünberg *et al.*, 2020, Myers-Smith *et al.*, 2020, Niittynen *et al.*, 2020).

# 627 Implications for microclimate warming

Our results highlight clear biome-specific differences in mean annual temperature between 628 629 air and soil temperatures, as well as a significant reduction in the spatial variation in temperature in the soil or near the soil surface, especially in cold and cool biomes (Fig. 4). 630 631 These patterns remain even despite the presence of often strongly opposing monthly offset trends (Fig. 2). The observed correlation between long-term averages in macroclimatic 632 conditions and the annual temperature offset illustrates that soil temperature is unlikely to 633 634 warm at the same rate as air temperature when macroclimate warms. Indeed, one degree of 635 air temperature warming could result in either a bigger or smaller soil temperature change, depending on where along the macroclimatic gradient this is happening. These effects might 636 637 be seen in cold biome soils most strongly, as they not only experience the largest (positive) temperature offsets and reductions in climate range compared to air temperature (Fig. 4b, c), 638 but they are also expected to experience the strongest magnitude of macroclimate warming 639 (Cooper, 2014, Overland et al., 2014, Chen et al., 2021, GISTEMP Team, 2021). As a result, 640 641 mean annual temperatures in cold climate soils can be expected to warm slower than the 642 corresponding macroclimate as offsets shrink with increasing macroclimate warming.

643 Contrastingly, predicted climate warming in hot and dry biomes could be amplified in the topsoil, where we show soils to become increasingly warmer than the air at higher 644 temperatures. Similarly, changes in precipitation regimes – and thus soil moisture – can 645 646 significantly alter the relationship between air and soil temperature, with critical implications for soil moisture-atmosphere feedbacks, especially in hot biomes (Zhou et al., 2021). Indeed, 647 as precipitation decreases, offsets could turn more positive and soil temperatures might 648 warm even faster than the observed macroclimate warming. Therefore, future research 649 650 should not only use soil temperature data as provided here to study belowground ecological processes (De Frenne et al., 2013, Lembrechts et al., 2020), it should also urgently investigate 651 652 future scenarios of soil climate warming in light of changing air temperature and precipitation,

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at ecologically relevant spatial and temporal resolutions to incorporate the non-linearrelationships exposed so far (Lembrechts & Nijs, 2020).

#### 655 Within-pixel heterogeneity

656 We chose to use a 1-km<sup>2</sup> resolution spatial grid to model mismatches between soil and air 657 temperature, aggregating all values from different microhabitats within the same 1-km<sup>2</sup> grid cell (e.g., sensors in forested versus open patches) as well as all daily and diurnal variation 658 659 within a month. Additionally, we used coarse-grained free-air temperature rather than in-situ 660 measured air temperatures. We are aware that higher spatiotemporal resolutions would 661 likely reveal the importance of locally heterogeneous variables. Finer-scale factors that affect the local radiation balance and wind (e.g., topography, snow and vegetation cover, 662 urbanization) at the landscape to local scales and those that directly affect neighbouring 663 664 locations (e.g. topographic shading and cold-air drainage, Whiteman, 1982, Ashcroft & Gollan, 665 2012, Lembrechts et al., 2020) would probably have emerged as more important drivers at regional scales and with higher spatiotemporal resolutions than those used here 666 (Supplementary Material Fig. S12). The latter is illustrated by the multi-degree Celsius 667 difference in mean annual temperature between forested and non-forested locations within 668 the same biome (Supplementary Material Table S7), as well as the lower accuracy obtained 669 during winter months at high latitudes, where and when fine-scale spatial heterogeneity in 670 671 snow cover and depth probably lowers models' predictability at the 1-km<sup>2</sup> resolution. *In-situ* 672 measurements were largely from areas with a representative vegetation type, supporting the reliability of our predictions for the dominant habitat type within a pixel. However, improved 673 674 accuracy at high latitudes will depend on the future development of high-resolution snow depth and/or snow water equivalent estimates (Luojus *et al.*, 2010). 675

676 The SoilTemp database (Lembrechts et al., 2020) will facilitate the necessary steps towards mapping soil temperature at higher spatiotemporal resolutions in the future, with its 677 georeferenced time series of in situ measured soil and near-surface temperature and 678 679 associated metadata. Nevertheless, when compared to existing soil temperature products 680 such as those from ERA5L (Copernicus Climate Change Service (C3S), 2019), we emphasize that the increased resolution of our data products already provides a major technical 681 682 advance, even though substantial finer within-pixel variation is still lost through spatiotemporal aggregation. 683

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## 684 Conclusions

685 The spatial (biome-specific) and temporal (seasonally variable) offsets between air and soil 686 temperature quantified here likely bias predictions of current and future climate impacts on 687 species and ecosystems (Körner & Paulsen, 2004, Kearney et al., 2009, Cooper, 2014, Opedal et al., 2015, Graae et al., 2018, Zellweger et al., 2020, Bergstrom et al., 2021). Temperature 688 689 in the topsoil rather than in the air ultimately defines the distribution and performance of 690 most terrestrial species, as well as many ecosystem functions at or below the soil surface (Pleim & Gilliam, 2009, Portillo-Estrada et al., 2016, Hursh et al., 2017, Gottschall et al., 2019). 691 692 As many ecosystem functions are highly correlated with temperature (yet often non-lineary, 693 Johnston et al., 2021), soil temperature rather than air temperature should in those instances 694 be the preferred predictor for estimating their rates and temperature thresholds (Rosenberg 695 et al., 1990, Coûteaux et al., 1995, Schimel et al., 1996). Correcting for the non-linear 696 relationship between air and soil temperature identified here is thus vital for all fields 697 investigating abiotic and biotic processes relating to terrestrial environments (White et al., 2020). Indeed, soil temperature, macroclimate and land-use change will interact to define the 698 699 future climate as experienced by organisms, and high-resolution soil temperature data is 700 needed to tackle current and future challenges.

701 By making our global soil temperature maps and the underlying monthly offset data openly 702 available, we offer gridded soil temperature data for climate research, ecology, agronomy 703 and other life and environmental sciences. Future research has the important task of further 704 improving the spatial and temporal resolution of global microclimate products as 705 microclimate operates at much higher temporal resolutions, with temporal variation over 706 hours, days, seasons and years (Potter et al., 2013, Bütikofer et al., 2020), as well as to confirm 707 accuracy of predictions in undersampled regions in the underlying maps (Lembrechts et al., 708 2021). However, we are convinced that the maps presented here bring us one step closer to 709 having accessible climate data exactly where it matters most for many terrestrial organisms 710 (Kearney & Porter, 2009, Ashcroft et al., 2014, Pincebourde et al., 2016, Niittynen & Luoto, 2018, Lembrechts & Lenoir, 2019). We nevertheless highlight that there is still a long way to 711 712 go towards global soil microclimate data with an optimal spatiotemporal resolution. We 713 therefore urge all scientists to submit their microclimate time series to the SoilTemp database

- to fill data gaps and help to increase the spatial resolution until it matches with the scale at
- which ecological processes take place (Bütikofer *et al.*, 2020, Lembrechts *et al.*, 2020).

716

# 717 Data availability

- All monthly data to train the models and reproduce the figures, sampled covariate data, and
- 719 models are available at <u>https://doi.org/10.5281/zenodo.4558663</u>. Soil bioclim layers SBIO1-
- 720 11 are also directly available in Google Earth Engine under
- 721 projects/crowtherlab/soil\_bioclim/soil\_bioclim\_0\_5cm and
- 722 projects/crowtherlab/soil\_bioclim/soil\_bioclim\_5\_15cm.
- 723

# 724 Code availability

- All source code is available at <u>https://doi.org/10.5281/zenodo.4558663</u>.
- 726

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# 906 Affiliations

907 1) Research Group PLECO (Plants and Ecosystems), University of Antwerp, 2610 Wilrijk, Belgium, 2) Department of Environmental Systems 908 Science, Institute of Integrative Biology, ETH Zürich, Zürich, Switzerland, 3) Finnish Meteorological Institute, P.O. Box 503, FI-00101 909 Helsinki, Finland, 4) Department of Geosciences and Geography, Gustaf Hällströmin katu 2a, FIN-00014 University of Helsinki, Finland, 5) 910 Centre for Sustainable Ecosystem Solutions, School of Biological Sciences, University of Wollongong, Wollongong, Australia, 6) Australian 911 Museum, Sydney, Australia, 7) Forest & Nature Lab, Department of Environment, Ghent University, Geraardsbergsesteenweg 267, 9090 912 Melle-Gontrode, Belgium, 8) Geography Research Unit, University of Oulu, Oulu, Finland, 9) Institute of Botany of the Czech Academy of 913 Sciences, Zámek 1, CZ-25243, Průhonice, Czech Republic, 10) Faculty of Forestry and Wood Sciences, Czech University of Life Sciences 914 Prague, Kamýcká 129, CZ-165 21, Prague 6 - Suchdol, Czech Republic, 11) Dept of Geosciences and Geography, Gustaf Hällströmin katu 2a, 915 FIN–00014 Univ. of Helsinki, Finland, 12) Environment and Sustainability Institute, University of Exeter, Penryn Campus, Penryn, UK, TR10 916 9FE, 13) Department of Geography, York St John University, Lord Mayor's Walk, York, YO31 7EX, United Kingdom, 14) Department of Earth 917 and Environmental Sciences, KU Leuven, Celestijnenlaan 200E, 3001 Leuven, Belgium, 15) School of Natural Resources and Environment, 918 University of Florida, Gainesville, FL 32611, USA, 16) Smithsonian Environmental Research Center, Edgewater MD 21037 USA, 17) 919 Department of Wildlife Ecology and Conservation, University of Florida, Gainesville, FL 32611, USA, 18) Department of Natural Sciences 920 and Environmental Health, University of South-Eastern Norway, Gullbringvegen 36, NO-3800, Bø, Norway, 19) Alpine Ecosystems Research 921 Program, Institute of Ecology, Ilia State University, Tbilisi, Georgia, 20) Department of Range Management, Faculty of Natural Resources 922 and Marine Sciences, Tarbiat Modares University, Noor, Mazandaran Province, I. R. Iran, 21) Department of Ecological Science, Vrije 923 Universiteit Amsterdam, The Netherlands., 22) Royal Botanic Garden Edinburgh, 20A Inverleith Row, EH3 5LR, Edinburgh, UK, 23) 924 Environmental Science Center, Qatar University, Doha, Qatar, 24) Department of Environmental Systems Science, Institute of Integrative 925 Biology, ETH Zurich, Universitätsstrasse 16, CH-8092 Zürich, Switzerland, 25) Research group ECOBE, University of Antwerp, 2610 Wilrijk, 926 Belgium, 26) Department of Agroecology and Environment, Agroscope Research Institute, Reckenholzstrasse 191, 8046 Zürich, 927 Switzerland, 27) Department of Environmental Systems Science, ETH Zurich, Universitaetstrasse 2, 8092 Zurich, Switzerland, 28) UK Centre 928 for Ecology & Hydrology, Bush Estate, Penicuik, Midlothian, EH26 0QB, United Kingdom, 29) Department of Physical Geography and 929 Ecosystem Science, Lund University, Sölvegatan 12, 223 62 Lund, Sweden, 30) European Commission, Joint Research Centre (JRC), Ispra, 930 Italy, 31) Siberian Federal University, 660041 Krasnoyarsk, Russia, 32) Instituto Argentino de Nivología, Glaciología y Ciencias Ambientales 931 (IANIGLA), CONICET, CCT-Mendoza; Facultad de Ciencias Exactas y Naturales, Universidad Nacional de Cuyo, 33) Instituto Argentino de 932 Nivologiá, Glaciologiá y Ciencias Ambientales (IANIGLA), CONICET, CCT-Mendoza, 34) Natural History Museum, University of Oslo, 0318, 933 Oslo, Norway, 35) Section for Ecoinformatics & Biodiversity, Department of Biology, Aarhus University, Aarhus C, Denmark, 36) Center for 934 Biodiversity Dynamics in a Changing World, Department of Biology, Aarhus University, Aarhus C, Denmark, 37) Ecological Plant Geography, 935 Faculty of Geography, University of Marburg, Deutschhausstrasse 10, 35032, Marburg, Germany, 38) Institute of Landscape Ecology Slovak 936 Academy of Sciences, Štefánikova 3, 81499 Bratislava, Slovakia, 39) Faculty of Environmental and Forest Sciences, Agricultural University of 937 Iceland, Árleyni 22, 112 Reykjavík, Iceland, 40) Instituto Argentino de Nivología, Glaciología y Ciencias Ambientales (IANIGLA), CONICET, 938 CCT-Mendoza, 41) Isotope Bioscience Laboratory - ISOFYS, Ghent University, Coupure Links 653, 9000 Gent, Belgium, 42) Université de 939 Rennes, CNRS, EcoBio (Ecosystèmes, biodiversité, évolution) - UMR 6553, F-35000 Rennes, France, 43) Department of Sustainable Agro-940 ecosystems and Bioresources, Research and Innovation Centre, Fondazione Edmund Mach, Via E. Mach 1, 38010 San Michele all'Adige, 941 Italy, 44) Forest Research, Alice Holt Lodge, Wrecclesham, Farnham, UK, 45) Department of Ecology, Pontificia Universidad Javeriana, 942 Bogota, Colombia, 46) Jolube Consultor Botánico. C/Mariano R de Ledesma, 4. E-22700 Jaca, Huesca, SPAIN, 47) Institute of Landscape and 943 Plant Ecology, Department of Plant Ecology, University of Hohenheim, Ottilie-Zeller\_weg 2, 70599 Stuttgart, Germany, 48) Disturbance 944 Ecology, BayCEER, University of Bayreuth, Universitätsstr. 30, 95447 Bayreuth, Germany, 49) Norwegian Institute for Nature Research, 945 FRAM - High North Research Centre for Climate and the Environment, P.O. Box 6606 Langnes, N-9296 Tromsø, Norway, 50) Department of 946 Earth Sciences, University of Gothenburg, P.O. Box 460, SE-40530 Gothenburg, Sweden, 51) Gothenburg Global Biodiversity Centre, P.O. 947 Box 461, SE-405 30 Gothenburg, Sweden, 52) Department of Biological and Environmental Sciences, University of Gothenburg, P.O. Box 948 461, 43 Gothenburg SE-405 30, Sweden, 53) Department of Environmental Science, Policy, and Management, University of California,

