



35 studies in northwest Europe. This is an important problem because RSL fluctuations during the LIG  
36 are portrayed as warning signs for how polar ice sheets may behave in a future, warmer than  
37 present, world. Here we review the evidence for RSL change during the LIG using stratigraphic data  
38 from the best resolved highstand records that exist in the near-field of northwest Europe, from a  
39 range of settings that include lagoonal, shallow marine, tidal flat, salt marsh and brackish-water  
40 fluvial environments. Consideration of previously published stratigraphic records from two sites  
41 in the Eemian coastal-marine embayment that existed in the central Netherlands, yields no clear  
42 indications for abrupt RSL change during the attainment of the near-field highstand. Nor do we find  
43 any such indications in other records from countries bordering the North Sea, the Baltic Sea and the  
44 White Sea. Two modelling experiments that explore the global signal of hypothetical sea-level  
45 oscillations caused by partial collapse and regrowth of either the Greenland or Antarctic LIG ice-  
46 sheet, show that the North Sea region is relatively insensitive to mass changes sourced from  
47 Greenland but should clearly register events with an Antarctic origin, especially those that occur late  
48 in the LIG. The lack of evidence for abrupt sea-level fluctuations at this time in northwest Europe  
49 concurs with a lack of clear near-field evidence for ice sheet collapse.

50

51 Keywords: sea level; interglacial; ice sheet collapse; Eemian; MIS 5e; glacio-isostatic adjustment

52

## 53 1. Introduction

54 Several studies, mainly from low latitude sites suggest that polar ice-sheet collapse late in  
55 the Last Interglacial caused an already high sea level to jump abruptly by a further 2 to 6 m (e.g.  
56 Chappell, 1974; Bloom et al., 1974; Stein et al., 1993; Stirling et al., 1998; Thompson and Goldstein,  
57 2005; Hearty et al., 2007; Rohling et al., 2008; Blanchon et al., 2009; O’Leary et al., 2013; Dabrio et  
58 al., 2013). Although not all studies document such variability (e.g. Muhs et al., 2002; Muhs et al.,  
59 2011), reports of multi-meter scale sea-level jumps raise concern regarding the potential instability  
60 of polar ice sheets in what remains of the current interglacial. Such a collapse would have generated  
61 a distinctive geometry, or fingerprint, of global sea-level rise that palaeo sea-level studies can aim to  
62 detect (e.g. Mitrovica et al., 2001; Hay et al., 2014). However, no conclusive evidence for such a  
63 jump has been reported in relative sea-level (RSL) or ice-sheet records from higher latitude settings  
64 (e.g. Zagwijn, 1983; Funder et al., 2002; Lambeck et al., 2006) and, as a result, the global fingerprint  
65 of this event is not known. This has led to uncertainty as to whether the jump is indeed global in  
66 nature – i.e. triggered by polar ice-sheet collapse - or simply the product of regional (e.g. climate) or  
67 local (e.g. tectonic) processes.

68 The aim of this paper is to review the European near-field evidence for abrupt relative sea-  
69 level (RSL) change during the LIG highstand (as defined and dated in the far-field) by scrutinising  
70 previously published records from northwest Europe, where the LIG is known as the ‘Eemian’. We  
71 start by reviewing RSL evidence from sites located distant to the MIS 6 ice sheets, to identify key  
72 features that characterise many of these studies. Next, we consider how such abrupt oscillations  
73 might be preserved in the near-field stratigraphic record, drawing in part on sedimentary principles  
74 that have been established from working on equivalent Holocene sequences (e.g. Vis et al., 2015).  
75 We focus on evidence from the coastal deposits preserved in the Amsterdam and Amersfoort glacial  
76 basins (central Netherlands), which have a superbly preserved depositional record of the LIG  
77 transgression and RSL highstand. The region is the bio- and chronostratigraphical type region of the  
78 ‘Eemian’ (Harting, 1874; 1875, Zagwijn, 1961; Turner, 2002) and data from here, as well as from  
79 other sites in the Netherlands, are critical for constraining the LIG RSL highstand in northwest Europe  
80 (e.g. Zagwijn, 1983; 1996; Streif, 1990).

81 We find no compelling signs of abrupt RSL oscillations during the near-field highstand in the  
82 coastal deposits of the Netherlands, or from elsewhere in northwest Europe. We note, however,  
83 that due to solid earth deformation and equatorial syphoning the attainment of this highstand may  
84 be several thousand years later to that observed in far- and intermediate-field sites, meaning that if  
85 fluctuations occurred early in the interglacial, these could have happened before the near-field  
86 highstand was attained. The lack of evidence for abrupt sea-level fluctuations during the latter part

87 of the LIG in northwest Europe concurs with a lack of clear near-field evidence for ice sheet collapse  
88 at this time.

89

## 90 **2. Sea-level changes during the Last Interglacial**

91 Evidence for higher-than-present sea level during the LIG is recorded in emergent landforms  
92 and sediments that include coral reef tracts, bioerosional notches as well as nearshore deposits (e.g.  
93 Szabo et al., 1994; Stirling et al., 1998; Muhs et al., 2002; Bruggermann et al., 2004; Hearty et al.,  
94 2007; Rohling et al., 2008; Thompson et al., 2011; O’Leary et al., 2013; Dutton et al., 2015).  
95 Consideration of four RSL records selected from different parts of the world exemplify elements of  
96 LIG sea-level behaviour that are common to many far- and intermediate-field RSL records that  
97 record sea-level oscillations within the LIG highstand (Figure 1).

98 From the Seychelles, Dutton et al. (2015) report evidence from raised coral deposits of an  
99 early “rapid collapse” of a polar ice sheet, likely part of Antarctica, by  $128.6 \pm 0.8$  k yr ago that  
100 pushed sea level to at least  $+5.9 \pm 1.7$  m above present (Figure 1A). This was followed by a slower  
101 rise of a further c. 2 m, at a rate of c.  $0.22 \pm 0.04$  m/k yr that is attributed to partial melt of the  
102 Greenland Ice Sheet, thermal expansion and the loss of mountain glaciers. Peak eustatic sea level of  
103 c.  $7.6 \pm 1.7$  m was reached at c. 125 k yr ago, after which sea level fell and the Seychelles RSL record  
104 ends.

105 Our second example is based on the calibration to sea level of a stable oxygen isotope  
106 record of sea surface temperatures obtained from planktonic foraminifera extracted from core KL11,  
107 located in the central part of the Red Sea (Rohling et al., 2008) (Figure 1B). The record contains two  
108 main and one subsidiary sea-level fluctuation of 4 to 10 m. Sedimentological observations from  
109 adjacent Red Sea coastlines are cited by Rohling et al. (2008) as supporting evidence of a fluctuating  
110 RSL during the LIG highstand (e.g. Orszag-Sperber et al., 2001; Bruggermann et al., 2004).

111 The next two examples document sea-level changes from the middle and towards the end of  
112 the LIG. From the Yucatán Peninsula (Mexico), an intermediate field location relative to the former  
113 Laurentide Ice Sheet, Blanchon et al. (2009) identify a 2 to 3 m RSL jump that they attribute to a  
114 short-lived (c. 1500 yr) interval of ice sheet instability that is ‘tentatively’ dated to c. 121 k yr ago  
115 (Figure 1C). The evidence for this sea-level jump is the sudden demise of an outer lower reef  
116 (developed to a sea-level at +3 m) that coincided with backstepping and accretion of an inner patch-  
117 reef (to a c. 3 meter elevation; see also Blanchon, 2010). Finally, in far-field western Australia,  
118 O’Leary et al. (2013; see also O’Leary et al., 2008) use stratigraphic and geomorphic mapping with U-  
119 series dating of fossil coral reefs to identify two architecturally distinct LIG highstand units that are  
120 separated by an unconformity contact and a palaeosol. These coral reef units are interpreted as

121 indicating two phases to the LIG highstand; the former forming over the period 127 to 119 k yr ago  
122 under slow 3 m RSL fall from an initial highstand and the second unit deposited late in the  
123 interglacial, dated to  $118.1 \pm 1.4$  k yr ago. The two units are observed at c. 2.5 m and c. 5-6 m above  
124 present MSL. O’Leary et al. (2013) include a glacial isostatic adjustment (GIA) modelling analysis as  
125 part of their study and propose a broadly stable “eustatic” sea level (i.e. ice-volume equivalent) of c.  
126 +3 to 4 m between 127 and 119 k yr ago and one at +9 m for c. 118 k yr ago, implying a c. 5 m sea-  
127 level jump late in MIS 5e (Figure 1E). Although the tectonic stability of the region has been  
128 challenged (Whitney and Hengesh, 2015), the sea-level jump component has not.

129 From these studies we identify the following as important elements of sea-level change  
130 during the LIG highstand:

131

132 1. LIG sea-level oscillations are recorded at the start, during the middle and towards the end of the  
133 LIG highstand, and are often interpreted as evidence for polar ice-sheet collapse. The North  
134 American and European ice masses are typically assumed to have only contributed to RSL rise early  
135 in MIS 5e (up to 127 k yr ago: Rohling et al., 2008; O’Leary et al., 2013). Chronological differences  
136 reflect, in part, the duration of the different records and the interval of time when sea level was  
137 higher than present, which is when sedimentary records are preserved. Several far-field sites record  
138 unconformities and palaeosols within double-reef architectures, from which meter-scale RSL  
139 oscillations are reconstructed. In the intermediate field, younger reef elements bury older elements  
140 and a marked sea-level jump is inferred against a broadly stable background RSL trend.

141

142 2. It is claimed by some studies that there was at least one interval during the main LIG highstand  
143 when RSL fell and then rose, possibly abruptly, to re-attain a LIG highstand, several meters higher  
144 than before. In the coral reef examples discussed above, the RSL fluctuation during the MIS 5e  
145 highstand was c. 3 to 5 m in amplitude. The fluctuation seems best characterised as a multi-millennia  
146 period of relatively stable sea level that terminated with a short-lived period of relatively rapid rise,  
147 the midpoint of which is dated to between c. 121 and 118 k yr ago.

148

149 3. Rates of sea-level change calculated from one of the Red Sea isotope records (core KL11), suggest  
150 RSL fluctuations with magnitudes of up to 10 m in 1-2 k yr or less, with peak rates of rise of 2.1 cm yr  
151 and fall of -1.8 cm yr during the LIG highstand (Rohling et al., 2008). The size of the Red Sea RSL  
152 fluctuations are approximately twice as large as those inferred from most reef records.

153

154            These different elements of the LIG sea level records form a basis for considering how near-  
155 field, low energy depositional environments might have responded to such changes.

156

### 157 **3. Relative sea-level change and coastal evolution on near-field coasts in northwest Europe**

158            The LIG near-field coasts of northwest Europe differ to their far- and intermediate-field  
159 counterparts in several ways. Because of their proximity to the Saalian (MIS 6) ice sheets, RSL  
160 records from this region were strongly impacted by GIA. In some peripheral locations, such as the  
161 Netherlands, forebulge collapse meant that the background (millennial) rate of RSL rise was  
162 relatively stable late into the LIG (Lambeck et al., 2006). This contrasts with the situation in far- and  
163 some intermediate-field sites where RSL was already falling by this time (Figure 2).

164

165            A second difference is that the LIG highstand coastal deposits can be well preserved in  
166 sedimentary depocenters, some of which have recorded post depositional subsidence, and are  
167 relatively complete based on known biostratigraphical changes established from northwest  
168 European terrestrial pollen zones (Müller, 1974). Most paleo-coastal records occur as infills of deep  
169 Saalian depressions that provided accommodation space for deposition as well as a good chance of  
170 preservation during RSL fall at the end of the LIG. Many depressions also experienced post  
171 depositional subsidence which further aided preservation.

172

173            A third difference is the abundance of fine-grained minerogenic and organic sediments that  
174 have accumulated in North Sea estuaries, tidal inlets, and coastal lagoons. These environments  
175 provide excellent archives from which to develop RSL records and reconstruct system-wide  
176 responses to abrupt or gradual sea-level forcing. Where these sequences have remained  
177 waterlogged since their deposition and include fossiliferous successions from intertidal to supratidal  
178 and terrestrial facies, the strong vertical zonation of coastal plant and animal communities with  
179 respect to tidal inundation can yield well-constrained sea-level data.

180

181            For many of the above reasons, near-field coastal systems are intensively studied for  
182 Holocene sea level; in northwest Europe they have been used to identify short-term (centennial)  
183 early-Holocene sea-level jumps of 1 to 3 m (e.g. Hijma and Cohen, 2010), instantaneous events such  
184 as tsunami and storms (e.g. Dawson et al., 1988), as well as gradual multi-century to millennial-scale  
185 RSL fluctuations (e.g. Denys and Bateman, 1995; Beets and Van der Spek, 2000; Kiden et al., 2002;  
186 Waller and Long, 2003). There are differences between Holocene and LIG coastal environments (e.g.  
187 Cohen et al., 2014), largely due to variations in substrate topography and composition, due to the

188 configuration of the main rivers (Busschers et al., 2007; 2008; Peeters et al. 2015) and, potentially,  
189 due to differences between the Holocene and LIG deglaciation, GIA and RSL histories. Nevertheless  
190 the methods and understanding developed from Holocene RSL studies are instructive when  
191 considering potential LIG RSL changes.

192

#### 193 **4. A conceptual model of Last Interglacial-type sea-level oscillations recorded in the near-field** 194 **coastal architecture of northwest Europe**

195 An abrupt RSL rise would cause the landward and upwards migration of an estuary and  
196 associated environments; indeed entirely new estuaries might develop in previously freshwater  
197 valley settings with freshwater swamps rapidly replaced by fluvial-tidal open-water environments.  
198 Hijma and Cohen (2010; 2011) record such a change at the time of an early Holocene sea-level jump  
199 that preceded the 8.2 ka BP cold event, when RSL in the Netherlands rose by c. 4 m in a few  
200 centuries due in part to the sudden drainage of Lakes Agassiz and Ojibway on the margins of the  
201 Laurentide Ice Sheet. In the Rhine-Meuse palaeovalley the coastline moved landward by c. 30 km  
202 and fluvial-deltaic wetland forests and open reed and sedge wetlands were transformed into tidal  
203 flats (Hijma and Cohen, 2011; Bos et al., 2012).

204

205 Within transgressed valleys, wind and wave energy would increase as surface area and  
206 water depth also increase (e.g. as happened in the mid to late Holocene within the Delaware  
207 Estuary, USA; Fletcher et al., 1990). Extensive flood tidal deltas would develop during the initial  
208 establishment of tidal inlets and barrier systems, and would migrate landwards (depending on  
209 sediment availability; Beets and Van der Spek, 2000), potentially over tidal flat and lagoonal  
210 environments. Alternatively, if sediment supply were deficient or sea-level rise very rapid, coastal  
211 barriers might be unable to migrate landwards and become drowned, with new barriers forming  
212 inland (e.g. Hijma et al., 2010). Eventually, the position of the barrier system would stabilize. In the  
213 lagoonal settings that result from that stage, an event of abrupt sea-level rise would potentially  
214 erode and rework marginal unconsolidated basin deposits.

215

216 From a lithostratigraphic perspective, an abrupt rise in RSL would see the development of  
217 extensive transgressive overlaps that mark the facies change from predominately fresh/brackish-  
218 water to marine deposits. In lagoons, any fall in salinity stratification might see the accumulation of  
219 more massive clastic deposits and a reduction in anoxic-tolerant biota, whilst reworking of marginal  
220 basin deposits might disrupt trends in litho- and biostratigraphy associated with more gradual  
221 sedimentation.

222

223           From a highstand, a rapid RSL fall (such as that associated with one of the Red Sea RSL  
224 oscillations; Rohling et al., 2008) would reverse many of the above processes. Mid-estuary tidal  
225 flood basins would be replaced by fluvial channel environments and migrate downstream, replacing  
226 former salt marsh and tidal flats. A raised estuarine terrace might develop, creating a depositional  
227 environment for freshwater floodplain sediments. Gravel and sand barriers would consolidate as  
228 increased sediment is made available due to shallower near-shore waters. In shallow coastal  
229 lagoons, the frequency of tidal inundation would fall sharply, and a freshening of the lagoon be  
230 accompanied by enhanced salinity stratification. Tidal inlets would likely constrict when sea level in  
231 the lagoon behind it falls and their tidal prism is reduced.

232

233           From a stratigraphic perspective a fall in sea level would be manifest in a variety of ways;  
234 there may be a hiatus or abrupt lithological change that marks a cessation or abrupt change in  
235 sediment accretion, or the widespread development of (partly erosive) regressive overlaps or soil  
236 horizons that mark the lithostratigraphical change between predominately brackish-water and fresh-  
237 water deposits (e.g. Peeters et al., 2015). Regressive overlaps would extend seawards and fall in  
238 elevation. An increase in salinity stratification within deeper lagoons might coincide with the  
239 accumulation of finer-grained, potentially laminated, organic-rich sediments and a fall in the  
240 occurrence of open coast biota.

241

242           Multiple meter-scale oscillations in sea-level in a few millennia (or less), would be more  
243 disruptive. In deep depositional centres they could create stacked sequences of estuarine and  
244 freshwater deposits characterised by multiple changes in sedimentary type, and along fringes leave  
245 complex erosional contacts and sedimentary hiatuses. Also the architectural properties of the  
246 record (i.e. the spatial continuity and uniformity of stratigraphic bounding surfaces), would be  
247 affected by the propensity for erosion to destroy older sequences deposited during earlier episodes  
248 of sea-level change.

249

## 250 **5. Evidence for near-field Last Interglacial relative sea-level changes in northwest Europe**

251           The higher than present sea levels of the LIG had a profound impact on the geometry of  
252 northwest European coasts. In the Southern North Sea, LIG shorelines were established inland of  
253 the present coastline, with shallow inner shelf seas developing in the Netherlands, Belgium and  
254 Germany (e.g. Jelgersma et al., 1979; Zagwijn, 1983; 1989; Paepe and Baeteman, 1979; Behre et al.,  
255 1979; De Gans et al., 2000; Streif, 2004; Peeters et al., 2015) (Figure 3). In England, coastal LIG

256 deposits (Ipswichian stage) are preserved in parts of Lincolnshire and Cambridgeshire (e.g. Gaunt et  
257 al., 1974; Gao and Boreham, 2011), within the Thames Estuary (e.g. Gibbard, 1985; Preece, 1999;  
258 Bridgland, 1994), and at several sites along the English Channel coast (e.g. Bates et al., 1997; 2010;  
259 West and Sparks, 1960; Preece et al., 1990). Based on their pollen assemblages, most of these  
260 coastal sequences preserve sediments that span only a relatively short interval of the LIG RSL record.  
261

262 Thicker, more complete LIG sequences are recorded in the Baltic and White Sea, and also on  
263 the North Sea coast of Germany and the Netherlands. LIG sequences occur in the White Sea region  
264 (reviewed by Grøsfjeld et al., 2006), the eastern Baltic at Peski (Karelian Isthmus, Russia; e.g.  
265 Miettinen et al., 2002; Miettinen et al., 2014) as well as in the western Baltic in the Lower Vistula  
266 region (Poland; e.g. Marks et al., 2014) and at Mommark (Denmark; e.g. Eriksson et al., 2006). On  
267 the North Sea coast of Germany, thick (c. 20 m) LIG records are described from Dagebüll (e.g. Temler  
268 et al., 1995; Win et al., 2000). Notwithstanding these long records, the most complete near-field  
269 sedimentary record for LIG RSL changes in northwest Europe is preserved in the Saalian glacial basins  
270 of the Netherlands.  
271

272 Lobes of the Saalian Scandinavian ice sheet reached their maximum southward expansion at  
273 c. 160 k yr ago (within MIS 6) (Busschers et al., 2008) and in the central Netherlands excavated a  
274 series of deep glacial basins up to -100 m NAP (Dutch Ordnance Datum = approximately mean sea  
275 level). These became depositional traps during the deglaciation stages (Drenthe and Warthe  
276 Substages of the Saalian), the Eemian (before and after the marine transgression), and the last  
277 glacial (Weichselian) (de Gans et al., 1987; Busschers et al. 2007; 2008; Peeters et al. 2015) (Figure  
278 4). The main Amsterdam and (adjacent) Amersfoort Basins have a sill in the base of the in situ  
279 Eemian sediments with a present height of c. -40 m that separated them from the Eemian Rhine  
280 Valley towards the north (Cleveringa et al., 2000; Gunnink et al., 2012; Peeters et al., 2015). This sill  
281 controlled the timing of initial marine inundation and influenced tidal mixing in each lagoon (Van  
282 Leeuwen et al., 2000), at least until sediment accumulation exceeded their heights.  
283

284 Many hundreds of boreholes exist from the Amsterdam Basin (Figure 5). A key site is the  
285 Amsterdam Terminal borehole (Figure 4) (Bosch et al., 2000; van Leeuwen et al., 2000; de Gans et  
286 al., 2000; Beets and Beets, 2003; Beets et al., 2003; Busschers et al., 2007; Peeters et al., 2015). The  
287 Eemian fill comprises a c. 30 m thick sequence of clay-rich sediments that are overlain by a wedge of  
288 shell-rich sand that is up to 20 m thick in the north of the basin and which disappears to the south.  
289 The deposits are mainly shallow marine in origin and span six distinct pollen zones (NW European

290 pollen zones E1 to E6) (Zagwijn, 1983; de Gans et al., 2000) that include a brief initial temperate  
291 terrestrial stage (E1, E2), the main RSL transgression (in E3-E4a), the highstand (during E4b, E5) and  
292 the start of the subsequent regression and RSL fall at the end of the LIG (E6). A broadly similar  
293 sedimentary fill is found in the Amersfoort Basin. Here, the key cored sites (including the stratotype  
294 site Amersfoort-1) (Zagwijn, 1961; Cleveringa et al., 2000), are from a more marginal, shallower  
295 setting (Figure 6). Importantly, the upper part of the Amersfoort-1 LIG record preserves a fining-  
296 upwards sequence that formed as tidal flats were replaced by salt marsh and then freshwater peat  
297 (Zagwijn, 1961; Cleveringa et al., 2000).

298

299 Zagwijn's (1983; 1996) RSL graph of Dutch Eemian RSL changes (Figure 1F) uses index points  
300 that have ages inferred from pollen biostratigraphy and altitudes established with respect to high  
301 tide level. The graph shows RSL rise from c. -25 m NAP to a highstand of c. -8 m NAP during E5,  
302 before falling during E6. A preceding part of the curve, that spans the initial rise from -45 to -25 m, is  
303 based on the inundation of the Amsterdam basin and on offshore (North Sea) cores. The highstand  
304 sea-level index points used by Zagwijn (1983) and Streif (1990) are collected from several sites  
305 scattered along the fringe of the Eemian embayment in the central Netherlands. They have been  
306 influenced differentially by the effects of long-term sediment compaction, tectonics and isostatic  
307 motions, the latter including GIA. Kooi et al. (1998) estimate that the long-term net effect of these  
308 processes varies between c. 12 and 18 cm/k yr. Recent re-analysis of the Zagwijn (1983) data,  
309 updated by Streif (1990), corrects for each of these effects (Figure 1F; Lambeck et al., 2006; Kopp et  
310 al., 2009).

311

312 The approximate duration of the European regional pollen zones is known from annually  
313 varved lake deposits in northwest Germany (Müller, 1974). The key pollen zones for the purposes of  
314 this study, those that span the culmination of the LIG RSL transgression and highstand in the  
315 Netherlands, are E4b (duration c. 1-1.2 k yrs) and E5 (duration c. 4 k yrs). Although the duration of  
316 the transgression and highstand is reasonably well-known, there is debate regarding its numerical  
317 age. Zagwijn (1983) originally dated the highstand peak to late in the LIG (c. 120 k yr ago), based on  
318 correlation with the deep sea oxygen isotope record of MIS 5e (Shackleton and Opdyke, 1973). A  
319 late LIG highstand (c. 123 k yr ago) is also proposed by more recent correlation of the NW European  
320 Eemian to tied terrestrial pollen and marine  $\delta^{18}\text{O}$  records and implies a lag between the start of MIS  
321 5e and the far-field LIG highstand of c. 4-5 k yr (Sanchez-Goñi et al., 1999 ; Shackleton et al., 2002;  
322 2003). A relatively late LIG sea-level highstand compared to that seen in the far-field is compatible  
323 with what is seen during the Holocene.

324

325 In contrast, Funder et al. (2002) correlated the highstand with that recorded by far-field  
326 coral reefs to  $128 \pm 1$  k yr ago (McCulloch and Esat, 2000; Shackleton, 2000). Beets et al. (2006) also  
327 prefer this timing, correlating the Eemian transgression to the main rise in benthic  $\delta^{18}\text{O}$  at the start  
328 of MIS 5e. Kopp et al. (2009) establish a numerical age for the Zagwijn (1983) sea-level curve by  
329 tuning this against the global oxygen isotope stack of Lisiecki and Raymo (2005). From this they  
330 propose a broad, flat peak to the highstand that lasted between c. 127-121 k yr ago.

331

332 More recently, it has been proposed that the start of the terrestrial Eemian in northwest  
333 Europe is even younger than that proposed by Sanchez-Goñi et al. (1999) and Shackleton et al.  
334 (2002; 2003) (c. 123 k yr ago). A key line of supporting evidence comes from palaeomagnetic  
335 investigations of the Blake Event, a prominent dipole moment low (Singer et al., 2014). In Neumark  
336 Nord 2, Germany, it is identified in the earliest part of the LIG pollen sequence (Sier et al., 2011). A  
337 second study of the Blake Event (supported by luminescence dating) by Sier et al. (2015) from a core  
338 drilled at Rutten – in the area of the Zagwijn (1983) sea-level data - places the start of the Eemian  
339 (zone E1) to c. 121 k yr ago, and by implication the sea-level highstand (zone E4b) was attained at c.  
340 119 k yr ago (Figure 7). In this model, the end of the nearfield highstand (end of zone E5) correlates  
341 with the end of MIS 5e, whilst the following early stage of RSL fall (zone E6) correlates with the start  
342 of MIS 5d (at c. 114 k yr ago).

343

## 344 **6. Evidence for last interglacial ('Eemian') sea-level oscillations or jumps in the Netherlands**

345 We now focus our attention on the evidence for RSL changes during the sea-level highstand  
346 of the Dutch Eemian sequences, starting with the lower-energy sequence recorded in Amersfoort-1.  
347 By the start of E5, sedimentation had already infilled the original basin to above its sill elevation and,  
348 as a result, the basin was operating as a tidal lagoon. Any large RSL oscillation during the LIG  
349 highstand should be manifest in the core between c. -23 m NAP and -10.5 m NAP (the latter  
350 representing the level at which freshwater peat accumulation begins). This interval spans pollen  
351 zones E4b to E6 (Figure 7).

352

353 Following the initial tidal inundation in E2, several proxies indicate that water depths and  
354 accommodation space increased due to RSL rise (Cleveringa et al., 2000). The lithology records an  
355 up-core transition from shallow-water, brackish clay-rich sands and shells to a sand with shells that  
356 coarsens upwards. The transition between these facies is dated to the end of E4b and the start of E5  
357 (c. -22.50 m NAP). Molluscs and foraminifera above this level point to a deepening and an increase

358 in salinity, with some of the latter taxa washed in from the North Sea inner shelf. Maximum water  
359 depths of c. 10 m are inferred between core depths of -18.8 m to -16.8 m NAP. Accommodation  
360 space decreased above this level as sedimentation outstripped the slowing rate of RSL rise. The  
361 coarsening-upwards trend in the sands seen in the second half of E5 indicates an increase in wave  
362 action and related hydrodynamic processes. Heavy mineral data and the setting of the site indicate  
363 the sands were likely derived from the local reworking of ice-pushed ridges close to the basin  
364 margins. Changes in the molluscan and ostracod assemblages point to local reworking within these  
365 sands, most likely owing to wave or current processes.

366

367         Towards the end of E5 (c. -14.50 m NAP) the shelly sands described above are abruptly  
368 overlain by c. 4 m of clays and sandy clays. Cleveringa et al. (2000) suggest a discontinuity in  
369 sedimentation (they use the term 'hiatus') between the shelly sands and overlying clays, also noting  
370 an abrupt change in molluscan and foraminiferal assemblages that indicate a change from a lagoon  
371 to an 'estuary-like water body surrounded by mudflats' (Figure 6). The uppermost fine-grained  
372 deposits were deposited under tidal mudflat and then salt marsh conditions. They record the final  
373 infilling of the basin during the broad RSL highstand that culminated in freshwater peat formation as  
374 RSL started to fall. The stratigraphic cross-section of the Amersfoort area compiled by Zagwijn  
375 (1983) (Figure 5) shows only a single regressive contact to the overlying freshwater peat that can be  
376 traced across the basin to its margins; there are no intercalations of organic or minerogenic  
377 sediments at the basin margins.

378

379         In the Amsterdam Terminal borehole, a thick sequence of clays and silts formed during the  
380 latter part of E4 and throughout E5 (van Leeuwen et al., 2000). Crumbly and then homogenous silty  
381 clays grade upwards into a laminated silty clay at c. -46 m NAP. Foraminifera and dinoflagellates  
382 indicate a steady deepening of the lagoon, and a strengthening of the exchange of water with the  
383 open sea that created a well-ventilated water column (van Leeuwen et al., 2000). An increase in  
384 tidal influence is also suggested by an increase in salt marsh pollen frequencies. This phase of  
385 unobstructed exchange of water with the open coast ended shortly after the start of E5 when the  
386 deposition of laminated finer-grained clays indicates the establishment of more restricted, lagoonal  
387 conditions. Drawing analogies with the Holocene development of coastal beach barriers - which in  
388 this region stabilised c. 6.5 k yr ago (e.g. Beets & Van de Spek, 2000), some 1.5-2 k yrs after Holocene  
389 RSL had begun to slow (Hijma & Cohen, 2010) - van Leeuwen et al. (2000) and de Gans et al. (2000)  
390 interpret these changes as recording a phase of sand barrier development in the sill area which  
391 restricted tidal inflow and lead to lagoonal stratification.

392

393           The final phase of the LIG highstand in the Amsterdam Terminal borehole above -38 m NAP  
394 comprises an upward coarsening trend with massive silty clays that contain increasingly thicker  
395 laminations of sand during the latter part of E5 (van Leeuwen et al., 2000). Foraminifera and  
396 molluscs suggest deposition in a large, tidally-affected lagoon that was shallowing, and which once  
397 had an unobstructed exchange of water with the open marine areas to the north. This may record  
398 the breakdown of a former barrier or spit (Van Leeuwen et al., 2000; de Gans et al., 2000). Van  
399 Leeuwen et al. (2000) propose reworking of these coarse sands by tidal channels as accommodation  
400 space fell further, as shown by a shell-rich lag deposit containing clay pebbles reworked from  
401 mudflats, encountered at c. -32 m NAP. However, Busschers et al. (2007) and Peeters et al. (2015)  
402 indicated that this lag is the base of a much younger (MIS 4) fluvial channel belt of the Rhine system.

403           In summary, the Amersfoort-1 and Amsterdam Terminal cores contain no evidence for  
404 abrupt oscillations in RSL during the LIG highstand (E4-E5). The discontinuities that do exist can all  
405 be explained by analogy to mid-Holocene coastal dynamics, without invoking vertical changes in RSL  
406 (other than RSL rise at the onset of the highstand and fall towards the end, Figure 4). The hiatuses  
407 seen in the Amersfoort-1 core are not seen within the finer-grained sediments that accumulated in  
408 the deeper-water setting of the Amsterdam Terminal core, away from the basin margins. Nor does  
409 the latter contain evidence for large-scale input of sand, organic material and/or pollen that might  
410 suggest sudden RSL rise and associated erosion of basin valley flanks by wave or tidal processes.

411

## 412 **7. A critical evaluation of the near-field relative sea-level record from northwest Europe**

### 413 *7.1 Issues of dating and vertical precision in near-field relative sea-level records*

414           It could be argued that the relatively poor vertical precision of the near-field LIG sea-level  
415 index points may hinder our ability to identify RSL oscillations/jumps; indeed we note that the  
416 highstand sediments in the Amsterdam Terminal core (but not the Amersfoort-1 core) are silty clays  
417 and coarse shelly sands that are not ideal for developing detailed RSL records. The chronology of  
418 these records is a floating one based on pollen zones and their registration in annually laminated  
419 lacustrine sediments elsewhere in NW Europe. The absolute age-models for the sequence vary  
420 between studies (see above) and this makes direct correlation with other, better-dated sequences  
421 from far- and intermediate-field settings difficult. The near-field highstand record is also  
422 considerably shorter than the far-field highstand, such that the former does not overlap all of the  
423 latter (i.e. the near-field Eemian sea-level record cannot overlap *both* the Seychelles *and* the Yucatán  
424 records of MIS 5e). We also note that during the main Eemian transgression, when sea level rose  
425 rapidly from c. -45 m NAP (the lowest sea-level indicator) to c. -15 m NAP (the start of the local

426 Dutch highstand), we cannot currently rule out the possibility of a sea-level jump within the earliest  
427 part of the Eemian (during E1-E3). However, when considering the sedimentary architectures from  
428 which these index-points are derived (by analogy to the reef studies reviewed in section 2) the  
429 resolution is similar and we identify no evidence for sea-level jumps in the time interval considered.

430

### 431 *7.2 Local and regional signals in the last interglacial sea level records of northwest Europe*

432 Looking beyond the Netherlands, we see no evidence for abrupt RSL changes in the adjacent  
433 Eemian sequences described by Temler (1995) and Winn et al. (2000) from North Germany. LIG  
434 marine deposits from south and east England typically only span brief time intervals and also record  
435 no evidence for synchronous abrupt RSL changes (e.g. West and Sparks, 1960; West et al., 1970;  
436 Holyoak and Preece, 1985). In Trafalgar Square (London), LIG deposition occurred at the landward  
437 limit of tidal influence. This setting should have been sensitive to abrupt RSL changes in the LIG  
438 highstand, yet only gradual changes are observed (Gibbard, 1985; Preece, 1999).

439

440 Elsewhere in northwest Europe, two extensive reviews from the Baltic/White Sea regions  
441 also find no evidence for such events during the RSL highstand (Funder et al., 2002; Lambeck et al.,  
442 2006). Indeed, of the examples of gradual changes in LIG RSL inferred from detailed litho- and  
443 biostratigraphy we highlight the study of Grøsfjeld et al. (2006) from the White Sea region of  
444 northwest Russia and the multiproxy examination of displaced marine sediments in the Mommarck  
445 area of southern Denmark (e.g. Haila et al., 2006; Kristensen and Knudsen, 2006) as two further  
446 records which document gradual RSL rise and then fall over the LIG highstand.

447

448 It would be misleading, though, to suggest that the possibility of RSL fluctuations in the LIG  
449 have not been made from northwest European sites. For example, Hollin (1977) interpreted  
450 evidence from the Thames Estuary as evidence for a two-phase RSL record in the LIG, including a  
451 jump from c. 0 m to +16 m at the end of the interglacial that they attribute to an Antarctic 'surge'.  
452 More recently, two studies suggest salinity variations during the LIG RSL highstand from the White  
453 Sea. Diatoms from Peski, on the Karelian Isthmus, suggest a brief mid-Eemian lowering in salinity  
454 that Miettinen et al. (2014) speculate could be related to a mid-LIG (c. 126 k yr ago in their age  
455 model) sea-level fall. At Cierpieta, in the Lower Vistula region of Poland, Marks et al. (2014) report a  
456 short-lived 'transgressive pulse' that happened about 6.2-6.6 k yrs into the interglacial (during the  
457 end of E5 or start of E6) and was associated with an increase in tidal and littoral diatoms. However,  
458 each of these studies is equivocal in one or more ways; the study of Hollin (1977) is now recognised  
459 as flawed because all of the high level sites they thought were from the LIG are now attributed to

460 MIS 7 or older (e.g. Bridgland, 1994; Penkman et al., 2011), whilst the Baltic sites are sensitive to  
461 climate-driven changes in freshwater discharge and changes in tidal dynamics that accompanied the  
462 evolution of the LIG White Sea (Miettinen et al., 2014; Marks et al., 2014).

463

464 In summary, the resolution of the near-field RSL reconstructions from the LIG is not as  
465 detailed as those from the Holocene. Nevertheless the size of the sea-level signal that we are testing  
466 for is much larger than any RSL variation during the Holocene highstand. Moreover, we would  
467 expect a large oscillation/jump in RSL to be a regional signal, with evidence preserved in many of the  
468 near-field coastal sequences described above. We see no evidence for regionally synchronous  
469 changes of this nature (other than the initial rapid transgression) and observe that synchronous  
470 regional sea-level fall is first recorded at the onset of zone E6, in the last millennia of the Eemian.

471

## 472 **Section 8.1**

### 473 *8.1 Sea-level signals of Last Interglacial ice sheet instability*

474 The size of the LIG sea-level oscillations/jumps reconstructed in several far- and  
475 intermediate-field RSL studies implicates either Greenland and/or Antarctica as potential sources,  
476 although other ice masses may have contributed too (see below). Sea-level fingerprinting provides a  
477 means to assess such a possibility, as previous studies predict spatially variable sea-level changes  
478 consequential to ice sheet collapse events. Hay et al. (2014) show that the ability to identify rapid  
479 collapse events in sea-level records should be more robust than with current gradual melting  
480 because the sea-level fingerprint of a sudden event is less sensitive to uncertainties in the geometry  
481 of the melt region and will dominate over the signal from steric effects.

482

483 One hypothesis for the lack of obvious sea-level oscillation/jump in the Netherlands (and  
484 elsewhere in northwest Europe) is that this reflects the geographical proximity of these sites to the  
485 LIG Greenland Ice Sheet. To test this hypothesis, we present a pair of experiments in which we  
486 modified global ice sheet deglaciation histories (ICE-5G; Peltier, (2004) and HUY-2; Simpson et al.  
487 (2009)) to simulate hypothetical episodes of ice sheet collapse/regrowth in Greenland or Antarctica.  
488 The combined ICE-5G-HUY-2 deglaciation history detailed in Simpson et al. (2009) focuses on sea-  
489 level change since the Last Glacial Maximum with the sea level calculation initiated at 122 k yr ago.  
490 In this model, sea level reached +8 m above present at 122 k yr ago until an abrupt fall at 116 k yr  
491 ago.

492

493 We modify the ICE-5G-HUY-2 deglaciation history by adding several synthetic eustatic sea-  
494 level fluctuations between 127 and 116 k yr ago. We limit the mass fluxes from Greenland  
495 (experiment G1) to the equivalent of  $\pm 2$ -3 m eustatic sea level, and the mass fluxes from Antarctica  
496 (experiment A1) equal to  $\pm 5$  m (Figure 8, Table S1). Note that the timing and number of fluctuations  
497 differ in each experiment and that our interest is in the relative size of the sea-level signal between  
498 sites and not their absolute values. Moreover, although these models are purely illustrative, we  
499 present a 'realistic' ice sheet history by allowing the geometry of the Greenland Ice Sheet to oscillate  
500 between states representative of 122 k yr ago and 110 k yr ago during the period of interest. For  
501 Antarctica, we construct oscillations in mass balance by applying scaling factors for ice thickness  
502 uniformly across the ice sheet, which we consider adequate for the purposes of this experiment (see  
503 Mitrovica et al. (2009) for details of the influence of more complex melt patterns). The scaling  
504 factors and details of ice sheet deglaciation for the period 127-117 k yr ago used in both scenarios  
505 are detailed in Supplementary Information (Table S1). Global ice extent outside of the target areas  
506 of each experiment (Greenland and Antarctica) is fixed at its 122 k yr ago extent during the LIG  
507 maximum.

508

509 We input the resulting loading histories into the sea-level model presented in Mitrovica and  
510 Milne (2003) and Kendall et al. (2005), and use this to predict RSL changes at the Netherlands,  
511 Seychelles, Red Sea, the Yucatán Peninsula, Bahamas and Western Australia (Figure 8). In the sea-  
512 level model, we do not include calculation of the Earth response to pre-LIG changes in global ice and  
513 ocean loads since the aim of this experiment is only to test sensitivity for recording jump-events and  
514 not to accurately reproduce any specific global or regional sea-level history. The sea-level change  
515 associated with pre-LIG glaciation-deglaciation cycles will not alter the magnitude of the applied LIG  
516 oscillations in each experiment but rather superimpose them on a long-term secular trend. We  
517 present the results for experiments A1 and G1 in which Antarctica and Greenland, respectively, are  
518 subjected to growth-collapse periods during the LIG.

519

520 Model results for an Antarctica-only source for the sea-level fluctuations show that all sites  
521 track the modified eustatic curve reasonably closely, especially late in the LIG (after 123 ky ago),  
522 when the Northern Hemisphere has almost completely deglaciated (Figure 8). This overall pattern is  
523 in agreement with other sea-level fingerprint experiments (Mitrovica et al., 2001; Hay et al., 2014)  
524 and shows that all of these sites should be sensitive to sea-level oscillations sourced from  
525 Antarctica. An Antarctic-sourced sea level oscillation seen in the Yucatán Peninsula, for example,  
526 should be manifest at the same time, and with a broadly similar amplitude, in each of the other

527 sites. Of course, whether the signal is actually preserved in coastal stratigraphic data would depend  
528 on the particular local RSL history (including background rates of tectonic- and GIA- and  
529 hydroisostasy-induced subsidence, e.g. in Yucatán, in the Netherlands). It is conceivable that in  
530 near-field sites, such as the Netherlands, RSL was potentially below present at the time of both an  
531 early and a late interglacial sea-level jump recorded in far- and intermediate-field settings, i.e. after  
532 c. 128 k yr (Seychelles) and before c. 119 k yr ago (Yucatán Peninsula).

533

534 Model results for a Greenland-only source show more considerable differences between the  
535 study sites (Figure 8). Because of the proximity of the Netherlands to Greenland and a deglaciating  
536 Eurasian Ice Sheet, the predicted sea-level record for this site is effectively ‘blind’ to the oscillations  
537 ( $\pm 2$  m) recorded in each of the far-field sites (see also Mitrovica et al. (2001)). This experiment raises  
538 the possibility that the lack of obvious sea-level oscillations/jumps in the Netherlands is because  
539 abrupt RSL changes observed in far-field sites were sourced from Greenland.

540

#### 541 *8.2 Near-field records of ice-sheet instability*

542 The experiments above show that the Netherlands is potentially sensitive to abrupt sea-level  
543 changes sourced from Antarctica, an advantage over other far-field RSL sites that are potentially  
544 unable to discriminate between Antarctic and Greenland contributions. Marine diatoms and  
545 cosmogenic isotopes from beneath the West Antarctic Ice Sheet (WAIS) indicate exposure to marine  
546 conditions at an unspecified interglacial (Scherer et al., 1993; Scherer et al., 1998), but they cannot  
547 be ascribed with any confidence to the LIG. Bryozoan diversity data suggest an open seaway  
548 between the Weddell and Ross Sea at some time in the late Quaternary (Barnes and Hillenbrand,  
549 2010; Vaughn et al., 2011). Rapid collapse of WAIS should result in massive discharges of icebergs,  
550 meltwater and significant IRD (Hillenbrand et al., 2009), yet no such evidence is recorded in sea bed  
551 cores from sites close to Antarctica (e.g. O’Cofaigh et al., 2001; Hillenbrand et al., 2002). A review of  
552 17 marine records from the wider Southern Ocean by Capron et al. (2014) reveals no evidence of  
553 synchronous abrupt warming or cooling events during the main part of the LIG that might trigger, or  
554 record, abrupt Antarctic instability.

555

556 Ice-core records from Antarctica provide an important source for identifying abrupt changes  
557 in climate and associated changes in ice sheet surface elevation that could be related to sea-level  
558 changes, although as Bradley et al. (2012) show, the location of many ice cores, mainly in East  
559 Antarctica, is not ideal for testing models of WAIS collapse. Common to most Antarctic ice cores is  
560 an early (c. 128 k yr ago, using the AICC2012 timescale; Bazin et al., 2013) maximum in LIG air

561 temperatures, 1-4 °C warmer than present and lasting c. 3 k yrs, that was followed by a multi-  
562 millennial scale plateau (126-122 k yr ago; Masson-Delmotte et al., 2011) of stable temperatures  
563 (Jouzel et al., 2007). This was followed by a cooling into the glacial inception at c. 110 k yr ago  
564 (Capron et al., 2012).

