

1 **Recent progress in understanding climate thresholds: ice sheets,**
2 **the Atlantic meridional overturning circulation, tropical forests and**
3 **responses to ocean acidification**

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22 **Abstract**

23

24 This article reviews recent scientific progress, relating to four major systems that could
25 exhibit threshold behaviour: ice sheets, the Atlantic meridional overturning circulation
26 (AMOC), tropical forests and ecosystem responses to ocean acidification. The focus is on
27 advances since the Intergovernmental Panel on Climate Change Fifth Assessment Report
28 (IPCC AR5). The most significant developments in each component are identified by
29 synthesizing input from multiple experts from each field. For ice sheets, some degree of

30 irreversible loss (time-scales of millennia) of part of the West Antarctic Ice Sheet (WAIS)
31 may have already begun, but the rate and eventual magnitude of this irreversible loss is
32 uncertain. The observed AMOC overturning has decreased from 2004-2014, but it is unclear
33 at this stage whether this is forced or is internal variability. New evidence from experimental
34 and natural droughts has given greater confidence that tropical forests are adversely affected
35 by drought. The ecological and socio-economic impacts of ocean acidification are expected
36 to greatly increase over the range from today's annual value of around 400, up to 650 ppm
37 CO₂ in the atmosphere (reached around 2070 under RCP8.5), with rapid development of
38 aragonite undersaturation at high latitudes affecting calcifying organisms. Tropical coral
39 reefs are vulnerable to the interaction of ocean acidification and temperature rise, and the
40 rapidity of those changes, with severe losses and risks to survival at 2K warming above pre-
41 industrial. Across the four systems studied, however, quantitative evidence for a difference
42 in risk between 1.5K and 2K warming above pre-industrial levels is limited.

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44

45 **I Introduction**

46

47 While some aspects of climate change can be viewed as becoming proportionately larger with
48 increasing forcings, other aspects may feature more complex, nonlinear behaviour (e.g.
49 Lenton et al., 2008). This can include abrupt and/or irreversible change, which may be
50 associated with key thresholds. Such behaviour must be considered differently in
51 assessments of the potential benefits of mitigation: it implies, for example, that certain
52 impacts could be significantly different above certain levels of anthropogenic interference,
53 although the impacts may not always be negative (Lenton, 2013). Further, thresholds involve
54 systems moving out of the limits of currently observed behaviour, so they require deep

55 process understanding based on a broad range of observations (Kopp et al., 2016). In some
56 cases, early warning of an approaching threshold may be possible (Lenton, 2011).

57

58 Clear evidence of threshold behaviour in the earth system is seen in the paleoclimate record
59 (e.g. McNeall et al., 2011). For example, central Greenland temperatures inferred from ice
60 cores show abrupt changes of 10°C within 100 years (Guillevic et al., 2013) during the last
61 ice age, even though the forcing over this period has evolved smoothly. These changes are
62 thought in part to be associated with changes in the ocean's thermohaline circulation
63 (Broecker, 2003). Strong paleoclimate evidence also exists for threshold behaviour in major
64 ice sheets and methane reservoirs (McNeall et al., 2011). Various different forms of abrupt
65 shifts have been found in current climate models (Drijfhout et al., 2015), although primarily
66 for different processes than explored in the present review.

67

68 This study focuses on four major systems that may feature threshold behaviour: ice sheets,
69 the Atlantic Meridional Overturning Circulation (AMOC), tropical forests, and ecosystem
70 responses to ocean acidification. The risk of significant change in these systems is not
71 necessarily linked to large-scale warming alone. Patterns of precipitation can be important
72 for the AMOC and tropical forests; tropical forests are also strongly affected by
73 anthropogenic land-use and the direct effect of carbon dioxide, while ocean acidification
74 arises directly from increased atmospheric CO₂ (although its impacts combine with those of
75 ocean warming) and West Antarctic Ice Sheet (WAIS) stability is influenced by changes in
76 ocean circulation.

77

78 Consequences of change in these systems range from amplified global warming through
79 altered climate patterns, elevated sea-level and direct loss of biodiversity and ecosystem

80 services (see individual sections below for details). These systems can in principle interact
81 with each other (Lenton et al., 2008), although this is explored in only a few studies.

82

83 Here we report primarily on new literature subsequent to that presented in the IPCC Fifth
84 Assessment Report, AR5. As found by O’Neil et al. (2017) in an updated review of IPCC
85 Reasons for Concern, the headline conclusions of AR5 still broadly hold, but there have been
86 considerable advances in understanding. We also briefly consider (in the Conclusions) the
87 difference in risk between 1.5K and 2K global mean warming above pre-industrial levels.