949 Berkeley, CA 94720 USA, 54) Alfred Wegener Institute Helmholtz Center for Polar and Marine Research, Telegrafenberg A45, 14473 950 Potsdam, Germany, 55) Geography Department, Humboldt-Universität zu Berlin, Germany, 56) Pós-Graduação em Ciências de Florestas 951 Tropicais, Instituto Nacional de Pesquisas da Amazônia, Manaus, Brasil, CEP: 69060-001, 57) UMR ECOSYS INRAE, AgroParisTech, 952 Uinversité Paris Saclay, France, 58) Biological Dynamics of Forest Fragments Project, BDFFP, Instituto Nacional de Pesquisas da Amazônia, 953 Av. André Araujo, 2936 - Petrópolis, Manaus, Amazonas, 69067-375, Brazil, 59) Department of Forest Sciences, Federal University of 954 Lavras, 37.200-900, Lavras, MG, Brazil, 60) Faculty of Arts and Sciences, Department of Molecular Biology and Genetics, Ordu University, 955 52200, Ordu, Turkey, 61) Ecological Plant Geography, Faculty of Geography, University of Marburg, Deutschhausstrasse 10, 35032, 956 Marburg, Germany., 62) Plant Ecology Group, Department of Evolution and Ecology, University of Tübingen, Auf der Morgenstelle 5, 72076 957 Tübingen, Germany, 63) Department of Science and High Technology, Insubria University, Via Valleggio 11, 22100 Como, Italy, 64) 958 Department of Chemistry, Life Sciences and Environmental Sustainability, University of Parma, Parco Area delle Scienze 11/A, 43124 959 Parma, Italy, 65) Department of Evolutionary Biology, Ecology and Environmental Sciences, Biodiversity Research Institute (IRBio), 960 University of Barcelona, 08028 Barcelona, Catalonia, Spain, 66) CREAF, E08193 Bellaterra (Cerdanyola del Vallès), Catalonia, Spain, 67) 961 Laboratorio de Ecofisiología Vegetal y Cambio Climático, Laboratorio de Ecofisiología Vegetal y Cambio Climático, Departamento de 962 Ciencias Veterinarias y Salud Pública, Universidad Católica de Temuco, Campus Luis Rivas del Canto and Núcleo de Estudios Ambientales 963 (NEA), Facultad de Recursos Naturales, Universidad Católica de Temuco, Temuco, 4780000, Chile, 68) German Centre for Integrative 964 Biodiversity Research (iDiv) Halle-Jena-Leipzig, Leipzig, Germany, 69) Institute of Biology, Leipzig University, Leipzig, Germany, 70) 965 Laboratory of Bioclimatology, Department of Ecology and Environmental Protection, Poznan University of Life Siences, ul. Piatkowska 94, 966 60-649, Poznan, Poland, 71) Univ. Grenoble Alpes, Univ. Savoie Mont Blanc, CNRS, LECA, F-38000 Grenoble, France, 72) Univ. Grenoble 967 Alpes, Univ. Savoie Mont Blanc, CNRS, LTSER Zone Atelier Alpes, F-38000 Grenoble, France, 73) Securing Antarctica's Environmental 968 Future, School of Biological Sciences, Monash University, Victoria 3800, Australia, 74) Forest Ecology and Conservation Group, Department 969 of Plant Sciences, University of Cambridge, Cambridge CB23EA, UK, 75) Faculty of Ecology and Environmental Sciences, Technical 970 University in Zvolen, T. G. Masaryka 24, 960 01 Zvolen, Slovakia, 76) Sub-Antarctic Biocultural Conservation Program, Universidad de 971 Magallanes, Pdte. Manuel Bulnes 01855, Punta Arenas, Magallanes y la Antártica Chilena, 77) Núcleo Milenio de Salmónidos Invasores, 972 INVASAL, Concepción, Chile, 78) British Antarctic Survey, NERC, High Cross, Madingley Road, Cambridge CB3 0ET, United Kingdom, 79) 973 Department of Arctic and Marine Biology, Faculty of Biosciences Fisheries and Economics, UIT-The Arctic University of Norway, N-9037 974 Tromsø, Norway, 80) Climate Change Unit, Environmental Protection Agency of Aosta Valley, Italy, 81) Department of Biological Sciences, 975 University of Notre Dame, Notre Dame, IN 46556, USA, 82) Department of Science, University of Roma Tre, 00146 Rome, Italy, 83) 976 Department of Ecology, Environment and Plant Sciences and Bolin Centre for Climate Research, Stockholm University, 106 91 Stockholm, 977 Sweden, 84) the County Administrative Board of Västra Götaland, SE-403 40 Gothenburg, Sweden, 85) School of GeoSciences, University 978 of Edinburgh, King's Buildings, Edinburgh, EH9 3FF, United Kingdom, 86) Department of Geology, Geography and Environment. University 979 of Alcalá. 28805 Alcalá de Henares, Madrid, Spain., 87) Chair of Geoinformatics, Technische Universität Dresden, Dresden, Germany, 88) 980 Vegetation Ecology, Institute of Natural Resource Sciences (IUNR), ZHAW Zurich University of Applied Sciences, Grüentalstr. 14, 8820 981 Wädenswil, Switzerland, 89) Plant Ecology, Bayreuth Center of Ecology and Environmental Research (BayCEER), University of Bayreuth, 982 Universitätsstr. 30, 95447 Bayreuth, Germany, 90) VITO-TAP, Boeretang 200, 2400-Mol, Belgium, 91) Swiss Federal Research Institute WSL, 983 8903 Birmensdorf, Switzerland, 92) Majella Seed Bank, Majella National Park, Colle Madonna, 66010 Lama dei Peligni, Italy, 93) 984 Department of Life, Health and Environmental Sciences, University of L'Aquila, Piazzale Salvatore Tommasi 1, 67100 L'Aquila, Italy, 94) 985 Grupo de Ecología de Poblaciones de Insectos, IFAB (INTA - CONICET), Modesta Victoria 4450, Bariloche, Argentina, 95) Department of 986 Biology and Biochemistry, University of Houston, Houston, Texas, 77204 USA, 96) Faculty of Science, Department of Botany, University of 987 South Bohemia, Na Zlaté Stoce 1, 37005 České Budějovice, Czech Republic, 97) Climate Impacts Research Centre, Department of Ecology 988 and Environmental Science, Umeå University, Abisko, Sweden, 98) Global Change Research Institute, Academy of Sciences of the Czech 989 Republic, 99) School of Biological Sciences, The University of Western Australia, Crawley, WA 6009, Australia, 100) Kings Park Science, 990 Department of Biodiversity, Conservation & Attractions, Kings Park, 6005 WA, Australia, 101) Department of Botany, Faculty of Biology, 991 University of Innsbruck, Sternwartestraße 15, 6020 Innsbruck, Austria, 102) Imperial College London, Silwood Park Campus, Ascot SL5 7PY, 992 UK, 103) Operation Wallacea, Wallace House, Old Bolingbroke, Lincolnshire, PE23 4EX, UK, 104) INRAE, Bordeaux Sciences Agro, UMR 993 1391 ISPA, F-33140 Villenave d'Ornon, France, 105) Department of Life and Environmental Sciences, University of Cagliari, Viale 994 Sant'Ignazio da Laconi 13, 09123, Cagliari, Italy., 106) Department of Botany, University of Granada, 18071, Granada, Spain, 107) IMIB – 995 Biodiversity Research Institute, University of Oviedo, Mieres, Spain, 108) Institute for Plant Science and Microbiology, University of 996 Hamburg, Ohnhorststr. 18, 22609 Hamburg, Germany, 109) Dartmouth College, Hanover, NH, USA, 110) Ecosystems and Global Change 997 Group, Department of Plant Sciences, University of Cambridge, Cambridge, CB2 3EA, United Kingdom, 111) WSL Institute for Snow and 998 Avalanche Research SLF, 7260 Davos, Switzerland, 112) Climate Change, Extremes and Natural Hazards in Alpine Regions Research Center 999 CERC, Flüelastrasse 11, 7260 Davos Dorf, Switzerland, 113) Swiss Federal Research Institute for Forest, Snow and Landscape Research WSL, 1000 8903 Birmensdorf, Switzerland, 114) Laboratorio de Invasiones Biológicas (LIB), Facultad de Ciencias Forestales, Universidad de 1001 Concepción, Concepción, Chile, 115) School of Education and Social Sciences, Adventist University of Chile, Chile, 116) Instituto de Ecología 1002 y Biodiversidad (IEB), Santiago, Chile, 117) Pyrenean Institute of Ecology (CSIC), Av. Montañana 1005, 50059 Zaragoza, Spain, 118) 1003 Biodiversity and Landscape, TERRA research centre, Gembloux Agro-Bio Tech, University of Liège, Gembloux, 5032, Belgium ; Research 1004 Group PLECO (Plants and Ecosystems), University of Antwerp, 2610 Wilrijk, Belgium, 119) Department of Geo-information in 1005 Environmental Management, Mediterranean Agronomic Institute of Chania, PO Box 85, 73100 Chania, Greece, 120) Georgian Institute of 1006 Public Affairs, department of Environmental management ad policy, Tbilisi, Georgia, 121) Flemish Institute for Technological Research, 1007 2400 Mol, Belgium, 122) Department of Earth and Environmental Science, Faculty of BioScience Engineering, KULeuven, Belgium, 123) 1008 Max Planck Institute for Biogeochemistry, Department of Biogeochemical Signals, Jena, Germany, 124) Sustainable Agricultural Sciences 1009 Department, Rothamsted Research, Harpenden, AL5 2JQ, UK, 125) Department of Biology, Norwegian University of Science and 1010 Technology, 7034 Trondheim, Norway, 126) Biodiversity, Wildlife and Ecosystem Health, Biomedical Sciences, University of Edinburgh, 1011 Edinburgh, EH8 9JZ, UK, 127) Department of Ecology, Swedish University of Agricultural Sciences, Box 7042, S-750 07 Uppsala, 128) School 1012 of Biological Sciences, The University of Hong Kong, Pok Fu Lam Road, Hong Kong SAR, China, 129) Department of Theoretical and Applied 1013 Sciences, Insubria University, Via Dunant 3, 21100 Varese, Italy, 130) CIRAD, UMR Eco&Sols, 34060 Montpellier, France, 131) Eco&Sols, 1014 Univ Montpellier, CIRAD, INRAE, IRD, Montpellier SupAgro, 34060 Montpellier, France, 132) Senckenberg Research Institute and Natural