565

566 Steig et al. (2015) note that the  $\delta^{18}\text{O}$  data from Mt. Moulton (West Antarctica) show a  
567 statistically significant smaller difference between LIG and Holocene values than is seen in East  
568 Antarctic ice cores. This is consistent with the differences predicted by a climate model that includes  
569 WAIS collapse that was well underway by ~130 k yr ago. Steig et al. (2015) note several caveats to  
570 this study; it assumes a fixed linear correlation between temperature and the ice core isotope data,  
571 whilst the Mt Moulton record implies relatively low and high  $\delta^{18}\text{O}$  values in MIS 5e and the coldest  
572 part of the glacial period (~20-40 k yr ago) compared to other ice core records. Moreover, there is  
573 no strong evidence that WAIS was significantly thicker during this period. Model experiments by  
574 Holden et al. (2010) of the impact of meltwater on the Antarctic Meridional Overturning Circulation  
575 and associated air temperatures over Antarctica also require WAIS melting so as to satisfactorily  
576 predict the observed peak temperatures in the early part of the LIG.

577

578 These data-model studies suggest, contrary to the geological evidence, that the early LIG  
579 warmth may have been associated with ice-sheet retreat that contributed to the higher than present  
580 sea levels observed during the LIG. But the relationship between atmospheric warmth and mass  
581 change in Antarctica is complex; indeed a period of warmth in the mid Holocene in Antarctica  
582 between c. 4.5 and 2 k yr ago (Bentley et al., 2009; Sterkin et al., 2012) coincided with a 30%  
583 increase in ice accumulation over West Antarctica (Siegert and Payne, 2004), whilst Frieler et al.  
584 (2015) use a combination of ice core data and modelling to derive a continental-scale increase in  
585 accumulation of approximately  $5 \pm 1 \text{ \% K}^{-1}$ , that would have competed with surface melting and  
586 dynamical loss at the ice sheet margin by other processes. So, although it is overly-simplistic to  
587 assume warmer air temperatures equate directly to ice-sheet mass loss, the main observation we  
588 draw from the Antarctic ice cores is that there is only one interval of significant warmth, and that  
589 this happened early in the LIG.

590

591 Turning to Greenland, evidence for LIG ice sheet behaviour is provided by sea-bed cores that  
592 contain proxies for Greenland vegetation cover, run-off and IRD. Pollen and spores from a sea-bed  
593 core located off southeast Greenland (core HU-90-014-013) indicate single peaks in relative  
594 abundances that implicate only a single interval of reduced ice cover in southern Greenland (de

595 Vernal and Hillaire-Marcel, 2008). Sr-Nd-Pb isotope ratios of silt-sized sediment in a core offshore of  
596 the southern tip of Greenland also indicate that the southern part of the ice sheet underwent  
597 continued retreat throughout much of the last interglacial (Coleville et al., 2011), although there is  
598 no evidence for multiple phases of ice-sheet collapse followed by regrowth. Records of Ti and Fe  
599 present in core MD99-2227, from the same region, suggest high but steady run-off throughout the  
600 main part of the LIG (Carlson et al., 2008; Carlson and Windsor, 2012).

601

602 Several records from the North Atlantic and Baffin Bay have evidence for one or more  
603 cooling episodes during the broad oxygen isotope plateau of the LIG that may record ice-sheet  
604 instability, although not necessarily all are attributed to Greenland (e.g. Maslin et al., 1998; Bauch,  
605 2011; Schwab et al., 2014). Sea surface temperature reconstructions and foraminiferal assemblages  
606 from core MD03-2664 (southwest Greenland) (Irvani et al., 2012), indicate a warm early start to MIS  
607 5e that was interrupted at c. 126 k yr ago by a 1.6 k yrs cooling event. This was accompanied by a  
608 significant freshening that is attributed to enhanced Greenland run-off. Galassen et al. (2014)  
609 identify multiple such transient cooling events that they relate to variations in North Atlantic Deep  
610 Water (NADW) production. Some of these are reminiscent of the 8.2 k yr cold event (see also Bauch,  
611 2011); indeed one is described in detail by Nicholl et al. (2012) who trace a red carbonate layer along  
612 the Labrador and Greenland margins of the Labrador Sea that they source to a large outburst flood  
613 from a LIG glacial Lake Agassiz, dated to c. 124.5 k yr ago. Variations in NADW during the latter part  
614 of the LIG are also identified by Galassen et al. (2014) at c. 119.5 and 116.8 k yr ago that are  
615 attributed to (unspecified) non-Greenland forcing.

616

617 The North Greenland Eemian Ice Drilling core (NEEM) records a maximum in air  
618 temperatures of  $+8 \pm 4$  °C (compared to the average of the last millennium) during the early Eemian  
619 (c. 126 k yr ago) (NEEM Members, 2013). Regular summer melt events occur between 127 and  
620 118.3 k yr ago, and the NEEM Members (2013) conclude that the GIS contributed no more than c. 2  
621 m to the LIG highstand. The NEEM core records a cooling after c. 121 k yr ago that broadly tracks  
622 the decline in northern hemisphere insolation. In NGRIP there is an abrupt cooling assigned to c.  
623 119 k yr, followed by relatively stable depleted  $\delta^{18}\text{O}$  phase (Greenland stadial 26) until c. 115 k yr  
624 ago (NGRIP Community Members, 2004). If the NEEM CH<sub>4</sub> tuned age-model (NEEM Members, 2013)  
625 is applied to the lower part of the NGRIP core (from Greenland stadial 24 towards the LIG), these  
626 ages become 1.5-2 k yrs younger (Sier et al. 2015).

627

628 Other potential near-field sources capable of driving abrupt sea-level changes include the  
629 Laurentide and Fennoscandinavian ice sheets. These are normally excluded from discussions on LIG  
630 sea-level (and GIA modelling) since it is assumed that by the peak of MIS 5e they had melted, or  
631 were too small to drive meter-scale changes in global sea level. However, the MIS 6  
632 Fennoscandinavian Ice Sheet was significantly larger than its LGM counterpart, accounting for up to  
633 50% of the total northern hemisphere ice volume (Svendsen et al., 2004; Lambeck et al., 2006). It is  
634 conceivable that this ice sheet may have continued to melt late into the LIG, and the GIA response  
635 and RSL signal in its near-field delayed in comparison to Termination I. In the eastern and northern  
636 Nordic Seas, Van Nieuwenhove et al. (2013) note that peak ocean temperatures occurred relatively  
637 late in the LIG, and implicate meltwater run-off from a residual Norwegian ice mass until as late as c.  
638 122-120 k yr ago (see also Bauch and Erlenkeuser (2008)). The absence of significant Laurentide-  
639 sourced IRD-events during MIS 6 beyond the Labrador Sea could be because the configuration of the  
640 Laurentide Ice Sheet was different to that of other glacials (i.e. smaller), or that no ice stream existed  
641 in Hudson Strait in MIS 6 (Naafs et al., 2013), in which case the ice sheet may have been more stable.  
642 Indeed, as noted above, Nichol et al. (2012) identify an abrupt drainage event that they attribute to  
643 former glacial Lake Agassiz in the Hudson Bay region at c. 124.5 k yr ago. A smaller and more stable  
644 Laurentide Ice Sheet might, therefore, have still been retreating during much of the interval covered  
645 by the Seychelles LIG sea level record, which is interpreted to reflect only Antarctic and Greenland  
646 contributions (Dutton et al., 2015).

647

648 The above discussion highlights that ice sheets other than Greenland and Antarctica may  
649 have played a role in causing sea level to vary during the LIG highstand and that their configurations  
650 and retreat dynamics during Termination II and MIS 5e were probably different to those during  
651 Termination I and the Holocene. However, as yet there is no compelling evidence for significant ice-  
652 sheet collapse in Antarctica or for polar ice sheet *regrowth* during the LIG that is required to explain  
653 abrupt oscillations in sea level, other than at the end of the highstand.

654

## 655 **9. Reflections on Last Interglacial sea-level studies**

656 Our first reflection is that few far- and intermediate-field RSL studies attribute sea-level  
657 jumps to known periods of ice-sheet instability. There is a large body of literature that has  
658 reconstructed abrupt oscillations in RSL during the LIG from far- and intermediate-field settings. Our  
659 intention in this study is not to criticise these reconstructions, although we note that in several  
660 studies that assert there is more than one LIG highstand the deposits are poorly dated or undated.  
661 Rather we aim to highlight the lack of evidence for equivalent changes in the near-field in northwest

662 Europe, and the lack of any coherent evidence for significant collapse and regrowth events from  
663 Greenland or Antarctica during the peak LIG capable of driving abrupt, meter-scale RSL fluctuations.

664

665 A second reflection is on the importance of using uninterrupted stratigraphic sequences to  
666 test for sea-level oscillations/jumps. In the Netherlands, it is the elevation of the basin sills that  
667 connected them to the Rhine Valley and the North Sea which provides the primary elevation control  
668 during the initial LIG transgression, whilst during the highstand the vertical RSL constraint is obtained  
669 from intertidal and salt marsh deposits and overlying terrestrial peat. It is the stratigraphic  
670 successions preserved in these sequences that are the key to testing for large amplitude sea-level  
671 fluctuations/jumps. The value of stratigraphically constrained data applies equally in the far- and  
672 intermediate-field settings.

673

674 A third observation is that there are periods during the LIG when different RSL records have  
675 been used to invoke episodes of abrupt sea-level change when at other sites RSL is stable or  
676 changing little. Mindful of the fingerprint discussion above, rapid jumps in sea level at one site  
677 should be recorded at the same time at others (as supposed to gradual sea-level change resulting in  
678 a spatially and temporally variable highstand). Between-site variability therefore indicates that one  
679 or more of the records in conflict must reflect site- (or region-) specific processes. One example  
680 illustrates this point; Thompson et al. (2011) report that RSL on the Bahamas fell from c. +4 m at 123  
681 k yr ago to 0 m and then rose again to + 6 m by 119.2 k yr ago. But in this same 4 k yr interval, RSL  
682 on the Yucatán Peninsula was stable until a sudden 3 m jump tentatively dated at c. 121 k yr ago  
683 (Blanchon et al., 2009). Meanwhile in the Red Sea, using the Grant et al. (2012) chronology, there is  
684 a rather modest (c. 2 m) oscillation in this interval (Figure 1). It is difficult, if not impossible, to  
685 reconcile these records with each other.

686

687 The above example underscores our fourth reflection, namely the need for precise age  
688 constraints from LIG sea-level reconstructions so as to enable robust chronological comparisons  
689 between sites and with the ice-sheet records and models. In practice, dating LIG sea-level records is  
690 a major challenge. In the Netherlands, the duration of the RSL record is determined with respect to  
691 'floating' varve chronologies and associated pollen zones (e.g. Caspers et al., 2002; Funder et al.,  
692 2002; Sirocko et al., 2005). Indeed, in some instances chronologies for the near-field RSL records are  
693 developed by aligning records with a single interglacial sea-level highstand identified in the far-field,  
694 regardless of potential differential GIA between these regions (e.g. Funder et al., 2002). In the  
695 intermediate- and far-fields, U-Th dating provide the potential for precise and accurate dating

696 control, but age determinations require careful screening to ensure that only the most trustworthy  
697 dates are used in any analysis. In a recent LIG database exercise, Dutton and Lambeck (2012) discard  
698 about two thirds of available U-Th dates by following a robust screening procedure (see also  
699 Medina-Elizalde (2013)). The importance of different age models is also evident in the way that the  
700 Red Sea sea-level record has shifted c. 6 k yr (from the early to the middle part of the LIG, Figure 1B)  
701 since its original publication (Rohling et al., 2008), whilst we note also that single-site chronologies  
702 can also be uncertain; the timing of the Yucatán sea-level jump at c. 121 k yr ago is only ‘tentatively’  
703 dated (Blanchon et al., 2008).

704

705 In Europe, debate regarding the absolute age of the Eemian deposits means that correlation  
706 with other intermediate- and far-field records is unclear. The oldest suggested age for the Eemian  
707 highstand is c. 129 k yr ago (Funder et al., 2002) and, if correct, would mean that the initial sea-level  
708 jump recorded in the Seychelles by Dutton et al. (2015) could be recorded in the main Dutch  
709 transgression. The more conventional age-model, based on palynological correlation to the Iberian  
710 Peninsula and its offshore marine record (Sanchez-Goni et al., 1999; Shackleton et al., 2003) has the  
711 highstand (within E4 and E5) spanning from 124 to 119 k yr ago. If correct, it would mean that  
712 Seychelles sea-level jump happened when much of the Netherlands was dry land and the coastline  
713 well offshore in the southern North Sea. Conversely, the late LIG sea-level jumps recorded on the  
714 Yucatán Peninsula/Bahamas and Western Australia (c. 121 to 118 k yr ago) would have happened  
715 (but are not recorded) during the final phase the Dutch highstand. The youngest proposed  
716 chronology has the highstand spanning from 119 to 114 k yr ago (Sier et al., 2015). This model would  
717 mean that the proposed Yucatán and/or western Australia oscillations happened during the main  
718 transgression phase (E3). Close interval sampling of multiple cores taken across these key time  
719 intervals from the flooding surface in the Amersfoort Basin would potentially help test the latter two  
720 hypotheses.

721

722 A final observation is that well-constrained ice-margin histories mean that it is now possible  
723 to infer upper bounds to individual Holocene sea-level jumps, but this is not yet the case in the LIG.  
724 Indeed, the challenge in reconstructing the 8.2 k yr sea-level jump (Hijma and Cohen, 2010), which  
725 required excellent dating control and detailed sedimentological investigations, highlights the  
726 potential of this approach to LIG deposits but also the scale of the challenge ahead. We therefore  
727 encourage caution against seeking to link RSL data from single sites or regions to ice-sheet behaviour  
728 until such time that abrupt sea-level changes are replicated in near-, intermediate- and far-field

729 sites, or robust ice-sheet histories demonstrating unequivocal evidence for collapse and regrowth  
730 events are established.

731

## 732 **10. Conclusions and future research priorities**

733 Abrupt jumps or oscillations in RSL caused by the partial or full collapse and regrowth of an  
734 ice sheet should be recorded globally as distinct, spatially-variable, RSL signatures. For decades,  
735 investigations in far- and intermediate-field sites have identified evidence that sea level during the  
736 LIG was several meters higher than present. Additionally, there is also a large body of literature that  
737 argues for significant, meter-scale fluctuations in RSL during the LIG highstand, which have been  
738 attributed to largely unidentified ice-sheet collapse and regrowth events.

739

740 Near-field stratigraphic sequences from the Amsterdam and Amersfoort Basins (the  
741 Netherlands) provide data that imply stable or only gradually changing ice volumes during the LIG  
742 highstand of this part of northwest Europe. Reviews of other LIG marine records from northwest  
743 Europe yield no compelling evidence for widespread abrupt changes in RSL during the LIG highstand,  
744 changes that, based on our understanding of how coastal systems in this region typically respond to  
745 Holocene forcings, should be clearly identifiable.

746

747 The fact that the near-field LIG records lack clear evidence for abrupt RSL change requires  
748 explanation if we are to be confident in making predictions for how the polar ice sheets may behave  
749 in a future, warmer-than-present world. A lack of direct field evidence from Greenland and  
750 Antarctica for collapse/regrowth events within the LIG RSL highstand is a significant challenge to  
751 ongoing and future glaciological and palaeoceanographic research. Although sea-level modelling  
752 experiments predict that the Netherlands is relatively insensitive to Greenland ice-mass change, the  
753 latter had already lost significant mass at the time of the LIG sea-level highstand and cannot have  
754 been the sole source of large fluctuations/jumps in sea level during the LIG. Present approaches to  
755 reconstructing LIG ice mass sources are hamstrung by poor chronologies for the deglacial history of  
756 Antarctica, Greenland, as well as other northern hemisphere ice sheets, parts of which may have  
757 persisted late into the LIG.

758

759 The differences noted in this study provide directions for future research in near-,  
760 intermediate- and far-field RSL studies, in the LIG and other older interglacials, that may be used as  
761 analogues for future sea-level change. For sites in the near-field, a priority must be replication of  
762 further, more detailed investigation of fine-grained tidal flat and salt marsh deposits formed during

763 the LIG highstand, including with closer sampling intervals across key stratigraphic changes that span  
764 periods of hypothesised ice sheet collapse. One such period – spanning the hypothesised sea-level  
765 jump observed c. 121 k yr ago – can be tested for by detailed study of the period of rapid  
766 transgression that marks the early part of the Eemian in the Dutch RSL record.

767

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1150 **List of Figures**

1151 Figure 1: Illustrative relative sea-level records from the last interglacial: A: The Seychelles  
1152 (Dutton et al., 2015). B: The Red Sea, core KL11 using the age models of Rohling et al. (2008)  
1153 and Grant et al. (2012); C: Yucatán Peninsula (Mexico) (Blanchon et al., 2009); D: The  
1154 Bahamas (Thompson and Goldstein, 2005); E: Western Australia (O’Leary et al., 2013), the  
1155 dashed green line is an inferred sea-level curve based on a minimum coral palaeodepth; and  
1156 F: The Netherlands. The graph shows the original data of Zagwijn (1983) and an enlarged  
1157 dataset (Streif, 1990) which has been corrected for GIA, compaction and tectonics (Lambeck,  
1158 1996).

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Figure 2: Schematic relative sea-level histories recorded at near- and far-field sites during the last interglacial (modified from Cohen et al., 2012). Note that the highstand occurs earlier in the far-field compared to the near-field. A fluctuating sea-level history is depicted for illustrative purposes to show how such a record might be recorded in each setting.

Figure 3: Location map of Northwest Europe showing last interglacial site names referred to in the text.

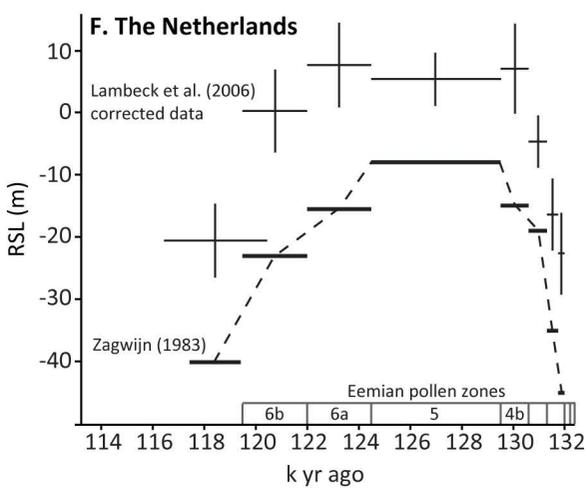
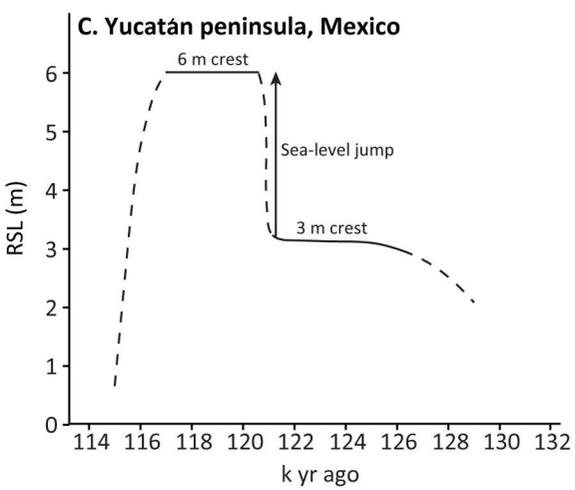
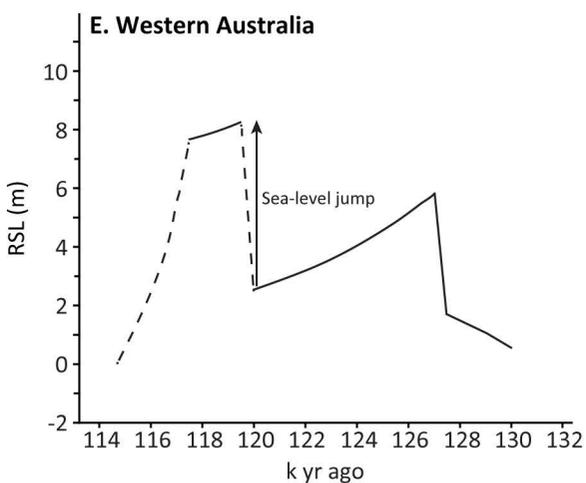
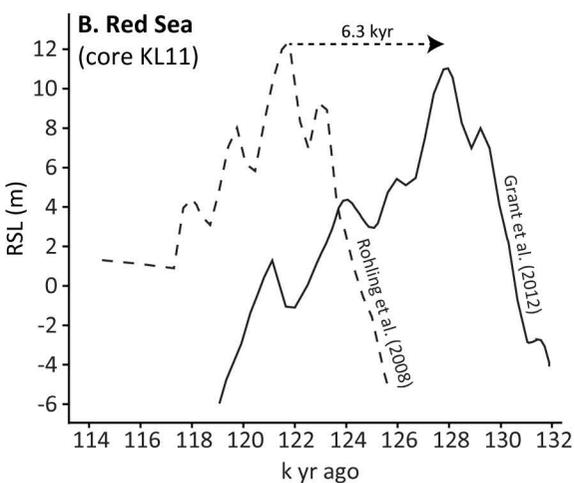
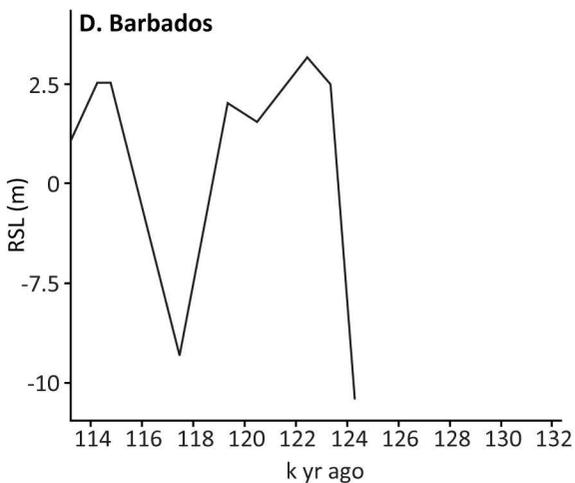
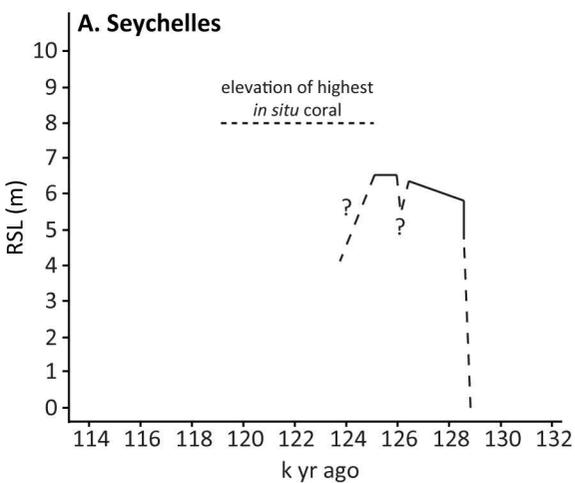
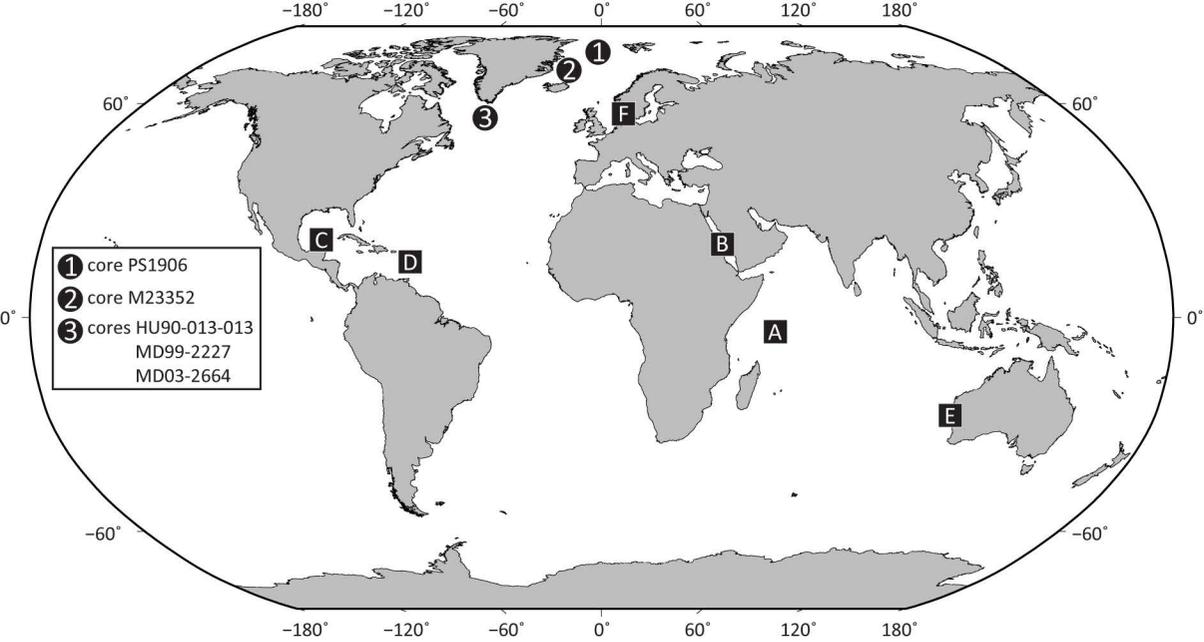
Figure 4: Reconstruction depicting (A) the Netherlands, (B) the Last Interglacial coastal configuration in the Netherlands during the sea-level highstand (modified from Peeters et al. (2015)), (B) the glacial basins in the Netherlands (modified from Cleveringa et al. (2000)).

Figure 5: Simplified lithostratigraphy in the Amersfoort Basin (modified from Zagwijn (1983)). Position of pollen zones are approximate, based upon Zagwijn (1983).

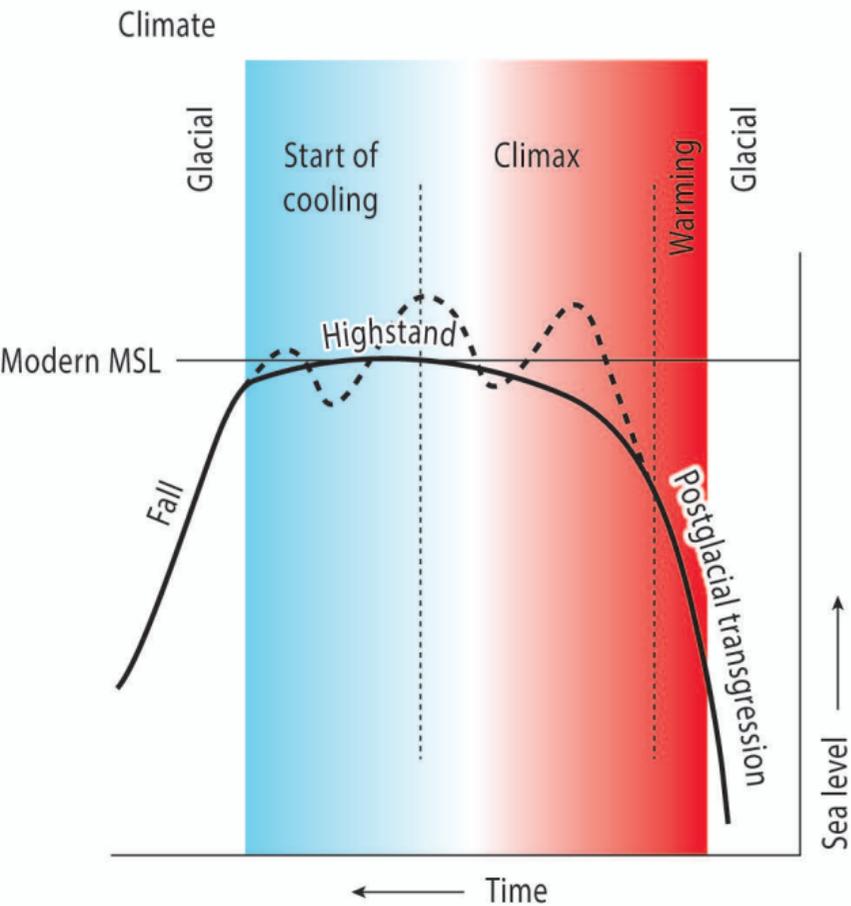
Figure 6: Summary lithology and selected biostratigraphic data from the Amersfoort-1 and Amsterdam Terminal boreholes (modified from Van Leeuwen et al. (2000) and Cleveringa et al. (2000))

Figure 7: The Zagwijn (1983) Netherlands sea level curve plotted against the established pollen zone chronology (in black) and Sier et al. (2015) chronology (in grey). The approximate timing of the Seychelles and Western Australia sea level jumps (as shown in Figure 1) are marked.

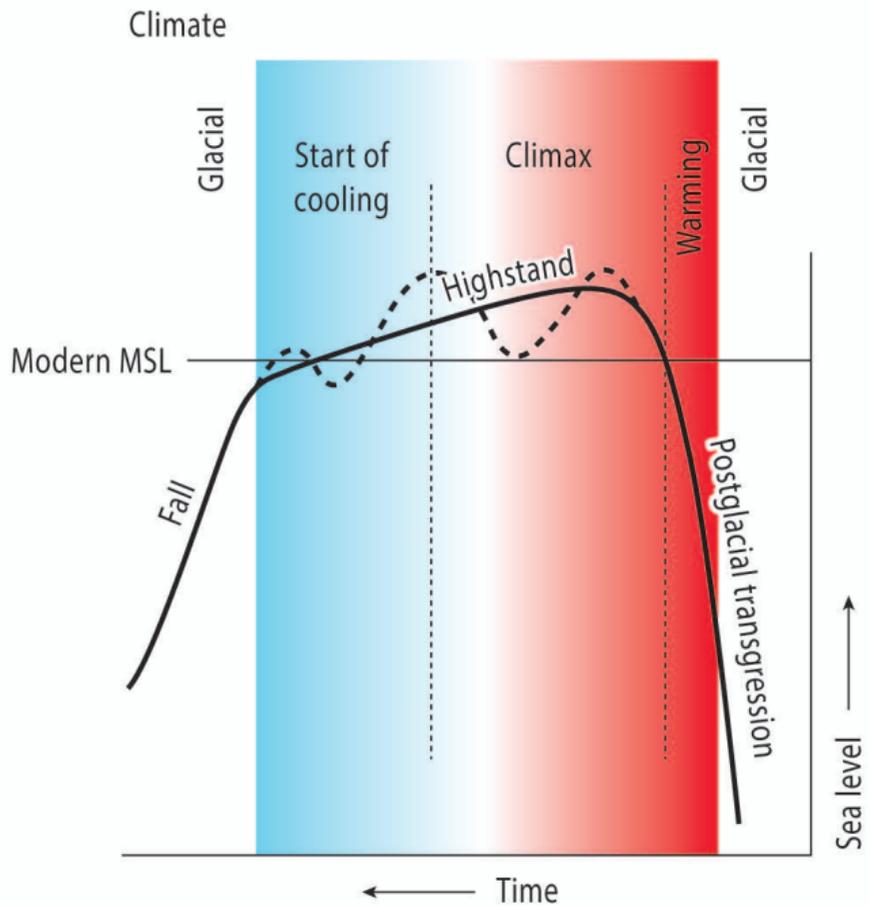
Figure 8: Sea-level predictions associated with hypothetical collapse/regrowth cycles of the Greenland and Antarctic ice sheets. Experiments G1 and A1 assume that Greenland and Antarctica, respectively, are subjected to growth-collapse cycles with amplitudes that vary between c.  $\pm$  2-3 and 5 m. A) Predictions assuming a Greenland-only origin for sea-level oscillations/jumps; B) Predictions assuming an Antarctic-only origin. Two periods of mass loss are shown for Greenland (122-121 k yr, 120-119 k yr) and three for Antarctica (123-122 k yr, 120-119 k yr, 118-117 k yr). Details of experimental set-up associated with each model are provided in Supplementary Information Table S1.

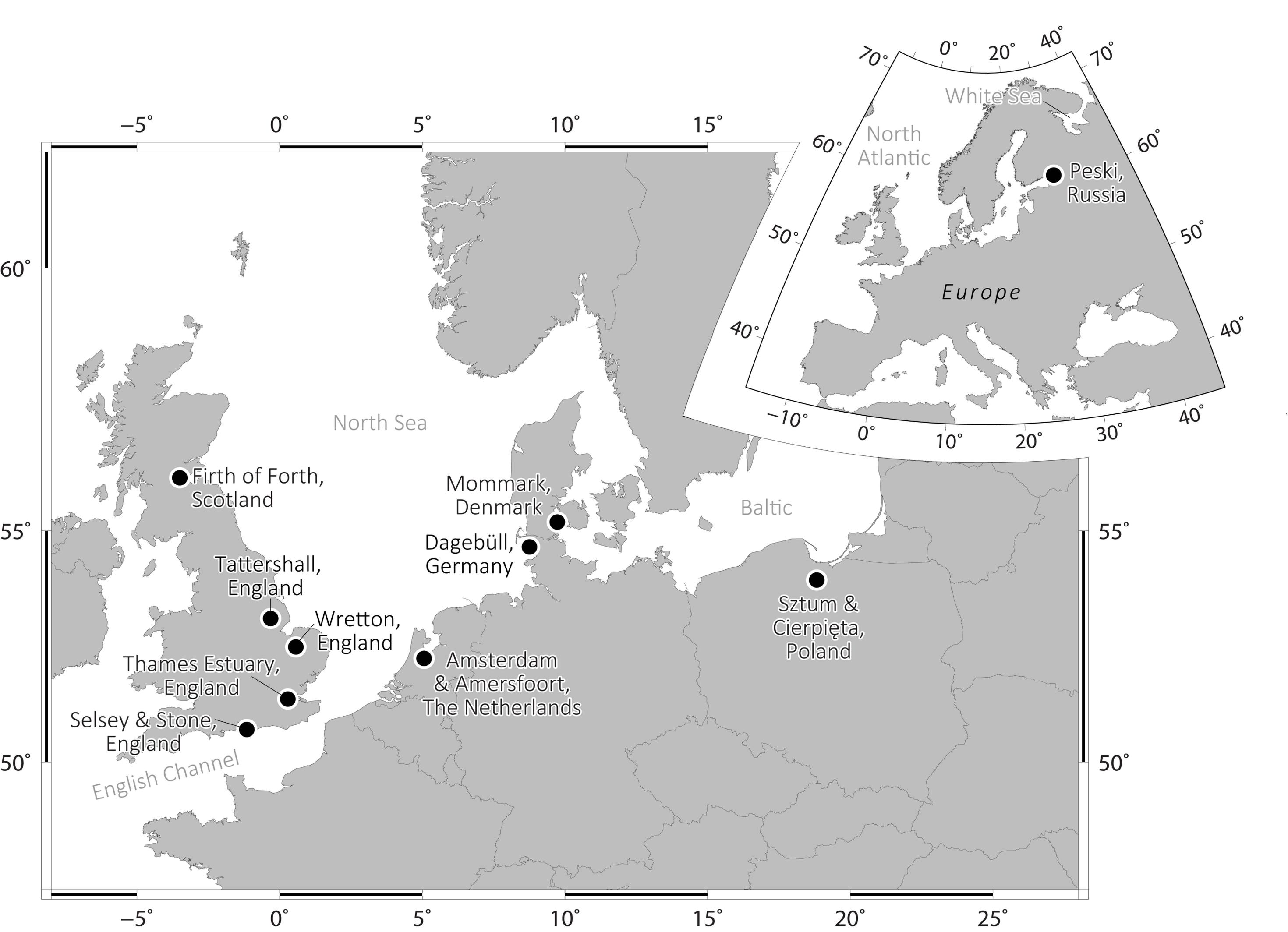


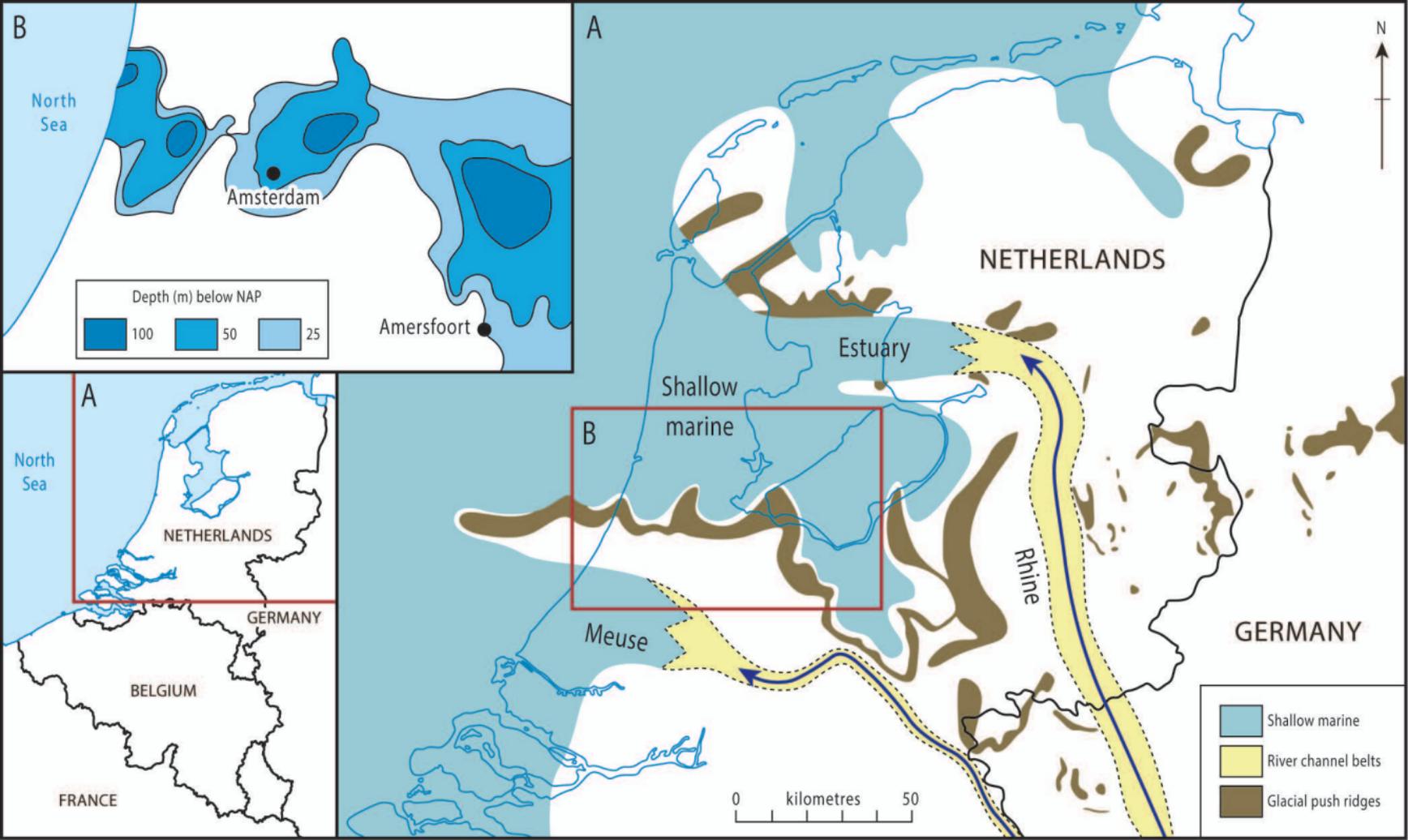
'Near field' sea-level rise e.g. North Sea region



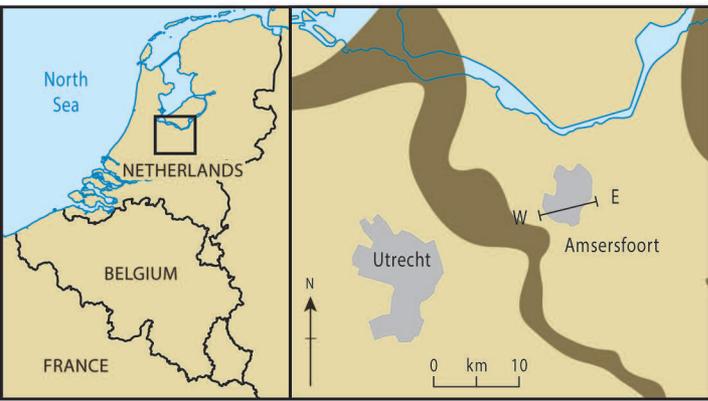
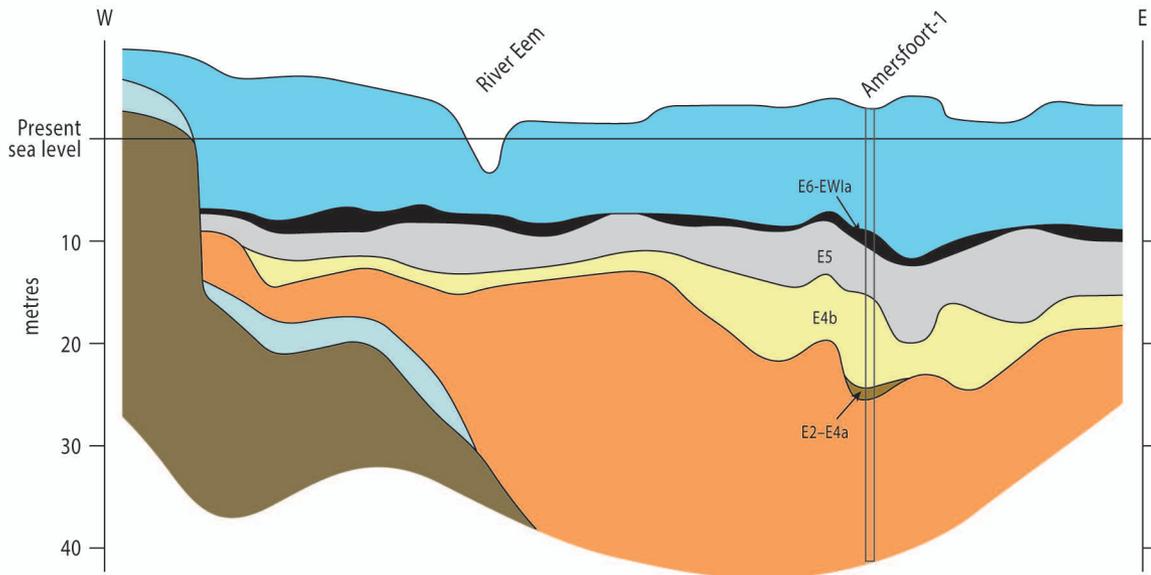
'Far field' sea-level rise e.g. tropical Asia / Pacific