88 This review was prepared using an iterative approach, by specialists both within and external
89 to the Met Office Hadley Centre. Initial drafts of each section were prepared by the Met
90 Office, then sent to external experts for review and editing (except that for Ocean
91 Acidification; prepared by experts in Plymouth Marine Laboratory and University of East
92 Anglia). The sections were revised accordingly by the Met Office, then sent to the external
93 experts for a second review.

94

95 Each system is addressed in a separate section below, each with the following subsections:
96 Introduction (the key issues for that system); Observations (relevant real-world observations);
97 Potential for significant change (literature addressing the question of how likely substantial
98 change is); Consequences (of significant change); Cautions (key scientific uncertainties); and
99 Comparison with AR5. The key conclusions are summarised in Table 1.

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105 **II Ice Sheets**

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107 **1 Introduction**

108

109 Ice-sheet mass loss, from the Greenland and Antarctic ice sheets, is of concern due to its
110 potential impact on global and local sea level (Alley et al., 2005; Shepherd, 2012), and
111 potential amplification of global warming over long timescales as low-albedo land surface is
112 exposed (Hansen et al., 2008). The ice sheets are the largest potential source of future sea
113 level rise on decadal to millennial timescales. Greenland and Antarctica respectively contain
114 enough ice to raise mean sea level by 7.4 m and 58.3m (Vaughan et al., 2013). In addition,
115 rapid mass loss may have an influence on ocean circulation through a change in salinity and
116 hence density gradients (Yang et al., 2016).

117

118 In a state of equilibrium, an ice sheet loses mass (through surface melting, and the calving
119 and submarine melting of its outlet glaciers and ice shelves, Rignot et al., 2010; Depoorter et
120 al., 2013), at the same rate as it gains mass (through the accumulation of snowfall, Alley et
121 al., 2005). Increased ice sheet mass loss occurs through two main mechanisms. Increased
122 surface melt is largely driven by higher air temperatures and currently affects Greenland and
123 the Antarctic Peninsula. ‘Dynamic thinning’ (i.e. losses due to increased solid ice discharge
124 into the ocean) involves glacier acceleration and consequent increases in iceberg calving for
125 marine-terminating glaciers and ice shelves, occurring at the fringes of Greenland (Pritchard
126 et al., 2009), the Antarctic Peninsula (Wouters et al., 2015) and West Antarctica (Pritchard et
127 al., 2009; Bingham et al., 2012). This may be induced by increased temperature of the water
128 beneath floating ice or at the glacial terminus (Gille, 2014).

129

130 Ice shelves, the floating portions of outlet glaciers, play a key role in modulating the mass
131 balance of the Antarctic ice sheet. They buttress the inland glaciers, controlling the rate of
132 ice leaving the continent and entering the ocean (Dupont and Alley, 2005). Ice shelves are
133 exposed to the underlying ocean and may weaken (Furst et al., 2016) as ocean temperatures
134 rise (Depoorter et al., 2013). If they melt rapidly or break away, ice flow can accelerate,
135 causing net ice-sheet mass loss (De Angelis and Skvarca, 2003; Nick et al., 2009) which adds
136 to sea level rise. This impact of ice shelf break-up occurred on Larsen-B on the Antarctic
137 Peninsula in 2002, with the consequent acceleration of the glaciers as buttressing was
138 removed (De Rydt et al., 2015). Much of the West Antarctic Ice Sheet (WAIS) is grounded
139 on bedrock below sea level on retrograde slopes (deeper inland). This configuration is
140 inherently unstable and sensitive to small changes at the grounding line (where the ice begins
141 to float; Mercer, 1968; Schoof, 2007; Durand et al., 2011; Gudmundsson et al., 2012). A
142 small retreat of the grounding line resting on a retrograde slope thickens the ice at the
143 grounding line, in turn increasing the ice flux and inducing further retreat, and so on, until a
144 prograde slope is reached. Hence local thresholds exist (where the grounding line retreats to
145 a retrograde slope). This is known as the Marine Ice Sheet Instability (MISI), and
146 simulations have shown that this is a mechanism for rapid collapse of the WAIS (Gladstone
147 et al., 2012; Cornford et al., 2015; DeConto and Pollard, 2016; Arthern and Williams, 2017).