1015 History Museum Frankfurt, 63571 Gelnhausen, Germany, 133) Faculty of Biology, University of Duisburg-Essen, 45141 Essen, Germany, 1016 134) Institute of Biology / Geobotany and Botanical Garden, Martin Luther University Halle-Wittenberg, Halle (Saale), Germany, 135) 1017 Department of Biological Sciences and Bjerknes Centre for Climate Research, University of Bergen, N-5020 Bergen, Norway, 136) Centre 1018 for Biodiversity & Taxonomy, Department of Botany, University of Kashmir, Srinagar - 190006, J&K, India, 137) Department of Ecology, 1019 University of Innsbruck, 6020 Innsbruck, Austria, 138) INRAE, Univ. Bordeaux, BIOGECO, F-33610 Cestas, France, 139) The Heathland 1020 Centre, Alver, Norway, 140) TERRA Teaching and Research Center, Faculty of Gembloux Agro-Bio Tech, University of Liege, Passage des 1021 déportés, 2, 5030 Gembloux, Belgium, 141) UK Centre for Ecology & Hydrology, Penicuik, EH26 0QB, Scotland, UK, 142) Vegetation 1022 Ecology, Institute of Natural Resource Sciences, ZHAW Zurich University of Applied Sciences, Grüental, 8820 Wädenswil, Switzerland, 143) 1023 Institute for Botany, University of Natural Resources and Life Sciences Vienna (BOKU), Gregor-Mendel-Straße 33/I, 1180 Vienna, Austria, 1024 144) Centre for Agrometeorological Research (ZAMF), German Meteorological Service (DWD), Bundesallee 33, 38116 Braunschweig, 1025 Germany, 145) Dept of Biology, Memorial University, St. John's, NL, A1B 3X9. Canada, 146) Department of Biological Sciences, Simon 1026 Fraser University, Burnaby, BC, V5A 1S6, Canada, 147) Department of Geography, University of Zaragoza, Pedro Cerbuna 12, 50009 1027 Zaragoza, Spain, 148) Faculty of Resource Management, HAWK University of Applied Sciences and Arts, 37077 Göttingen, Germany, 149) 1028 Plant Ecology, Albrecht-von-Haller-Institute for Plant Sciences, Georg-August University of Goettingen, Untere Karspuele 2, 37073 1029 Goettingen, Germany, 150) Department of Ecoscience and Arctic Research Centre, Aarhus University, Grenåvej 14, 8410 Rønde, Denmark, 1030 151) Department of Geography, Masaryk University, Faculty of Science, Kotlarska 2, 611 37, Brno, Czech Republic, 152) Department of 1031 Environmental Science, Shinshu University, Matsumoto, Japan, 153) Department of Ecoscience and Arctic Research Centre, Aarhus 1032 University, Frederiksborgvej 399, 4000 Roskilde, Denmark, 154) INRAE, University of Bordeaux, BIOGECO, F-33610 Cestas, France, 155) 1033 Department of Forest Ecology and Management, Swedish University of Agricultural Sciences, 90183 Umeå, Sweden, 156) Forest Research 1034 Institute, Department of Silviculture and Forest Tree Genetics, Braci Lesnej Street, No 3, Sekocin Stary, 05-090 Raszyn, Poland, 157) 1035 Bayreuth Center of Ecology and Environmental Research, 158) ARAID/IPE-CSIC, Pyrenean Institute of Ecology, Avda. Llano de la Victoria, 1036 16, Jaca 22700, Spain, 159) Life and Environmental Sciences, University of Iceland, Sturlugata 7, 102 Reykjavík, Iceland, 160) School of 1037 Biological Sciences, University of Bristol, Bristol, United Kingdom, 161) Biological and Environmental Sciences, Faculty of Natural Sciences, 1038 University of Stirling, Scotland, FK9 4LA, 162) Faculty of Environmental Sciences, Czech University of Life Sciences Prague, Kamýcká 129, 1039 165 21 Prague 6 - Suchdol, Czech Republic, 163) Centre for Environmental and Climate Science, Lund University, Sölvegatan 37, 223 62, 1040 Lund, Sweden, 164) University of Goettingen, Bioclimatology, Büsgenweg 2, 37077 Göttingen, Germany., 165) Environment Agency 1041 Austria, Spittelauer Lände 5, 1090 Vienna, Austria, 166) Centre for Ecological Research, Institute of Ecology and Botany, H-2163 Vácrátót, 1042 Alkotmány út 2-4., Hungary, 167) Experimental Plant Ecology, Institute of Botany and Landscape Ecology, University of Greifswald, D-1043 17487 Greifswald, Germany, 168) GLORIA Coordination, Institute for Interdisciplinary Mountain Research, Austrian Academy of Sciences 1044 (ÖAW) & Department of Integrative Biology and Biodiversity Research, University of Natural Resources and Life Sciences, Vienna (BOKU), 1045 Silbergasse 30/3, 1190 Vienna, Austria, 169) Department of Arctic Biology, The University Centre in Svalbard (UNIS), 9171 Longyearbyen, 1046 Svalbard, Norway, 170) Department of Land Resources and Environmental Sciences, Montana State University, Bozeman MT, USA, 59717, 1047 171) Climate Impacts Research Centre, Department of Ecology and Environmental Sciences, Umeå University, Vetenskapens väg 38, 98107 1048 Abisko, Sweden, 172) Centre for Polar Ecology, Faculty of Science, University of South Bohemia, Na Zlaté Stoce 3, 370 05, České 1049 Budějovice, Czech Republic, 173) School of Biological Sciences, Monash University, Victoria 3800, Australia, 174) Terrestrial Ecology Unit, 1050 Dept. of Biology, Ghent University, B-9000 Gent, Belgium, 175) Finnish Meteorological Institute, Climate System Research, PoB503, 00101 1051 Helsinki, Finland, 176) INAR Institute for Atmospheric and Earth System Research/Physics, Faculty of Science, POBox 68 FI-00014 1052 University of Helsinki, Finland, 177) Interuniversity Institute for Earth System Research, University of Granada, Granada 18006 Spain, 178) 1053 CNR Institute for Agricultural and Forestry Systems in the Mediterranean, P.le Enrico Fermi 1 - Loc. del Granatello, 80055, Portici (Napoli) 1054 Italy, 179) Faculty of Forestry, Technical University in Zvolen, T.G.Masaryka 24, 960 01 Zvolen, Slovakia, 180) CNR Insitute for Agricultural 1055 and Forestry Systems in the Mediterranean, P.le Enrico Fermi 1 - Loc. del Granatello, 80055, Portici (Napoli) Italy, 181) School of Pure & 1056 Applied Sciences, Environmental Conservation & Management Programme Open University of Cyprus, PO Box 12794, 2252 Latsia, Nicosia, 1057 182) Department of Biology - Aquatic Biology, Aarhus University, Ole Worms Allé 1, 8000 Aarhus C, Denmark, 183) Aarhus Institute of 1058 Advanced Studies, AIAS Høegh-Guldbergs Gade 6B, 8000 Aarhus, Denmark, 184) Department of Forest Botany, Dendrology and 1059 Geobiocoenology, Faculty of Forestry and Wood Technology, Mendel University in Brno, Zemedelska 1, 613 00 Brno, Czech Republic, 185) 1060 Regional Centre for Integrated Environmental Monitoring, Odesa National I.I. Mechnikov University, 7 Mayakovskogo lane, 65082 Odesa, 1061 Ukraine, 186) Department of Agroecology, Aarhus University, 20 Blichers Allé, 8830 Tjele, Denmark, 187) NGO New Energy, 11 Bakulina 1062 str., 61166 Kharkiv, Ukraine, 188) Biological Dynamics of Forest Fragments Project, Coordenação de Dinâmica Ambiental, Instituto 1063 Nacional de Pesquisas da Amazônia, Manaus, AM CEP 69067-375, Brazil., 189) Swiss Federal Institute for Forest, Snow and Landscape 1064 Research (WSL), CH-8903 Birmensdorf, Switzerland., 190) Department of Biology, University of Antwerp, Universiteitsplein 1, 2610 Wilrijk, 1065 Belgium, 191) Department of Botany and Biodiversity Research Centre, University of British Columbia, Vancouver, BC, Canada, 192) 1066 Department of Environment, Province of Antwerp, Koningin Elisabethlei 22, 2018 Antwerpen, Belgium, 193) Institute of Plant and Animal 1067 Ecology of Ural Division of Russian Academy of Science, 8 Marta st., 202, Ekaterinburg, Russia, 194) Department of Earth and 1068 Environmental Sciences, University of Pavia, Via S. Epifanio 14, Pavia, Italy, 195) Faculty of Science and Technology, Free University of 1069 Bolzano, Piazza Università 5, 39100 Bolzano, Italy, 196) Climate Change Unit, Environmental Protection Agency of Aosta Valley, Loc. La 1070 Maladière, 48, 11020 Saint-Christophe, Italy, 197) University of Freiburg, Chair of Geobotany, Schänzlestrasse 1, 79104 Freiburg, Germany, 1071 198) Environment and Sustainability Institute, University of Exeter, Penryn Campus, Cornwall TR10 9FE, United Kingdom, 199) Centre for 1072 Ecosystem Science, School of Biological, Earth and Environmental Sciences, UNSW Sydney, NSW 2052, Sydney, Australia, 200) Department 1073 of Plant Biology and Ecology, University of Seville , 41012 Seville, Spain, 201) Department of Biology, Washington University in St. Louis, St. 1074 Louis, MO 63130, USA., 202) Department of Animal Biology, Institute of Biology, University of Campinas, Campinas, SP, CEP 13083-862, 1075 Brazil, 203) CNR Institute of BioEconomy, Via Gobetti 101, 40129 Bologna, Italy, 204) National Wildlife Research Centre, Environment and 1076 Climate Change Canada, Carleton University, 1125 Colonel by Drive, Ottawa, ON K1A 0H3, Canada, 205) School of Life and Environmental 1077 Sciences, Deakin University, Burwood, Victoria, Australia, 3125, 206) Institute for Alpine Environment, Eurac Research, Viale Druso 1, 1078 39100 Bozen/Bolzano, Italy, 207) Institute of Biology, Dept. of Molecular Botany, University of Hohenheim, 70599 Stuttgart, Germany, 1079 208) Instituto de Matemática Aplicada San Luis, IMASL, CONICET and Universidad Nacional de San Luis, Ejército de los Andes 950, 1080 D5700HHW San Luis, Argentina, 209) Cátedra de Climatología Agrícola (FCA-UNER), Ruta 11, km 10, Oro Verde, Entre Ríos, Argentina, 210)