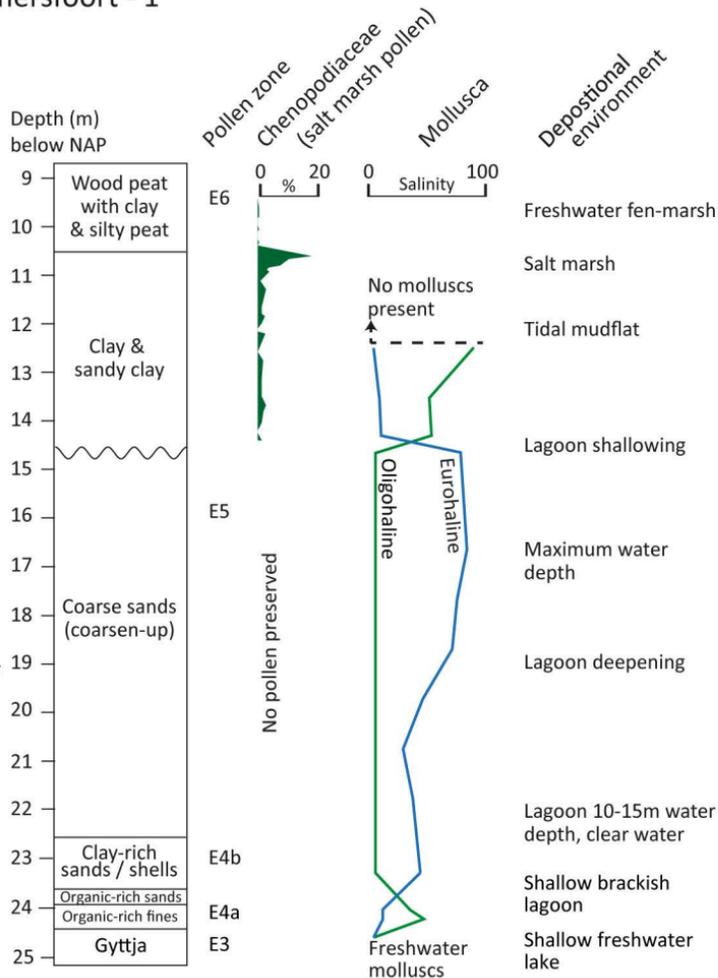


# Amersfoort

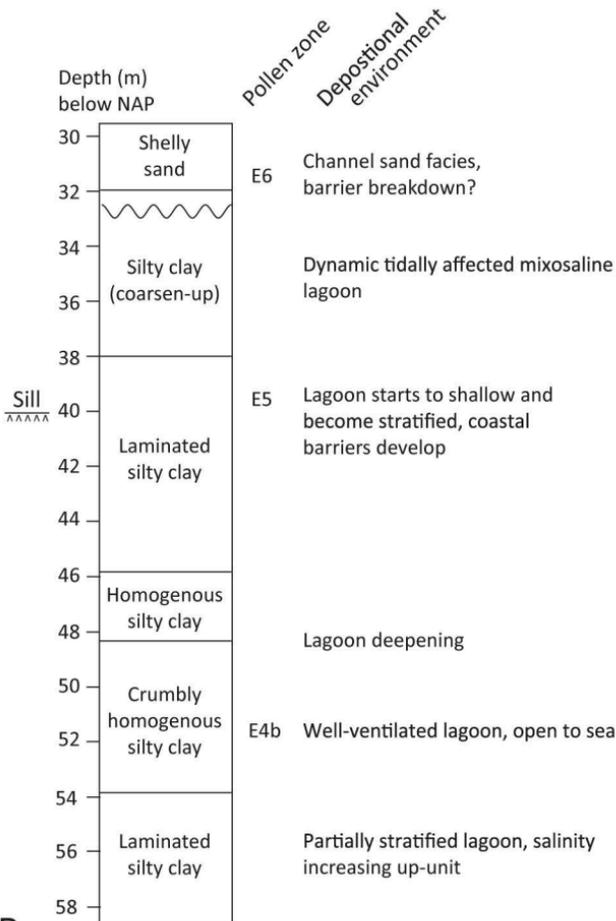


- Periglacial local deposits (Twente Fm)
  - Upper peat-member (freshwater)
  - Clay-member (marine and brackish)
  - Sand-member (marine)
  - Lower peat-member (freshwater and marine)
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# Amersfoort - 1

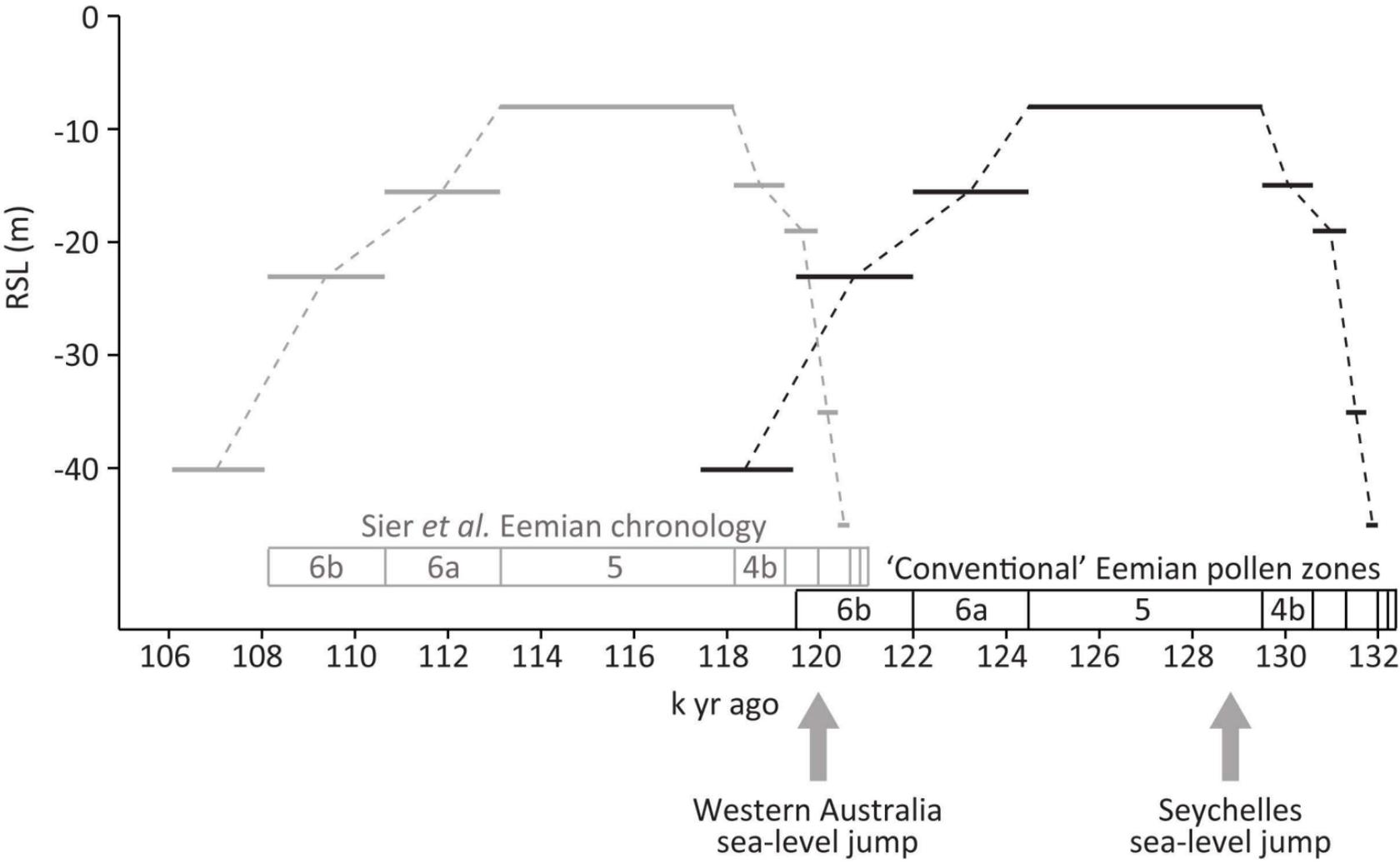


# Amsterdam Terminal

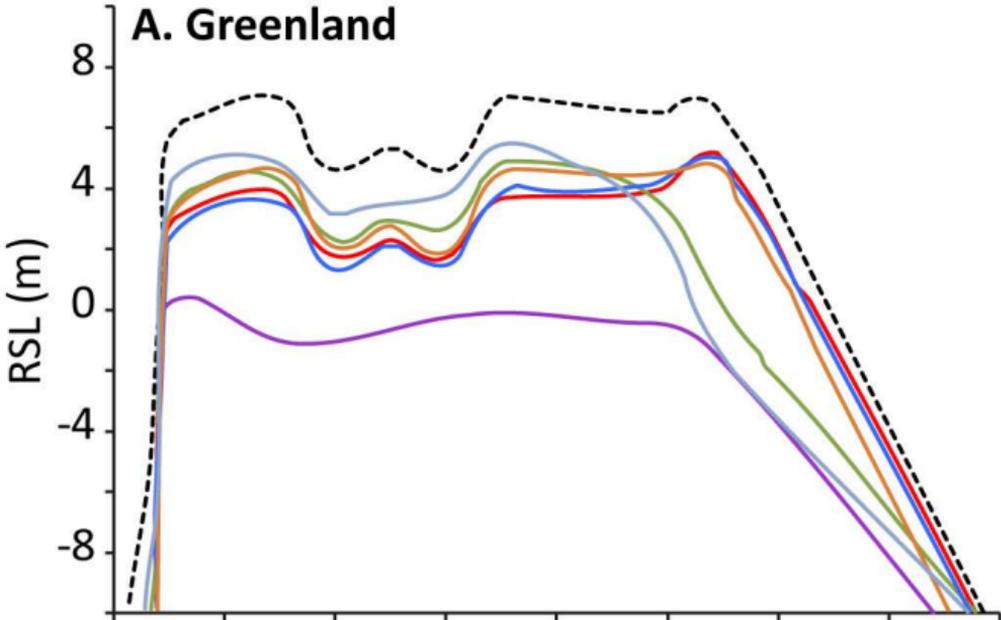


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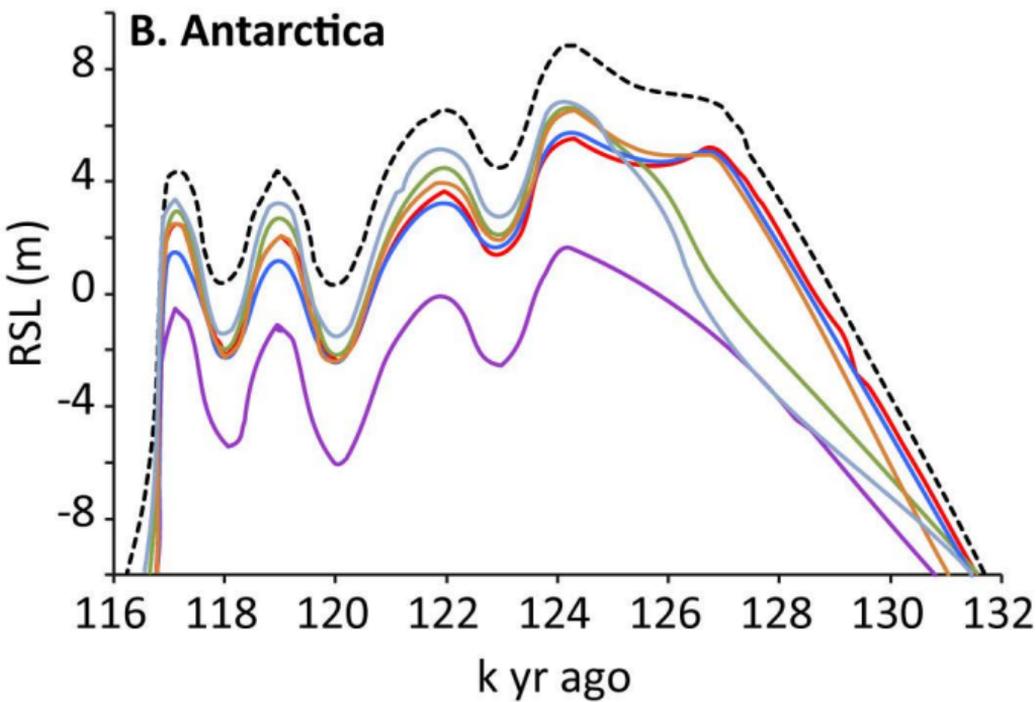
B



### A. Greenland



### B. Antarctica



--- Eustatic sea level

— Red Sea

— Netherlands

— Yucatan

— Western Australia

— Seychelles

— Bahamas

## Quaternary Science Reviews 2015

### Near-field sea-level variability in northwest Europe and ice sheet stability during the Last Interglacial

Long, A.J., Barlow, N.L.M., Busschers, F.S., Cohen, K.M., Gehrels, W. R. & Wake, L.M.

#### **Supplementary information Table S1**

Model parameters governing deglaciation scenarios A1 and G1. Columns 2,4 and 6 titled 'Extent' refer to the time period (kyr ago) which is used to represent the extent of global ice sheets applied at each time interval. Columns 3,5 and 7 denote the scaling factor applied uniformly to the regional ice distributions at each time step. For example, at 127.00 kyr BP in model A1, Antarctica, Greenland and the Rest of the World display ice cover identical to that at 122.00 kyr BP in ICE5G - HUY2 deglaciation history (Peltier, 2004; Simpson et al. 2009). These ice extents are scaled by a factor of 1.

Loading Step (kyr ago BP)	Antarctica		Greenland		Rest of World	
	Extent	Scaling Factor	Extent	Scaling Factor	Extent	Scaling Factor
132.00	116.00	0.980	116.00	0.980	116.00	0.980
127.00	122.00	1.000	122.00	1.000	122.00	1.000
126.00	122.00	1.000	122.00	1.000	122.00	1.000
125.00	122.00	0.999	122.00	1.000	122.00	1.000
124.00	122.00	0.998	122.00	1.000	122.00	1.000
123.00	122.00	1.020	122.00	1.000	122.00	1.000
122.00	122.00	1.000	122.00	1.000	122.00	1.000
121.00	122.00	1.020	122.00	1.000	122.00	1.000
120.00	122.00	1.090	122.00	1.000	122.00	1.000
119.00	122.00	1.030	122.00	1.000	122.00	1.000
118.00	122.00	1.100	122.00	1.000	122.00	1.000
117.00	116.00	1.000	122.00	1.000	122.00	1.000
116.00	116.00	1.000	115.00	1.000	116.00	1.000
<b>Model G1</b>						
Loading Step (kyr ago BP)	Antarctica		Greenland		Rest of World	
	Extent	Scaling Factor	Extent	Scaling Factor	Extent	Scaling Factor

132.00	116.00	0.980	116.00	0.980	116.00	0.980
127.00	122.00	1.000	122.00	1.000	122.00	1.000
126.00	122.00	1.000	122.00	0.900	122.00	1.000
125.00	122.00	1.000	122.00	0.700	122.00	1.000
124.00	122.00	1.000	122.00	0.500	122.00	1.000
123.00	122.00	1.000	122.00	0.200	122.00	1.000
122.00	122.00	1.000	116.00	1.500	122.00	1.000
121.00	122.00	1.000	116.00	1.050	122.00	1.000
120.00	122.00	1.000	116.00	1.500	122.00	1.000
119.00	122.00	1.000	122.00	0.200	122.00	1.000
118.00	122.00	1.000	122.00	1.000	122.00	1.000
117.00	122.00	1.000	116.00	1.000	122.00	1.000
116.00	116.00	1.000	116.00	1.000	116.00	1.000

### Package 1

The figures contained in PDF packages **FigureS1.pdf** and **FigureS2.pdf** show plots of global relative sea level (RSL) and global sea level trends for the time period between 132 kyr BP and 116 kyr BP for deglaciation histories A1 and G1 respectively. Snapshots of global relative sea level are taken at the times listed in Table S1, whereas sea level trends correspond to the rate of change of relative sea level from the preceding time slice. For this reason, no corresponding sea-level trends accompany the plot of global relative sea level for 132 kyr BP.

**FigureS1.pdf: Global relative sea level and sea level trends for deglaciation history A1.**

**FigureS2.pdf: Global relative sea level and sea level trends for deglaciation history G1.**

### Package 2

PDF Package 2 contains plots of global ice extent and changes in ice load between the time steps detailed in Table S1. The ice loads applied at each time step in Table S1 correspond to the change in ice thickness with respect to the previous time step. For example, the ice load applied at 122 kyr BP reflects the change in global ice thickness from 123 kyr to 122 kyr BP. For this reason, no corresponding change in ice load accompanies the plot of ice extent for 132 kyr BP.

**FigureS3.pdf: Global ice extent and loading changes for deglaciation history A1.**

**FigureS4.pdf: Global ice extent and loading changes for deglaciation history G1.**

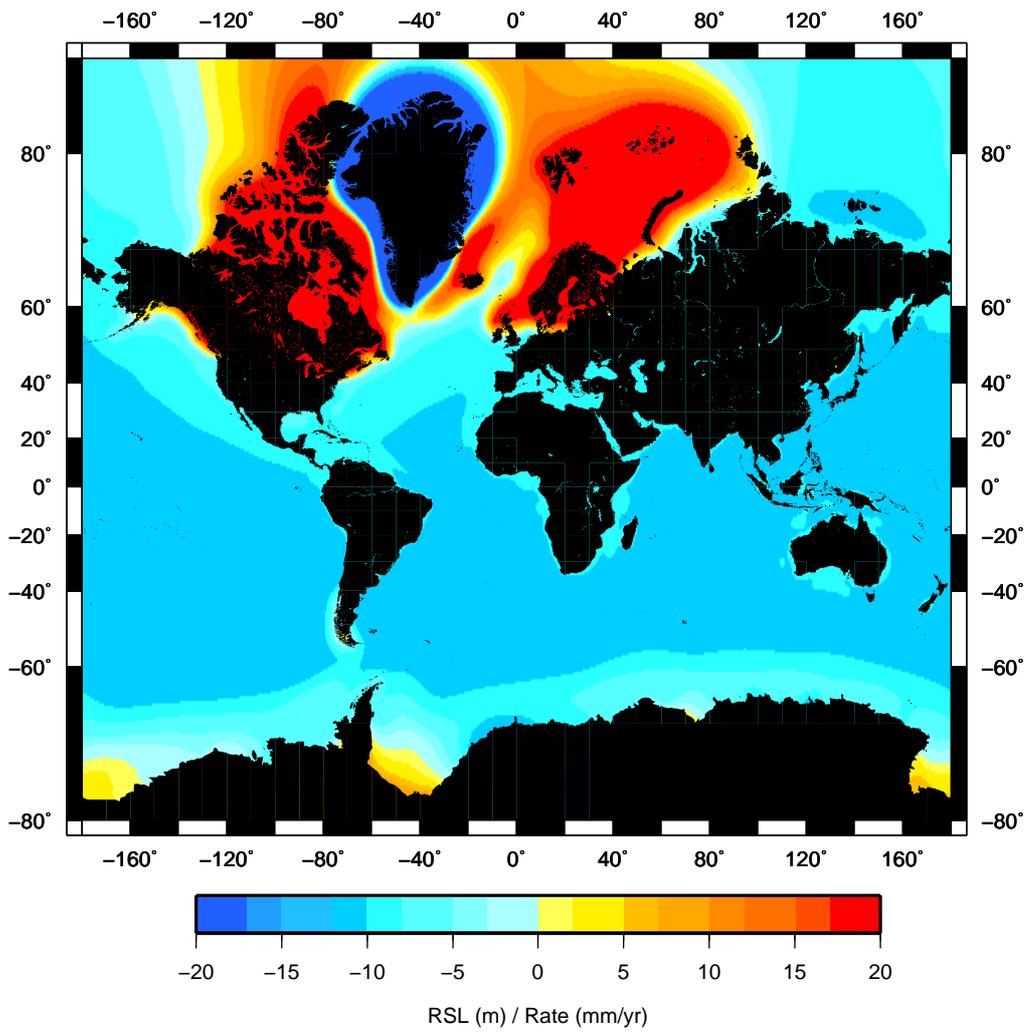
### Package 3

Package 3 contains two animations of global ice extent and corresponding relative sea level change for deglaciation histories A1 and G1.

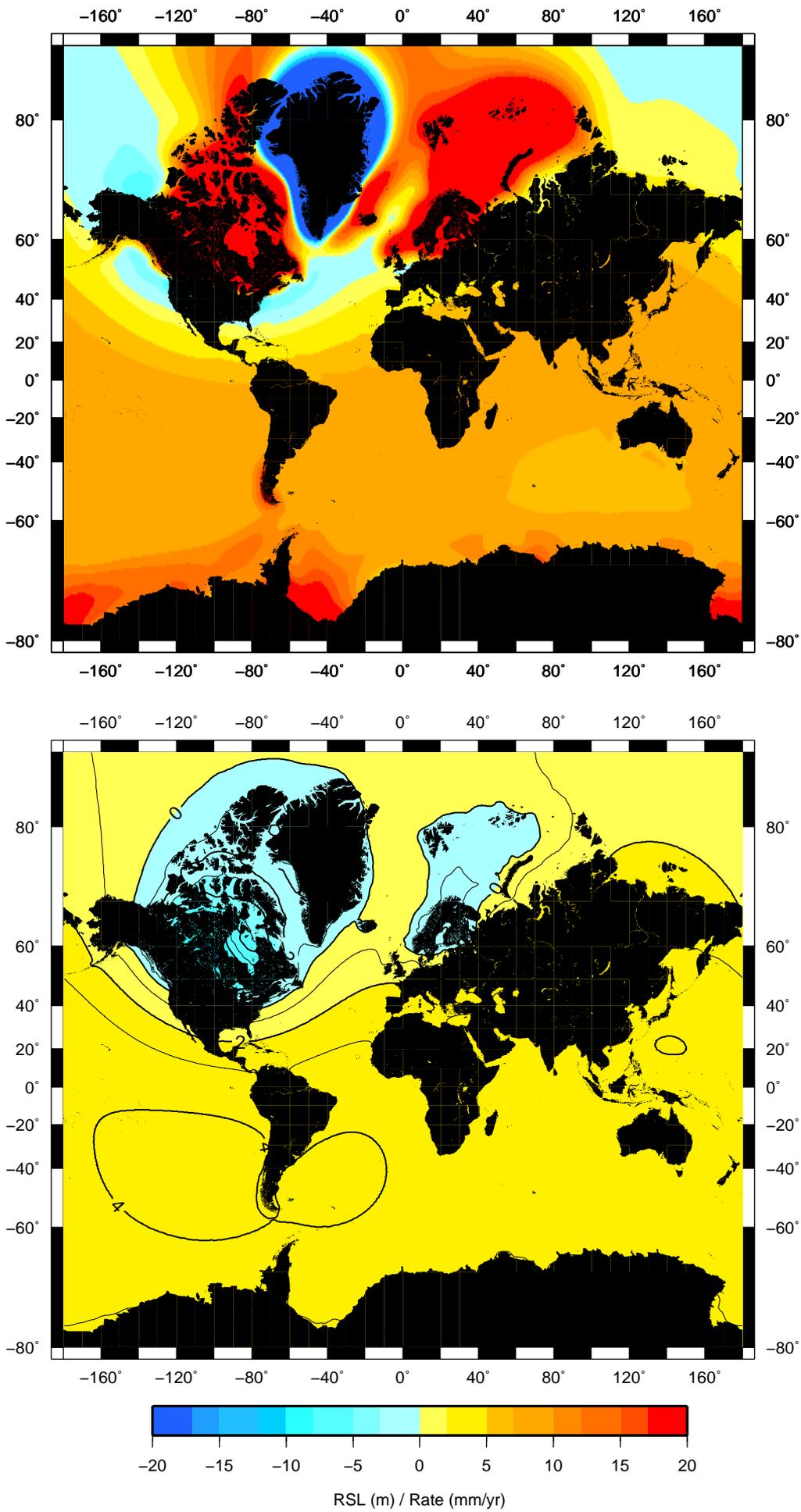
**AnimationS1.avi: Thirteen-frame Audio Video Interleaved (.avi) animation of global ice extent and relative sea level for the period 132 – 116 kyr BP for experiment A1.**

**AnimationS2.avi: Thirteen-frame Audio Video Interleaved (.avi) animation of global ice extent and relative sea level for the period 132 – 116 kyr BP for experiment G1.**

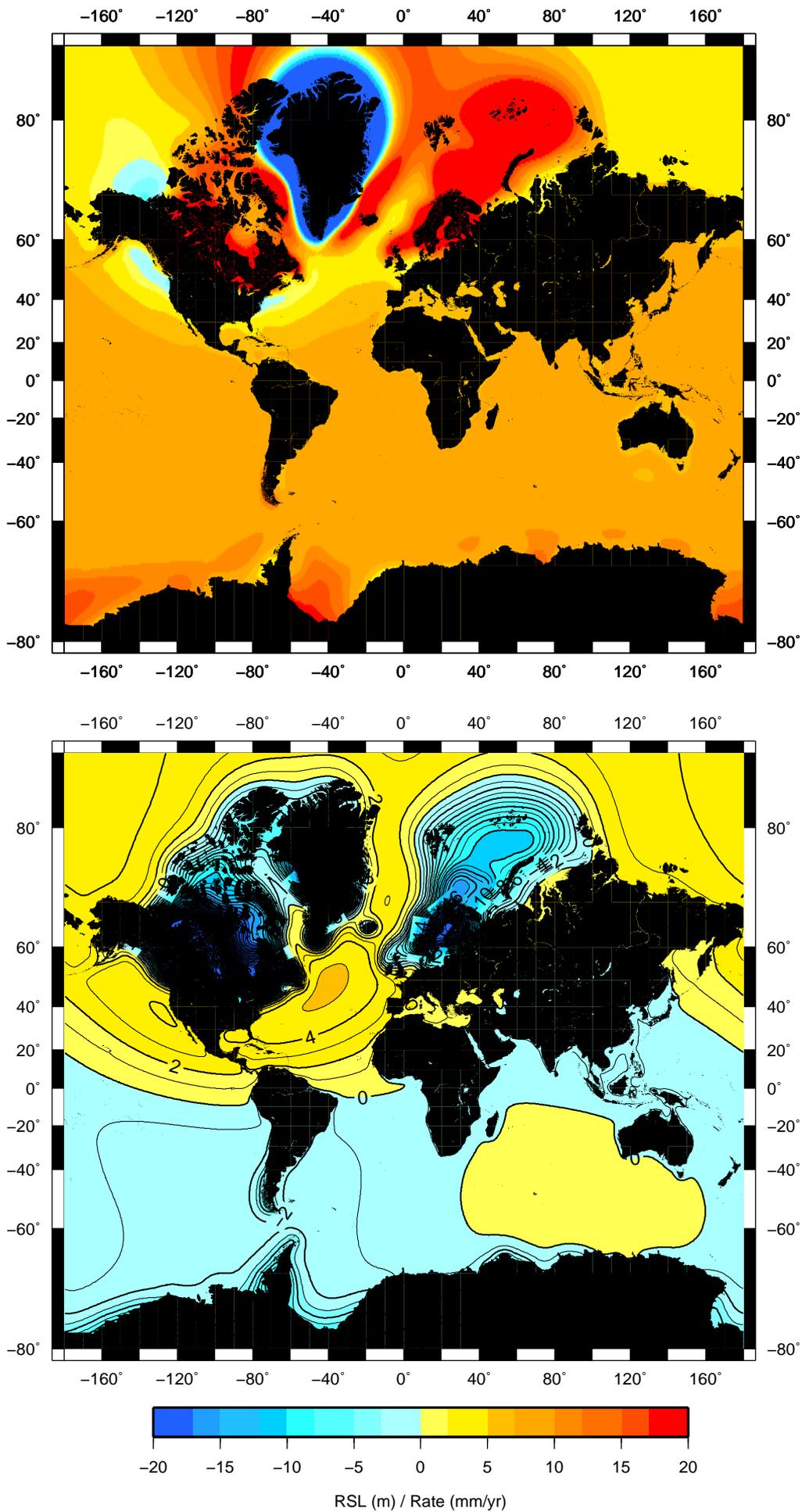
RSL at time 132 kyr BP for Antarctica experiment



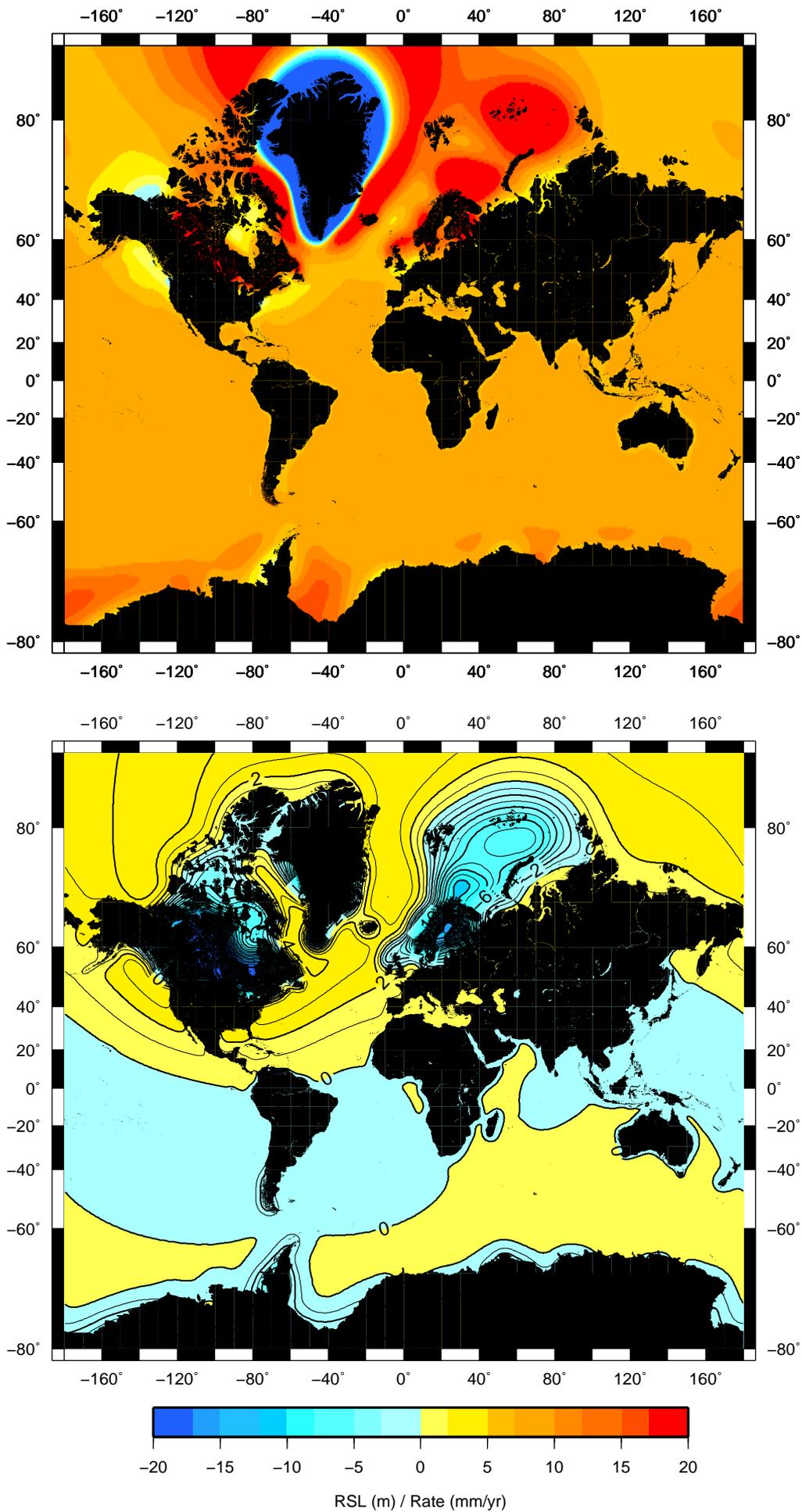
RSL (upper) and sea-level trend (lower) at time 127 kyr BP for Antarctica experiment



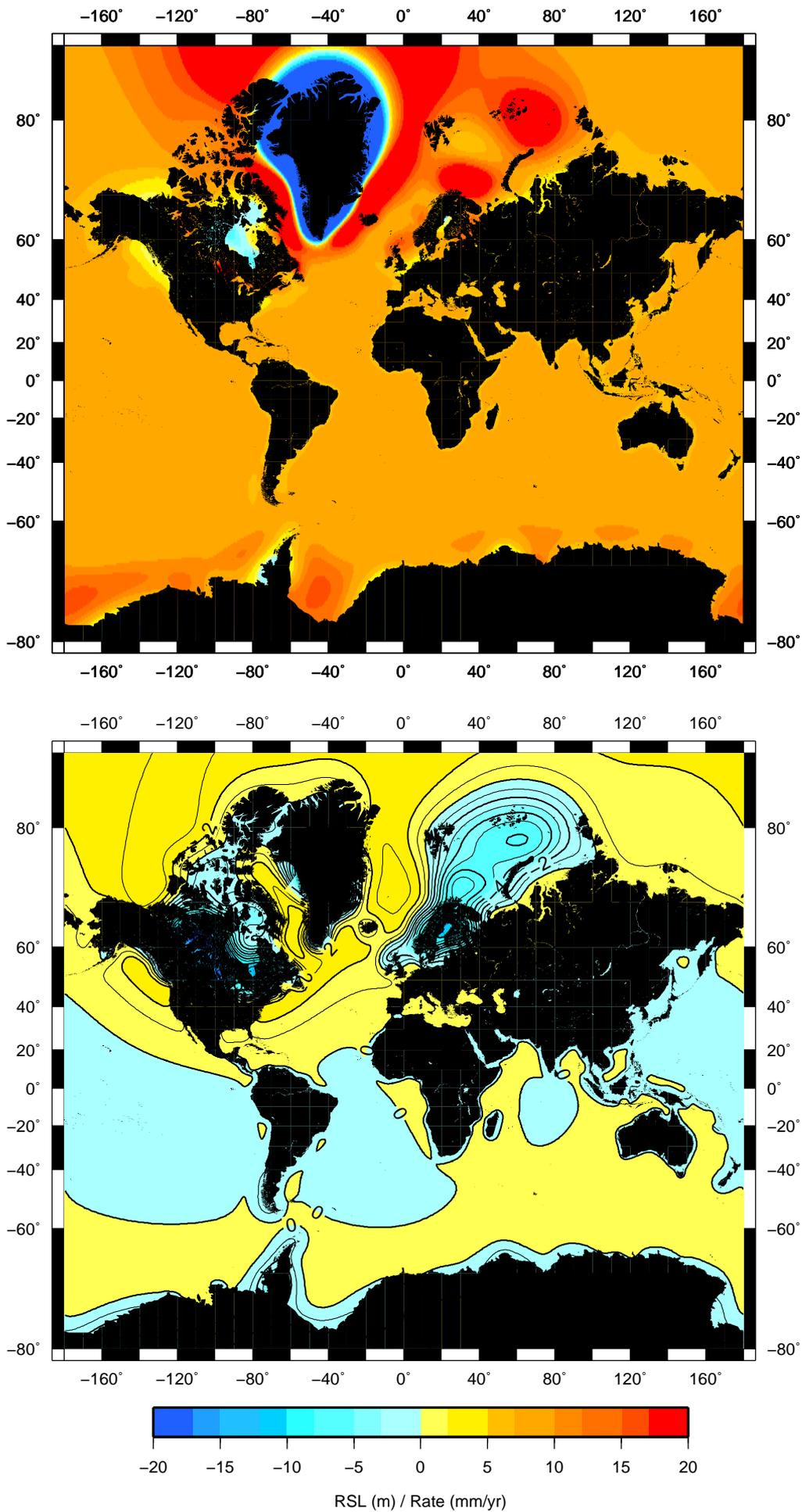
RSL (upper) and sea-level trend (lower) at time 126 kyr BP for Antarctica experiment



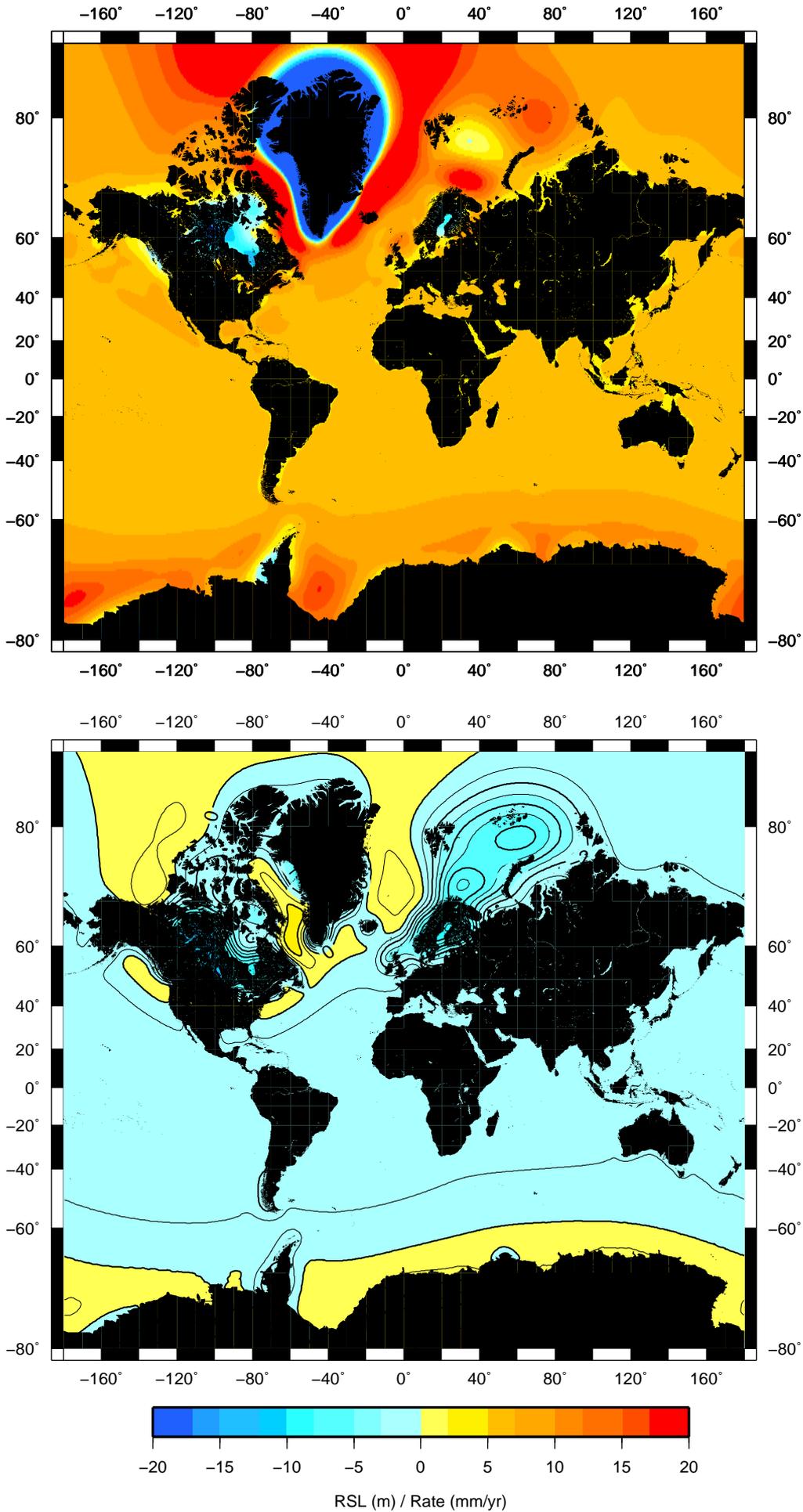
RSL (upper) and sea-level trend (lower) at time 125 kyr BP for Antarctica experiment



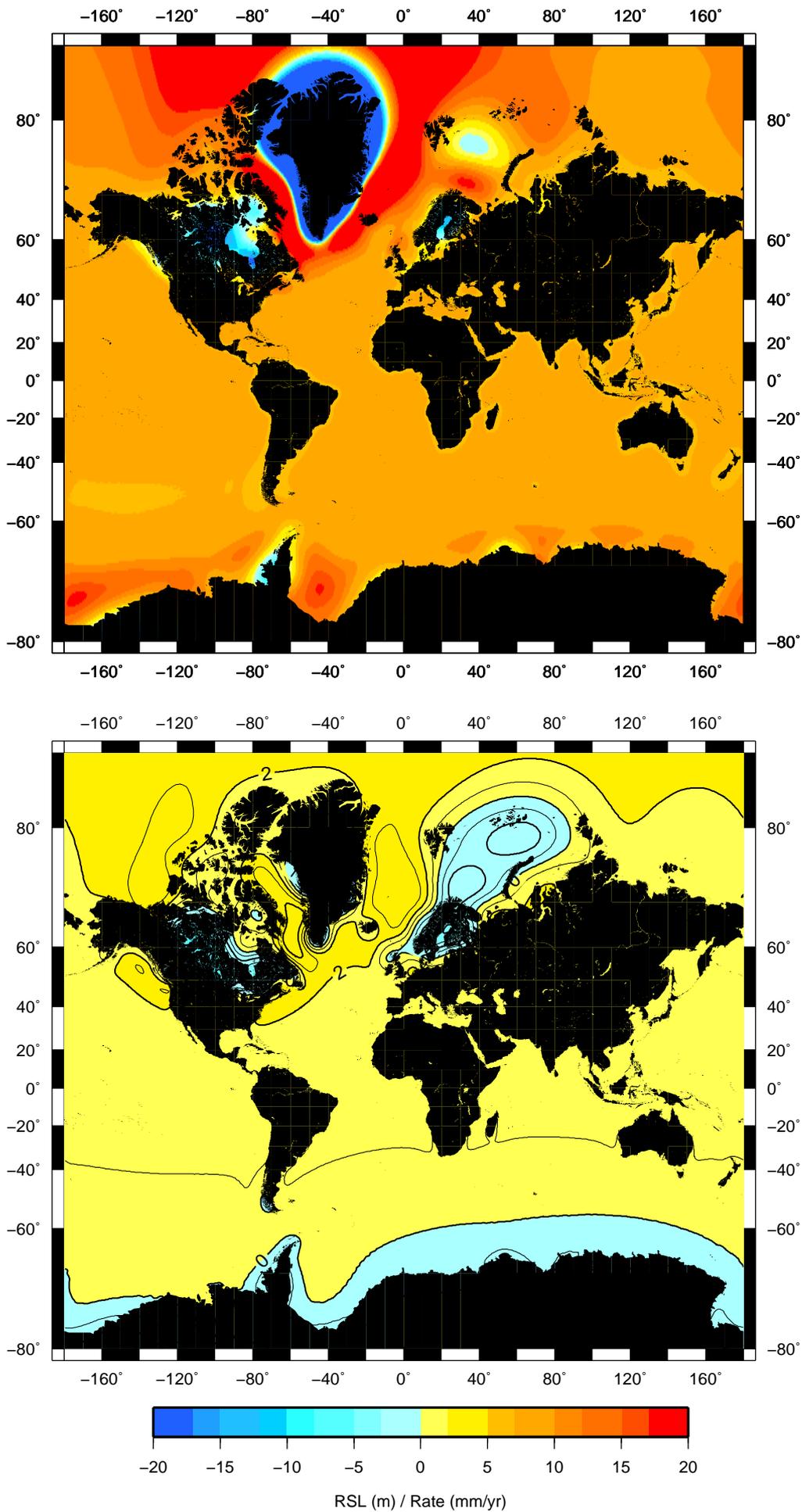
RSL (upper) and sea-level trend (lower) at time 124 kyr BP for Antarctica experiment



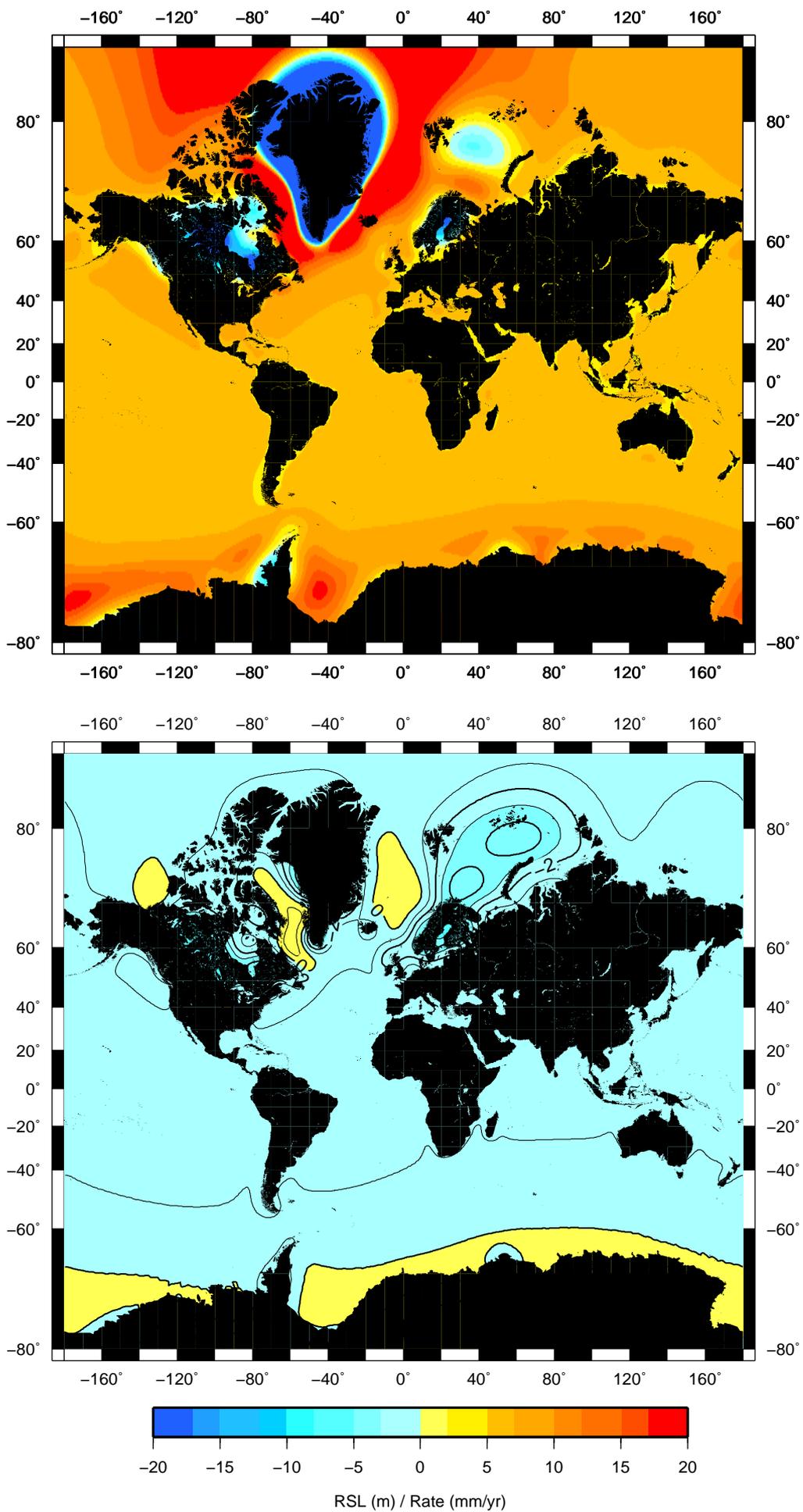
RSL (upper) and sea-level trend (lower) at time 123 kyr BP for Antarctica experiment