148

149 Iceberg calving has been implicated in the retreat and acceleration of glaciers where ice
150 shelves have disintegrated along the margins of the Greenland and Antarctic ice sheets,
151 indicating that they may be vulnerable to rapid ice loss through catastrophic disintegration
152 (Bassis and Jacobs, 2013). Processes such as fracture propagation in response to local stress
153 imbalances in the immediate vicinity of the glacier front; undercutting of the glacier terminus

154 by melting at or below the waterline; and bending at the junction between grounded and
155 buoyant parts of an ice tongue combine to generate a feedback which accelerates mass loss
156 through increased iceberg calving (Bassis and Walker, 2012). This is known as the Marine
157 Ice Cliff Instability (MICI).

158

159 Surface meltwater stored in ponds and crevasses can weaken and fracture ice shelves,
160 triggering their rapid disintegration (Scambos et al., 2004). This ice-shelf collapse results in
161 an increased flux of ice from adjacent glaciers. This mechanism has been included in one
162 model (Pollard et al., 2015), predicting a sea level rise from Antarctica of around 1m by 2100
163 (Deconto and Pollard, 2016). However, there is uncertainty in this process due to additional
164 effects from surface transport of meltwater onto, across and away from ice shelves. The net
165 result of this transport could either increase or decrease ice-shelf stability (Bell et al., 2017;
166 Kingslake et al., 2017).

167

168 It is thought that no ice sheet would grow in Greenland if the current one were to be removed,
169 even without human-induced warming, and hence it is a “relict” from the last Glacial Cycle
170 that ended about 12 thousand years ago. The altitude of the ice sheet interior maintains the
171 persistently cold temperatures required for the ice sheet to survive. There is a temperature
172 threshold above which the Greenland ice sheet is no longer viable (Gregory and Huybrechts,
173 2006; Robinson et al., 2012). This is because, as temperatures increase, so does the area of
174 summer melt, resulting in a lower surface elevation, causing further warming and increased
175 melt (atmospheric temperature decreases with altitude). This positive feedback is known as
176 the small ice cap instability, or melt-elevation feedback (e.g. Crowley and North, 1988;
177 Levermann and Winkelmann, 2016).

178

179 Key issues addressed by recent studies include: what the observed ice-sheet loss implies for
180 the rate of future global sea-level change, the potential long-term sea-level rise, and the
181 possibility of abrupt or irreversible changes on timescales of a few hundred years.

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184

185 **2 Observed recent changes**

186

187 Between 2002 and 2011 the West Antarctic Ice Sheet (WAIS) contributed 0.3 ± 0.1 mm yr⁻¹ to
188 global sea-level rise (Peng et al., 2016). The majority of this loss has come from basal melt of
189 ice shelves, and associated dynamical thinning, with half the basal melt arising from 10 small
190 ice shelves in the Bellingshausen and Amundsen seas (Rignot et al., 2013). Of these, Pine
191 Island glacier (Favier et al., 2014) and Thwaites glacier (Joughin et al., 2014) are the
192 principal outlets of the WAIS that have rapidly thinned, retreated, and accelerated since the
193 1990's. Recent assessments indicate that Thwaites is contributing ~ 0.1 mm per year to sea-
194 level rise, double that of the 1990s, and Pine Island glacier ~ 0.13 mm per year, however,
195 there has been no acceleration in mass loss since 2008 (Medley et al., 2014). The spatial
196 pattern of coincident changes in thickness across ice shelves of the Amundsen Sea suggests
197 that the loss of grounded ice is the direct result of increased basal melting of the ice shelf, as a
198 consequence of the inflow of warm water from the southern Pacific (Jacobs et al., 2011; Ha et
199 al., 2014). Multi-decadal warming at the seabed in the Bellingshausen and Amundsen seas is
200 linked to a shoaling of the mid-depth temperature maximum over the continental slope,
201 allowing warmer, saltier water greater access to the continental shelf in recent years
202 (Schmidtke et al., 2014). Before 2009, the glaciers of the Southern Antarctic Peninsula were
203 in equilibrium, but have since been contributing significantly to sea level rise at a near-

204 constant rate of $0.16 \pm 0.02 \text{ mm yr}^{-1}$ (Wouters et al., 2015). The onset of this sudden and rapid
205 mass loss appears to have a similar origin to that seen in the Amundsen Sea sector. The
206 retrograde bedrock configuration is such that the mass loss is likely to be sustained for years
207 to decades into the future, for this sector of Antarctica.