1081 Grupo de Ecología de Invasiones, INIBIOMA, CONICET/ Universidad Nacional del Comahue, Av. de los Pioneros 2350, Bariloche 8400, 1082 Argentina, 211) CSIC, Global Ecology Unit CREAF- CSIC-UAB, Bellaterra, 08193, Catalonia, Spain., 212) CREAF, E08193, Cerdanvola del 1083 Vallès, Catalonia, Spain, 213) Mountains of the Moon University, P.O Box 837, Fort Portal, Uganda, 214) National Agricultural Research 1084 Organisation, Mbarara Zonal Agricultural Research and Development Institute, P.O Box 389, Mbarara , Uganda, 215) Laboratory of 1085 Meteorology, Department of Construction and Geoengineering, Faculty of Environmental Engineering and Mechanical Engineering, 1086 Poznan University of Life Siences, ul. Piatkowska 94, 60-649, Poznan, Poland, 216) Department of Agroecology, Aarhus University, Blichers 1087 Allé 20, 8830 Tjele, Denmark, 217) Department of Biology, Lund University, SE-223 62 Lund, Sweden, 218) Department of Earth and 1088 Environmental Sciences, University of Pavia, Via S. Epifanio 14, 27100 Pavia, Italy, 219) Institute of Botany and Landscape Ecology, 1089 University Greifswald, D-17487 Greifswald, Germany, 220) V.N. Sukachev Institute of Forest SB RAS, Krasnoyarsk, Russia, 221) Institute of 1090 Ecology and Earth Sciences, University of Tartu, Liivi 2, Tartu 50409, Estonia, 222) Department of Biology, Aarhus University, Ole Worms 1091 Allé 1, 8000 Aarhus C, Denmark, 223) Department of Biology and Ecology Center, Utah State University, 5305 Old Main Hill, Logan, UT 1092 84322, USA, 224) Department of Life Sciences, Imperial College, Silwood Park Campus, Ascot, Berkshire SL5 7PY, UK, 225) Landscape 1093 Ecology, Institute of Terrestrial Ecosystems, Department of Environmental Systems Science, ETH Zürich, 8092 Zürich, Switzerland, 226) 1094 Unit of Land Change Science, Swiss Federal Research Institute WSL, 8903 Birmensdorf, Switzerland, 227) Department of Biology, 1095 Washington University in St. Louis, Campus Box 1137, 1 Brookings Drive, St. Louis, MO 63130 USA, 228) CREAF, Cerdanyola del Vallès 1096 08193, Catalonia, Spain, 229) School of Ecology and Environment Studies, Nalanda University, Rajgir, India, 230) School of Biosciences, 1097 University of Sheffield, Western Bank, Sheffield, S10 2TN, U.K., 231) CESAM & Department of Environment, University of Aveiro, 3810-193 1098 Aveiro, Portugal, 232) Department of Agronomy, Food, Natural resources, Animals and Environment - University of Padua, 35020 Legnaro, 1099 Italy, 233) Univ. Savoie Mont Blanc, CNRS, Univ. Grenoble Alpes, EDYTEM, F-73000 Chambéry, France, 234) Universitat Autònoma de 1100 Barcelona, E08193 Bellaterra (Cerdanyola del\r\nVallès), Catalonia, Spain, 235) Department of Ecology and Biogeography, Faculty of 1101 Biological and Veterinary Sciences, Nicolaus Copernicus University, Toruń, Poland, 236) Centre for Climate Change Research, Nicolaus 1102 Copernicus University, Toruń, Poland, 237) A. Borza Botanic Garden, Babeş-Bolyai University, Cluj-Napoca, Romania, 238) Faculty of 1103 Biology and Geology, Department of Taxonomy and Ecology, Babeş-Bolyai University, Cluj-Napoca, Romania, 239) E. G. Racoviță Institute, 1104 Babes-Bolyai University, Cluj-Napoca, Romania, 240) Centre for Sustainable Ecosystem Solutions, School of Earth, Atmospheric and Life 1105 Sciences, University of Wollongong, Wollongong, New South Wales, 2522, Australia, 241) University of Applied Sciences Trier, 1106 Environmental Campus Birkenfeld, 55761 Birkenfeld, Germany, 242) Institut Universitaire de France, 1 Rue Descartes, 75231 Paris cedex 1107 05, France, 243) Swiss Federal Institute for Forest, Snow and Landscape Research WSL, Zuercherstrasse 111, 8903 Birmensdorf, 1108 Switzerland, 244) Securing Antarctica's Environmental Future, School of Earth, Atmospheric and Life Sciences, University of Wollongong, 1109 2522 Australia, 245) Aquatic Ecology & Environmental Biology, Radboud Institute for Biological and Environmental Sciences Water and 1110 Wetland Research, Faculty of Science, Radboud University Nijmegen, 6525 NJ Nijmegen, The Netherlands., 246) University of Notre Dame, 1111 Department of Biological Sciences and the Environmental Change Initiative, 247) Swiss National Park, Chastè Planta-Wildenberg, 7530 1112 Zernez, Switzerland, 248) Remote Sensing Laboratories, Dept. of Geography, University of Zurich, Winterthurerstrasse 190, 8057 Zurich, 1113 Switzerland, 249) CIRAD, UMR Eco&Sols, BP1386, CP18524, Dakar, Senegal, 250) Eco&Sols, Univ Montpellier, CIRAD, INRAE, IRD, Institut 1114 Agro, Montpellier, France, 251) LMI IESOL, Centre IRD-ISRA de Bel Air, BP1386, CP18524, Dakar, Senegal, 252) Parc national des Ecrins -1115 Domaine de Charance - 05000 GAP - France, 253) Universidad Nacional de San Antonio Abad del Cusco, Cusco, Perú, 254) Centro de 1116 Investigación de la Biodiversidad Wilhelm L. Johannsen, Cusco, Perú, 255) Biological Dynamics of Forest Fragments Project, PDBFF, 1117 Instituto Nacional de Pesquisas da Amazônia, Av. André Araujo, 2936 - Petrópolis, Manaus, Amazonas, 69067-375, Brazil, 256) Department 1118 of Ecology and Environmental Science, Umeå University, 901 87 Umeå, Sweden, 257) Institute of Bio- and Geosciences (IBG-3): 1119 Agrosphere, Forschungszentrum Jülich GmbH, Jülich, Germany, 258) Chair of Soil Science and Geomorphology, Department of 1120 Geosciences, University of Tuebingen, 72070 Tuebingen, Germany, 259) Department of Geography, The University of British Columbia, 1121 Vancouver, BC V6T 122, 260) Department of Botany and Biodiversity Research, Rennweg 14, 1030 Vienna, 261) Princeton School of Public 1122 and International Affairs, Princeton University, Princeton, NJ 08540, USA, 262) Université de Lorraine, AgroParisTech, INRAE, Silva, 54000 1123 Nancy, France., 263) Department of Soil Science and Landscape Management, Faculty of Earth Sciences and Spatial Management, Nicolaus 1124 Copernicus University, Toruń, Poland, 264) Terra Nova National Park, Parks Canada Agency, Glovertown NL, A0G3Y0, 265) Universidade 1125 Estadual do Norte Fluminense Darcy Ribeiro, Campos dos Goytacazes, Rio de Janeiro, Brazil, 266) National Forest Centre, Forest Research 1126 Institute Zvolen, T. G. Masaryka 22, 96001 Zvolen, Slovakia, 267) Department of Biology, Norwegian University of Science and Technology, 1127 7491 Trondheim, Norway, 268) Department of Physical Geography, Stockholm University, 106 91 Stockholm, Sweden, 269) Department of 1128 Geography, University of British Columbia, 1984 West Mall, Vancouver, BC V6T 122, 270) Department of Earth and Environmental 1129 Sciences, Celestijnenlaan 200E, 3001 Leuven, Belgium, 271) Soil Science Department, Federal University of Viçosa, Prof. Peter Henry Rolfs 1130 Ave., 36570-900, Viçosa-MG, Brazil, 272) Universidade Federal da Paraíba, Departamento de Geociências. Cidade Universitária, João 1131 Pessoa - PB, CEP 58051-900, Brasil, 273) Goethe-Universität Frankfurt, Department of Physical Geography, Altenhöferallee 1, 60438 1132 Frankfurt am Main, Germany, 274) Department of Evolution, Ecology, and Organismal Biology, University of California Riverside, Riverside, 1133 CA, 92521, USA, 275) Department of Natural History, NTNU University Museum, Norwegian University of Science and Technology, NO-1134 7491 Trondheim Norway, 276) UMR 7058 CNRS 'Ecologie et Dynamique des Systèmes Anthropisés' (EDYSAN), Univ. de Picardie Jules 1135 Verne, Amiens, France, 277) EnvixLab, Dipartimento di Bioscienze e Territorio, Università degli Studi del Molise, Via Duca degli Abruzzi 1136 s.n.c., 86039 Termoli, Italy, 278) Institute of Meteorology and Climate Research (IMK), Department of Atmospheric Environmental 1137 Reserach (IFU), Karlsruhe Institute of Technology (KIT), Kreuzeckbahn Straße 19, 82467 Garmisch-Partenkirchen, Germany, 279) Swedish 1138 University of Agricultural Sciences, SLU Swedish Species Information Centre, Almas allé 8 E, 75651 Uppsala, Sweden, 280) University 1139 Duisburg-Essen, Faculty for Biology, Universitätsstr. 5, 45141 Essen, Germany, 281) Department of Geosciences and Natural Resource 1140 Management, University of Copenhagen, Øster Voldgade 10, DK-1350 Copenhagen, Denmark, 282) Experimental Plant Ecology, Institute 1141 of Botany and Landscape Ecology, University of Greifswald, partner in the Greifswald Mire Centre, D-17487 Greifswald, Germany, 283) 1142 Fondation J.-M. Aubert, 1938 Champex-Lac, Switzerland, 284) Département de Botanique et Biologie végétale, Université de Genève, Case 1143 postale 71, CH-1292 Chambésy, Switzerland, 285) Department of Geography and Earth Sciences, Aberystwyth University, Wales, UK, 286) 1144 Center for Systematic Biology, Biodiversity and Bioresources - 3B, Babes-Bolyai University, Cluj-Napoca, Romania, 287) Northern 1145 Environmental Geoscience Laboratory, Department of Geography and Planning, Queen's University, 288) Finnish Meteorological Inst., P.O. 1146 Box 503, FI-00101 Helsinki, Finland, 289) Graduate School of Life and Environmental Sciences, Osaka Prefecture University, 599-8531,

1147 Japan, 290) Nature Research Centre, Akademijos 2, 08412 Vilnius, Lithuania, 291) Institute of Biological Research Cluj-Napoca, National 1148 Institute of Research and Development for Biological Sciences, Bucharest, Romania, 292) CNR Institute for BioEconomy, Via Giovanni 1149 Caproni, 50144 Firenze, Italy, 293) The Ecosystem Management Research Group (ECOBE), University of Antwerp, 2610 Wilrijk 1150 (Antwerpen), Belgium, 294) Plant Conservation and Population Biology, Department of Biology, KU Leuven, Kasteelpark Arenberg 31, 3001 1151 Heverlee, Belgium, 295) A.N. Severtsov Institute of Ecology and Evolution, Russian Academy of Sciences, 119071, Leninsky pr.33, Moscow, 1152 Russia, 296) Netherlands Institute of Ecology, Droevendaalsesteeg 10, 6708 PB, Wageningen, the Netherlands, 297) Plant Ecology & 1153 Nature Conservation Group Wageningen University, Droevendaalse Steeg 3a 6708 PB Wageningen, 298) Centre for Integrative Ecology, 1154 School of Life and Environmental Sciences, Deakin University, Burwood, Victoria, Australia, 3125, 299) CAVElab - Computational and 1155 Applied Vegetation Ecology, Department of Environment, Ghent University, Coupure Links 653, 9000 Gent, Belgium, 300) Earth Surface 1156 Processes Team, Centre for Environmental and Marine Studies (CESAM), Dept. Environment and Planning, University of Aveiro, 3810-193, 1157 Aveiro, Portugal, 301) Instituto Pirenaico de Ecología, IPE-CSIC. Av. Llano de la Victoria, 16. 22700 Jaca (Huesca) Spain, 302) CNR - Institute 1158 for Agricultural and Forestry Systems in the Mediterranean, P.le Enrico Fermi 1- Loc. del Granatello, 80055, Portici, (Napoli), Italy, 303) 1159 Institute of Earth Surface Dynamics, Faculty of Geosciences and Environment, University of Lausanne, Géopolis, 1015 Lausanne, 1160 Switzerland, 304) Climate Impacts Research Centre, Department of Ecology and Environmental Sciences, Umeå University, Abisko, 1161 Sweden, 305) Forest Research, Northern Research Station, Roslin, EH25 9SY, UK, 306) Institute of Mountain Hazards and Environment, 1162 Chinese Academy of Sciences, Chengdu, P.R. China, 307) MOE Key Laboratory of Geographical Processes and Ecological Security in 1163 Changbai Mountains, School of Geographical Sciences, Northeast Normal University, Changchun, Jilin 130024, China, 308) Department of 1164 Earth and Environmental Sciences, Lehigh University, Bethlehem, PA 18015, United States, 309) High Meadows Environmental Institute, 1165 Princeton University, NJ 08544, USA, 310) Zhejiang Tiantong Forest Ecosystem National Observation and Research Station, School of 1166 Ecological and Environmental Sciences, East China Normal University, Shanghai 200241, China, 311) University of Bayreuth, Ecological-1167 Botanical Gardens, Universitaetsstr. 30, Bayreuth, Germany, 312) Key Laboratory of Geographical Processes and Ecological Security in 1168 Changbai Mountains, Ministry of Education, School of Geographical Sciences, Northeast Normal University, Changchun 130024, China.