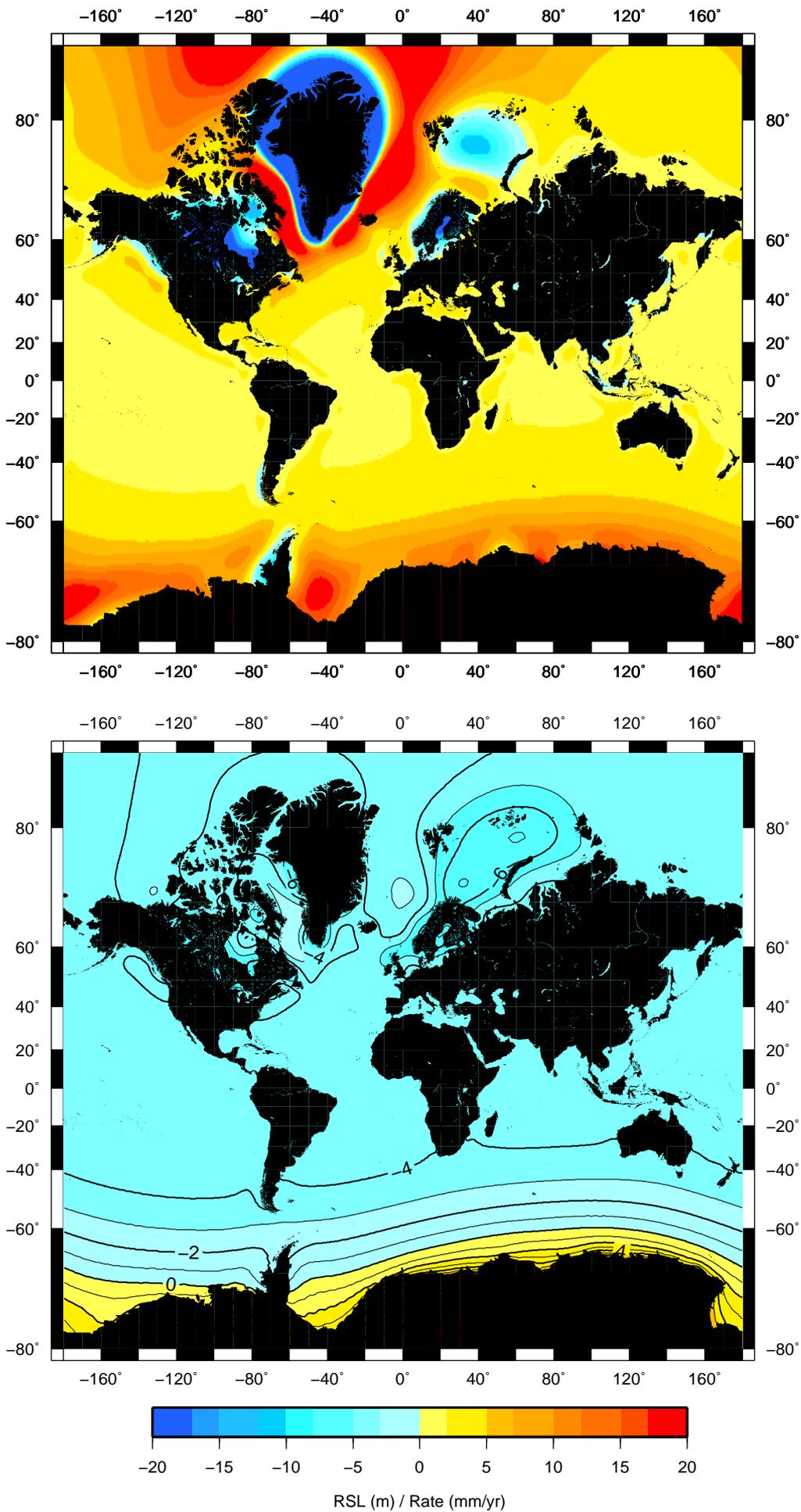
RSL (upper) and sea-level trend (lower) at time 122 kyr BP for Antarctica experiment



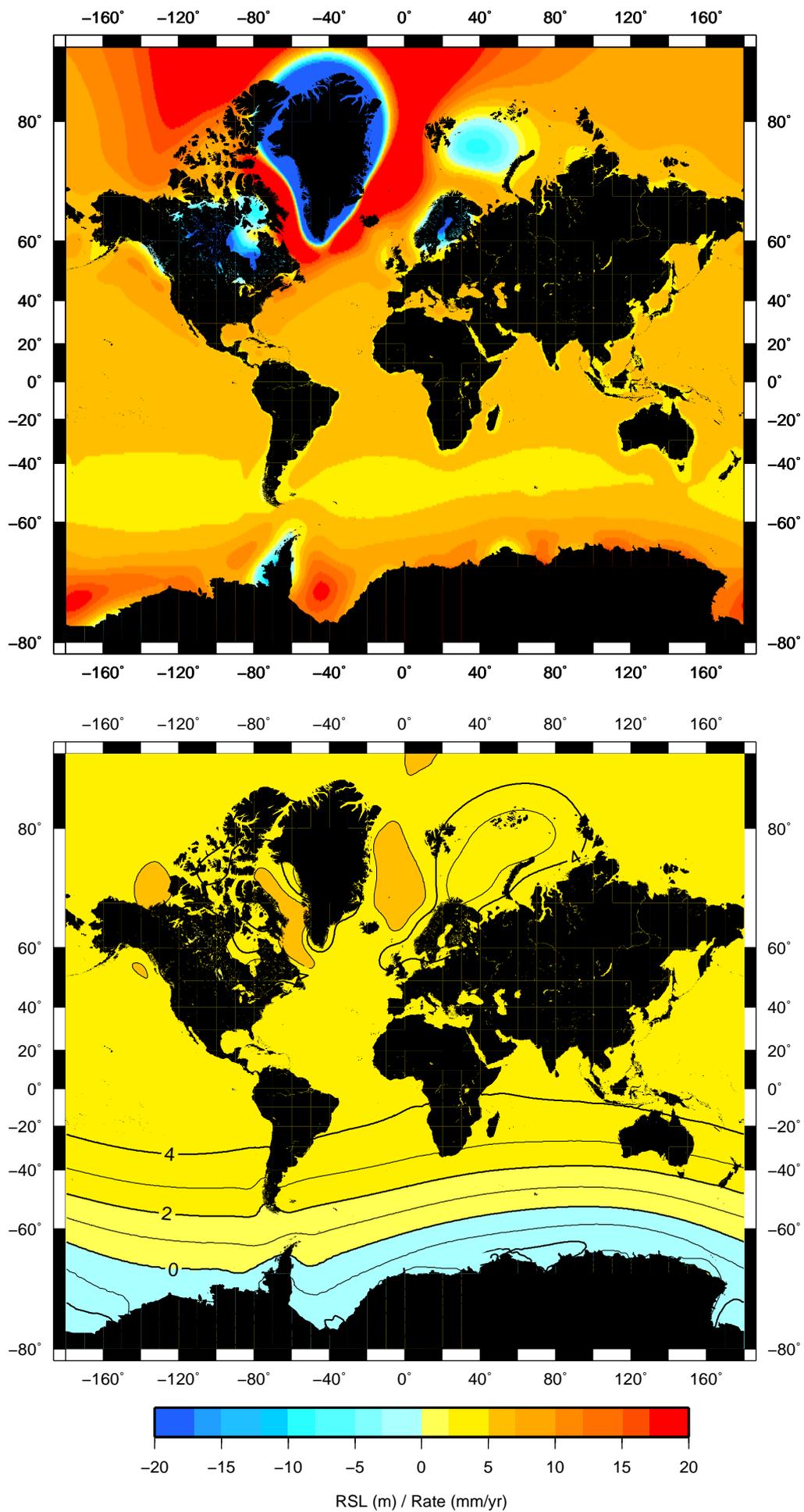
RSL (upper) and sea-level trend (lower) at time 121 kyr BP for Antarctica experiment



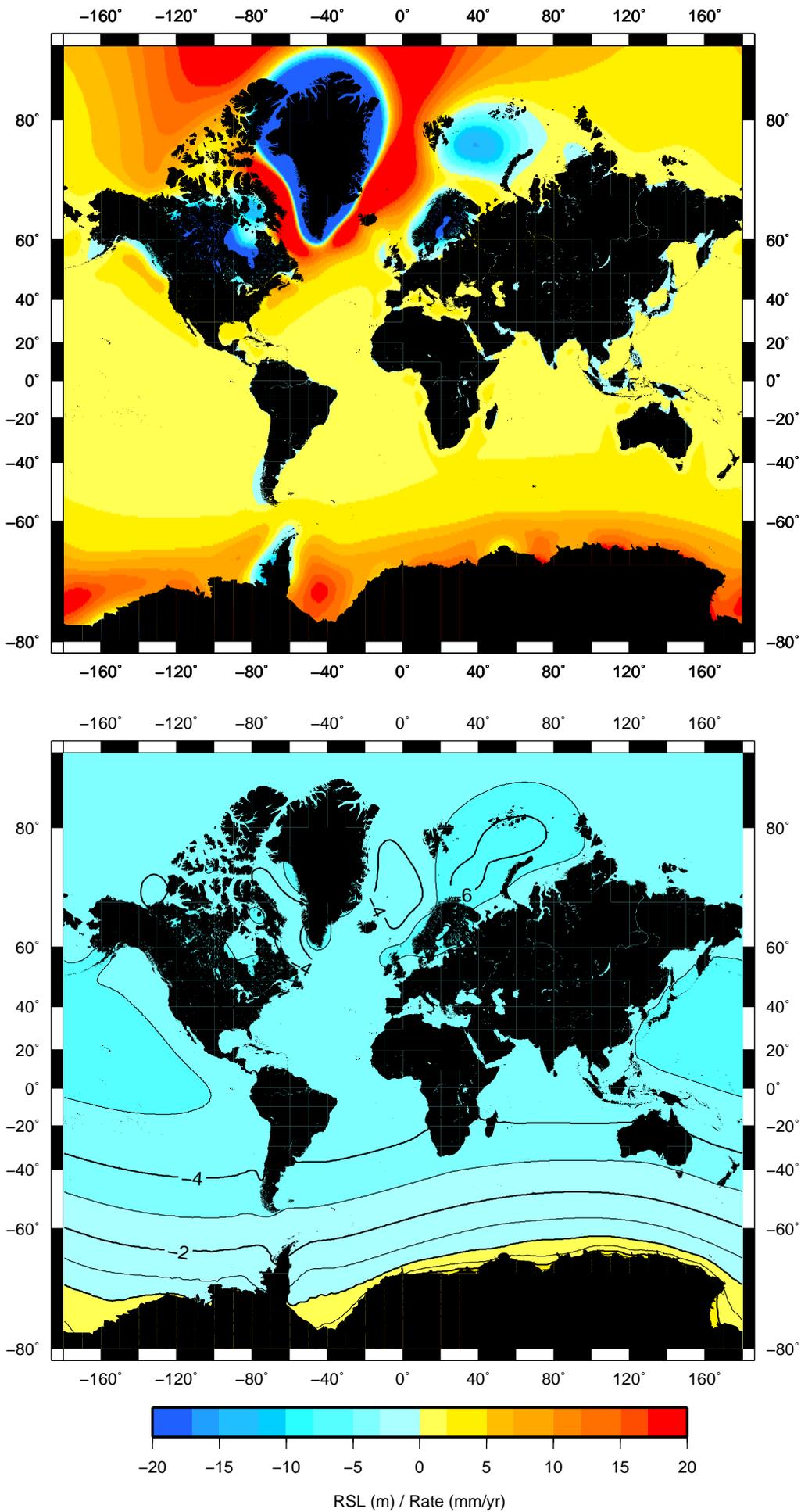
RSL (upper) and sea-level trend (lower) at time 120 kyr BP for Antarctica experiment



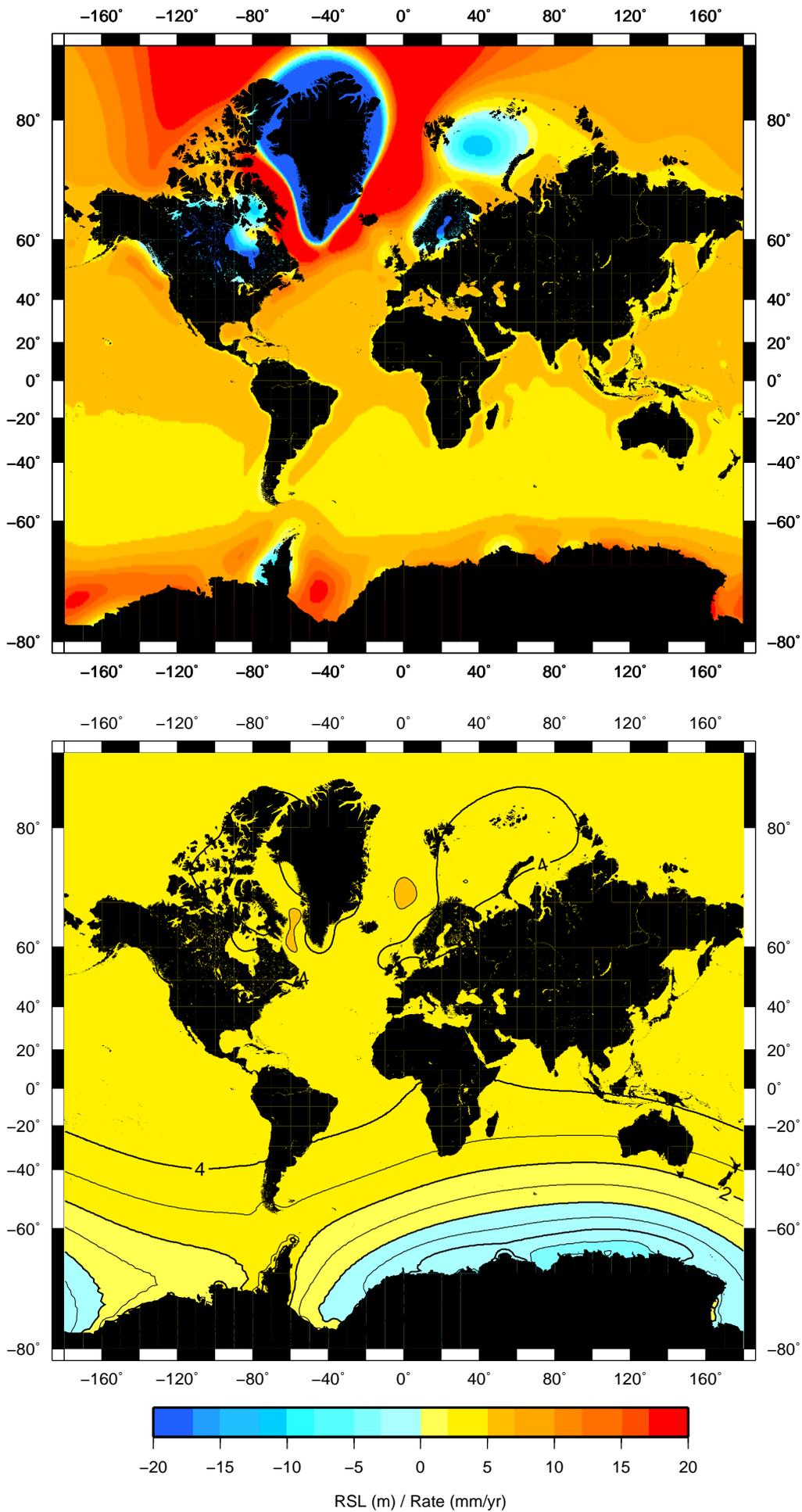
RSL (upper) and sea-level trend (lower) at time 119 kyr BP for Antarctica experiment



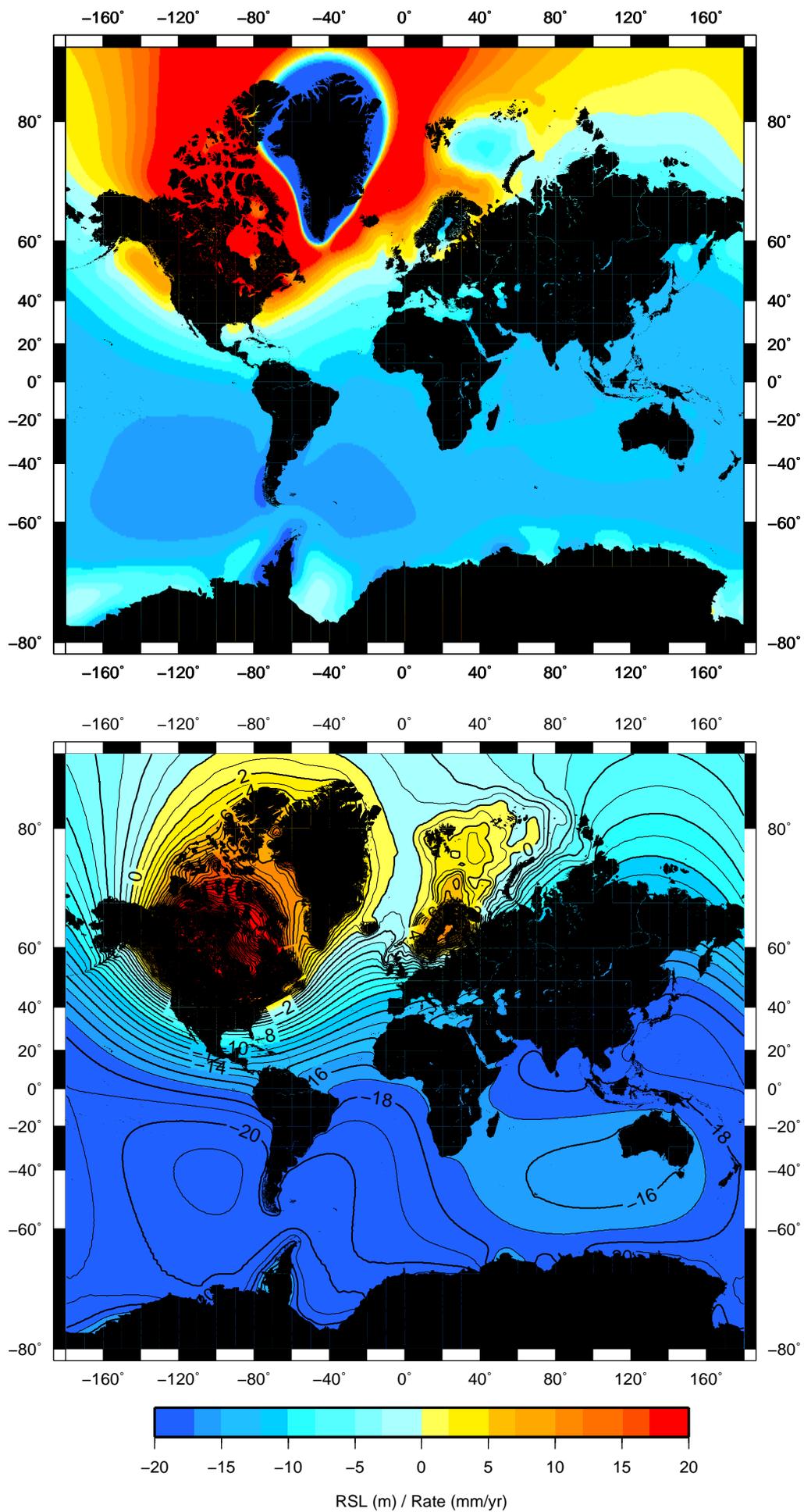
RSL (upper) and sea-level trend (lower) at time 118 kyr BP for Antarctica experiment



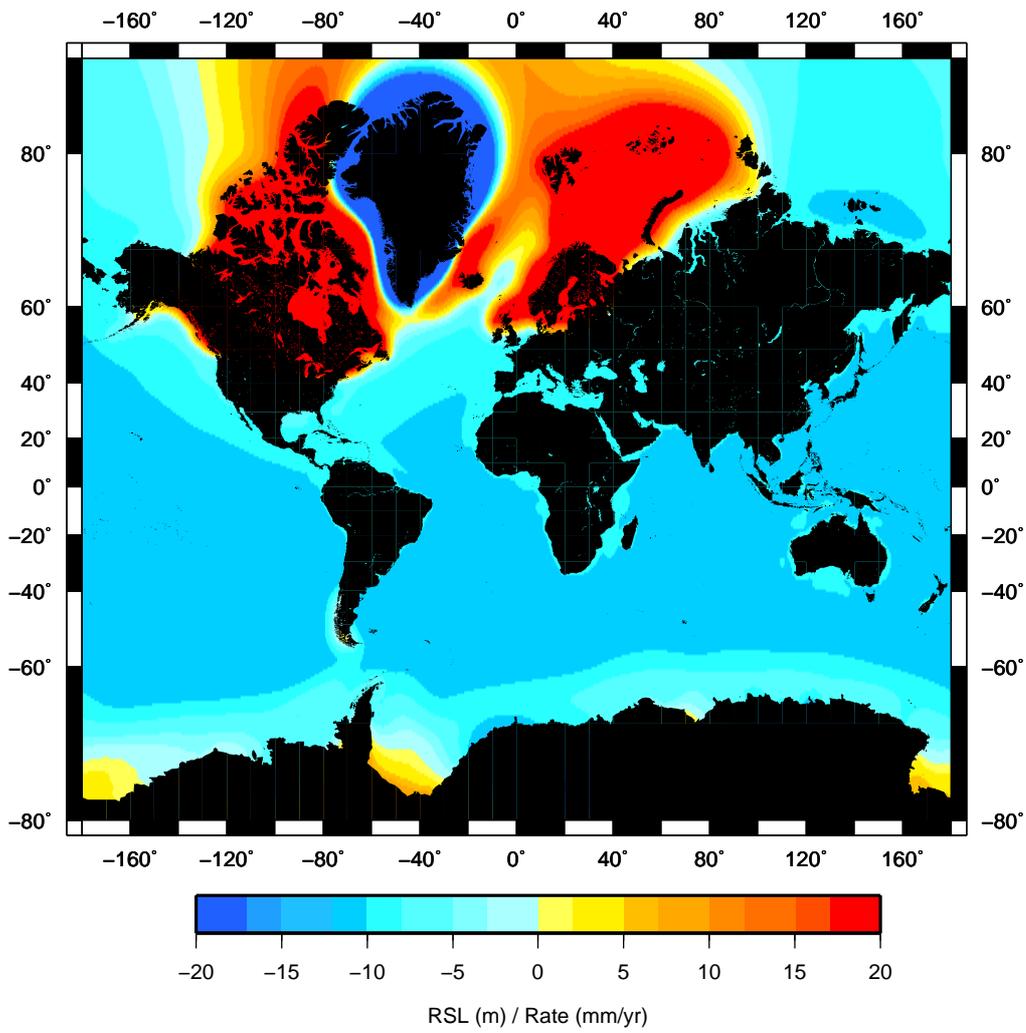
RSL (upper) and sea-level trend (lower) at time 117 kyr BP for Antarctica experiment



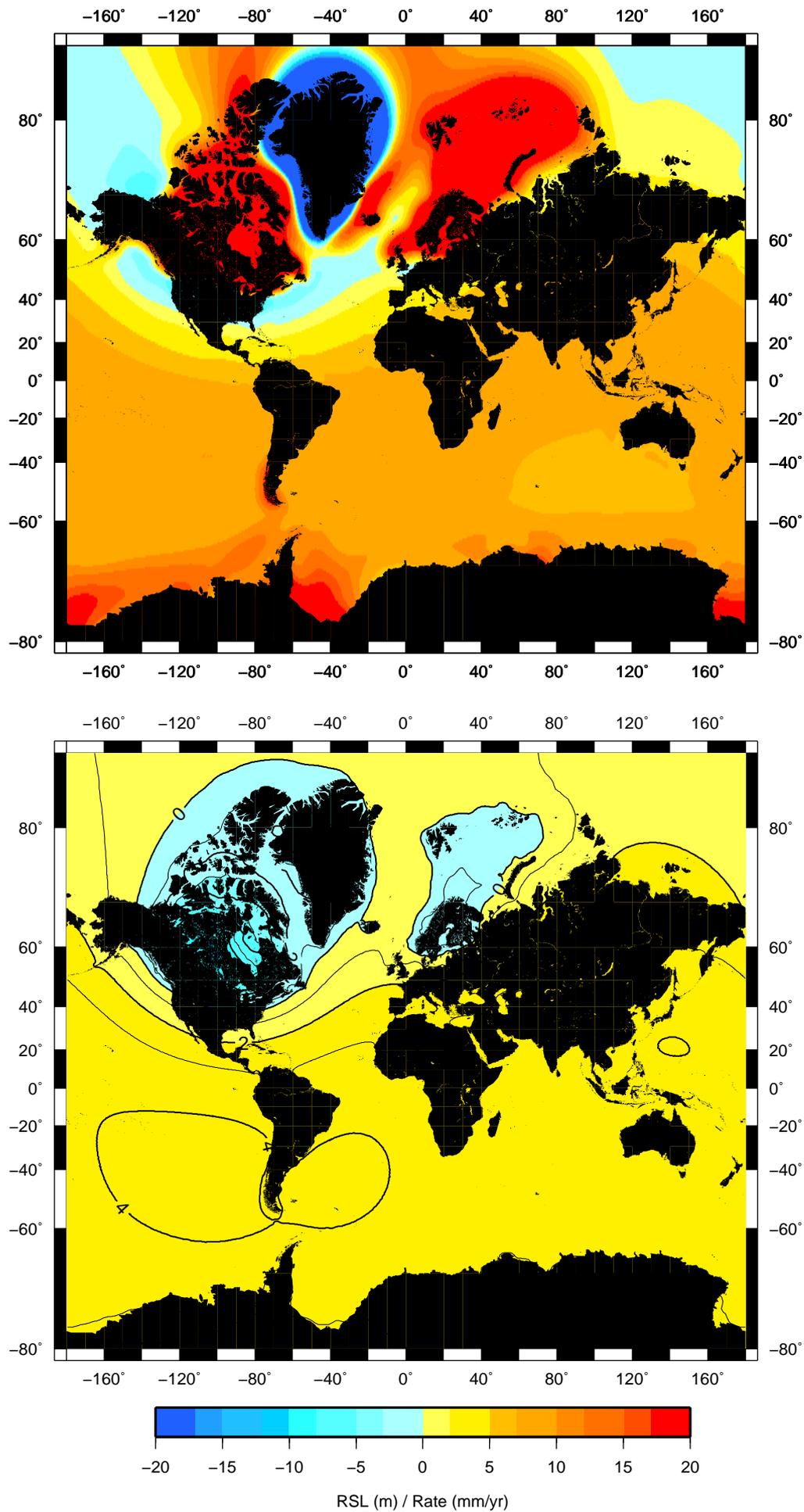
RSL (upper) and sea-level trend (lower) at time 116 kyr BP for Antarctica experiment



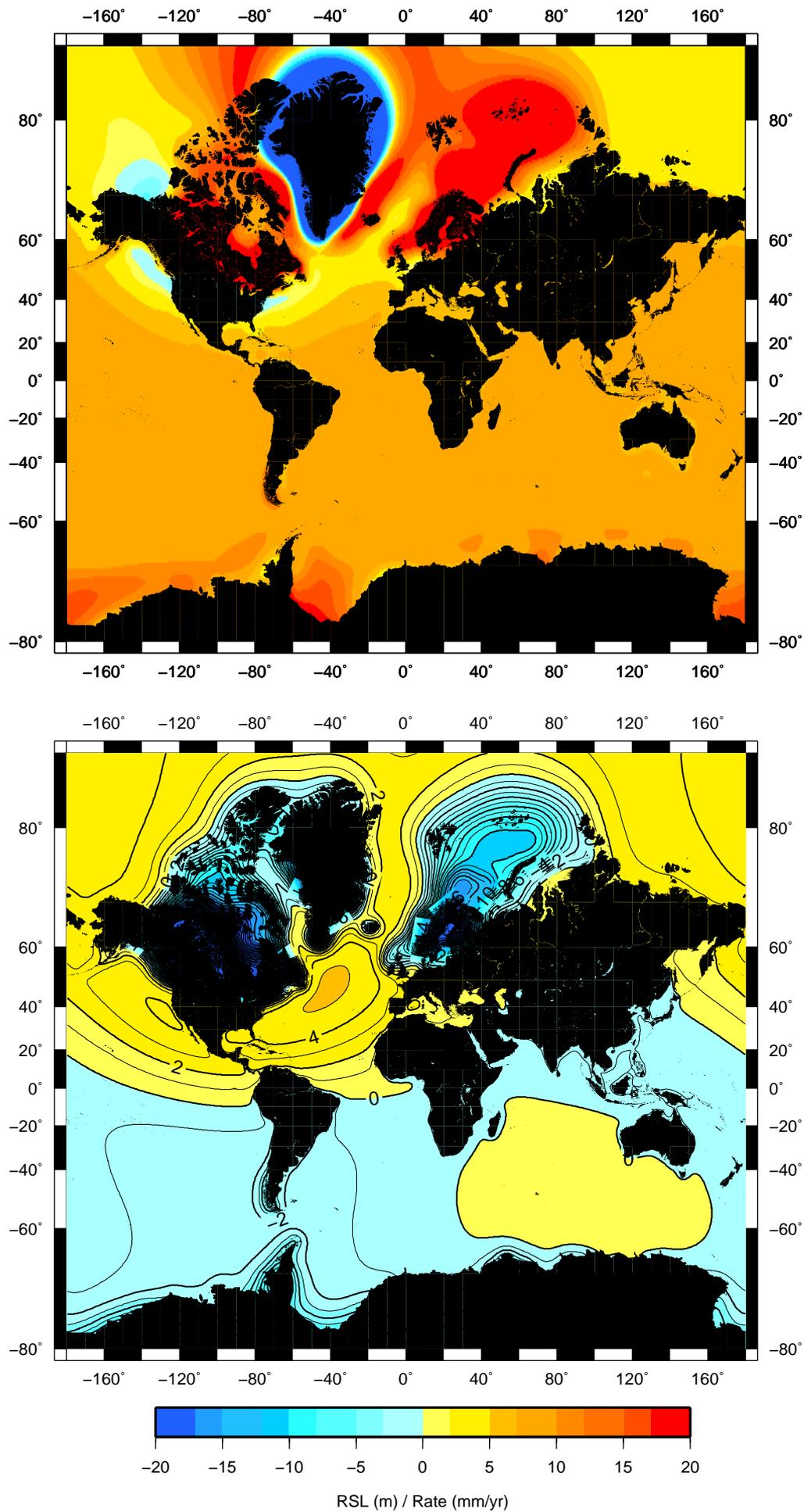
RSL at time 132 kyr BP for Greenland experiment



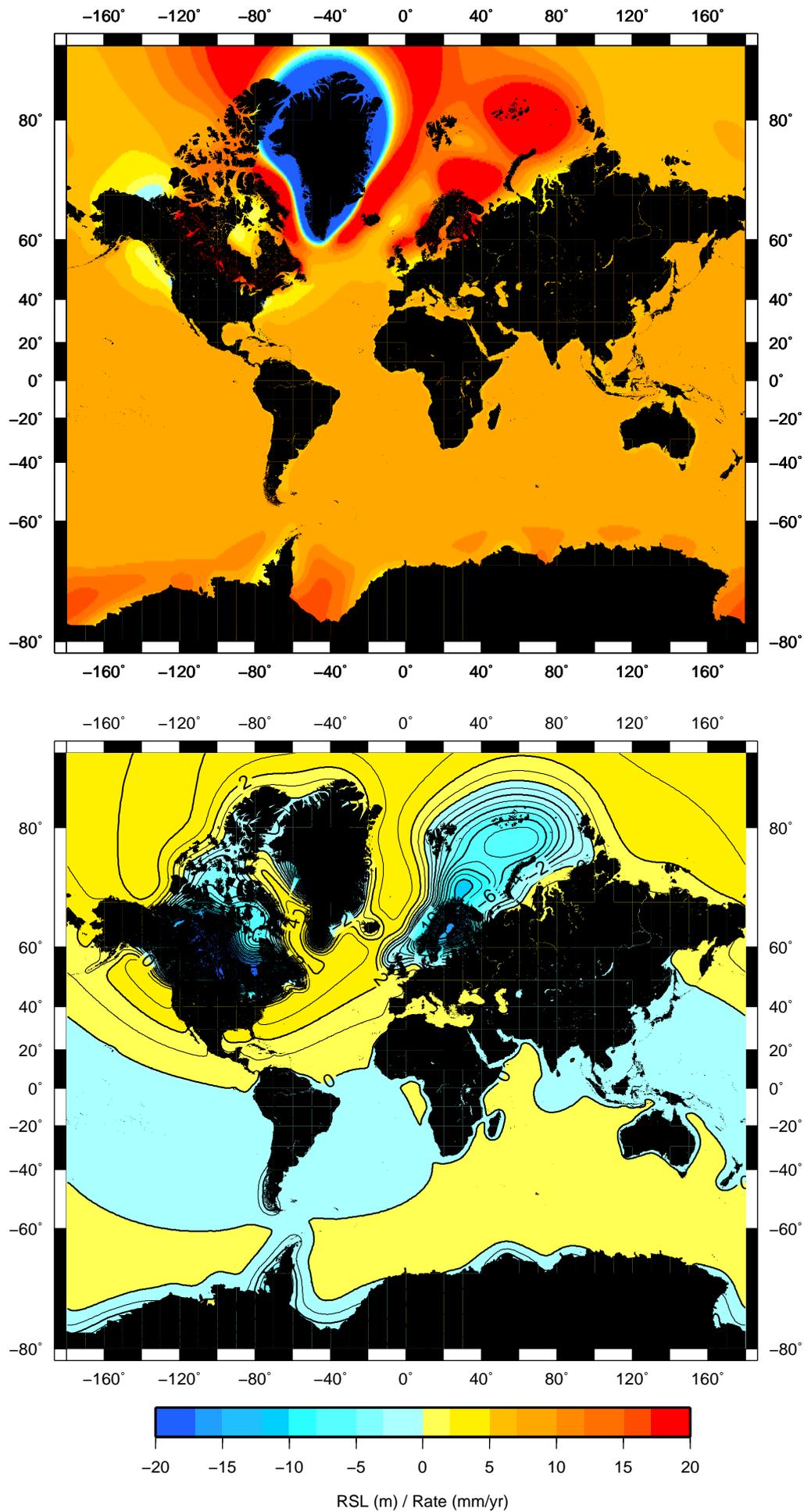
RSL (upper) and sea-level trend (lower) at time 127 kyr BP for Greenland experiment



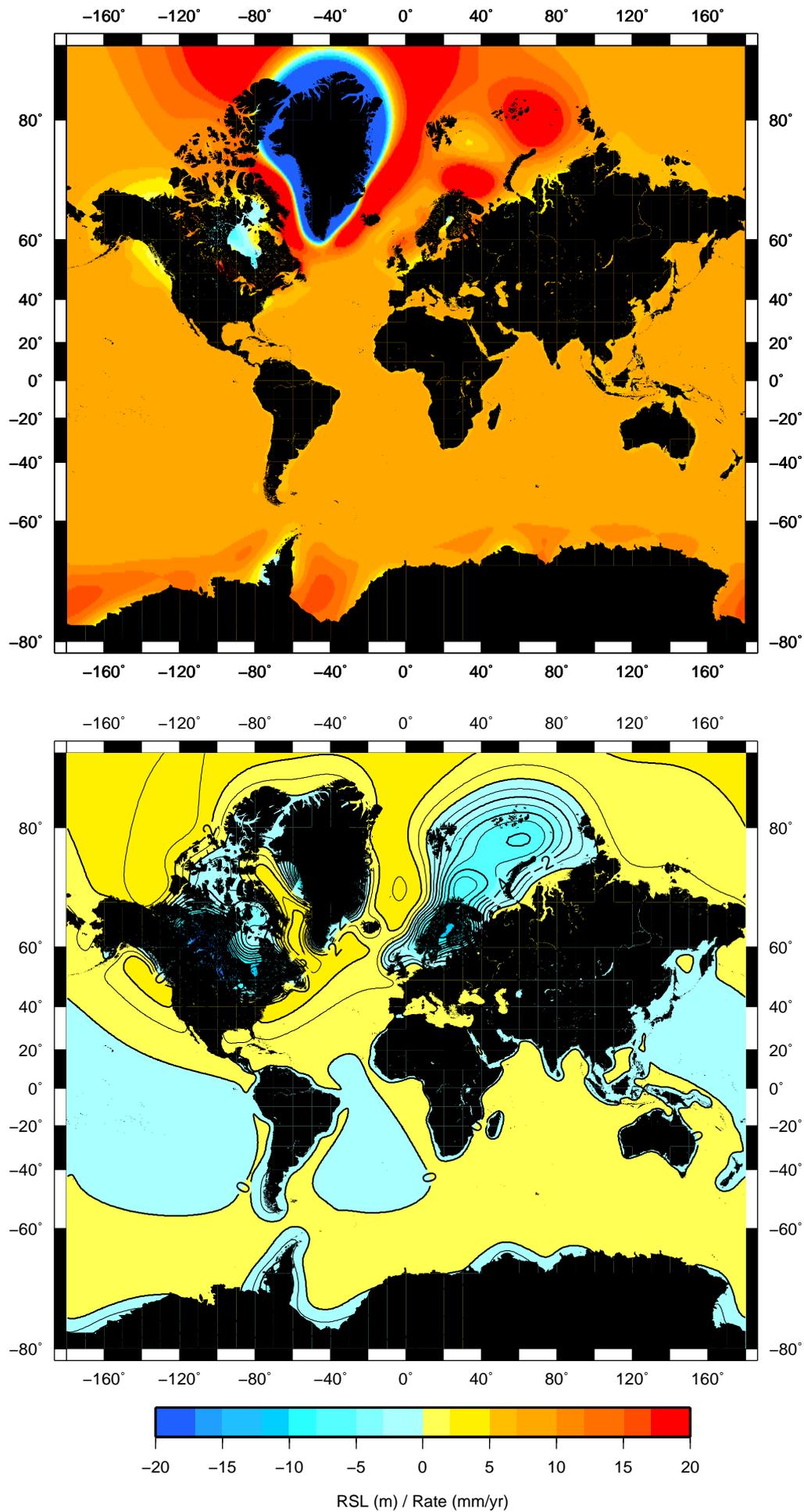
RSL (upper) and sea-level trend (lower) at time 126 kyr BP for Greenland experiment



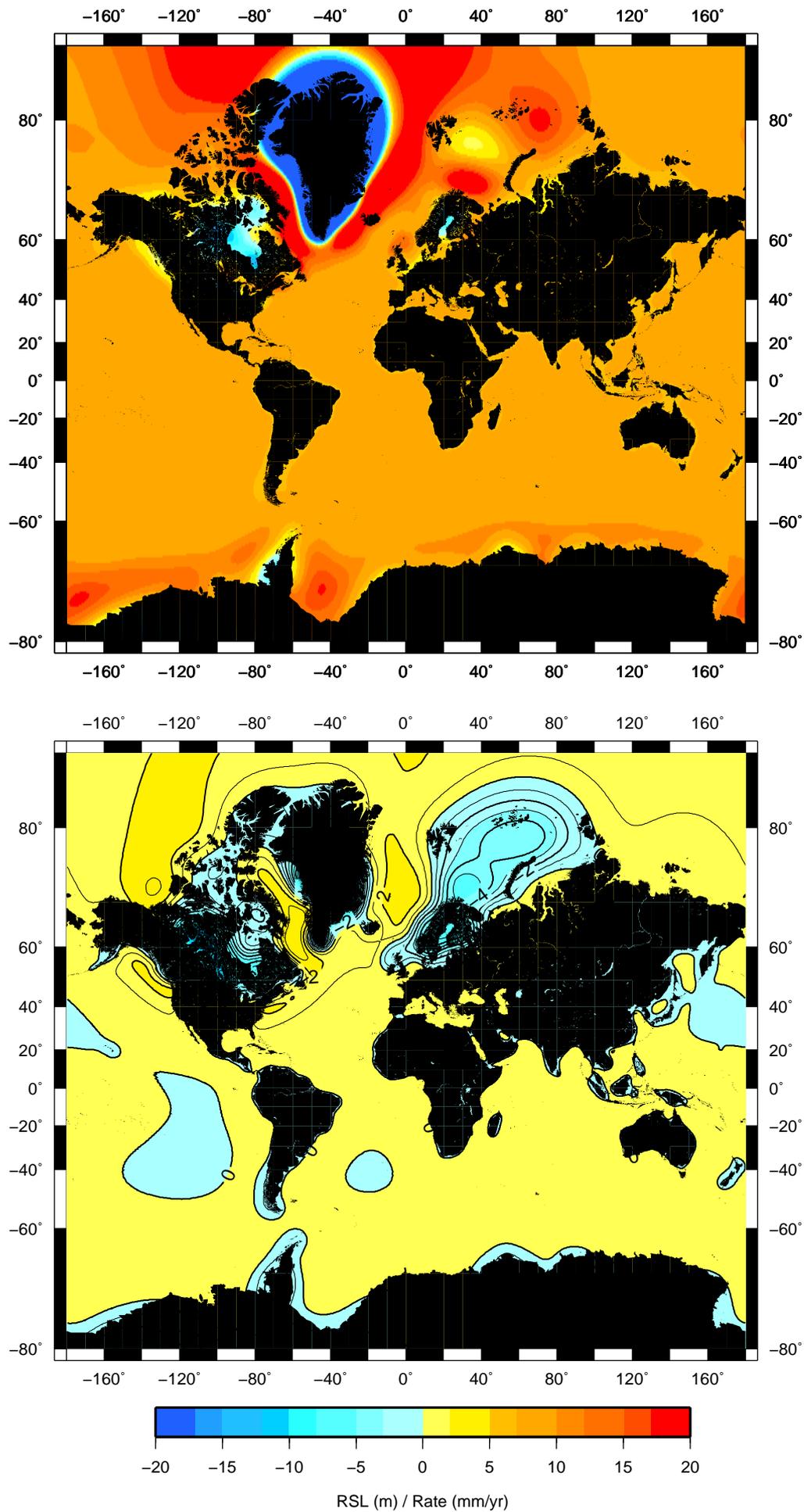
RSL (upper) and sea-level trend (lower) at time 125 kyr BP for Greenland experiment



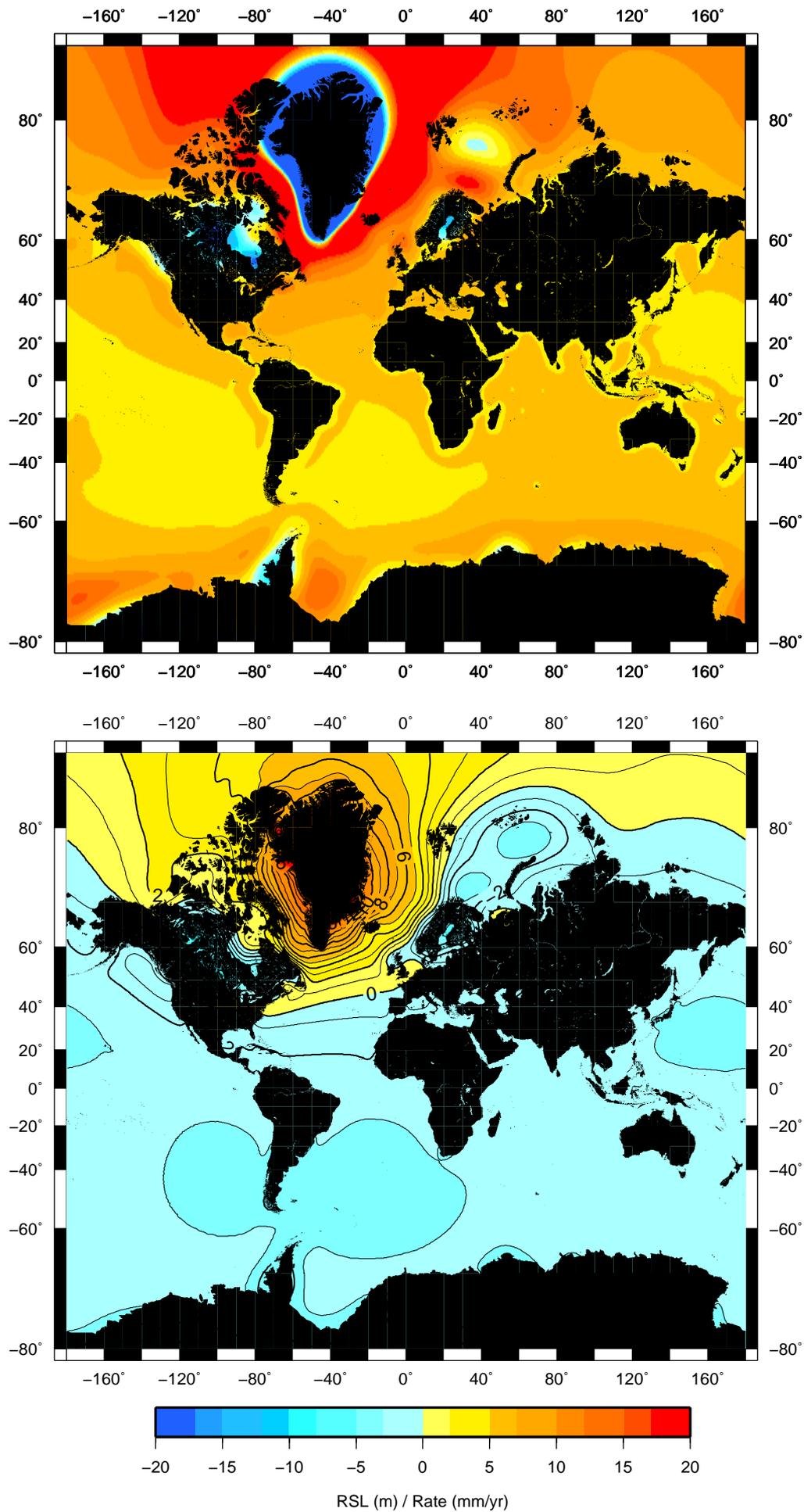
RSL (upper) and sea-level trend (lower) at time 124 kyr BP for Greenland experiment



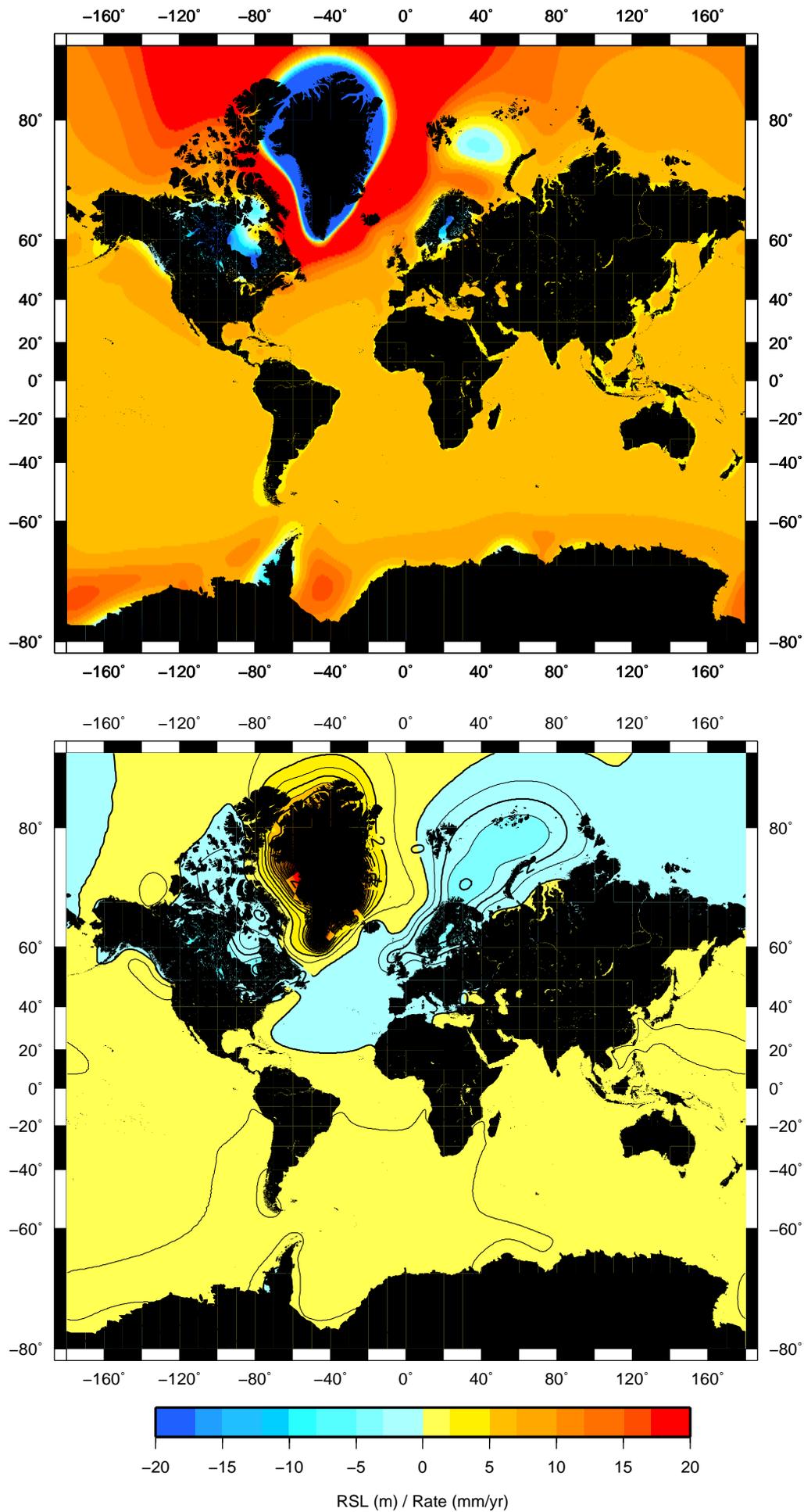
RSL (upper) and sea-level trend (lower) at time 123 kyr BP for Greenland experiment



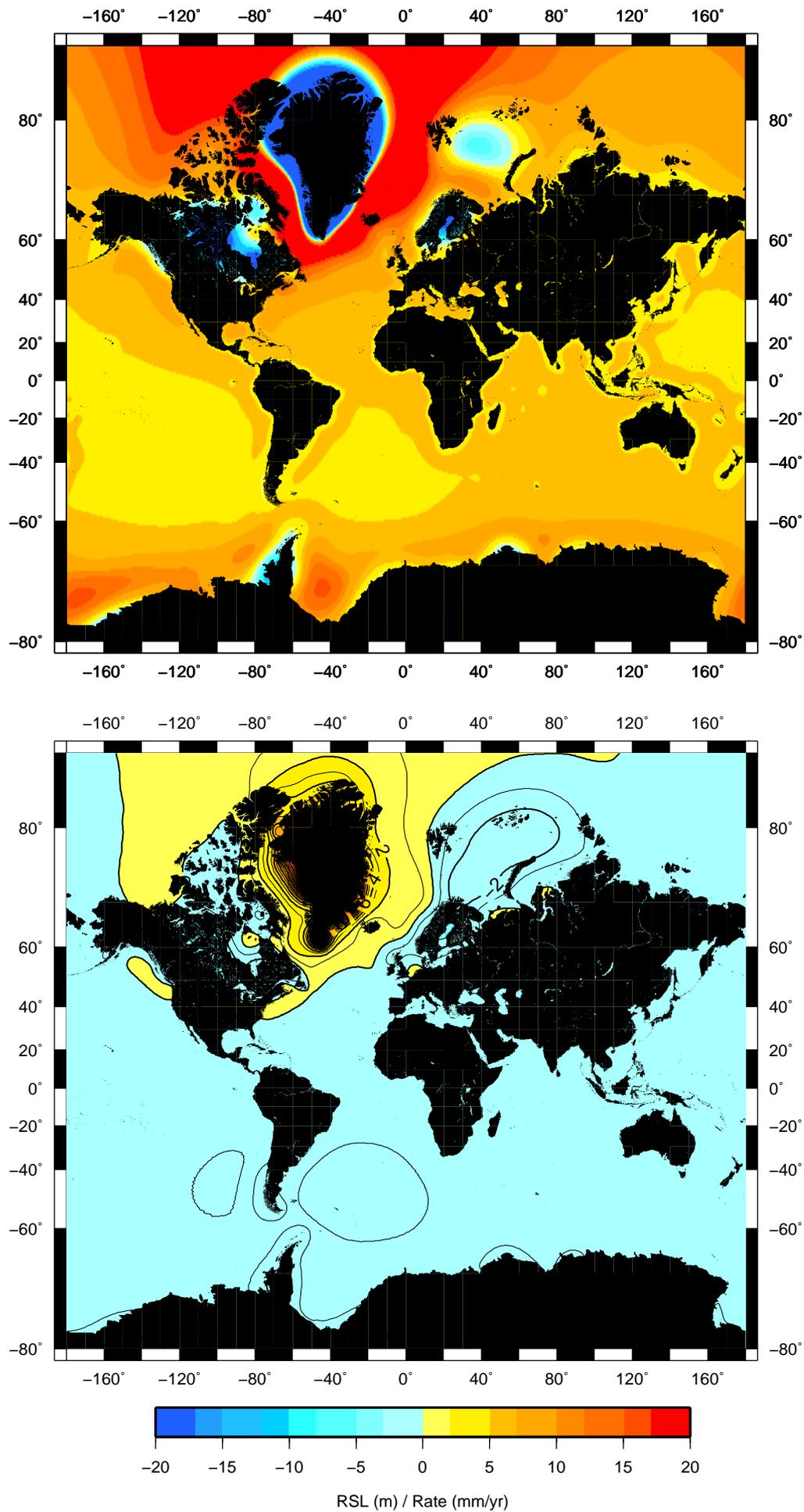
RSL (upper) and sea-level trend (lower) at time 122 kyr BP for Greenland experiment



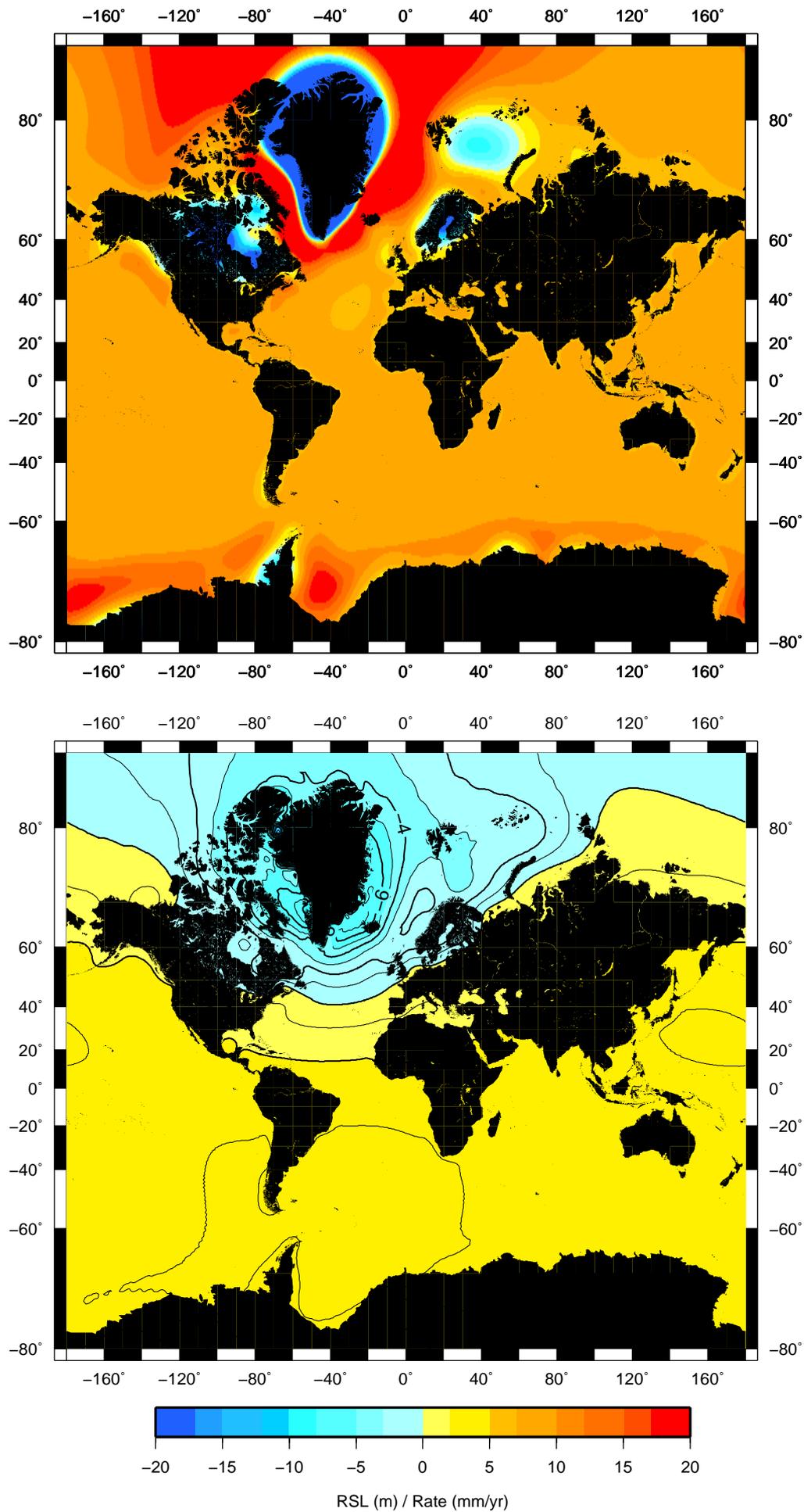
RSL (upper) and sea-level trend (lower) at time 121 kyr BP for Greenland experiment



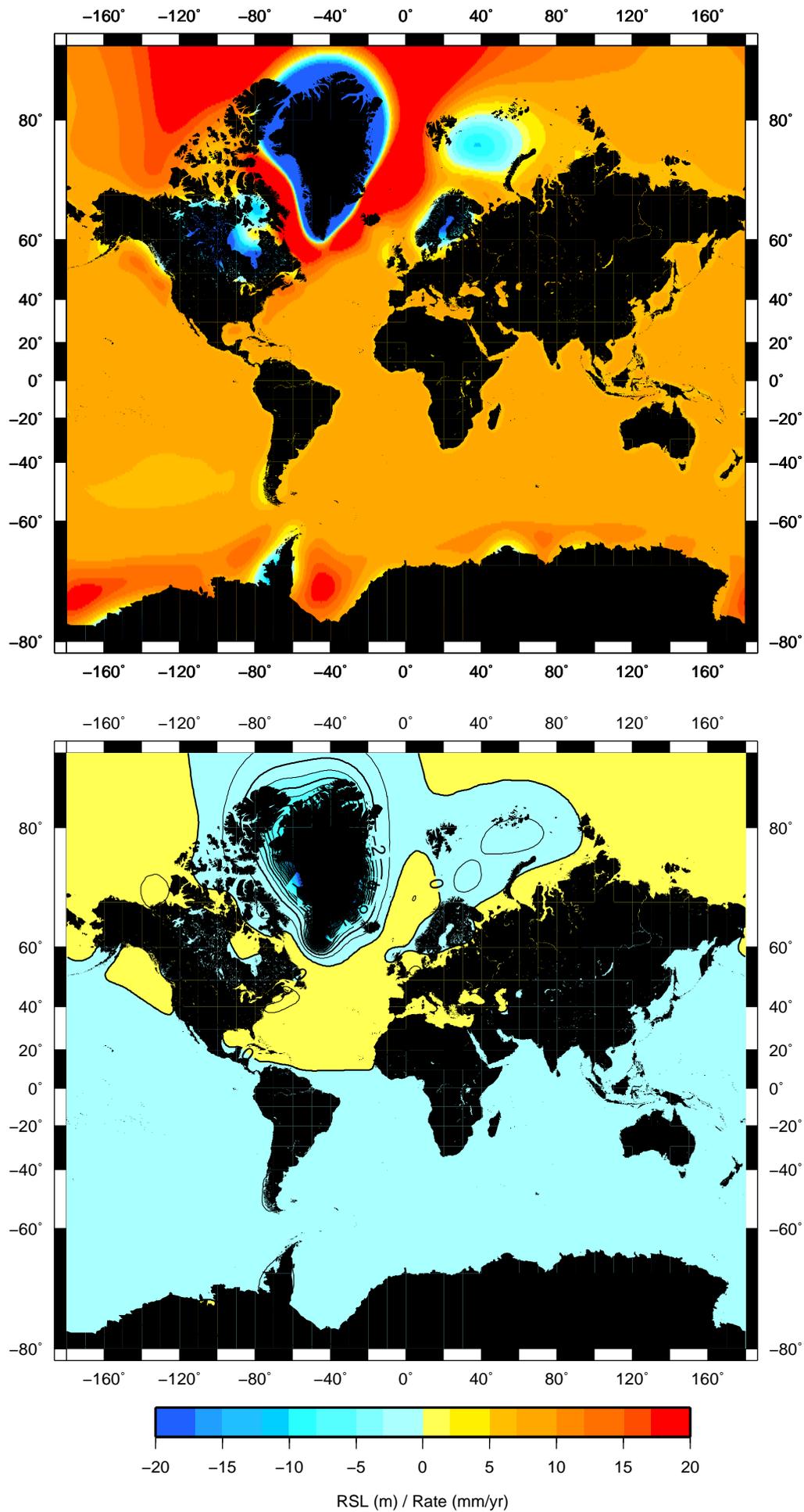
RSL (upper) and sea-level trend (lower) at time 120 kyr BP for Greenland experiment



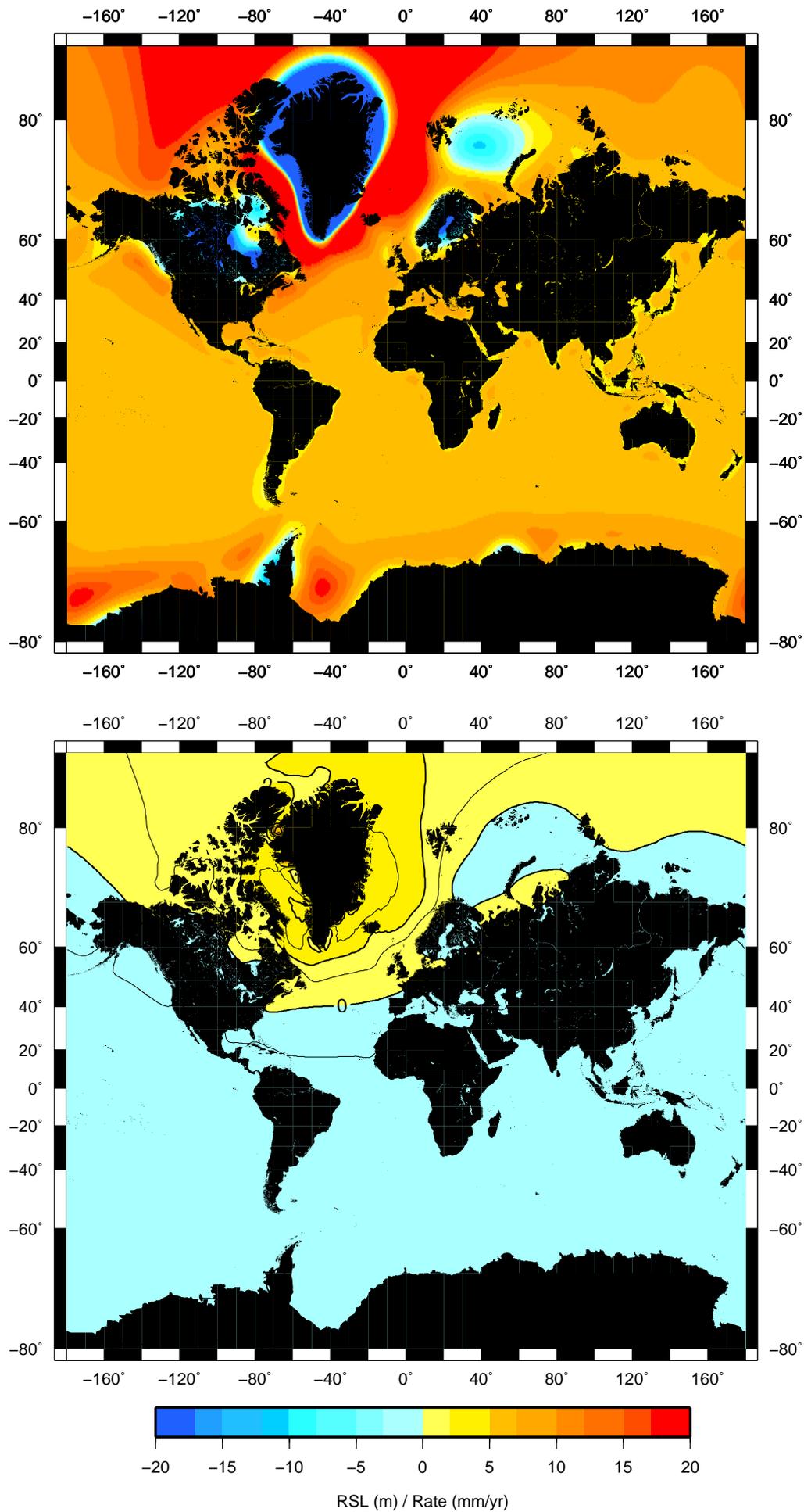
RSL (upper) and sea-level trend (lower) at time 119 kyr BP for Greenland experiment



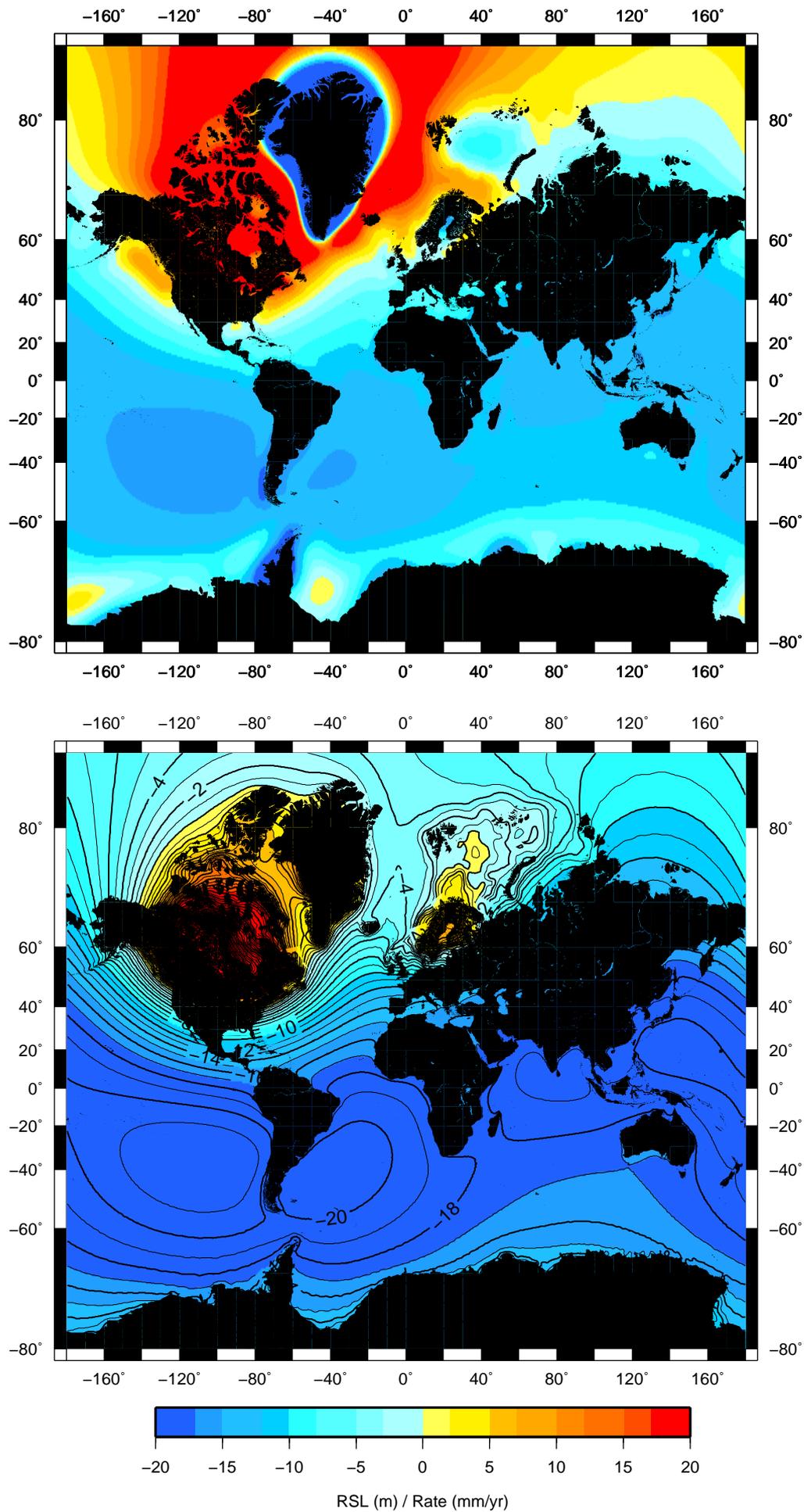
RSL (upper) and sea-level trend (lower) at time 118 kyr BP for Greenland experiment



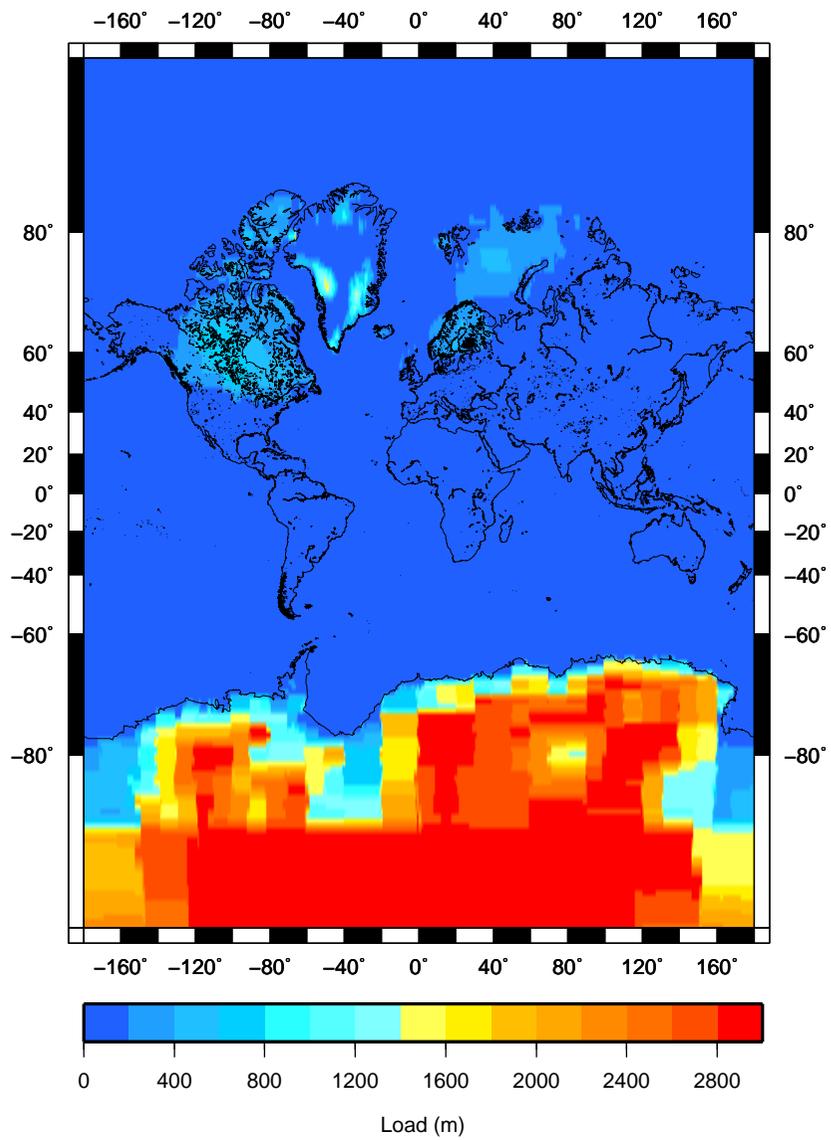
RSL (upper) and sea-level trend (lower) at time 117 kyr BP for Greenland experiment



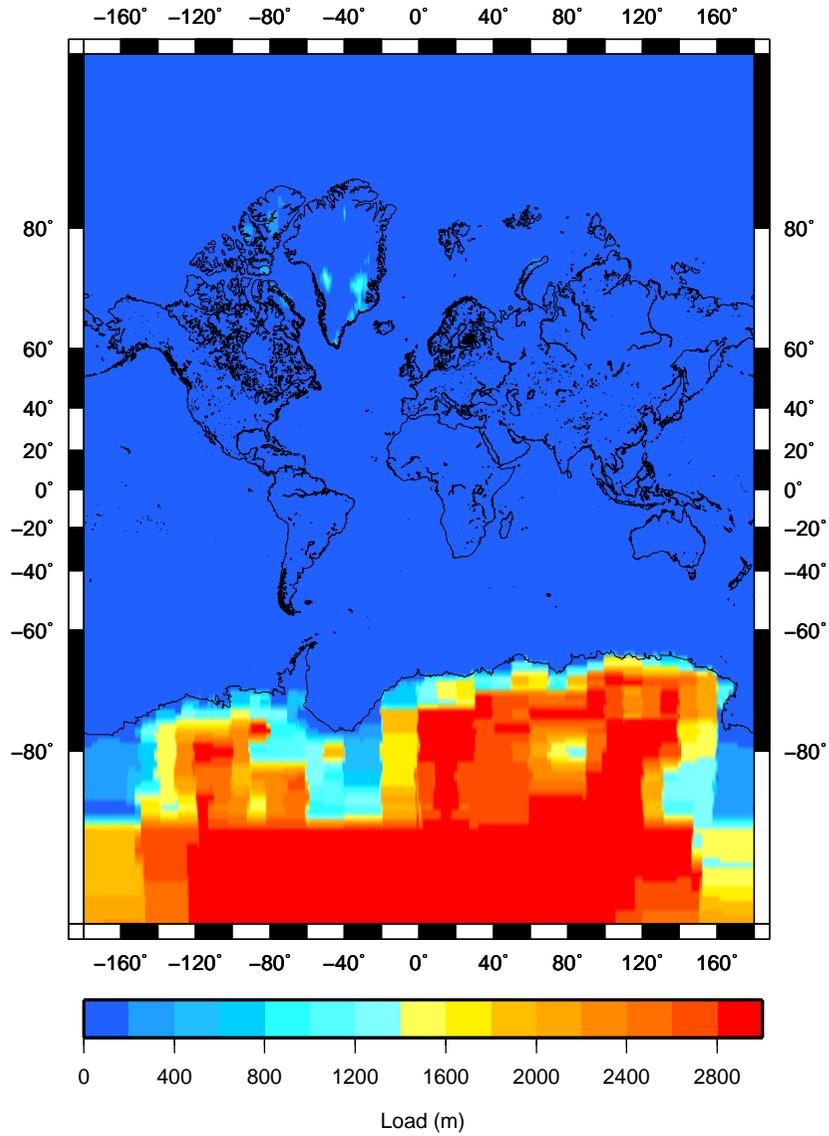
RSL (upper) and sea-level trend (lower) at time 116 kyr BP for Greenland experiment



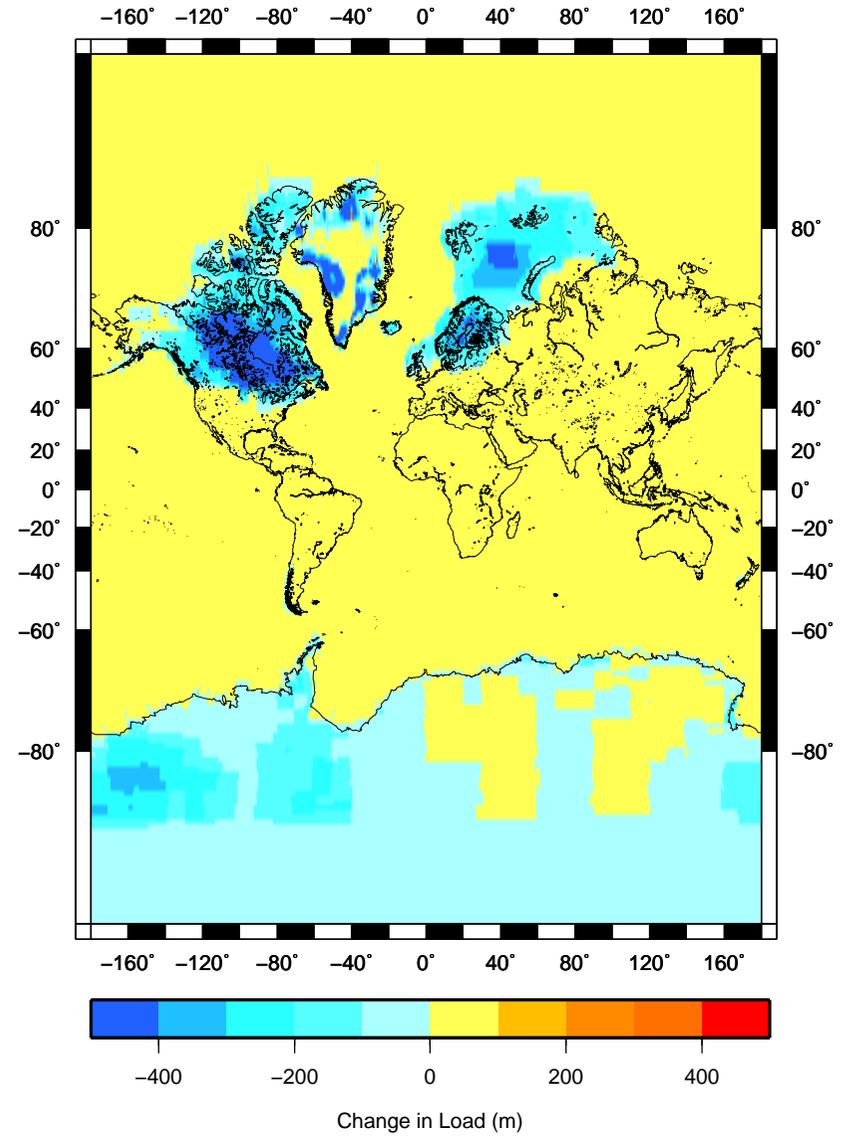
Ice load at 132 kyr BP for Antarctica experiment



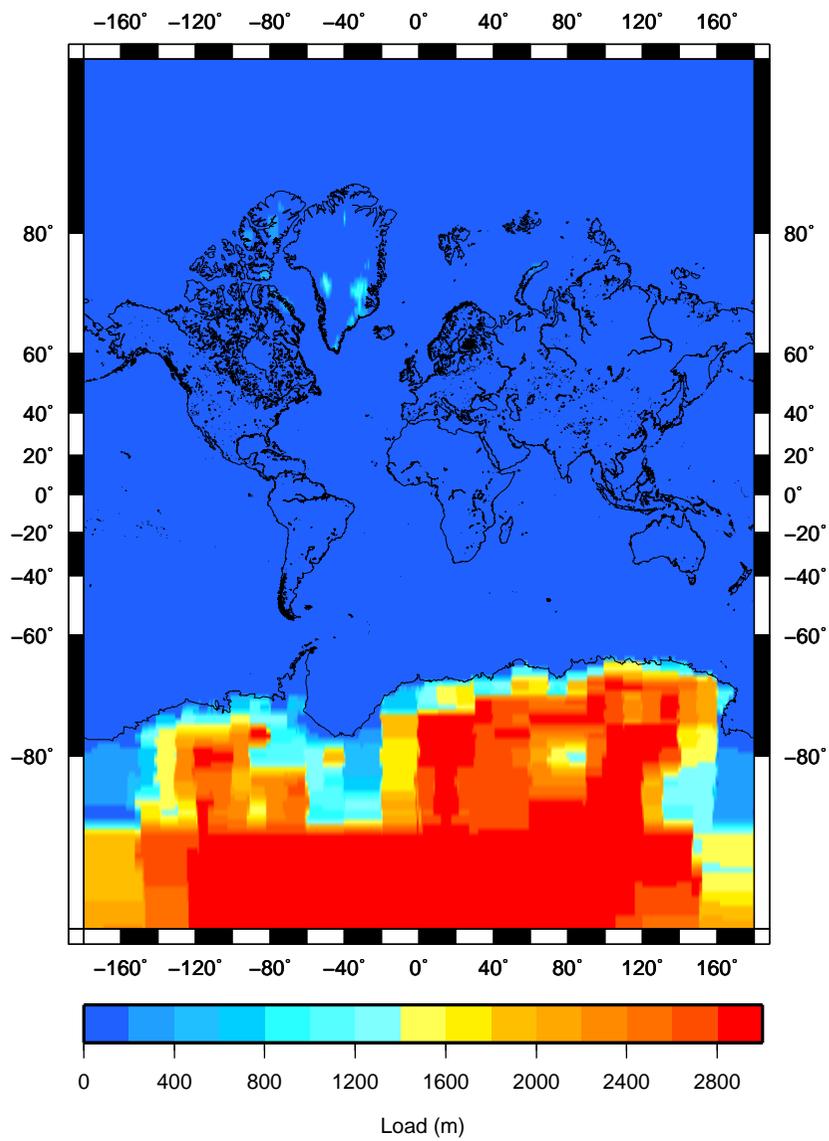
Ice load at 127 kyr BP for Antarctica experiment



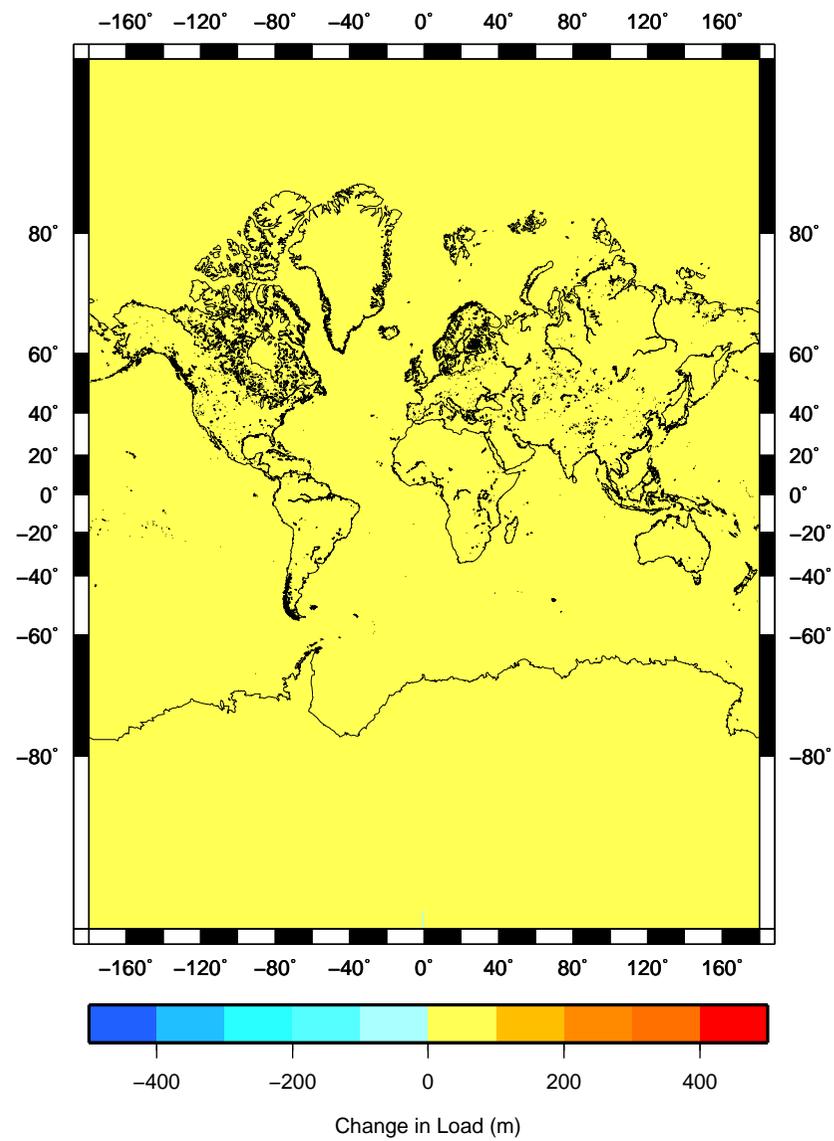
Change in ice load at 127 kyr BP for Antarctica experiment



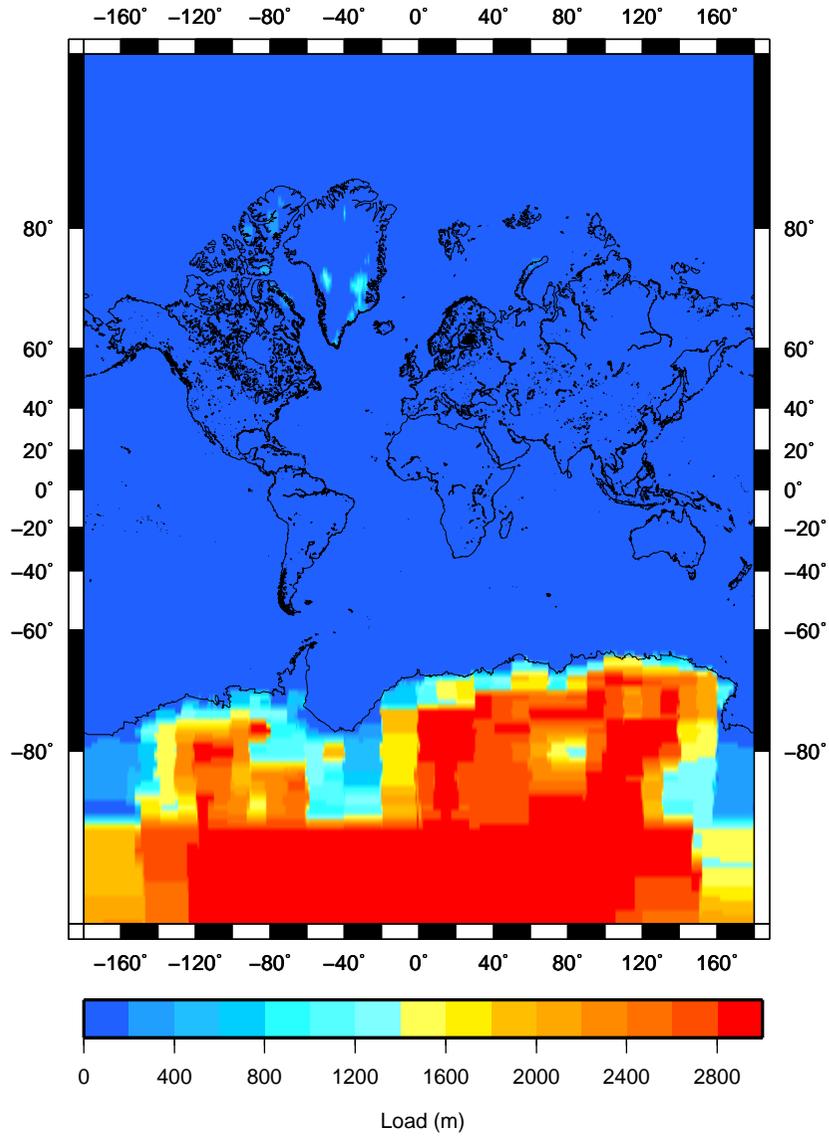
Ice load at 126 kyr BP for Antarctica experiment