# 1169 OrcIDs

1170 Jonas J. Lembrechts: https://orcid.org/0000-0002-1933-0750. Johan van den Hoogen: https://orcid.org/0000-0001-6624-8461. Juha Aalto: 1171 https://orcid.org/0000-0001-6819-4911. Pieter De Frenne: https://orcid.org/0000-0002-8613-0943. Julia Kemppinen: 1172 https://orcid.org/0000-0001-7521-7229. Martin Kopecký: https://orcid.org/0000-0002-1018-9316. Miska Luoto: https://orcid.org/0000-0002-108-9316. Miska Luoto: https://orcid.org/0000-0002-108-9316. Miska Luoto: https://orcid.org/000-0002-108-9316. Miska Luoto: https://orcid.org/000 1173 0001-6203-5143. Ilya M. D. Maclean: https://orcid.org/0000-0001-8030-9136. Thomas W. Crowther: https://orcid.org/0000-0001-5674-1174 8913. Joseph J. Bailey: https://orcid.org/0000-0002-9526-7095. Stef Haesen: https://orcid.org/0000-0002-4491-4213. David H. Klinges: 1175 https://orcid.org/0000-0002-7900-9379. Pekka Niittynen: https://orcid.org/0000-0002-7290-029X. Brett R. Scheffers: 1176 https://orcid.org/0000-0003-2423-3821. Koenraad Van Meerbeek: https://orcid.org/0000-0002-9260-3815. Peter Aartsma: 1177 https://orcid.org/0000-0001-5086-856X. Otar Abdalaze: https://orcid.org/0000-0001-8140-0900. Mehdi Abedi: https://orcid.org/0000-1178 0002-1499-0119. Rien Aerts: https://orcid.org/0000-0001-6694-0669. Negar Ahmadian: https://orcid.org/0000-0002-7427-7198. Antje 1179 Ahrends: https://orcid.org/0000-0002-5083-7760. Juha M. Alatalo: https://orcid.org/0000-0001-5084-850X. Jake M. Alexander: https://orcid.org/0000-0003-2226-7913. Camille Nina Allonsius: https://orcid.org/0000-0003-2599-9941. Jan Altman: 1180 1181 https://orcid.org/0000-0003-4879-5773. Christof Ammann: https://orcid.org/0000-0002-0783-5444. Christian Andres: 1182 https://orcid.org/0000-0003-0576-6446. Christopher Andrews: https://orcid.org/0000-0003-2428-272X. Jonas Ardö: 1183 https://orcid.org/0000-0002-9318-0973. Nicola Arriga: https://orcid.org/0000-0001-5321-3497. Alberto Arzac: https://orcid.org/0000-1184 0002-3361-5349. Valeria Aschero: https://orcid.org/0000-0003-3865-4133. Rafael L. Assis: https://orcid.org/0000-0001-8468-6414. Jakob 1185 Johann Assmann: https://orcid.org/0000-0002-3492-8419. Maaike Y. Bader: https://orcid.org/0000-0003-4300-7598. Khadijeh Bahalkeh: 1186 https://orcid.org/0000-0003-1485-0316. Peter Barančok: https://orcid.org/0000-0003-1171-2524. Isabel C. Barrio: https://orcid.org/0000-1187 0002-8120-5248. Agustina Barros: https://orcid.org/0000-0002-6810-2391. Edmund W. Basham: https://orcid.org/0000-0002-0167-7908. 1188 Marijn Bauters: https://orcid.org/0000-0003-0978-6639. Manuele Bazzichetto: https://orcid.org/0000-0002-9874-5064. Luca Belelli 1189 Marchesini: https://orcid.org/0000-0001-8408-4675. Michael C. Bell: https://orcid.org/0000-0002-3401-7746. Juan C. Benavides: 1190 https://orcid.org/0000-0002-9694-2195. José Luis Benito Alonso: https://orcid.org/0000-0003-1086-8834. Bernd J. Berauer: 1191 https://orcid.org/0000-0002-9472-1532. Jarle W. Bjerke: https://orcid.org/0000-0003-2721-1492. Robert G. Björk: https://orcid.org/0000-1192 0001-7346-666X. Mats P. Björkman: https://orcid.org/0000-0001-5768-1976. Katrin Björnsdóttir: https://orcid.org/0000-0001-7421-9441. 1193 Benjamin Blonder: https://orcid.org/0000-0002-5061-2385. Pascal Boeckx: orcid.org/https://orcid.org/0000-0003-3998-0010. Julia Boike: 1194 https://orcid.org/0000-0002-5875-2112. Stef Bokhorst: https://orcid.org/0000-0003-0184-1162. Bárbara N. S. Brum: 1195 https://orcid.org/0000-0002-8421-3200. Josef Brůna: https://orcid.org/0000-0002-4839-4593. Nina Buchmann: https://orcid.org/0000-1196 0003-0826-2980. José Luís Camargo: https://orcid.org/0000-0003-0370-9878. Otávio C. Campoe: https://orcid.org/0000-0001-9810-8834. 1197 Onur Candan: https://orcid.org/0000-0002-9254-4122. Rafaella Canessa: https://orcid.org/0000-0002-6979-9880. Nicoletta Cannone: 1198 https://orcid.org/0000-0002-3390-3965. Michele Carbognani: https://orcid.org/0000-0001-7701-9859. Jofre Carnicer: 1199 https://orcid.org/0000-0001-7454-8296. Angélica Casanova-Katny: https://orcid.org/0000-0003-3860-1445. Simone Cesarz: 1200 https://orcid.org/0000-0003-2334-5119. Bogdan Chojnicki: https://orcid.org/0000-0002-9012-4060. Philippe Choler: 1201 https://orcid.org/0000-0002-9062-2721. Steven L. Chown: https://orcid.org/0000-0001-6069-5105. Edgar F. Cifuentes: 1202 https://orcid.org/0000-0001-5918-5861. Marek Čiliak: https://orcid.org/0000-0002-6720-9365. Tamara Contador: https://orcid.org/0000-1203 0002-0250-9877. Peter Convey: https://orcid.org/0000-0001-8497-9903. Elisabeth J. Cooper: https://orcid.org/0000-0002-0634-1282. 1204 Edoardo Cremonese: https://orcid.org/0000-0002-6708-8532. Salvatore R. Curasi: https://orcid.org/0000-0002-4534-3344. Maurizio 1205 Cutini: https://orcid.org/0000-0002-8597-8221. C. Johan Dahlberg: https://orcid.org/0000-0003-0271-3306. Gergana N. Daskalova: 1206 https://orcid.org/0000-0002-5674-5322. Miguel Angel de Pablo: https://orcid.org/0000-0002-4496-2741. Stefano Della Chiesa: 1207 https://orcid.org/0000-0002-6693-2199. Jürgen Dengler: https://orcid.org/0000-0003-3221-660X. Patrice Descombes: 1208 https://orcid.org/0000-0002-3760-9907. Valter Di Cecco: https://orcid.org/0000-0001-9862-1267. Michele Di Musciano: 1209 https://orcid.org/0000-0002-3130-7270. Jan Dick: https://orcid.org/0000-0002-4180-9338. Jiri Dolezal: https://orcid.org/0000-0002-5829-1210 4051. Ellen Dorrepaal: https://orcid.org/0000-0002-0523-2471. Jiří Dušek: https://orcid.org/0000-0001-6119-0838. Nico Eisenhauer: 1211 https://orcid.org/0000-0002-0371-6720. Lars Eklundh: https://orcid.org/0000-0001-7644-6517. Todd E. Erickson: https://orcid.org/0000-1212 0003-4537-0251. Brigitta Erschbamer: https://orcid.org/0000-0002-6792-1395. Werner Eugster: http://orcid.org/0000-0001-6067-0741. 1213 Dan A. Exton: https://orcid.org/0000-0001-8885-5828. Nicolas Fanin: https://orcid.org/0000-0003-4195-855X. Fatih Fazlioglu: 1214 https://orcid.org/0000-0002-4723-3640. Iris Feigenwinter: https://orcid.org/0000-0001-7493-6790. Giuseppe Fenu: 1215 https://orcid.org/0000-0003-4762-5043. Olga Ferlian: https://orcid.org/0000-0002-2536-7592. Eduardo Fernández-Pascual: 1216 https://orcid.org/0000-0002-4743-9577. Manfred Finckh: https://orcid.org/0000-0003-2186-0854. Rebecca Finger Higgens: 1217 https://orcid.org/0000-0002-7645-504X. T'ai G. W. Forte: https://orcid.org/0000-0002-8685-5872. Erika C. Freeman: 1218 https://orcid.org/0000-0001-7161-6038. Esther R. Frei: https://orcid.org/0000-0003-1910-7900. Eduardo Fuentes-Lillo: 1219 https://orcid.org/0000-0001-5657-954X. Rafael A. García: https://orcid.org/0000-0002-0591-0391. María B. García: 1220 https://orcid.org/0000-0003-4231-6006. Charly Géron: https://orcid.org/0000-0001-7912-4708. Mana Gharun: https://orcid.org/0000-1221 0003-0337-7367. Dany Ghosn: https://orcid.org/0000-0003-1898-9681. Khatuna Gigauri: https://orcid.org/0000-0002-6707-0818. Anne 1222 Gobin: https://orcid.org/0000-0002-3742-7062. Ignacio Goded: https://orcid.org/0000-0002-1912-325X. Mathias Goeckede: 1223 https://orcid.org/0000-0003-2833-8401. Felix Gottschall: https://orcid.org/0000-0002-1247-8728. Keith Goulding: https://orcid.org/0000-1224 0002-6465-1465. Sanne Govaert: https://orcid.org/0000-0002-8939-1305. Bente Jessen Graae: https://orcid.org/0000-0002-5568-4759. 1225 Sarah Greenwood: https://orcid.org/0000-0001-9104-7936. Caroline Greiser: https://orcid.org/0000-0003-4023-4402. Achim Grelle: 1226 https://orcid.org/0000-0003-3468-9419. Benoit Guénard: https://orcid.org/0000-0002-7144-1175. Joannès Guillemot: 1227 https://orcid.org/0000-0003-4385-7656. Peter Haase: https://orcid.org/0000-0002-9340-0438. Sylvia Haider: https://orcid.org/0000-0002-1228 2966-0534. Aud H. Halbritter: https://orcid.org/0000-0003-2597-6328. Maroof Hamid: https://orcid.org/0000-0003-3406-5008. Albin 1229 Hammerle: https://orcid.org/0000-0003-1963-5906. Arndt Hampe: https://orcid.org/0000-0003-2551-9784. Siri V. Haugum: 1230 https://orcid.org/0000-0003-4958-7132. Lucia Hederová: https://orcid.org/0000-0003-1283-0952. Bernard Heinesch: 1231 https://orcid.org/0000-0001-7594-6341. Carole Helfter: https://orcid.org/0000-0001-5773-4652. Daniel Hepenstrick: 1232 https://orcid.org/0000-0003-1090-6888. Maximiliane Herberich: https://orcid.org/0000-0003-0716-1520. Luise Hermanutz:

1233 https://orcid.org/0000-0003-0706-7067. David S. Hik: https://orcid.org/0000-0002-8994-9305. Raúl Hoffrén: https://orcid.org/0000-0002-1234 9123-304X. Jürgen Homeier: https://orcid.org/0000-0001-5676-3267. Lukas Hörtnagl: https://orcid.org/0000-0002-5569-0761. Toke T. 1235 Høye: https://orcid.org/0000-0001-5387-3284. Filip Hrbacek: https://orcid.org/0000-0001-5032-9216. Kristoffer Hylander: 1236 https://orcid.org/0000-0002-1215-2648. Hiroki Iwata: https://orcid.org/0000-0002-8962-8982. Marcin Antoni Jackowicz-Korczynski: 1237 1238 0001-6317-660X. Szymon Jastrzębowski: https://orcid.org/0000-0003-1239-4847. Anke Jentsch: https://orcid.org/0000-0002-2345-8300. 1239 Juan J. Jiménez: https://orcid.org/0000-0003-2398-0796. Ingibjörg S. Jónsdóttir: https://orcid.org/0000-0003-3804-7077. Tommaso Jucker: 1240 https://orcid.org/0000-0002-0751-6312. Alistair S. Jump: https://orcid.org/0000-0002-2167-6451. Radoslaw Juszczak: 1241 https://orcid.org/0000-0002-5212-7383. Róbert Kanka: https://orcid.org/0000-0002-7071-7280. Vít Kašpar: https://orcid.org/0000-0002-1242 0879-0137. Julia Kelly: https://orcid.org/0000-0002-7370-1401. Anzar A. Khuroo: https://orcid.org/0000-0002-0251-2793. Leif 1243 Klemedtsson: https://orcid.org/0000-0002-1122-0717. Marcin Klisz: https://orcid.org/0000-0001-9486-6988. Natascha Kljun: 1244 https://orcid.org/0000-0001-9650-2184. Alexander Knohl: https://orcid.org/0000-0002-7615-8870. Johannes Kobler: 1245 https://orcid.org/0000-0003-0052-4245. Jozef Kollár: https://orcid.org/0000-0002-0069-4220. Martyna M. Kotowska: 1246 https://orcid.org/0000-0002-2283-5979. Bence Kovács: https://orcid.org/0000-0002-8045-8489. Juergen Kreyling: https://orcid.org/000-0002-8045-8489. Juergen Kreyling: https://orcid.org/000-000-0004-8489. Juergen Kreyling: https://orcid.org/000-000-8489. Juergen Kreyling: https://orcid.org/000-8489. Juergen Kreyling: https://orcid.org/000-8489. Juergen Kreyling: https://orcid.org/000-1247 0001-8489-7289. Andrea Lamprecht: https://orcid.org/0000-0002-8719-026X. Simone I. Lang: https://orcid.org/0000-0002-6812-2528. 1248 Christian Larson: https://orcid.org/0000-0002-7567-4953. Keith Larson: https://orcid.org/0000-0001-7089-524X. Kamil Laska: 1249 1250 0001-9672-625X. Luc Lens: https://orcid.org/0000-0002-0241-2215. Bengt Liljebladh: https://orcid.org/0000-0002-2998-5865. Annalea 1251 Lohila: https://orcid.org/0000-0003-3541-672X. Juan Lorite: https://orcid.org/0000-0003-4617-8069. Benjamin Loubet: 1252 https://orcid.org/0000-0001-8825-8775. Joshua Lynn: https://orcid.org/0000-0002-7190-7991. Martin Macek: https://orcid.org/0000-0000-7190-7991. Martin Macek: https://orcid.org/0000-7190-7991. Martin Macek: https://orcid.org/0000-7991. Martin Macek: https://orcid.org/000-7991. Martin Macek: Macek 1253 0002-5609-5921. Roy Mackenzie: https://orcid.org/0000-0001-6620-1532. Enzo Magliulo: https://orcid.org/0000-0001-5505-6552. Regine 1254 Maier: https://orcid.org/0000-0003-3158-4136. Francesco Malfasi: https://orcid.org/0000-0002-2660-8327. František Máliš: 1255 https://orcid.org/0000-0003-2760-6988. Matěj Man: https://orcid.org/0000-0002-4557-8768. Giovanni Manca: https://orcid.org/0000-1256 0002-9376-0310. Antonio Manco: https://orcid.org/0000-0002-3677-4134. Paraskevi Manolaki: https://orcid.org/0000-0003-3958-0199. 1257 Radim Matula: https://orcid.org/0000-0002-7460-0100. Sergiy Medinets: https://orcid.org/0000-0001-5980-1054. Volodymyr Medinets: 1258 https://orcid.org/0000-0001-7543-7504. Camille Meeussen: https://orcid.org/0000-0002-5869-4936. Sonia Merinero: 1259 https://orcid.org/0000-0002-1405-6254. Rita de Cássia Guimarães Mesquita: https://orcid.org/0000-0003-1746-3215. Katrin Meusburger: 1260 https://orcid.org/0000-0003-4623-6249. Filip J. R. Meysman: https://orcid.org/0000-0001-5334-7655. Sean T. Michaletz: 1261 https://orcid.org/0000-0003-2158-6525. Ann Milbau: https://orcid.org/0000-0003-3555-8883. Pavel Moiseev: https://orcid.org/0000-1262 0003-4808-295X. Andrea Mondoni: https://orcid.org/0000-0002-4605-6304. Leonardo Montagnani: orcid.org/https://orcid.org/0000-1263 0003-2957-9071. Mikel Moriana-Armendariz: https://orcid.org/0000-0001-8251-1338. Umberto Morra di Cella: https://orcid.org/0000-1264 0003-4250-9705. Martin Mörsdorf: https://orcid.org/0000-0002-3903-2021. Jonathan R. Mosedale: https://orcid.org/0000-0001-9008-1265 5439. Lena Muffler: https://orcid.org/0000-0001-8227-7297. Miriam Muñoz-Rojas: https://orcid.org/0000-0002-9746-5191. Jonathan A. 1266 Myers: https://orcid.org/0000-0002-2058-8468. Isla H. Myers-Smith: https://orcid.org/0000-0002-8417-6112. Marianna Nardino: 1267 https://orcid.org/0000-0001-9466-8340. Ilona Naujokaitis-Lewis: https://orcid.org/0000-0001-9504-4484. Lena Nicklas: 1268 https://orcid.org/0000-0002-9337-4153. Georg Niedrist: https://orcid.org/0000-0002-7511-6273. Mats B. Nilsson: https://orcid.org/0000-1269 0003-3765-6399. Signe Normand: https://orcid.org/0000-0002-8782-4154. Marcelo D. Nosetto: https://orcid.org/0000-0002-9428-490X. 1270 Yann Nouvellon: https://orcid.org/0000-0003-1920-3847. Martin A. Nuñez: https://orcid.org/0000-0003-0324-5479. Romà Ogaya: 1271 https://orcid.org/0000-0003-4927-8479. Jérôme Ogée: https://orcid.org/0000-0002-3365-8584. Joseph Okello: https://orcid.org/0000-1272 0003-4462-3923. Janusz Olejnik: https://orcid.org/0000-0001-5305-1045. Jørgen Eivind Olesen: https://orcid.org/0000-0002-6639-1273. 1273 Øystein Opedal: https://orcid.org/0000-0002-7841-6933. Simone Orsenigo: https://orcid.org/0000-0003-0348-9115. Andrej Palaj: 1274 https://orcid.org/0000-0001-7054-4183. Timo Pampuch: https://orcid.org/0000-0002-6290-9661. Meelis Pärtel: https://orcid.org/0000-1275 0002-5874-0138. Ada Pastor: https://orcid.org/0000-0002-7114-770X. Aníbal Pauchard: https://orcid.org/0000-0003-1284-3163. Harald 1276 Pauli: https://orcid.org/0000-0002-9842-9934. Marian Pavelka: https://orcid.org/0000-0002-7339-3410. William D. Pearse: 1277 https://orcid.org/0000-0002-6241-3164. Matthias Peichl: https://orcid.org/0000-0002-9940-5846. Rachel M. Penczykowski: 1278 https://orcid.org/0000-0003-4559-0609. Josep Penuelas: https://orcid.org/0000-0002-7215-0150. Matteo Petit Bon: 1279 https://orcid.org/0000-0001-9829-8324. Alessandro Petraglia: https://orcid.org/0000-0003-4632-2251. Shyam S. Phartyal: 1280 https://orcid.org/0000-0003-3266-6619. Gareth K. Phoenix: https://orcid.org/0000-0002-0911-8107. Casimiro Pio: https://orcid.org/0000-1281 0002-3531-8620. Andrea Pitacco: https://orcid.org/0000-0002-7260-6242. Camille Pitteloud: https://orcid.org/0000-0002-4731-0079. 1282 Roman Plichta: https://orcid.org/0000-0003-2442-8522. Francesco Porro: https://orcid.org/0000-0001-9855-2468. Miguel Portillo-Estrada: 1283 https://orcid.org/0000-0002-0348-7446. Jérôme Poulenard: https://orcid.org/0000-0003-0810-0308. Rafael Poyatos: 1284 https://orcid.org/0000-0003-0521-2523. Anatoly S. Prokushkin: https://orcid.org/0000-0001-8721-2142. Radoslaw Puchalka: 1285 https://orcid.org/0000-0002-4764-0705. Mihai Puşcaş: https://orcid.org/0000-0002-2632-640X. Dajana Radujković: 1286 https://orcid.org/0000-0003-4981-5879. Krystal Randall: https://orcid.org/0000-0003-2507-1000. Amanda Ratier Backes: 1287 https://orcid.org/0000-0002-7229-578X. David Renault: https://orcid.org/0000-0003-3644-1759. Anita C. Risch: https://orcid.org/0000-1288 1289 7130-9617. Bjorn J.M. Robroek: https://orcid.org/0000-0002-6714-0652. Adrian V. Rocha: https://orcid.org/0000-0002-4618-2407. 1290 Christian Rossi: https://orcid.org/0000-0002-4618-2407. Graziano Rossi: https://orcid.org/0000-0002-5102-5019. Olivier Roupsard: 1291 https://orcid.org/0000-0002-1319-142X. Alexey V. Rubtsov: https://orcid.org/0000-0002-9663-4344. Patrick Saccone: 1292 https://orcid.org/0000-0001-8820-593X. Jhonatan Sallo Bravo: https://orcid.org/0000-0001-9007-4959. Cinthya C. Santos: 1293 https://orcid.org/0000-0001-7042-5993. Judith M. Sarneel: https://orcid.org/0000-0001-6187-499X. Tobias Scharnweber: 1294 https://orcid.org/0000-0002-4933-5296. Marius Schmidt: https://orcid.org/0000-0001-5292-7092. Thomas Scholten: 1295 https://orcid.org/0000-0002-4875-2602. Max Schuchardt: https://orcid.org/0000-0003-3103-8063. Tony Scott: https://orcid.org/0000-1296 0002-6631-0672. Julia Seeber: https://orcid.org/0000-0003-0189-7377. Tim Seipel: https://orcid.org/0000-0001-6472-2975. Philipp 1297 Semenchuk: https://orcid.org/0000-0002-1949-6427. Rebecca A. Senior: https://orcid.org/0000-0002-8208-736X. Josep M. Serra-Diaz: 1298 https://orcid.org/0000-0003-1988-1154. Piotr Sewerniak: https://orcid.org/0000-0002-3071-3963. Ankit Shekhar: https://orcid.org/0000-0002-3071-3963. Ankit Shekhar: https://orcid.org/0000-0002-3071-3963.