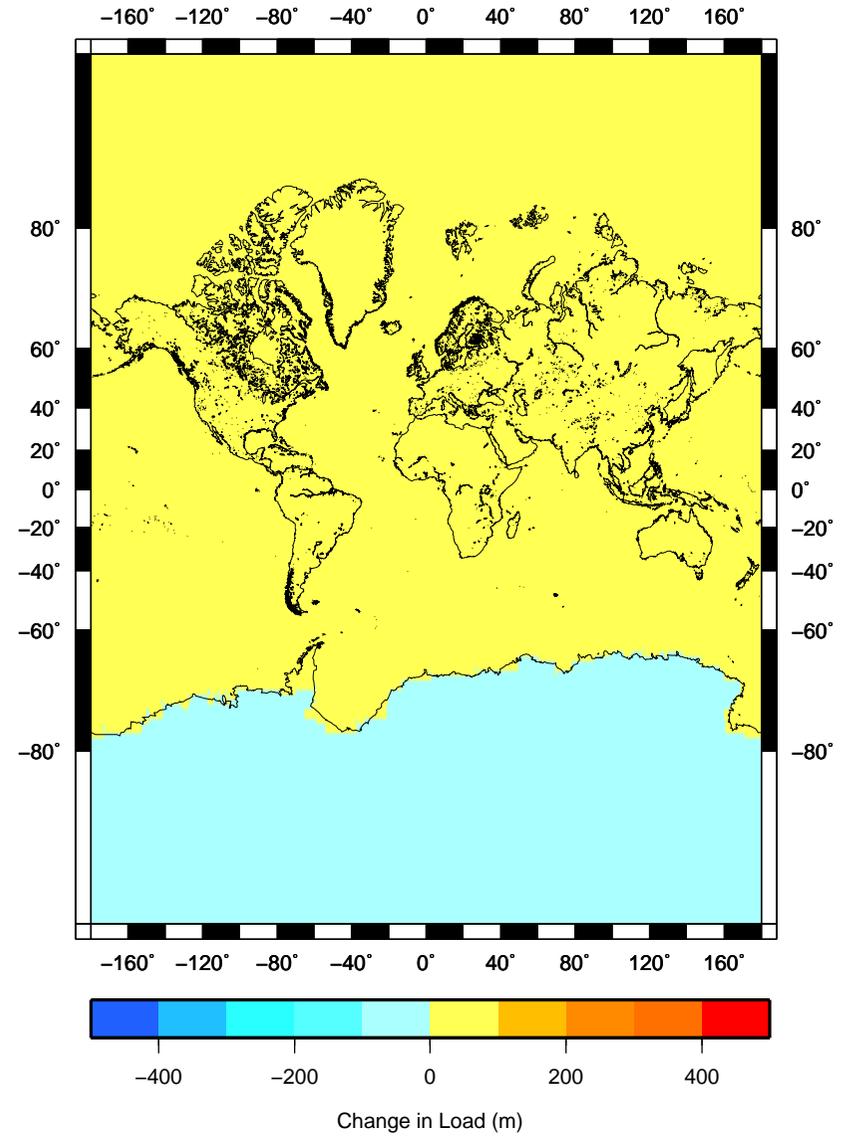
Change in ice load at 126 kyr BP for Antarctica experiment



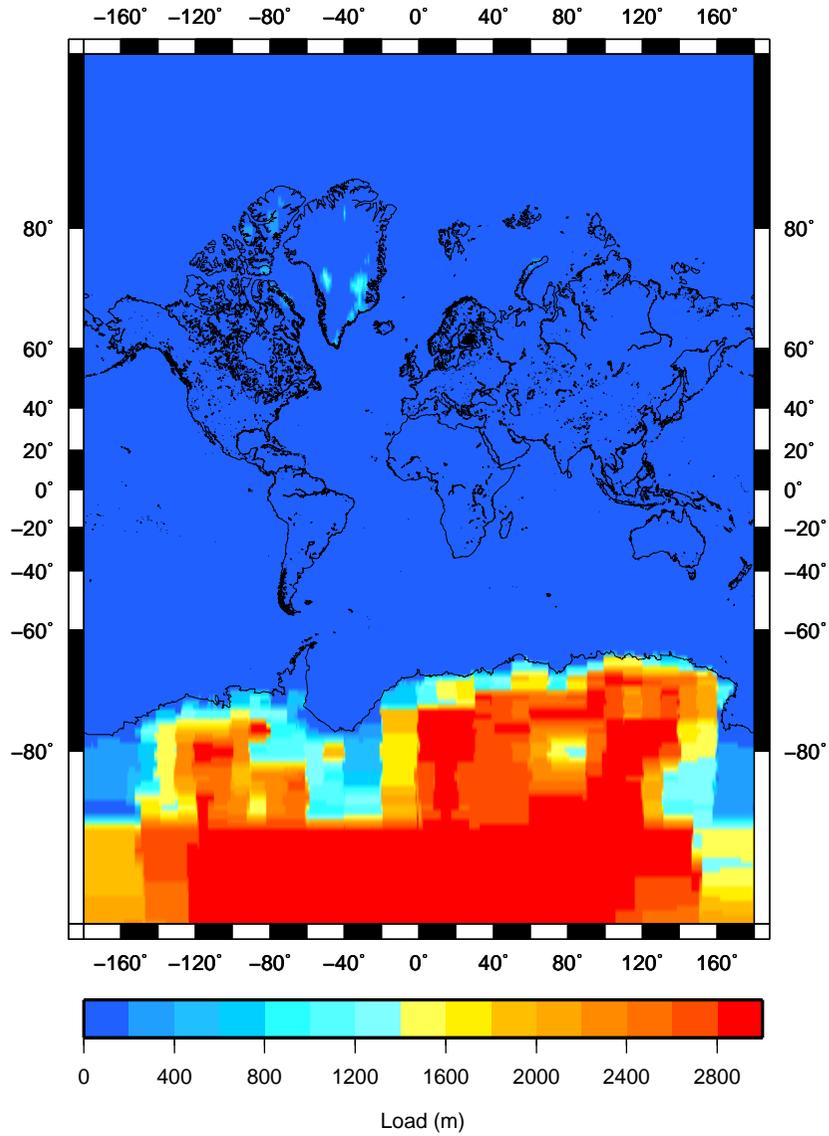
Ice load at 125 kyr BP for Antarctica experiment



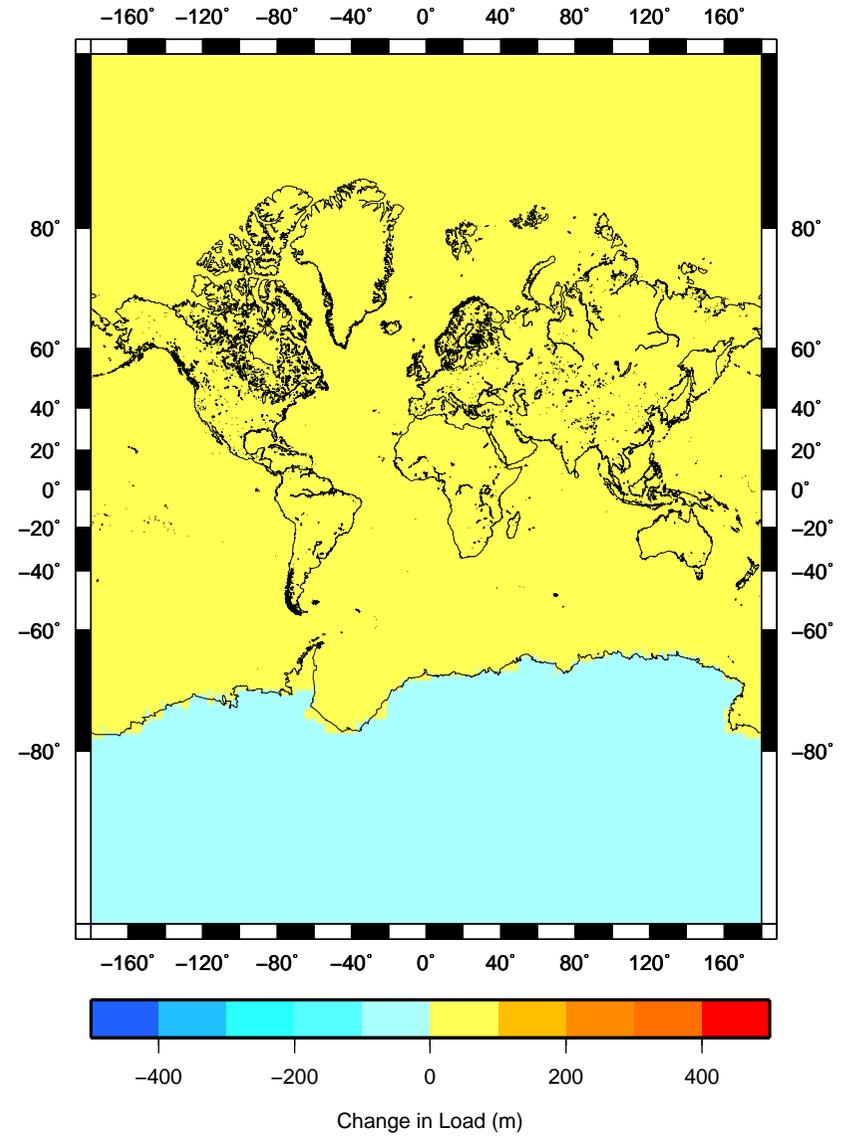
Change in ice load at 125 kyr BP for Antarctica experiment



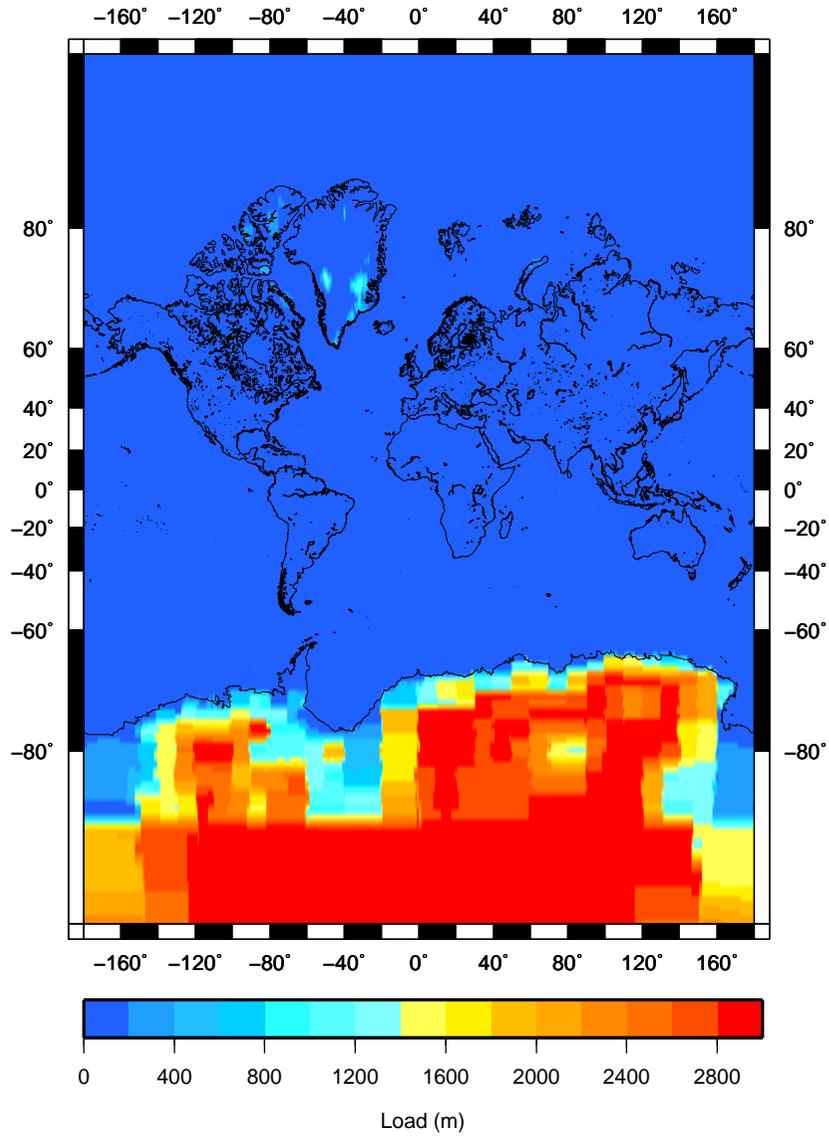
Ice load at 124 kyr BP for Antarctica experiment



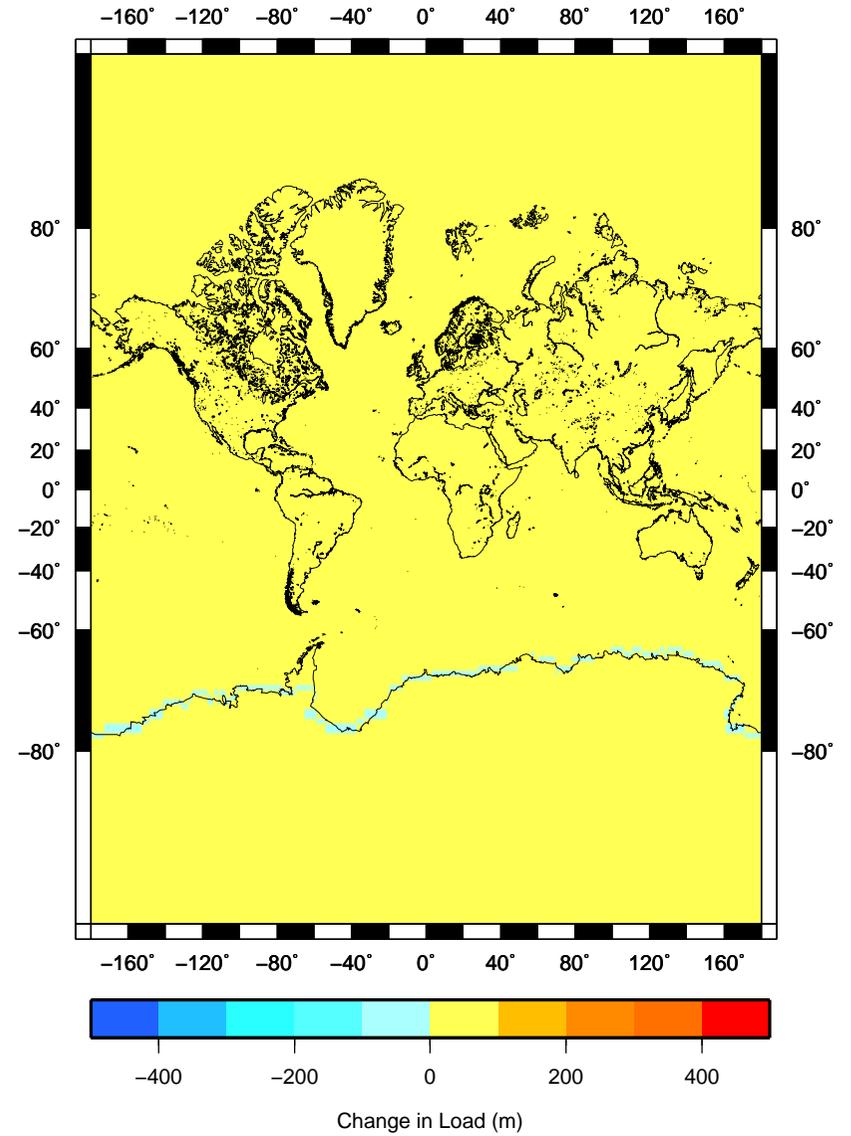
Change in ice load at 124 kyr BP for Antarctica experiment



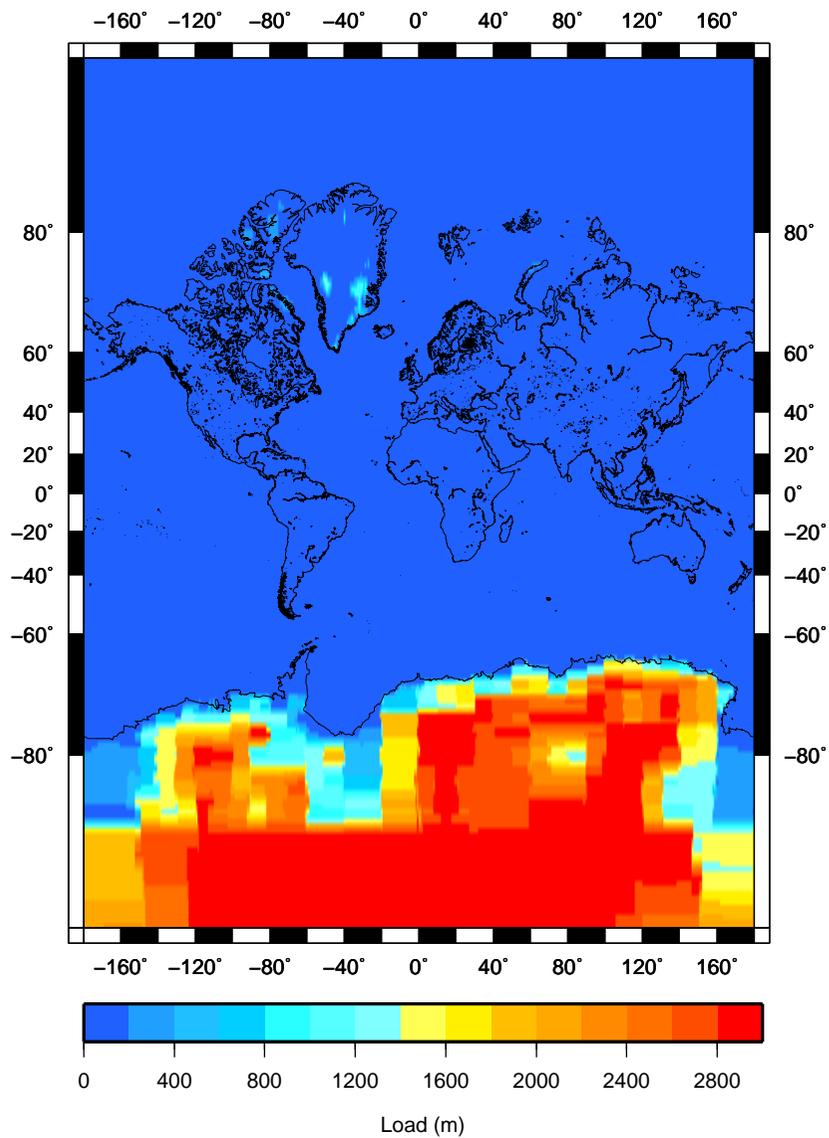
Ice load at 123 kyr BP for Antarctica experiment



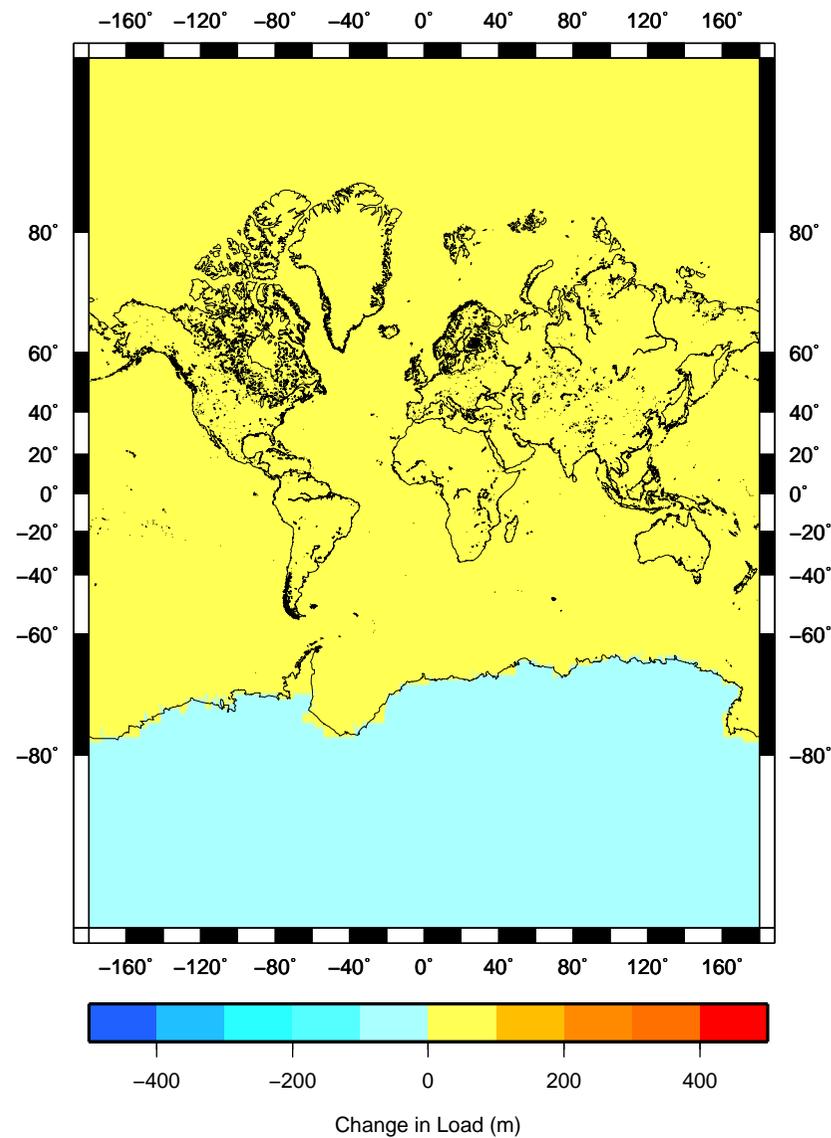
Change in ice load at 123 kyr BP for Antarctica experiment



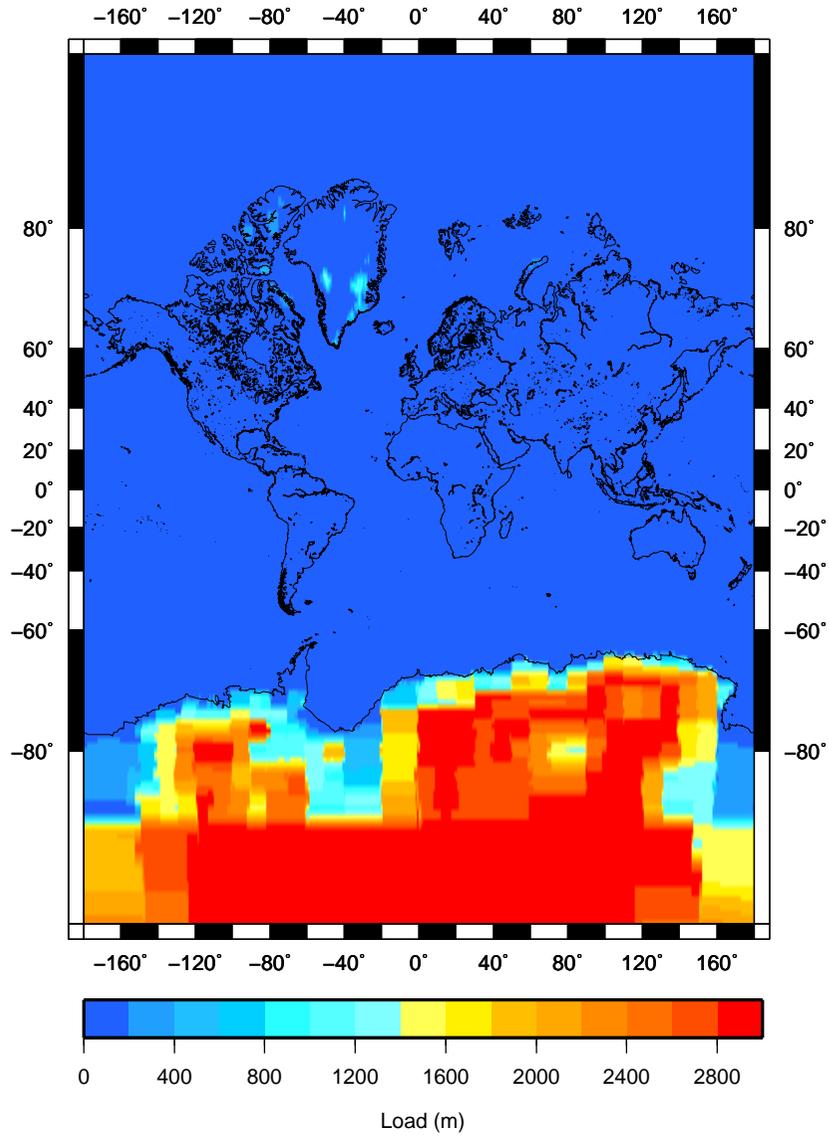
Ice load at 122 kyr BP for Antarctica experiment



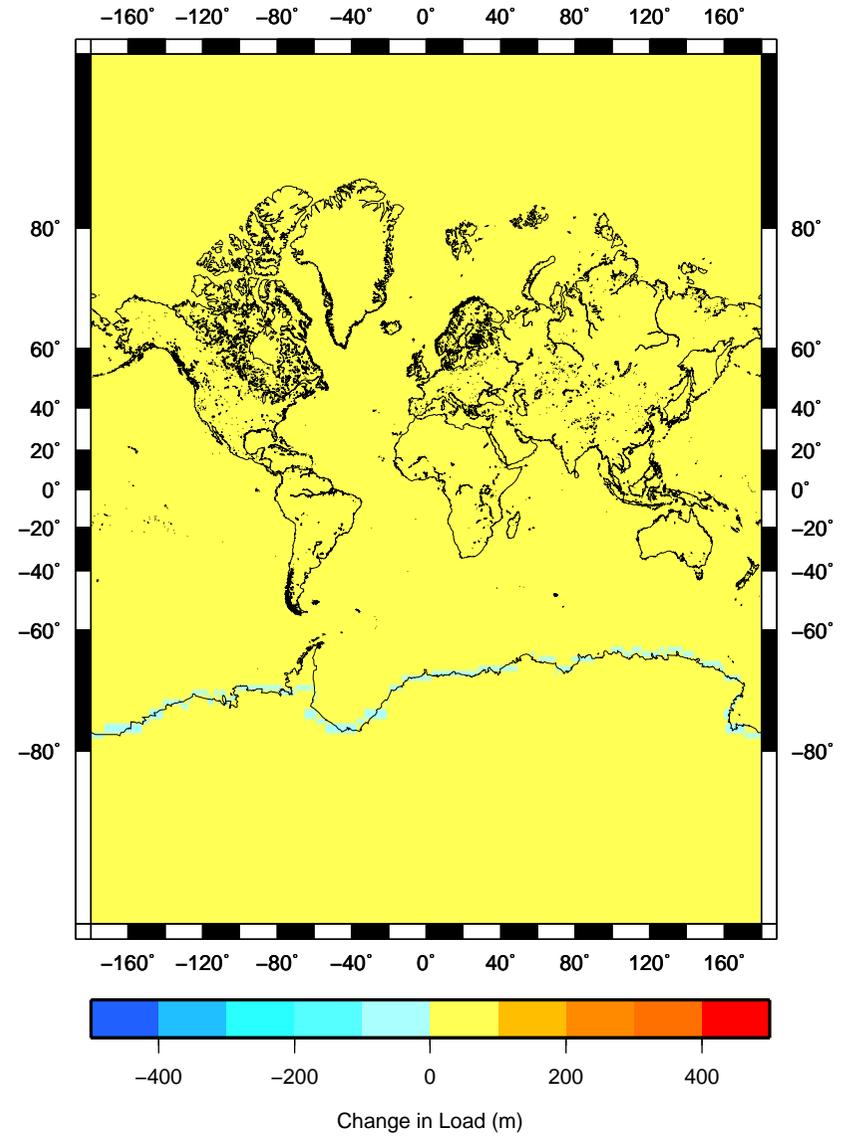
Change in ice load at 122 kyr BP for Antarctica experiment



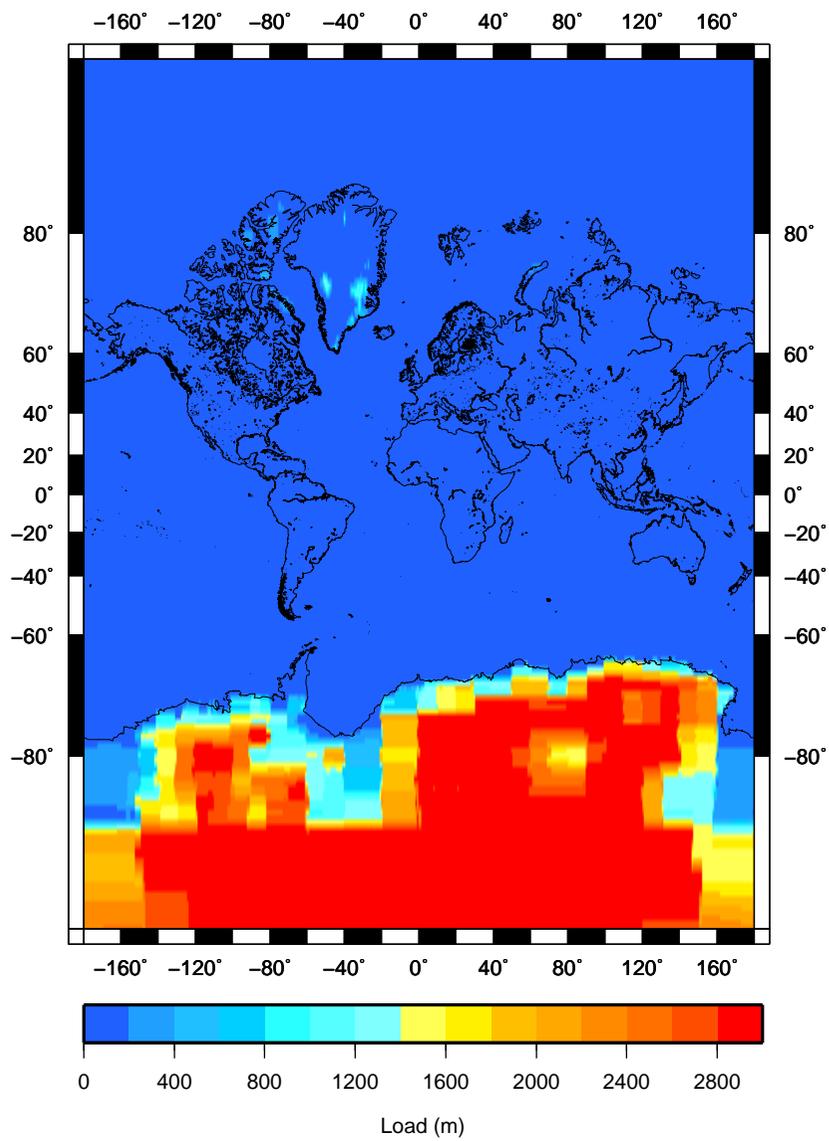
Ice load at 121 kyr BP for Antarctica experiment



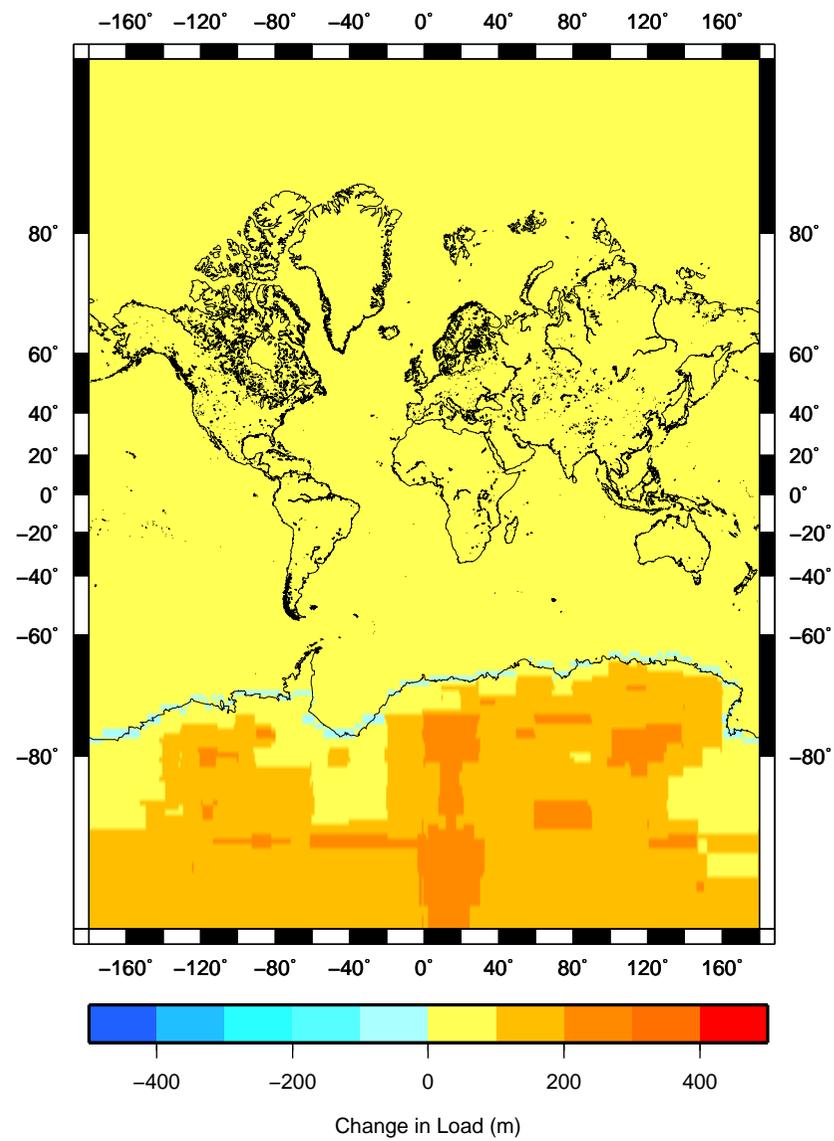
Change in ice load at 121 kyr BP for Antarctica experiment



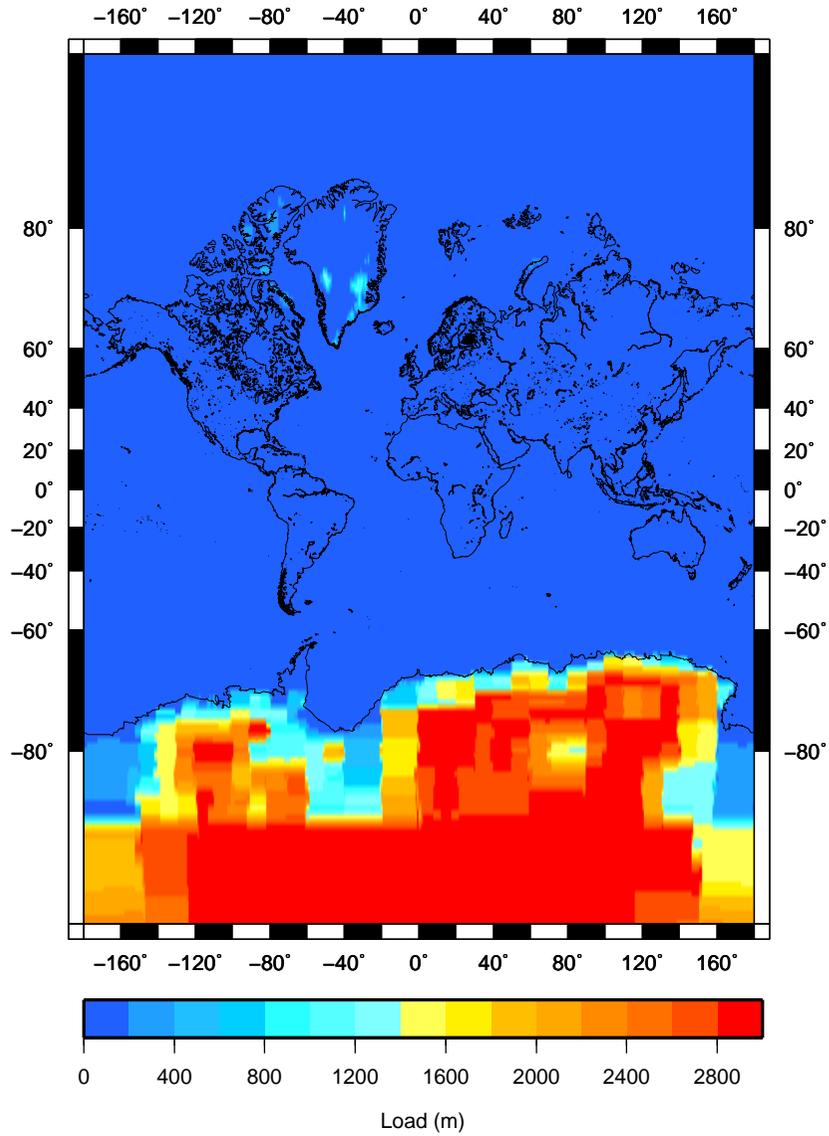
Ice load at 120 kyr BP for Antarctica experiment



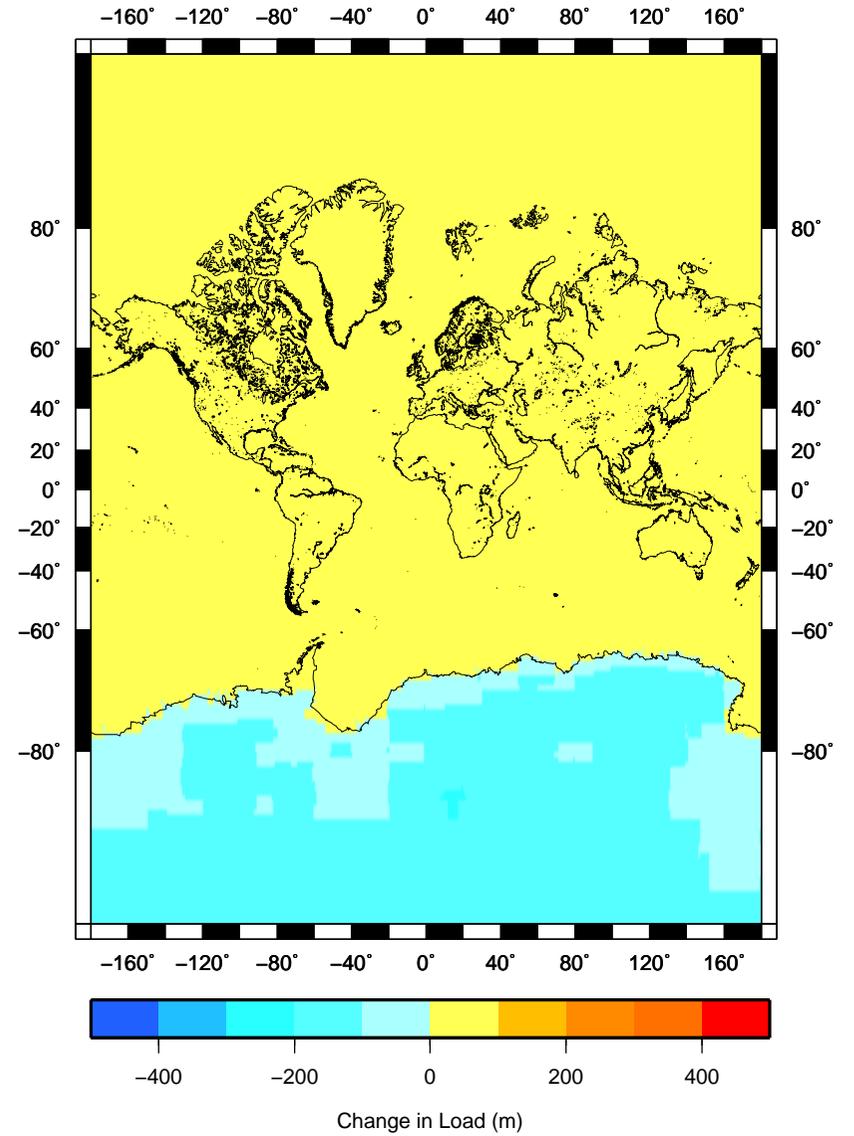
Change in ice load at 120 kyr BP for Antarctica experiment



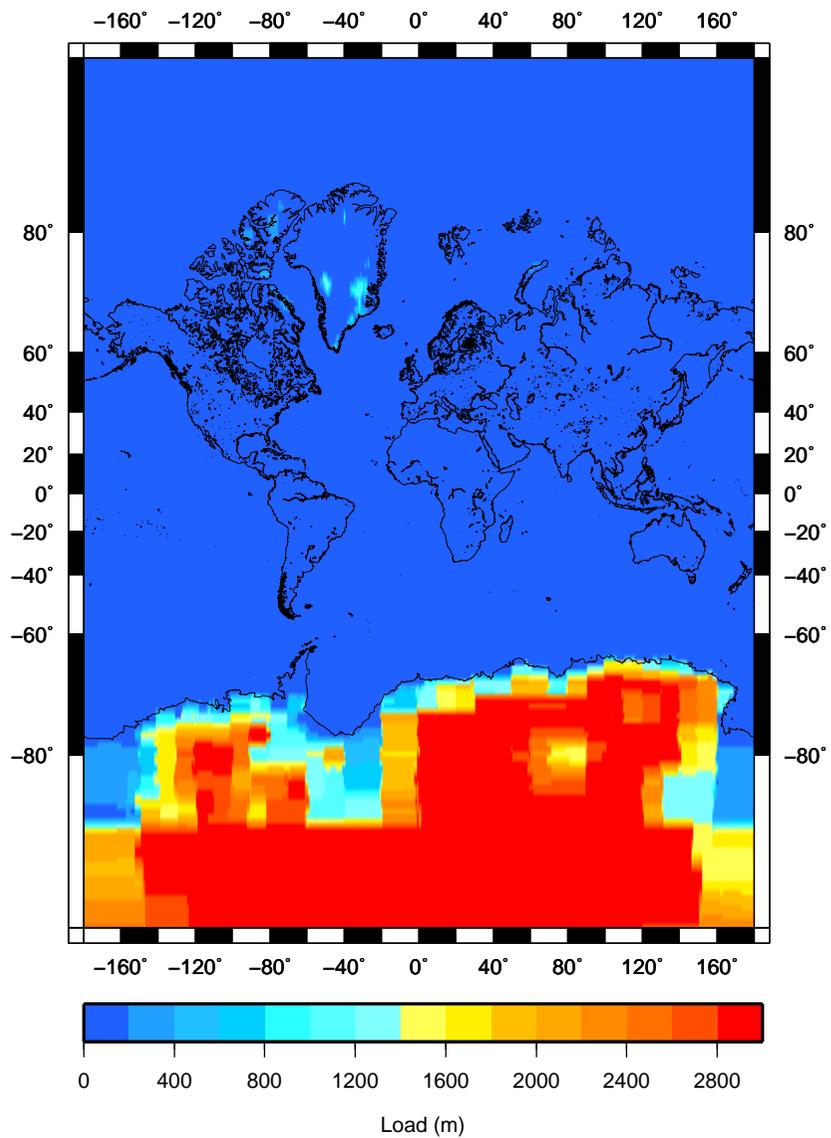
Ice load at 119 kyr BP for Antarctica experiment



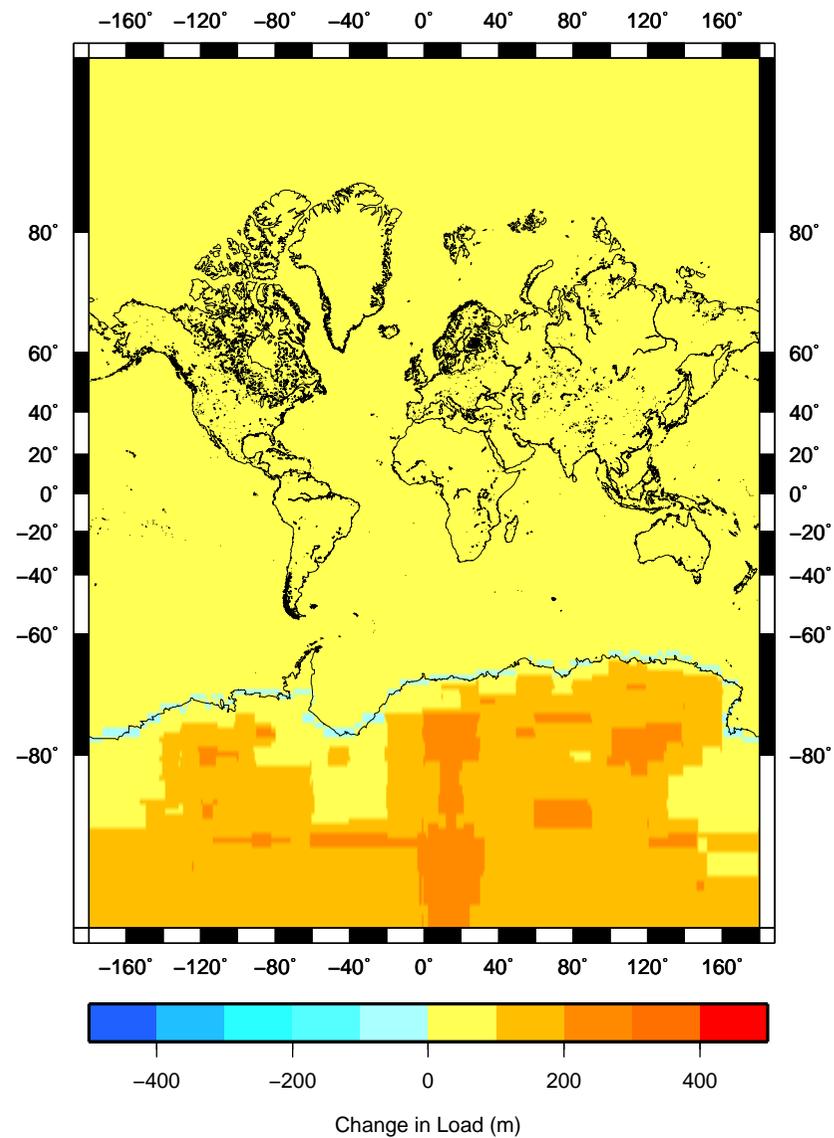
Change in ice load at 119 kyr BP for Antarctica experiment



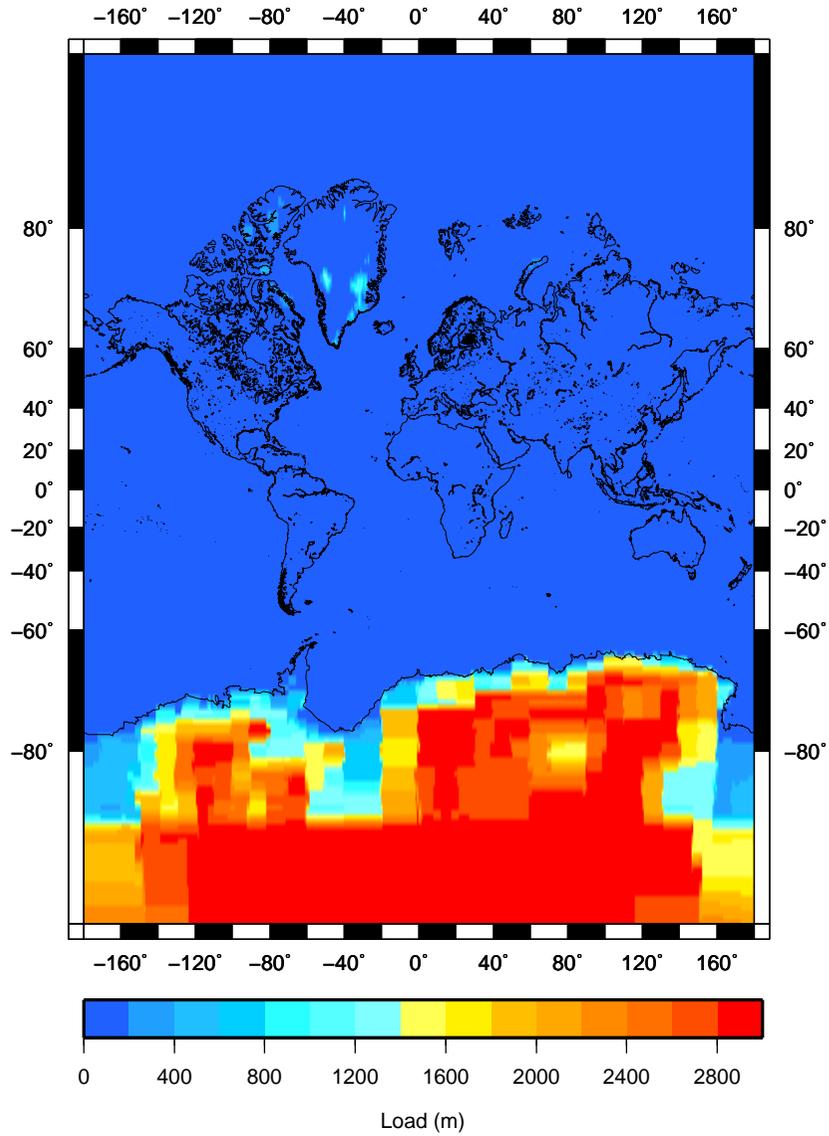
Ice load at 118 kyr BP for Antarctica experiment



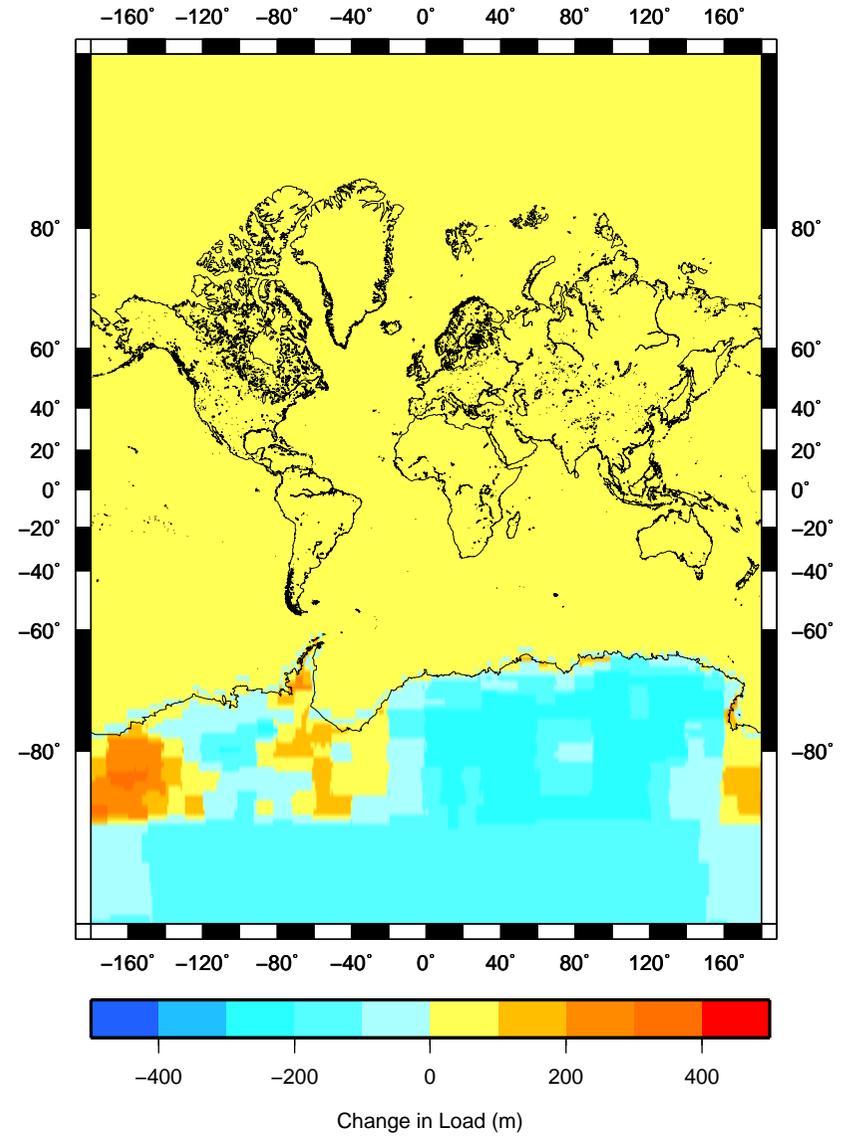
Change in ice load at 118 kyr BP for Antarctica experiment



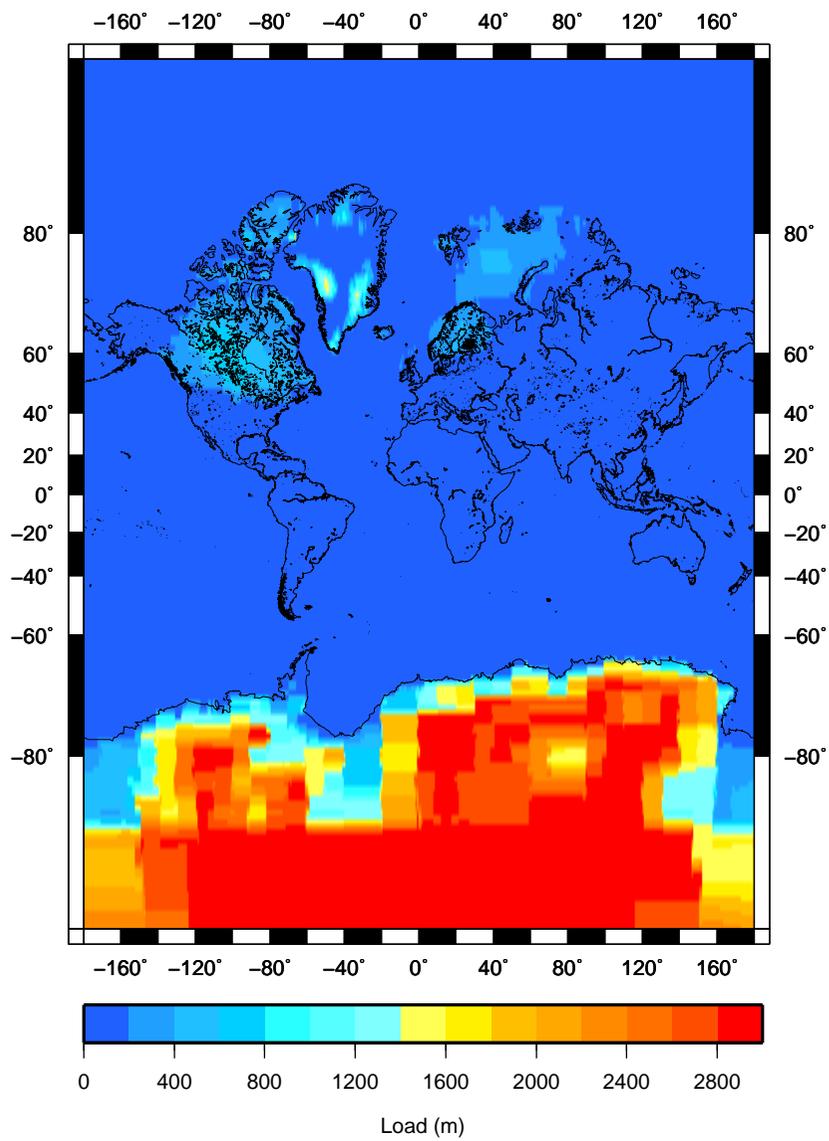
Ice load at 117 kyr BP for Antarctica experiment



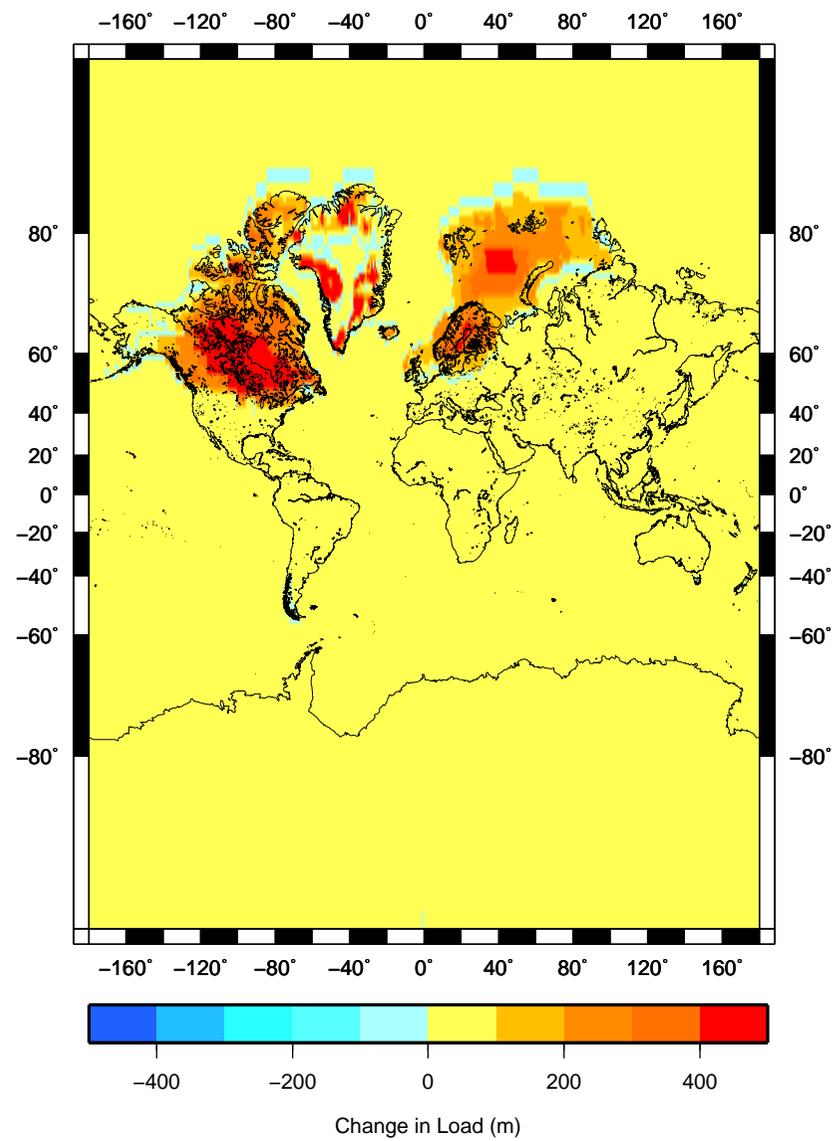
Change in ice load at 117 kyr BP for Antarctica experiment



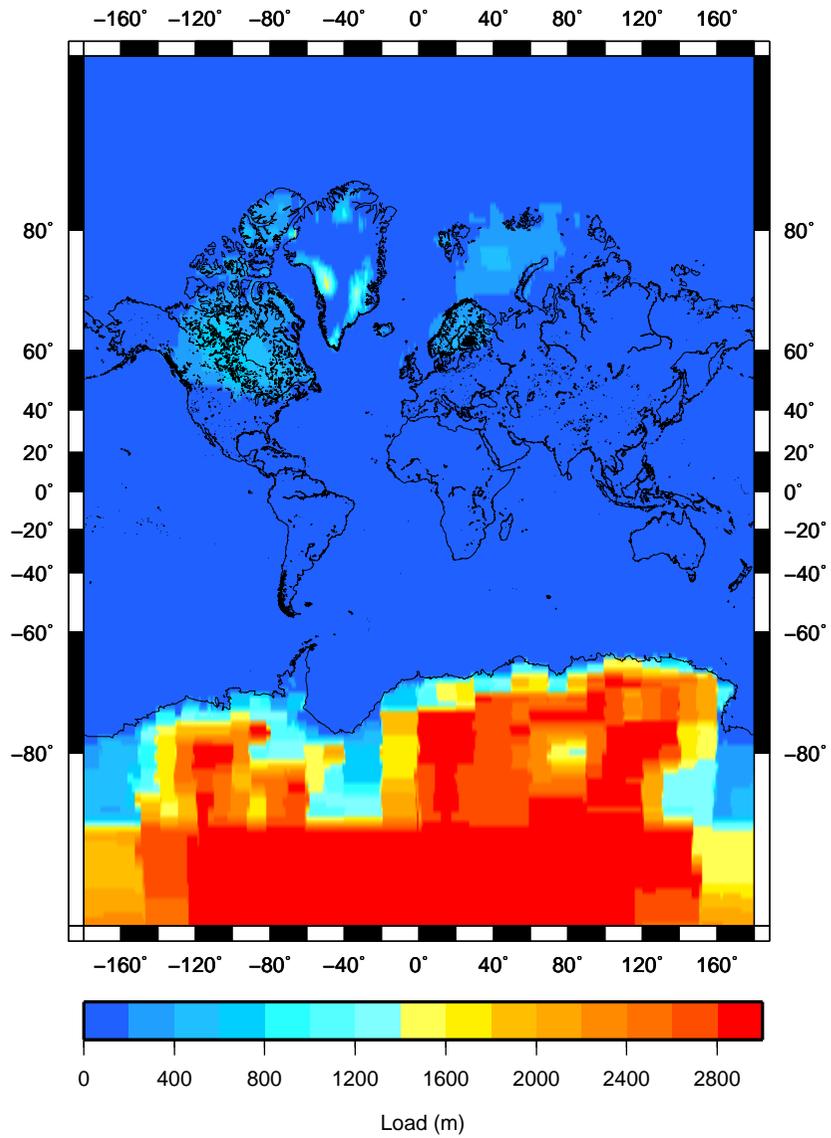
Ice load at 116 kyr BP for Antarctica experiment



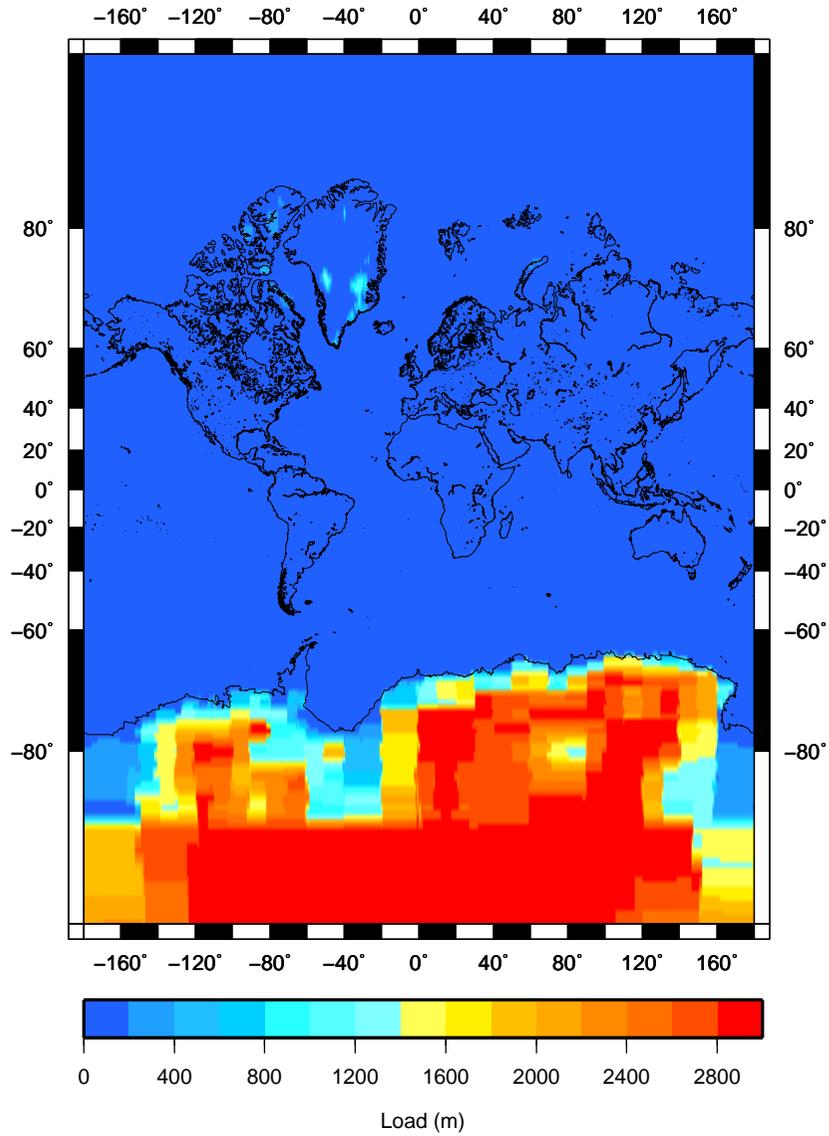
Change in ice load at 116 kyr BP for Antarctica experiment



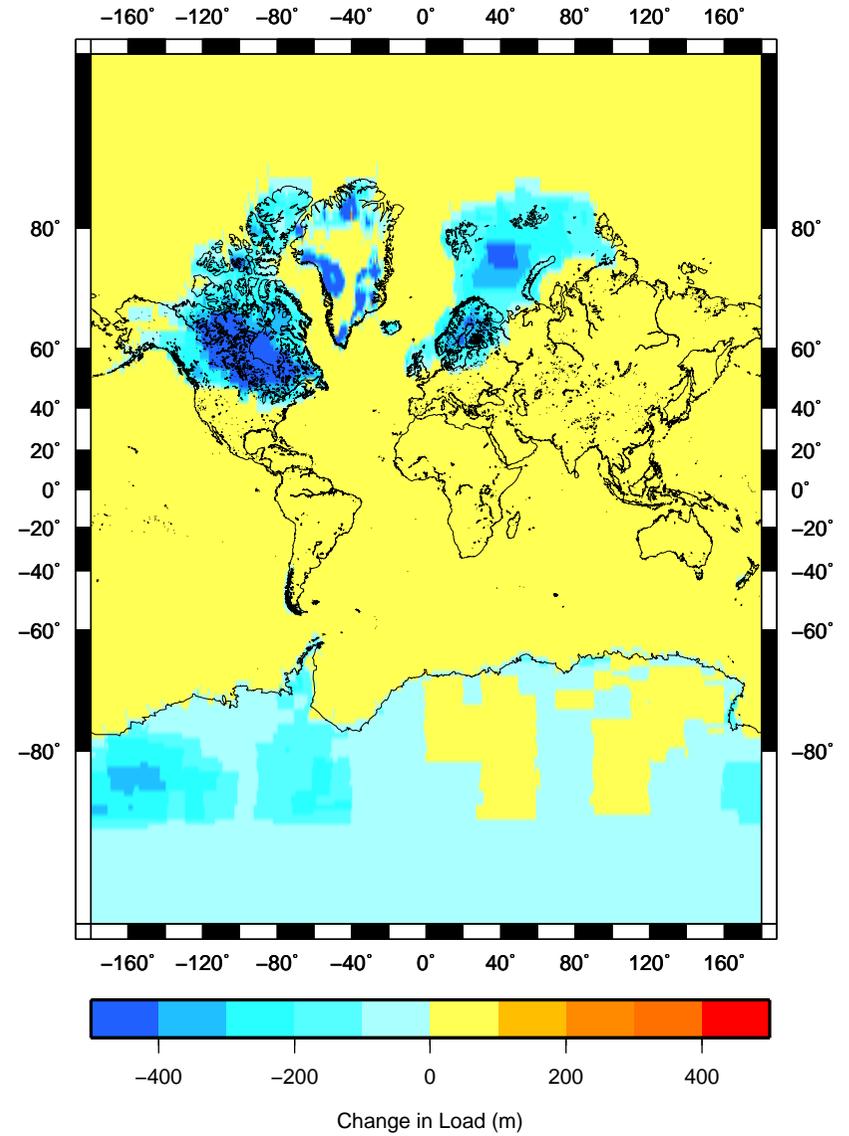
Ice load at 132 kyr BP for Greenland experiment



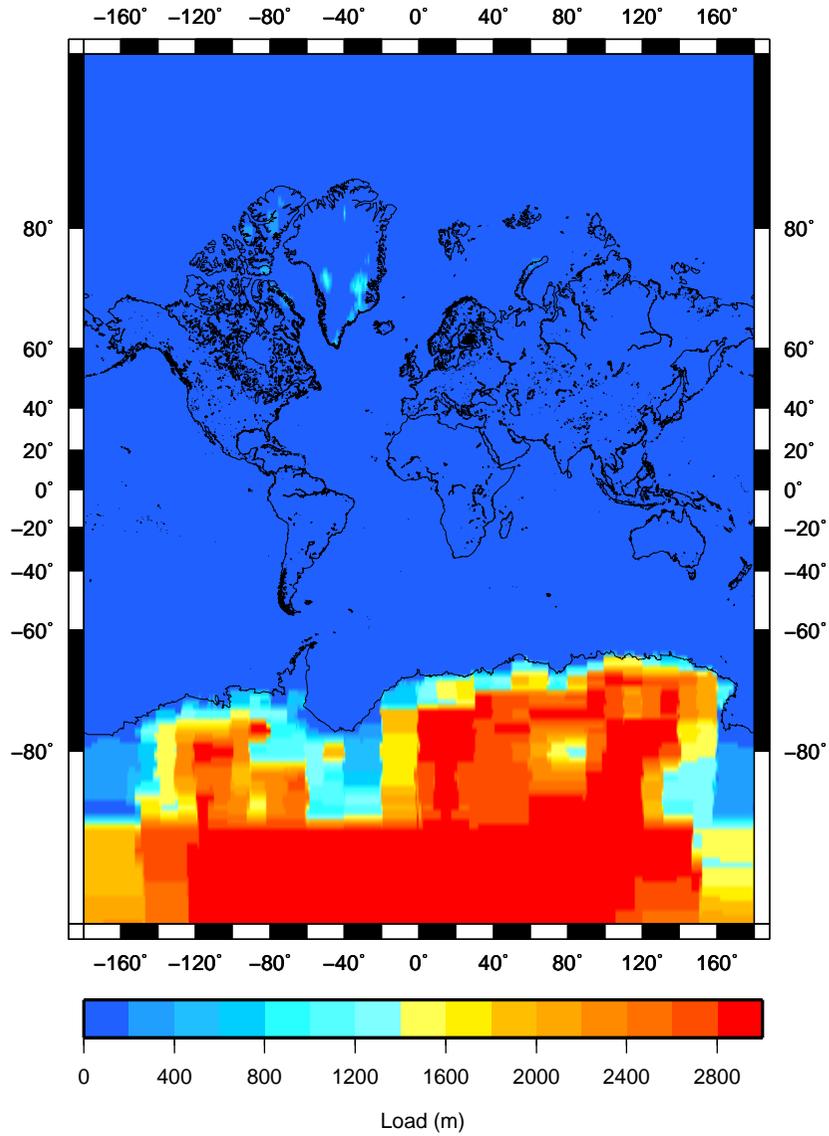
Ice load at 127 kyr BP for Greenland experiment



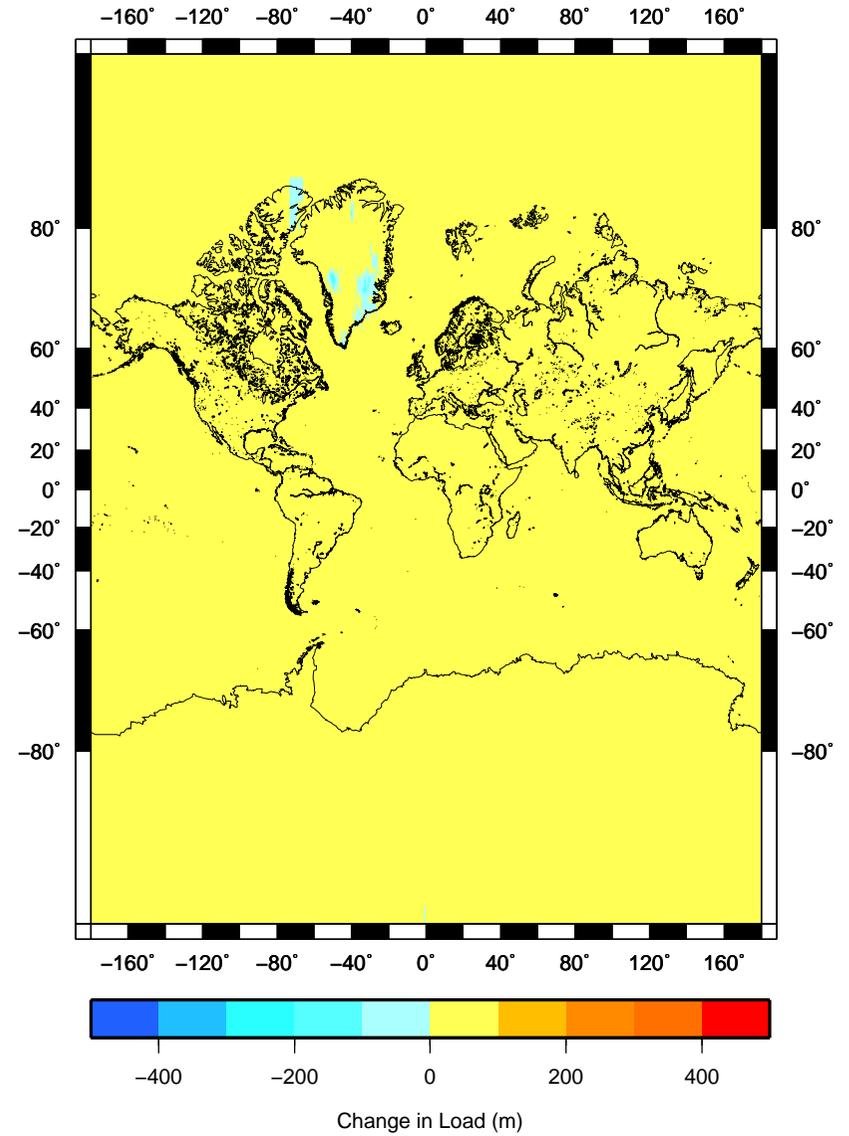
Change in ice load at 127 kyr BP for Greenland experiment



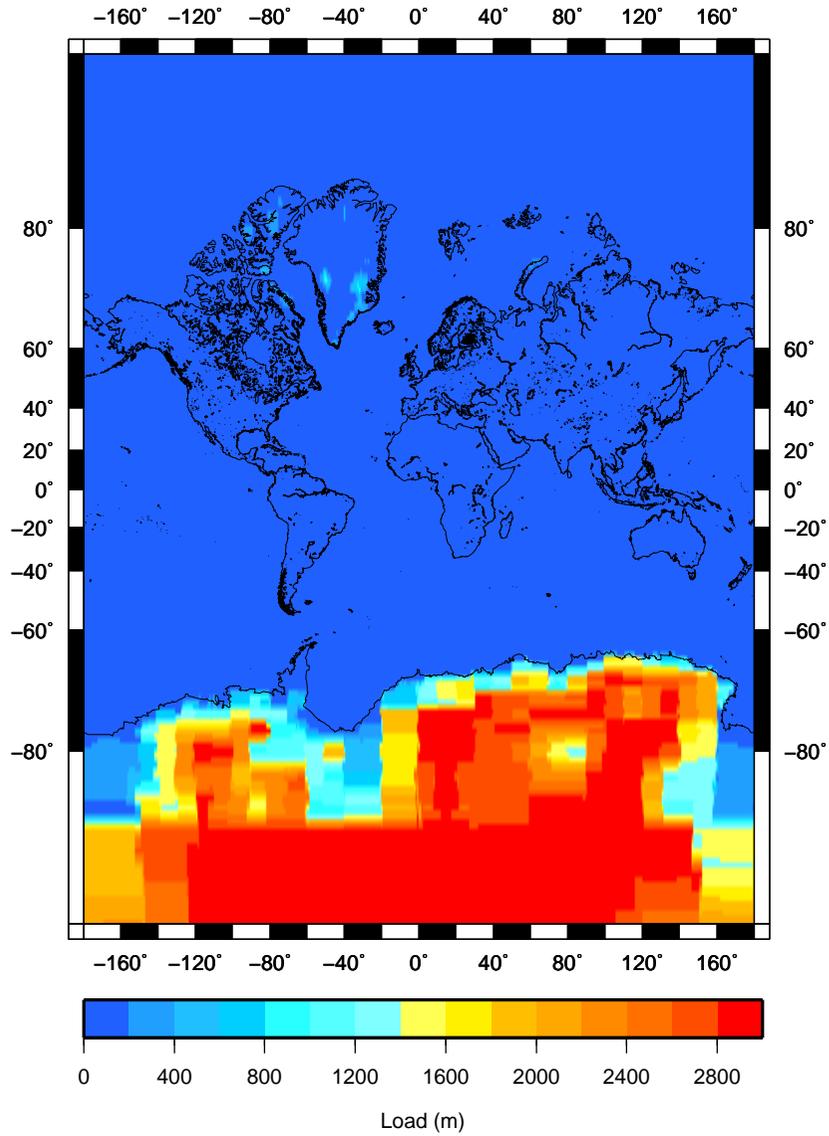
Ice load at 126 kyr BP for Greenland experiment



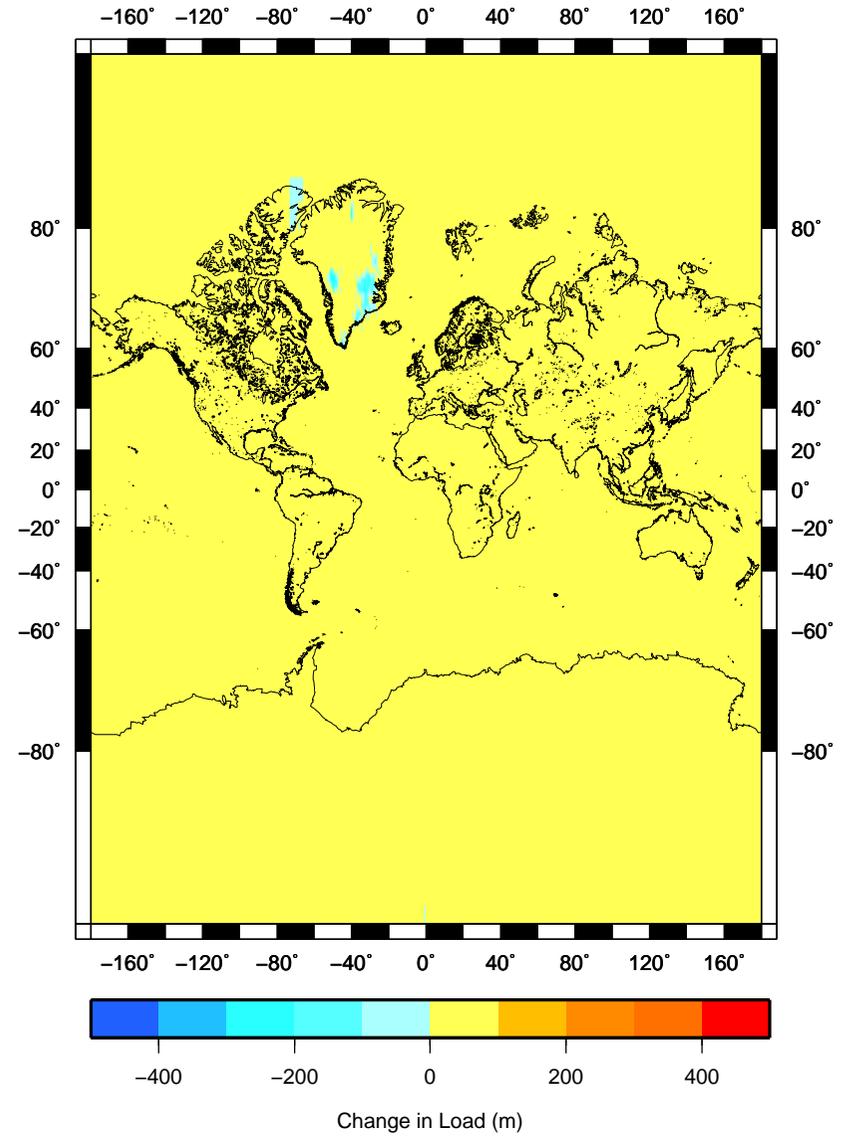
Change in ice load at 126 kyr BP for Greenland experiment



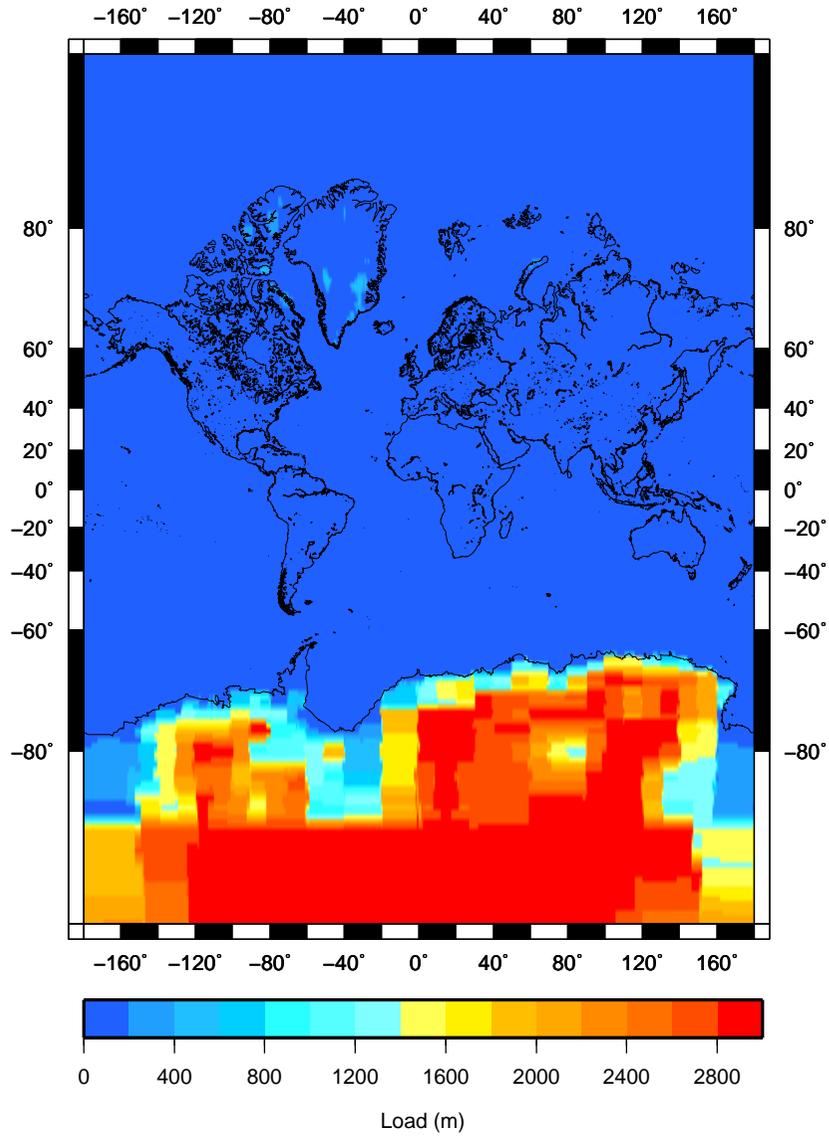
Ice load at 125 kyr BP for Greenland experiment



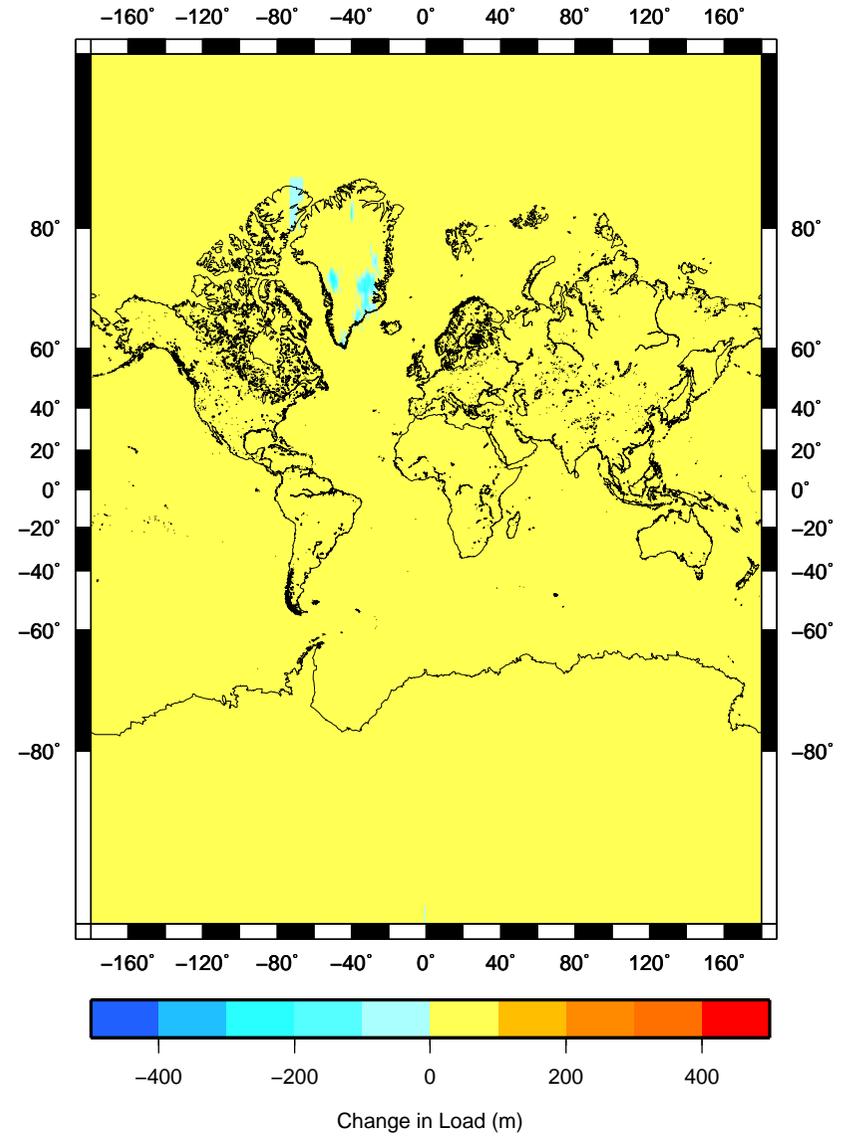
Change in ice load at 125 kyr BP for Greenland experiment



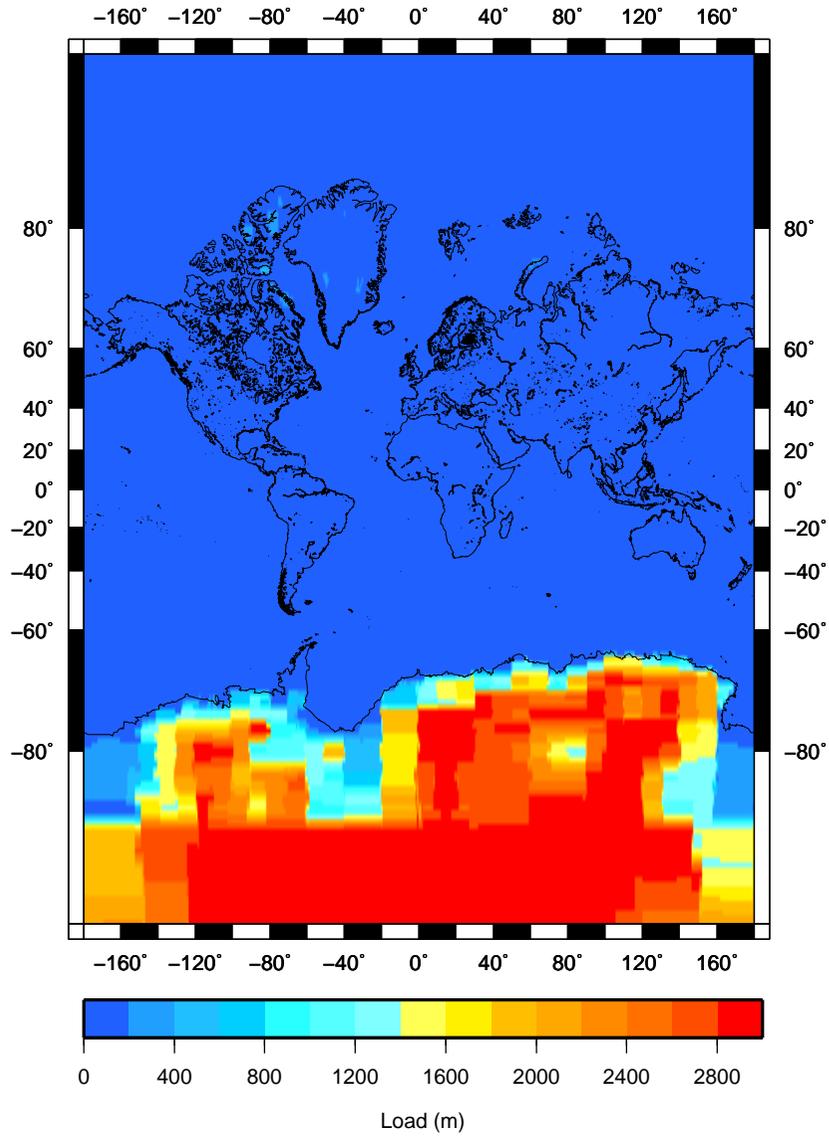
Ice load at 124 kyr BP for Greenland experiment