1299 0003-0802-2821. Laura Siegwart Collier: https://orcid.org/0000-0003-0985-9615. Elizabeth Simpson: https://orcid.org/0000-0002-6107-1300 0286. David P. Sigueira: https://orcid.org/0000-0002-0756-0153. Zuzana Sitková: https://orcid.org/0000-0001-6354-6105. Johan Six: 1301 https://orcid.org/0000-0001-9336-4185. Marko Smiljanic: https://orcid.org/0000-0002-2324-0723. Stuart W. Smith: 1302 https://orcid.org/0000-0001-9396-6610. Ben Somers: https://orcid.org/0000-0002-7875-107X. José João L. L. Souza: 1303 https://orcid.org/0000-0003-4670-6626. Bartolomeu Israel Souza: https://orcid.org/0000-0003-2173-8314. Arildo Souza Dias: 1304 https://orcid.org/0000-0002-5495-3435. Marko J. Spasojevic: https://orcid.org/0000-0003-1808-0048. James D. M. Speed: 1305 https://orcid.org/0000-0002-0633-5595. Fabien Spicher: https://orcid.org/0000-0002-9999-955X. Angela Stanisci: https://orcid.org/0000-1306 0002-5302-0932. Klaus Steinbauer: https://orcid.org/0000-0002-3730-9920. Rainer Steinbrecher: https://orcid.org/0000-0002-5931-4210. 1307 Michael Steinwandter: https://orcid.org/0000-0001-8545-6047. Michael Stemkovski: https://orcid.org/0000-0002-9854-887X. Jörg G. 1308 Stephan: https://orcid.org/0000-0001-6195-7867. Christian Stiegler: https://orcid.org/0000-0002-0130-2401. Stefan Stoll: 1309 https://orcid.org/0000-0002-3656-417X. Martin Svátek: https://orcid.org/0000-0003-2328-4627. Miroslav Svoboda: 1310 https://orcid.org/0000-0003-4050-3422. Torbern Tagesson: https://orcid.org/0000-0003-3011-1775. Andrew J. Tanentzap: 1311 https://orcid.org/0000-0002-2883-1901. Franziska Tanneberger: https://orcid.org/0000-0002-4184-9671. Jean-Paul Theurillat: 1312 https://orcid.org/0000-0002-1843-5809. Haydn J. D. Thomas: https://orcid.org/0000-0001-9099-6304. Andrew D. Thomas: 1313 https://orcid.org/0000-0002-1360-1687. Marcello Tomaselli: https://orcid.org/0000-0003-4208-3433. Urs Albert Treier: 1314 https://orcid.org/0000-0003-4027-739X. Mario Trouillier: https://orcid.org/0000-0001-9151-7686. Pavel Dan Turtureanu: 1315 https://orcid.org/0000-0002-7422-3106. Vilna A. Tyystjärvi: https://orcid.org/0000-0002-1175-5463. Masahito Ueyama: 1316 https://orcid.org/0000-0002-4000-4888. Karol Ujházy: https://orcid.org/0000-0002-0228-1737. Mariana Ujházyová: 1317 https://orcid.org/0000-0002-5546-1547. Domas Uogintas: https://orcid.org/0000-0002-3937-1218. Josef Urban: https://orcid.org/0000-1318 0003-1730-947X. Marek Urbaniak: https://orcid.org/0000-0002-1225-9170. Tudor-Mihai Ursu: https://orcid.org/0000-0002-4898-6345. 1319 Francesco Primo Vaccari: https://orcid.org/0000-0002-5253-2135. Stijn Van de Vondel: https://orcid.org/0000-0002-0223-7330. Liesbeth 1320 van den Brink: https://orcid.org/0000-0003-0313-8147. Maarten Van Geel: https://orcid.org/0000-0001-8688-6225. Vigdis Vandvik: 1321 https://orcid.org/0000-0003-4651-4798. Pieter Vangansbeke: https://orcid.org/0000-0002-6356-2858. Andrej Varlagin: 1322 https://orcid.org/0000-0002-2549-5236. G.F. (Ciska) Veen: https://orcid.org/0000-0001-7736-9998. Elmar Veenendaal: 1323 https://orcid.org/0000-0001-8230-2501. Susanna E. Venn: https://orcid.org/0000-0002-7433-0120. Hans Verbeeck: 1324 https://orcid.org/0000-0003-1490-0168. Erik Verbrugggen: https://orcid.org/0000-0001-7015-1515. Frank G.A. Verheijen: 1325 https://orcid.org/0000-0001-6741-4249. Luca Vitale: https://orcid.org/0000-0002-7637-264X. Pascal Vittoz: https://orcid.org/0000-0003-1326 4218-4517. Maria Vives-Ingla: https://orcid.org/0000-0003-4887-8392. Jonathan von Oppen: https://orcid.org/0000-0001-6346-2964. 1327 Josefine Walz: https://orcid.org/0000-0002-0715-8738. Runxi Wang: https://orcid.org/0000-0003-4902-169X. Yifeng Wang: 1328 https://orcid.org/0000-0003-2660-7874. Robert G. Way: https://orcid.org/0000-0003-4763-7685. Ronja E. M. Wedegärtner: 1329 https://orcid.org/0000-0003-4633-755X. Robert Weigel: https://orcid.org/0000-0001-9685-6783. Jan Wild: https://orcid.org/0000-0003-1330 3007-4070. Matthew Wilkinson: https://orcid.org/0000-0002-3858-553X. Martin Wilmking: https://orcid.org/0000-0003-4964-2402. Lisa 1331 Wingate: https://orcid.org/0000-0003-1921-1556. Manuela Winkler: https://orcid.org/0000-0002-8655-9555. Sonja Wipf: 1332 https://orcid.org/0000-0002-3492-1399. Georg Wohlfahrt: https://orcid.org/0000-0003-3080-6702. Georgios Xenakis: 1333 https://orcid.org/0000-0002-2950-4101. Yan Yang: https://orcid.org/0000-0003-0858-7603. Zicheng Yu: https://orcid.org/0003-0858-7603. Zicheng Yu: https://orcid.org/003-0858-7603. Zicheng Yu: https://orcid.org/003-0858-7603. Zicheng Yu: https://orcid.org/003-7603. Zicheng Yu: htt 1334 2358-2712. Kailiang Yu: https://orcid.org/0000-0003-4223-5169. Florian Zellweger: https://orcid.org/0000-0003-1265-9147. Jian Zhang: 1335 https://orcid.org/0000-0003-0589-6267. Peng Zhao: https://orcid.org/0000-0003-3289-5067. Klaudia Ziemblińska: https://orcid.org/0000-1336 0003-4070-6553. Reiner Zimmermann: https://orcid.org/0000-0002-8724-941x. Shengwei Zong: https://orcid.org/0000-0002-3583-6110. 1337 Viacheslav I. Zvryanov: https://orcid.org/0000-0002-1748-4801. Ivan Nijs: https://orcid.org/0000-0003-3111-680X. Jonathan Lenoir: 1338 https://orcid.org/0000-0003-0638-9582. Juha Aalto: https://orcid.org/0000-0001-6819-4911. Martin Kopecký: https://orcid.org/0000-0001-6819-4911. 1339 0002-1018-9316. David H. Klinges: https://orcid.org/0000-0002-7900-9379. Valeria Aschero: https://orcid.org/0000-0003-3865-4133. 1340 Jakob Johann Assmann: https://orcid.org/0000-0002-3492-8419. Bernd J. Berauer: https://orcid.org/0000-0002-9472-1532. Robert G. 1341 Björk: https://orcid.org/0000-0001-7346-666X. Mats P. Björkman: https://orcid.org/0000-0001-5768-1976. Julia Boike: 1342 https://orcid.org/0000-0002-5875-2112. Rafaella Canessa: https://orcid.org/0000-0002-6979-9880. Jofre Carnicer: https://orcid.org/0000-1343 0001-7454-8296. Simone Cesarz: https://orcid.org/0000-0003-2334-5119. 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Sylvia Haider: https://orcid.org/0000-0002-2966-0534. Siri V. Haugum: 1351 https://orcid.org/0000-0003-4958-7132. Jürgen Homeier: https://orcid.org/0000-0001-5676-3267. Marcin Antoni Jackowicz-Korczynski: 1352 https://orcid.org/0000-0002-6574-5703. Anke Jentsch: https://orcid.org/0000-0002-2345-8300. Vít Kašpar: https://orcid.org/0000-0002-1353 0879-0137. Kamil Laska: https://orcid.org/0000-0002-5199-9737. Guerric le Maire: https://orcid.org/0000-0002-5227-958X. Annalea 1354 Lohila: https://orcid.org/0000-0003-3541-672X. Juan Lorite: https://orcid.org/0000-0003-4617-8069. Paraskevi Manolaki: 1355 https://orcid.org/0000-0003-3958-0199. Radim Matula: https://orcid.org/0000-0002-7460-0100. Sergiy Medinets: https://orcid.org/0000-1356 0001-5980-1054. Miriam Muñoz-Rojas: https://orcid.org/0000-0002-9746-5191. Signe Normand: https://orcid.org/0000-0002-8782-4154. 1357 Marcelo D. Nosetto: https://orcid.org/0000-0002-9428-490X. Yann Nouvellon: https://orcid.org/0000-0003-1920-3847. Martin A. Nuñez: 1358 https://orcid.org/0000-0003-0324-5479. Romà Ogaya: https://orcid.org/0000-0003-4927-8479. Joseph Okello: https://orcid.org/0000-1359 0003-4462-3923. Ada Pastor: https://orcid.org/0000-0002-7114-770X. Aníbal Pauchard: https://orcid.org/0000-0003-1284-3163. William 1360 D. Pearse: https://orcid.org/0000-0002-6241-3164. Josep Penuelas: https://orcid.org/0000-0002-7215-0150. Matteo Petit Bon: 1361 https://orcid.org/0000-0001-9829-8324. Camille Pitteloud: https://orcid.org/0000-0002-4731-0079. Rafael Poyatos: 1362 https://orcid.org/0000-0003-0521-2523. Anatoly S. Prokushkin: https://orcid.org/0000-0001-8721-2142. Radoslaw Puchalka: 1363 https://orcid.org/0000-0002-4764-0705. Mihai Puşcaş: https://orcid.org/0000-0002-2632-640X. Amanda Ratier Backes: 1364 https://orcid.org/0000-0002-7229-578X. David Renault: https://orcid.org/0000-0003-3644-1759. Christian Rossi: https://orcid.org/0000-

- 1365 0002-4618-2407. Olivier Roupsard: https://orcid.org/0000-0002-1319-142X. Jhonatan Sallo Bravo: https://orcid.org/0000-0001-9007-
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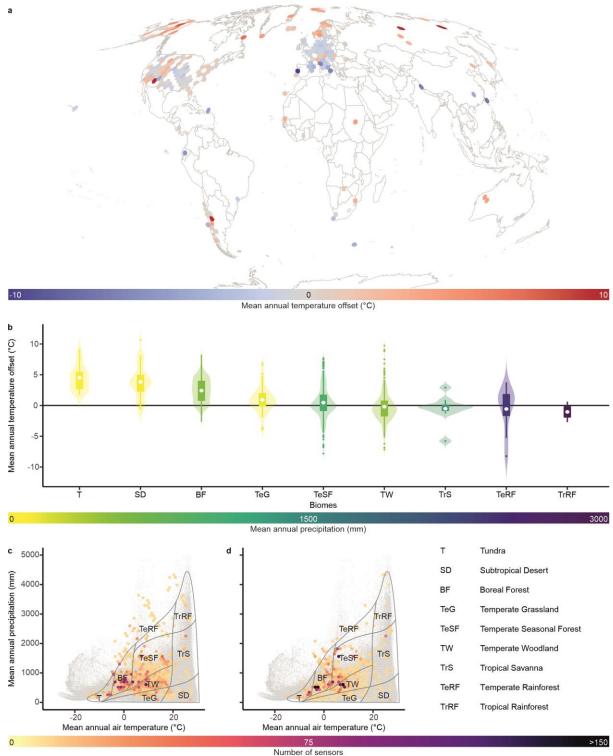
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## 1666 Tables