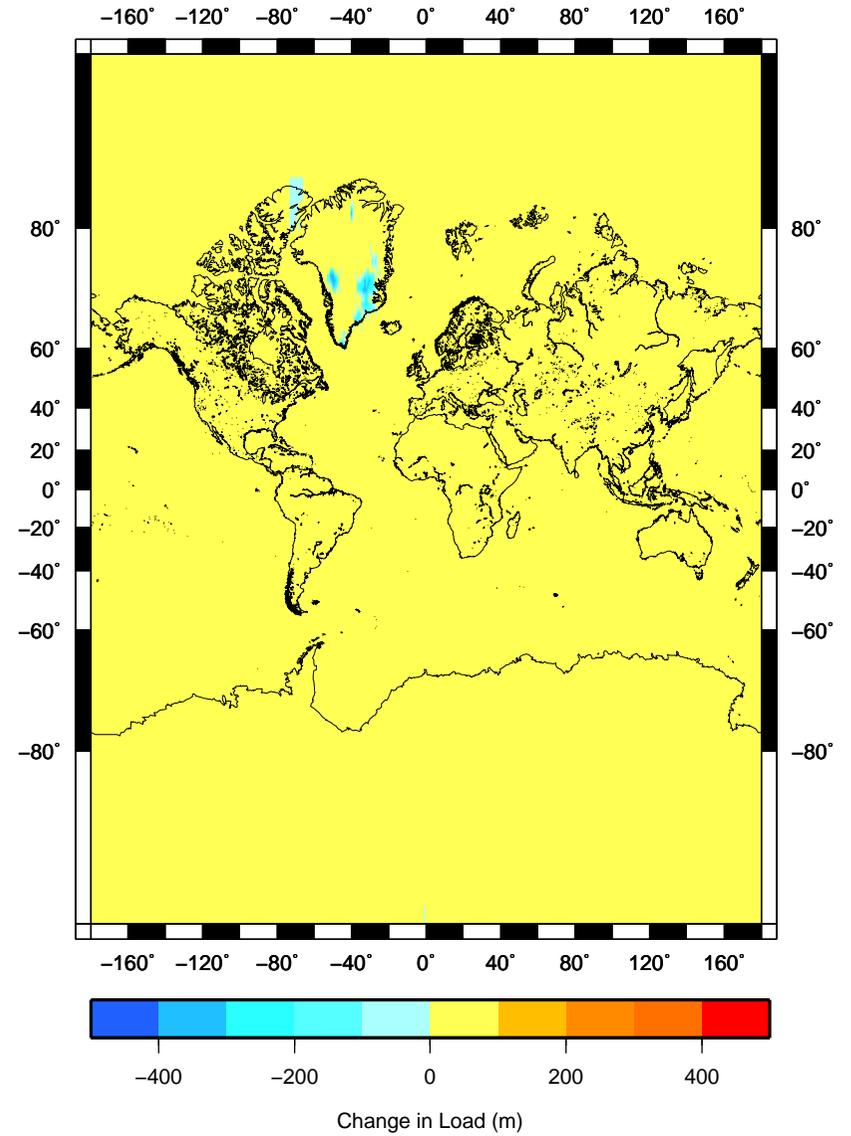
Change in ice load at 124 kyr BP for Greenland experiment



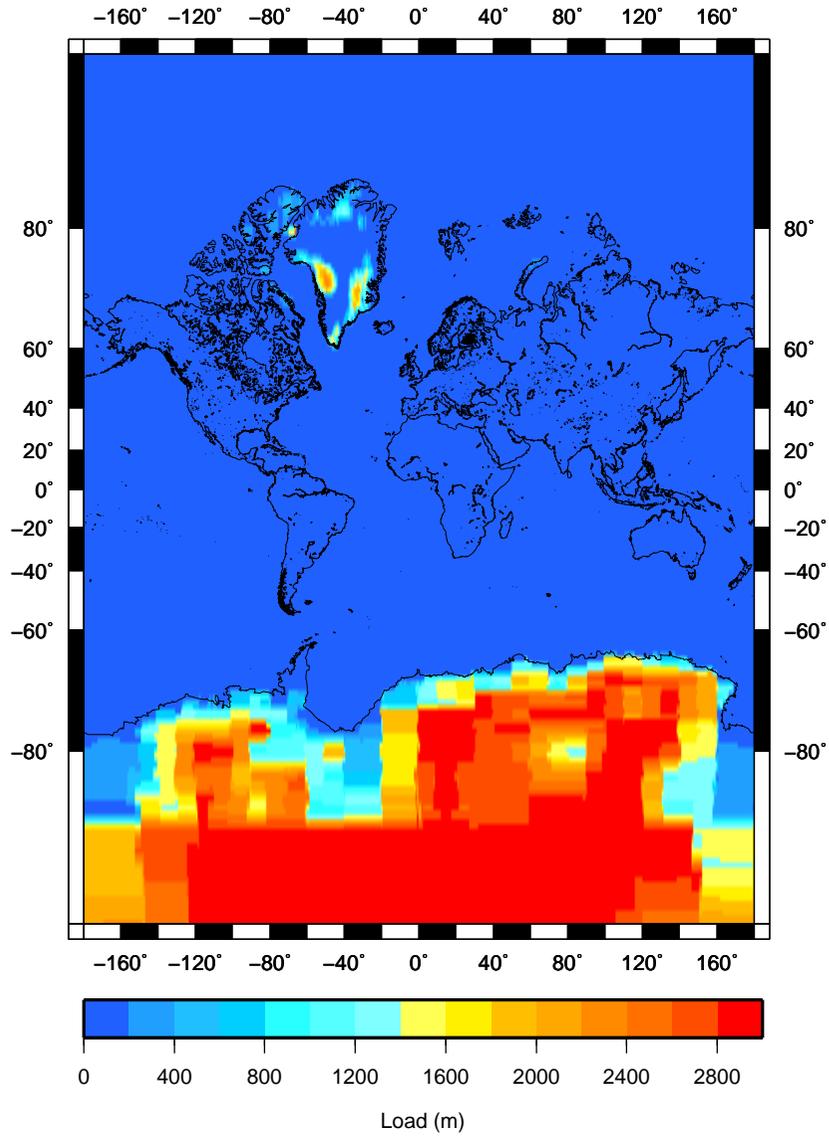
Ice load at 123 kyr BP for Greenland experiment



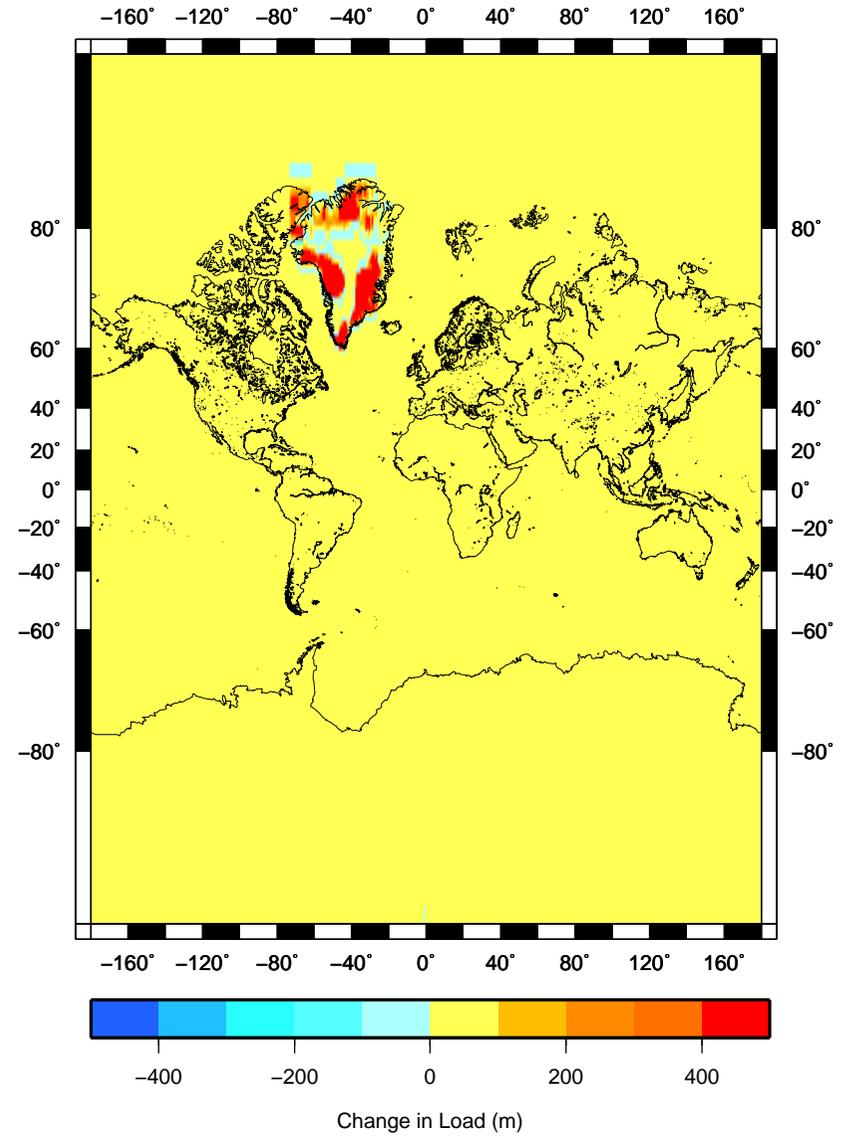
Change in ice load at 123 kyr BP for Greenland experiment



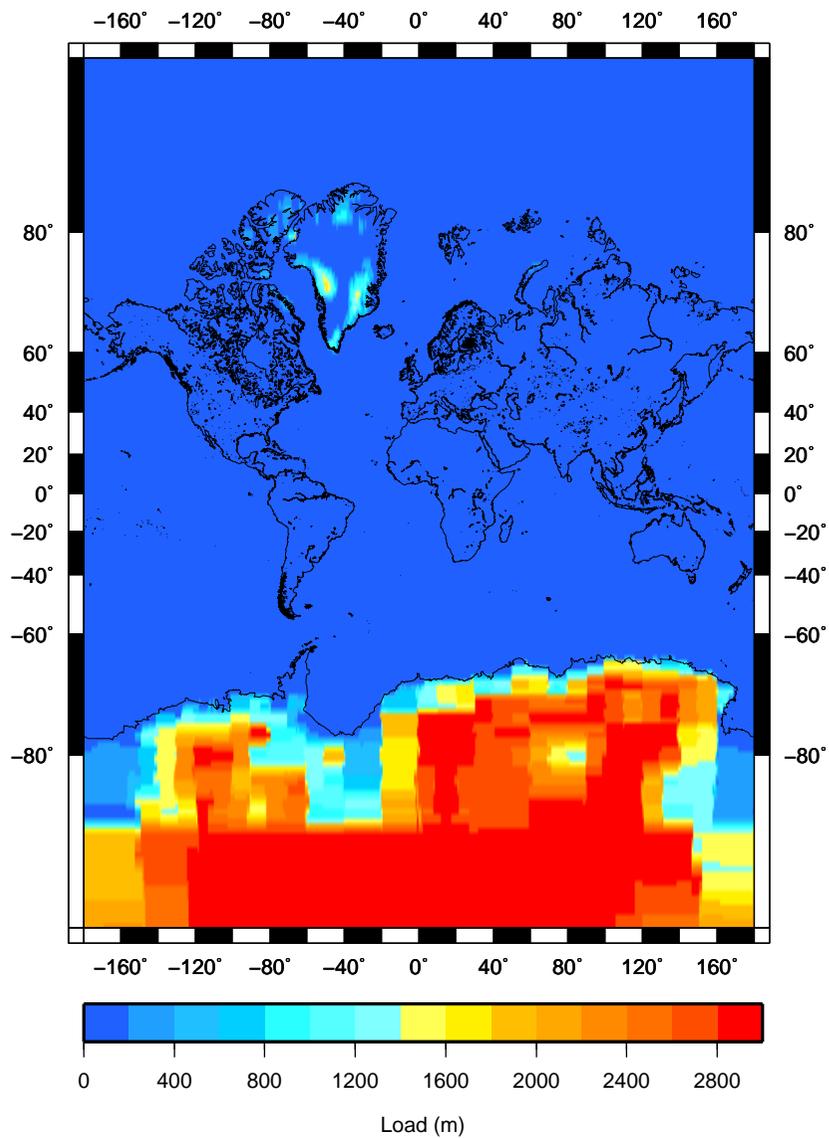
Ice load at 122 kyr BP for Greenland experiment



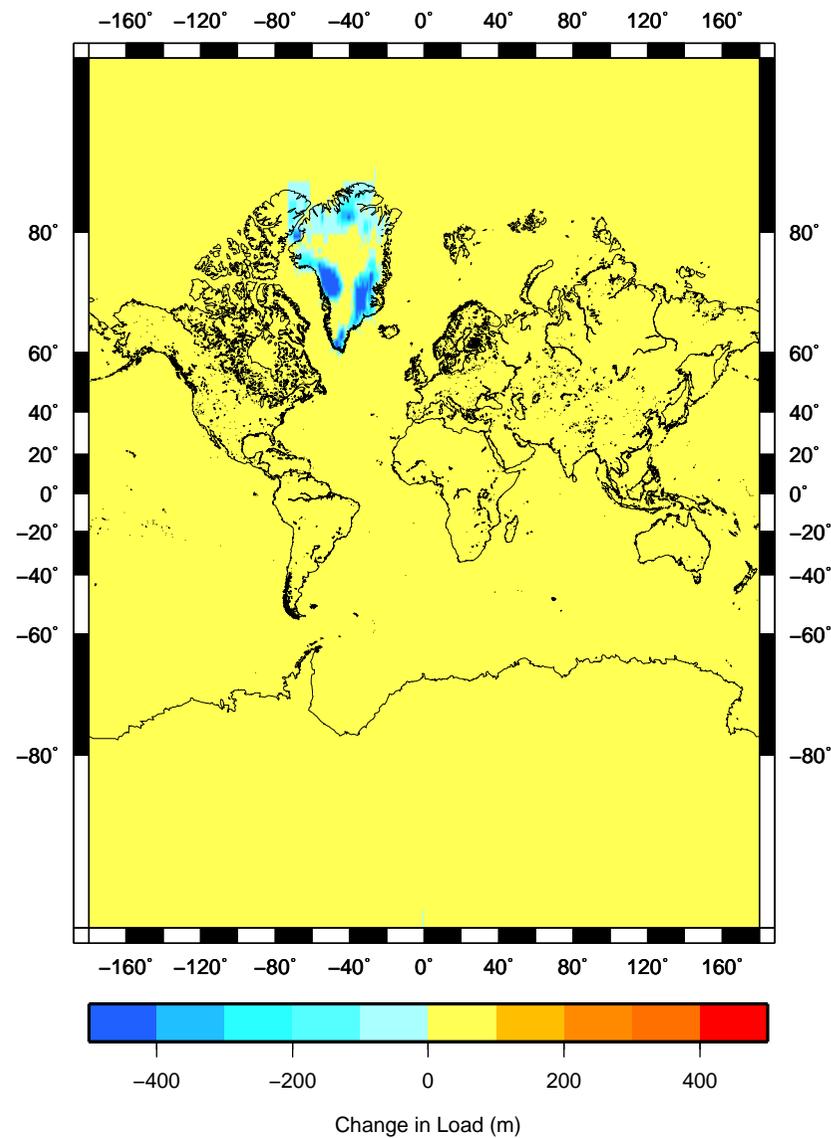
Change in ice load at 122 kyr BP for Greenland experiment



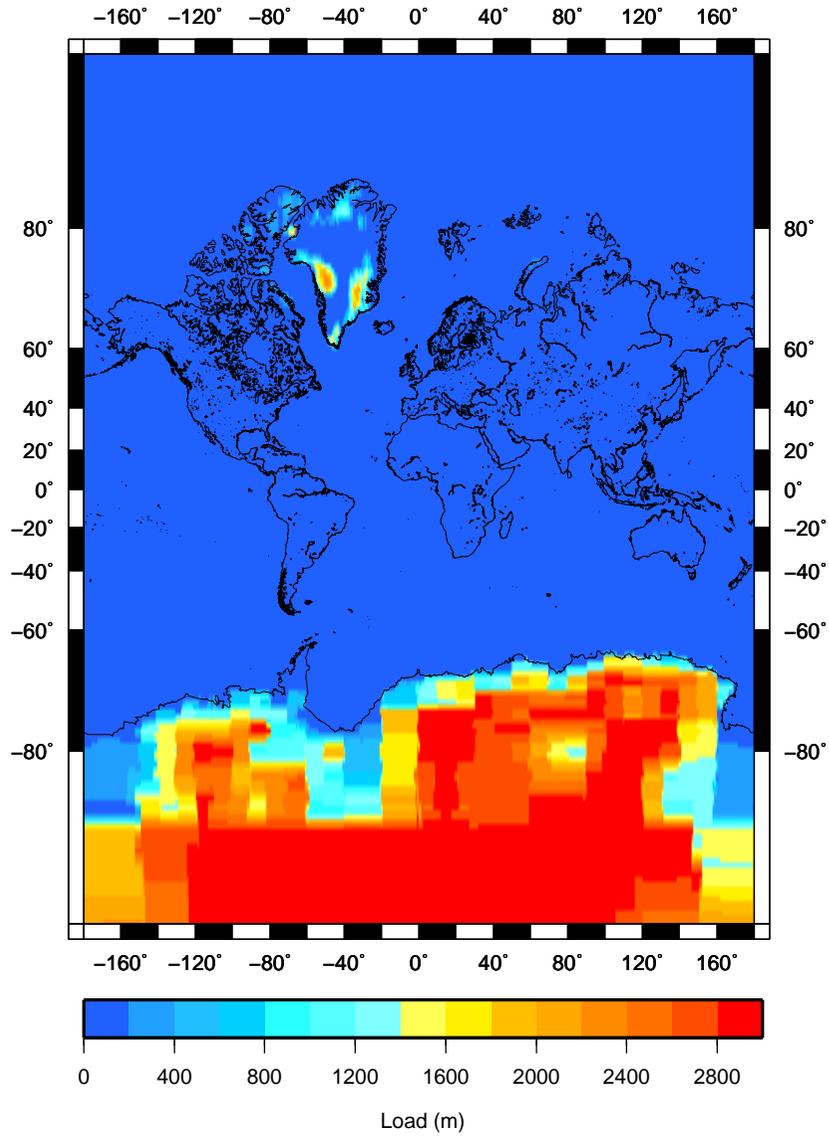
Ice load at 121 kyr BP for Greenland experiment



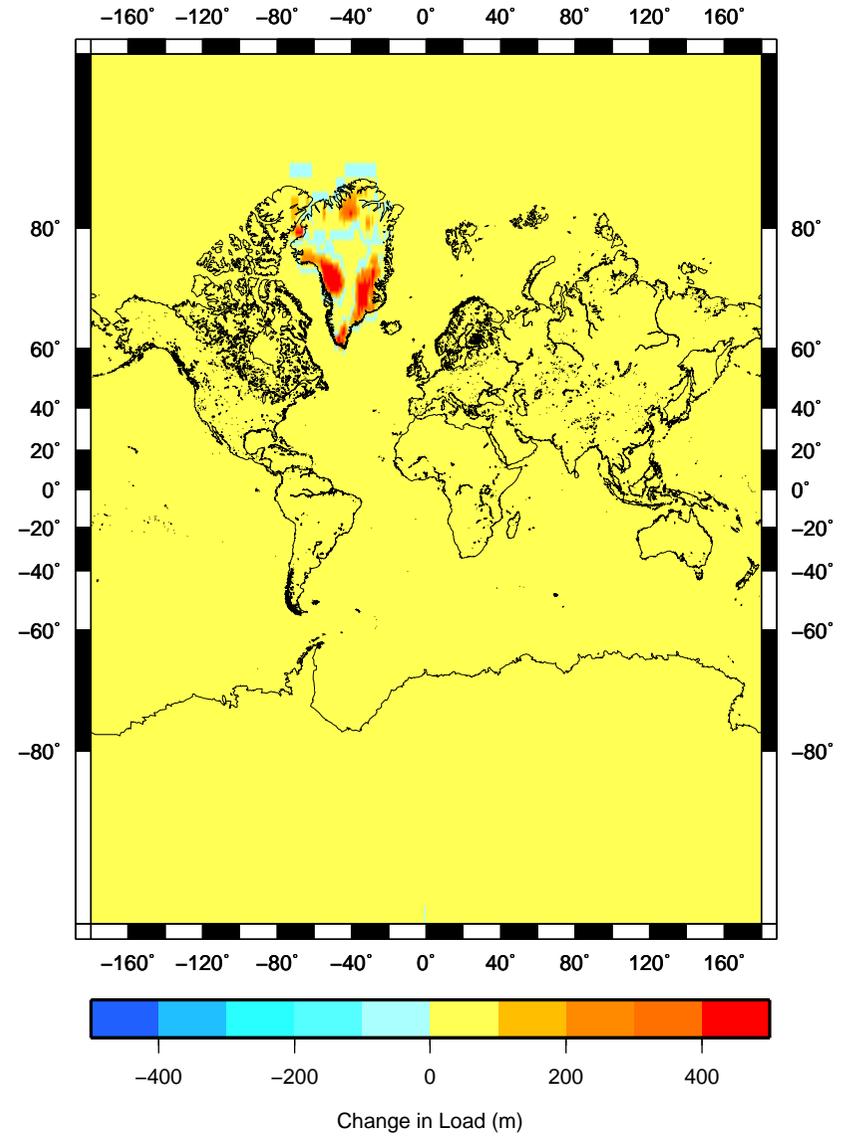
Change in ice load at 121 kyr BP for Greenland experiment



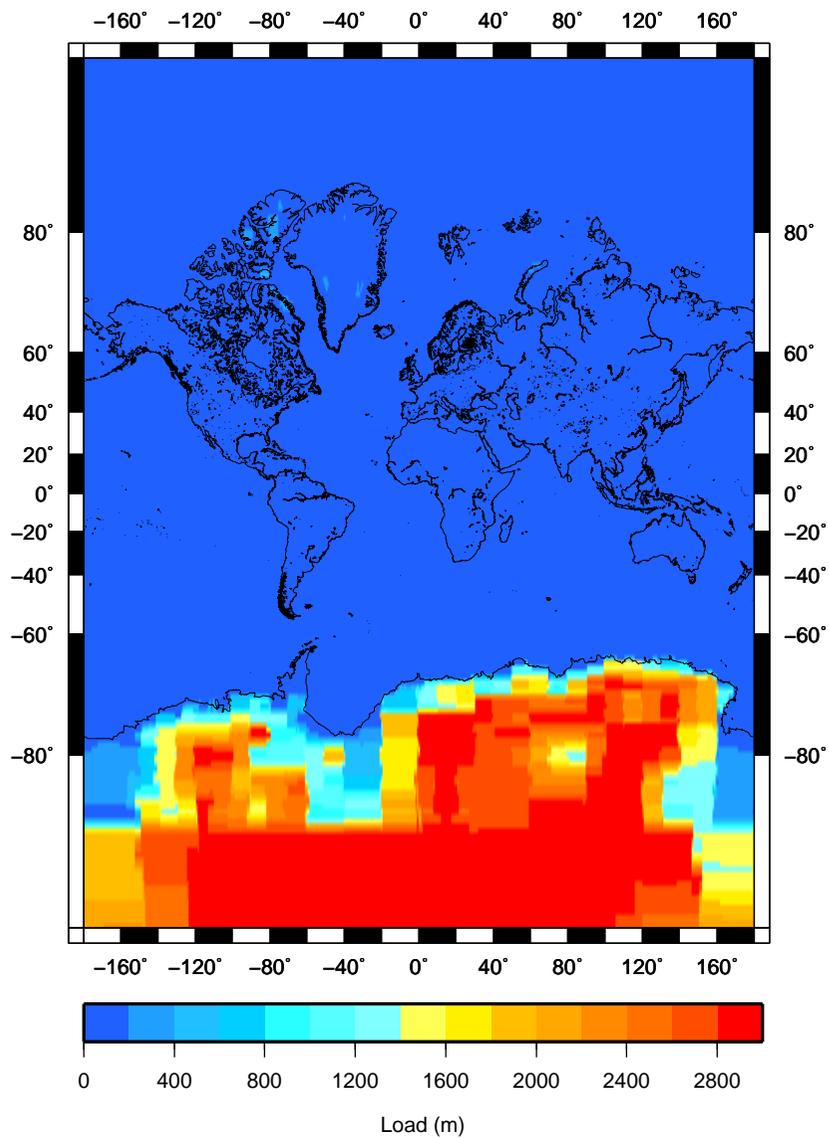
Ice load at 120 kyr BP for Greenland experiment



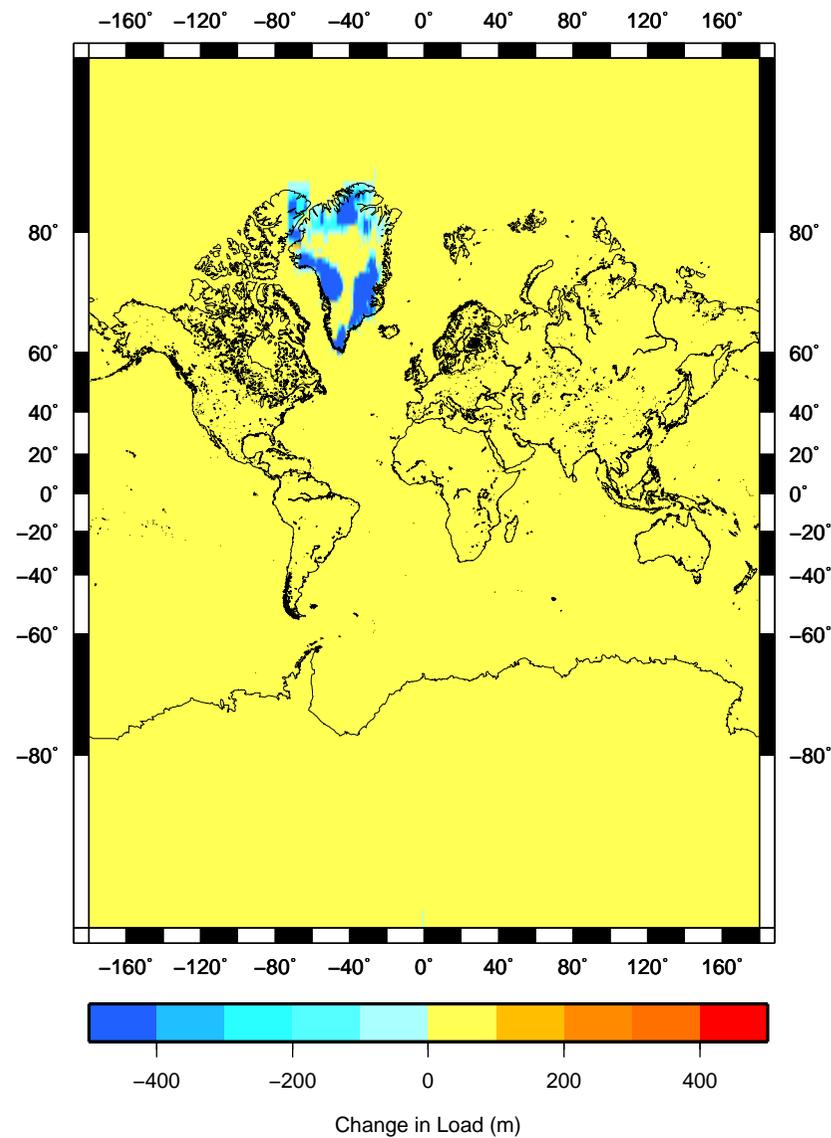
Change in ice load at 120 kyr BP for Greenland experiment



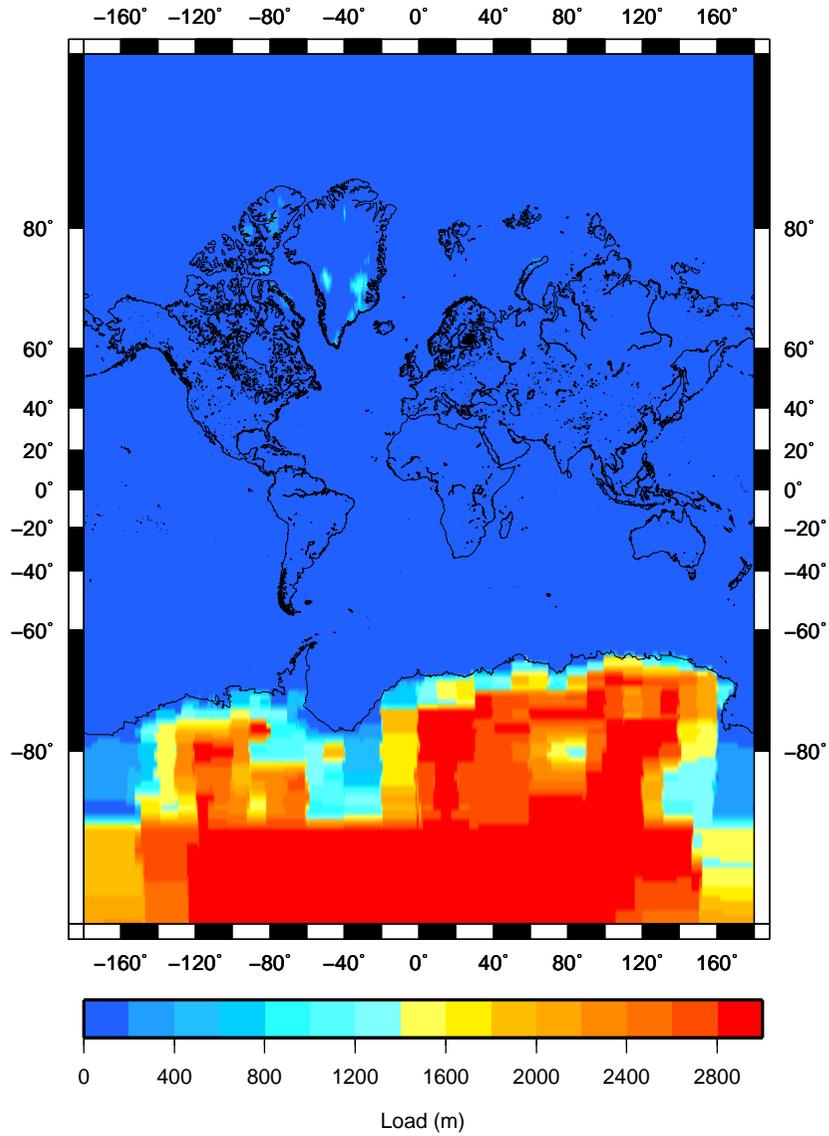
Ice load at 119 kyr BP for Greenland experiment



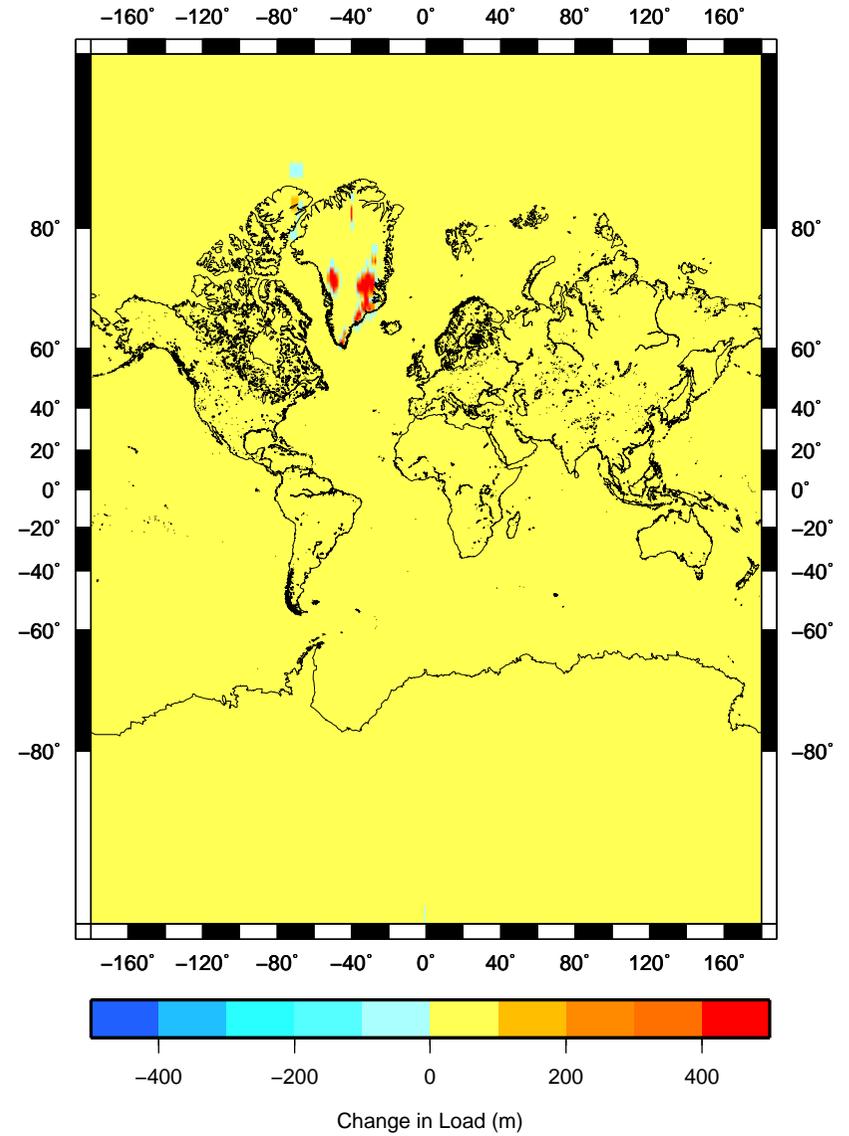
Change in ice load at 119 kyr BP for Greenland experiment



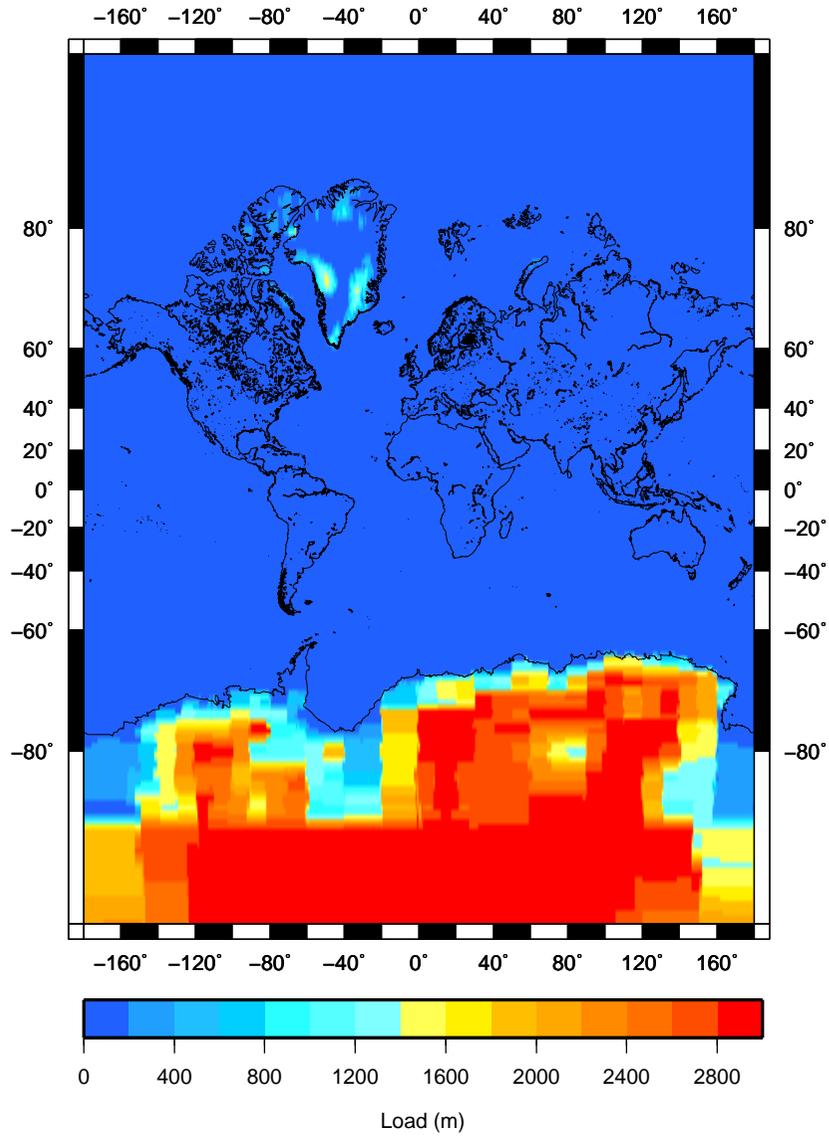
Ice load at 118 kyr BP for Greenland experiment



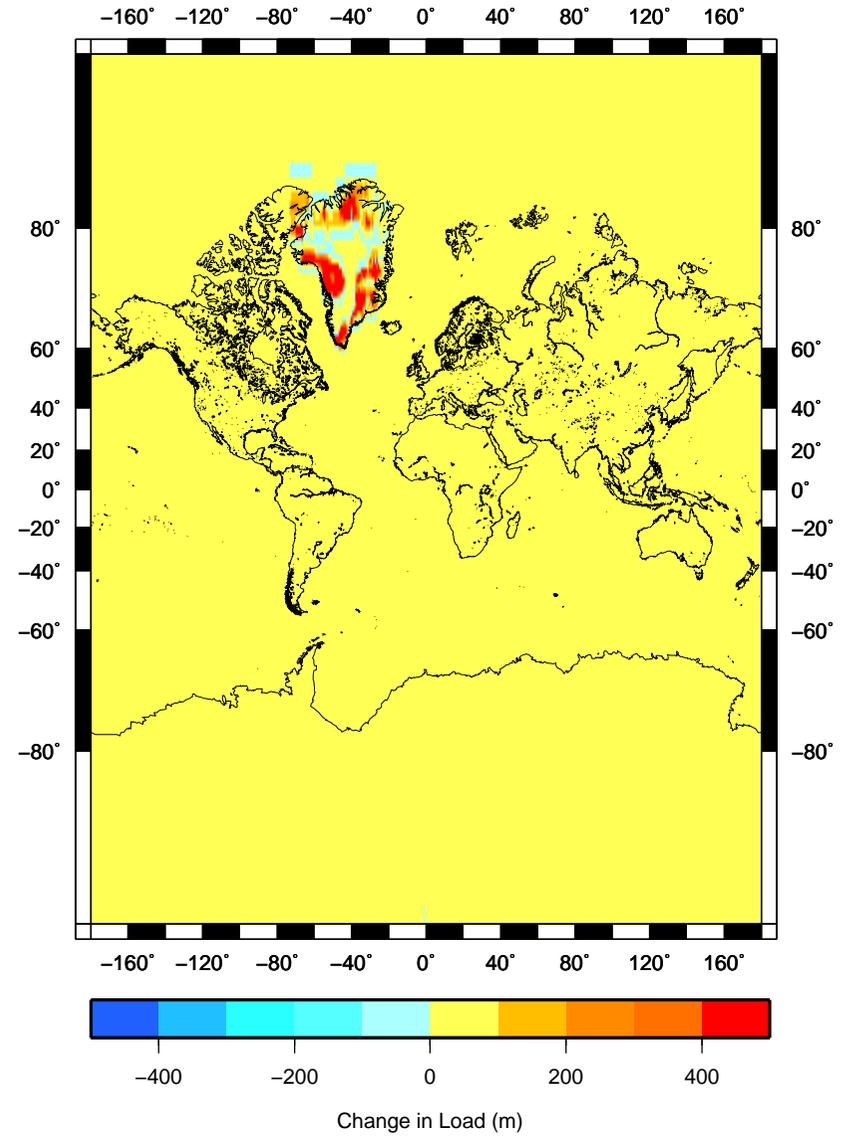
Change in ice load at 118 kyr BP for Greenland experiment



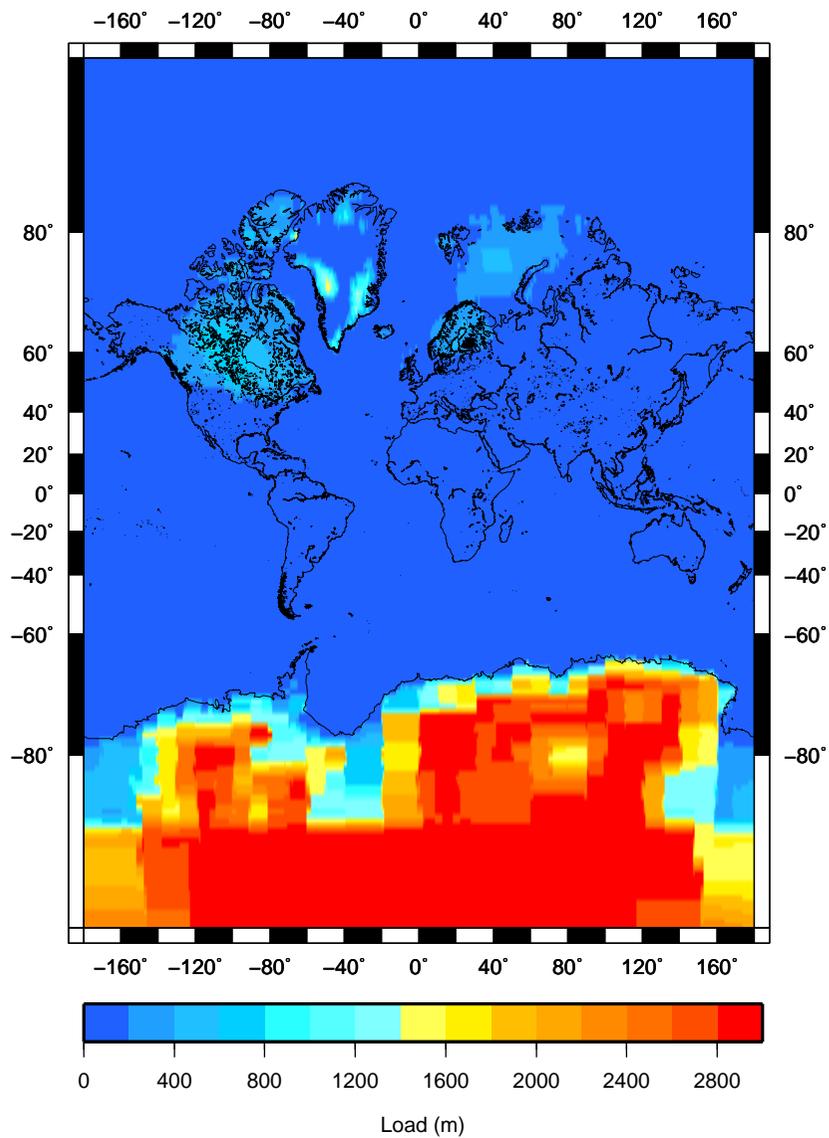
Ice load at 117 kyr BP for Greenland experiment



Change in ice load at 117 kyr BP for Greenland experiment



Ice load at 116 kyr BP for Greenland experiment



Change in ice load at 116 kyr BP for Greenland experiment