**Table 1:** Overview of soil bioclimatic variables as calculated in this study.

Bioclimatic variable	Meaning
SBIO1	annual mean temperature
SBIO2	mean diurnal range (mean of monthly (max temp - min temp))
SBIO3	isothermality (SBIO2/SBIO7) (×100)
SBIO4	temperature seasonality (standard deviation ×100)
SBIO5	max temperature of warmest month
SBIO6	min temperature of coldest month
SBIO7	temperature annual range (SBIO5-SBIO6)
SBIO8	mean temperature of wettest quarter
SBIO9	mean temperature of driest quarter
SBIO10	mean temperature of warmest quarter
SBIO11	mean temperature of coldest quarter

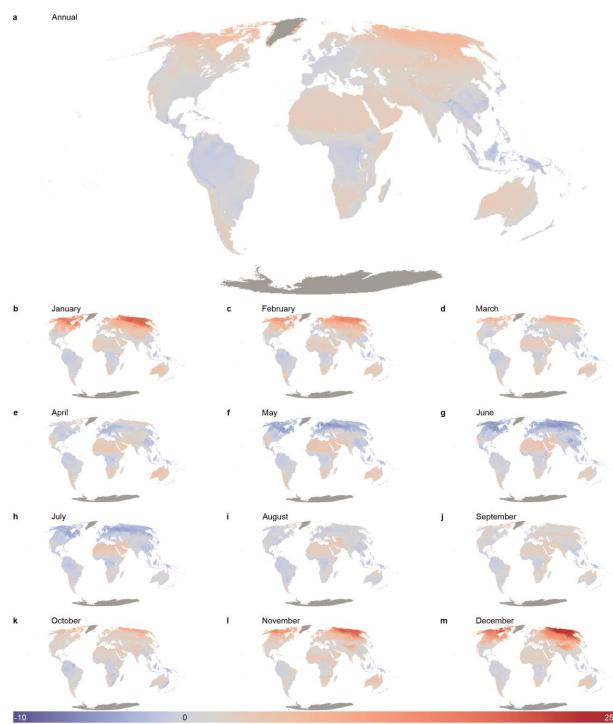
#### 1671 **Figure legends**



1672

1673 Figure 1: Temperature offsets between soil and air temperature differed significantly among 1674 biomes. (a) Distribution of in-situ measurement locations across the globe, coloured by the mean annual temperature offset (in °C) between in situ measured soil temperature (topsoil, 0–5 cm depth) 1675 1676 and gridded air temperature (ERA5L). Offsets were averaged per hexagon, each with a size of 1677 approximately 70,000 km<sup>2</sup>. Mollweide projection. (b) Mean annual temperature offsets per Whittaker 1678 biome (adapted from Whittaker 1970, based on geographic location of sensors averaged at 1 km<sup>2</sup>; 0-1679 5 cm depth), ordered by mean temperature offset and coloured by mean annual precipitation. (c–d)

- 1680 Distribution of sensors in 2D climate space for the topsoil (c, 0-5 cm depth, N = 4530) and the second
- layer (d, 5–15 cm depth, N = 3989). Colours of hexagons indicate the number of sensors at each climatic
   location, with a 40 × 40 km resolution. Grey dots in the background represent the global variation in
- 1682 climatic space (obtained by sampling 1,000,000 random locations from the CHELSA world maps).
- 1684 Overlay with grey lines depicts a delineation of Whittaker biomes.

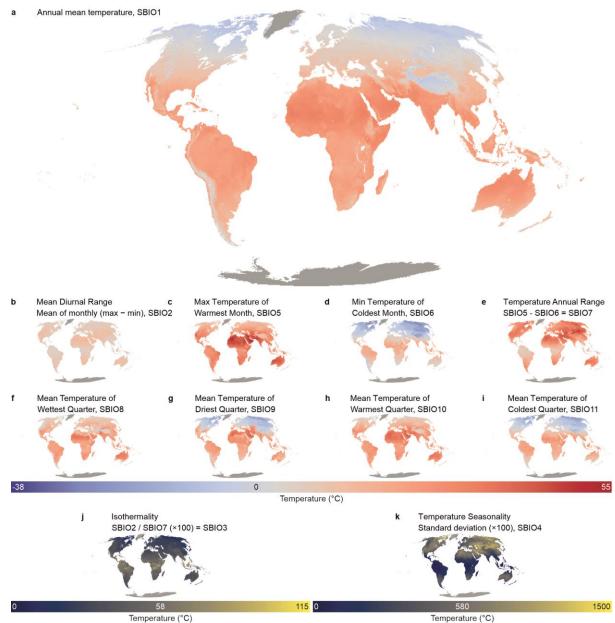


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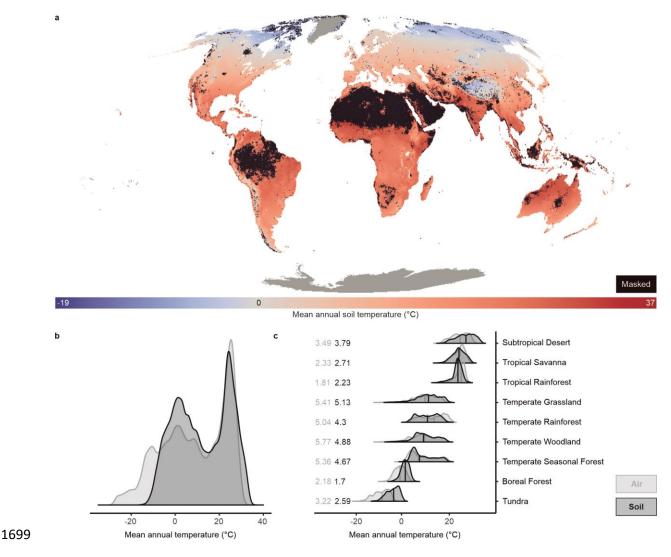
Mean temperature offset (°C)

1688Figure 2: Global modelled temperature offsets between soil and air temperature show strong1689spatiotemporal variation across months. Modelled annual (a) and monthly (b-m) temperature1690offset (in °C) between in situ measured soil temperature (topsoil, 0–5 cm) and gridded air

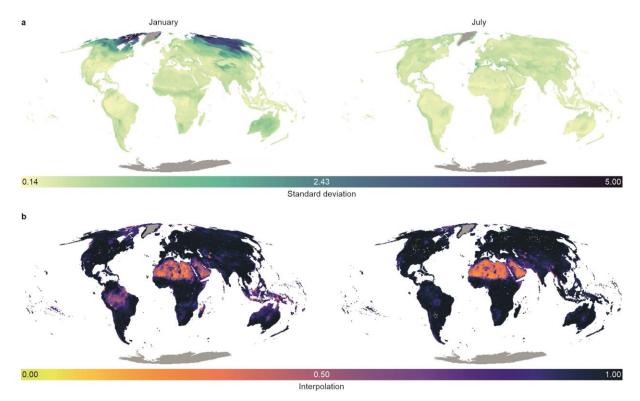
- 1691 temperature. Positive (red) values indicate soils that are warmer than the air. Dark grey represents
- 1692 regions outside the modelling area.



1694Temperature (°C)1695Figure 3: Soil bioclimatic variables. Global maps of bioclimatic variables for topsoil (0–5 cm depth)1696climate, calculated using the maps of monthly soil climate (see Fig. 2), and the bioclimatic variables for1697air temperature from CHELSA.

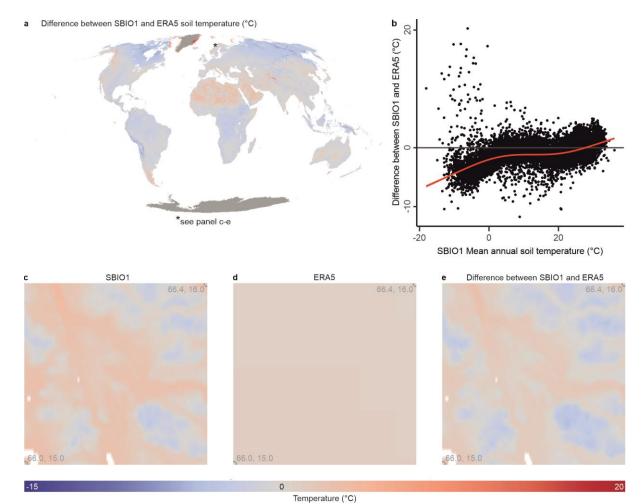


1700 Figure 4: Mean annual soil temperature shows significantly lower spatial variability than air temperature. (a) Global map of mean annual topsoil temperature (SBIO1, 0–5 cm depth, in °C), created 1701 1702 by adding the monthly offset between soil and air temperature for the period 2000–2020 (Fig. 2) to 1703 the monthly air temperature from CHELSA. A black mask is used to exclude regions where our models 1704 are extrapolating (i.e., interpolation values in Fig. 5 are < 0.9, 18% of pixels). Dark grey represents 1705 regions outside the modelling area. (b-c) Density plots of mean annual soil temperature across the 1706 globe (b) and for each Whittaker biome separately (c) for SBIO1 (dark grey, soil temperature), 1707 compared with BIO1 from CHELSA (light grey, air temperature), created by extracting 1 000 000 1708 random points from the 1-km<sup>2</sup> gridded bioclimatic products. The numbers in (c) represent the standard 1709 deviations of air temperature (light grey) and soil temperature (dark grey). Biomes are ordered 1710 according to the median annual soil temperature values (vertical black line) from the highest 1711 temperature (subtropical desert) to the lowest (tundra).



1713

1714 Figure 5: Models of the temperature offset between soil and air temperature have low standard 1715 deviations and good global coverage. Analyses for the temperature offset between in situ measured 1716 topsoil (0–5 cm depth) temperature and gridded air temperature. (a) Standard deviation (in °C) over the predictions from a cross-validation analysis that iteratively varied the set of covariates 1717 (explanatory data layers) and model hyperparameters across 100 models and evaluated model 1718 1719 strength using 10-fold cross-validation, for January (left) and July (right), as examples of the two most 1720 contrasting months. (b) The fraction of axes in the multidimensional environmental space for which 1721 the pixel lies inside the range of data covered by the sensors in the database. Low values indicate 1722 increased extrapolation.



1725 Figure 6: The mean annual soil temperature (SBIO1, 1 x 1 km resolution) modelled here is consistently cooler than ERA5L (9 x 9 km) soil temperature in forested areas. (a) Spatial 1726 1727 representation of the difference between SBIO1 based on our model and based on ERA5L soil 1728 temperature data. Negative values (blue colours) indicate areas where our model predicts cooler soil 1729 temperature. Dark grey areas (Greenland and Antarctica) are excluded from our models. Asterisk in Scandinavia indicates the highlighted area in panels d to f (see below). (b) Distribution of the difference 1730 between SBIO1 and ERA5L along the macroclimatic gradient (represented by SBIO1 itself) based on a 1731 1732 random subsample of 50 000 points from the map in a). Red line from a Generalized Additive Model 1733 (GAM) with k=4. (c-e) High-resolution zoomed panels of an area of high elevational contrast in Norway (from 66.0-66.4° N, 15.0-16.0° E) visualizing SBIO1 (c), ERA5L (d) and their difference (e), to highlight 1734 1735 the higher spatial resolution as obtained with SBIO1.