208

209 In addition there have been synchronous advances and retreats of the calving front of tide-
210 water glaciers of the East Antarctic Ice Sheet (EAIS) (Miles et al., 2013). These appear to be
211 associated with changes in the Southern Annular Mode (SAM), and consequently due to
212 natural climate variability. However, there is evidence of ocean warming causing thinning of
213 the Totten ice shelf, combined with a retreat of the grounding line (Silvano et al., 2016). A
214 substantial area of ice sheet inland of Totten is below sea level, equivalent to 3.5m of sea
215 level rise, and consequently the grounding line is potentially unstable due to the marine ice
216 sheet instability. Evidence provided by Silvano et al. (2016) that warm circumpolar water
217 does cross over the EAIS shelf break to cause rapid basal melt for several small ice shelves,
218 suggests that EAIS could be more vulnerable to ocean heat fluxes than previously thought.
219 Overall EAIS currently shows a gain in mass, through increased precipitation, with an
220 implied sea level fall of 0.32 mm per year over 2009-2011 (Boening et al., 2012).

221

222 Estimates of overall sea level changes associated with net ice loss from Antarctica have been
223 made by the gravity satellite, GRACE, between 2003 and 2014, at $0.25 \pm 0.2 \text{ mm per year}$
224 (Harig and Simons, 2015).

225

226 The Greenland ice sheet (GrIS) is losing mass as a result of both increased runoff due to
227 surface melting and increased ice discharge from marine-terminating outlet glaciers (Rignot
228 et al., 2008; Rignot et al., 2011; Sasgen et al., 2012; van den Broeke et al., 2009). The

229 Greenland mass loss over the period 2000-2005 contributed about 0.43 ± 0.09 mm yr⁻¹ of sea
230 level rise. The rate has, however, been accelerating, with estimates for 2009-2012 of
231 1.05 ± 0.14 mm yr⁻¹ (Figure 2; Enderlin et al., 2014), and for 2011-2014 of 0.74 ± 0.14 mm yr⁻¹
232 (McMillan et al., 2016). The relative contribution of ice discharge (dynamic thinning) to
233 total loss decreased from 58% before 2005 to 32% between 2009 and 2012. As such, 84% of
234 the increase in mass loss after 2009 was due to increased surface runoff, as opposed to
235 increased discharge (Enderlin et al, 2014). These observations support recent model
236 projections that changes in surface mass balance driven, primarily, by increases in air
237 temperature, rather than ice dynamics, will likely dominate the ice sheet's contribution to 21st
238 century sea level rise (e.g. Goelzer et al., 2013; Vizcaino et al., 2015).

239

240 The glaciers in the southeast and northwest of Greenland sped up between 2000 and 2005 and
241 have since stabilised or slowed (Enderlin et al, 2014). The slow down in the southeast has
242 been compensated for by the northeast Greenland ice stream, which extends more than 600
243 km into the interior of the ice sheet, and is now undergoing sustained dynamic thinning,
244 linked to regional warming, after more than a quarter of a century of stability (Khan et al.,
245 2014). This sector of the Greenland ice sheet is of particular interest, because the drainage
246 basin area covers 16% of the ice sheet, and numerical model predictions suggest no
247 significant mass loss for this sector, leading to a possible under-estimation of future global
248 sea-level rise (Khan et al, 2014). As for the Southern Antarctic Peninsula, the geometry of the
249 bedrock and monotonic trend in glacier speed-up and mass loss suggests that dynamic loss of
250 grounded ice in this region will continue in the near future (Khan et al., 2014).

251

252 **3 Potential for significant change**

253

254 The ice sheets can respond to climate change through accelerated discharge of freshwater to
255 the ocean, and the associated sea level rise may be irreversible. Accelerated discharge,
256 particularly from a marine ice sheet instability (MISI), has implications for the predictability
257 of future sea level rise. The IPCC AR5 projections of sea level rise (Church et al., 2013)
258 states that MISI may add tens of centimetres by 2100, but this mechanism was not quantified
259 in the summary projections due to a lack of understanding. New studies here have focussed
260 on the sea level contribution from the WAIS.

261

262 An ice flow model (Gagliardini et al., 2013) reveals that the Pine Island Glacier's grounding
263 line is probably engaged in an unstable 40 km retreat (Figure 1; Favier et al., 2014). The
264 associated mass loss increases substantially over the course of the simulations from an
265 average value of 0.05 mm yr^{-1} observed for the 1992-2011 period, up to and above 0.28 mm
266 yr^{-1} , equivalent to 3.5-10 mm mean sea-level rise over the next 20 years (Favier et al., 2014).
267 They find that mass loss remains elevated from then on, ranging from 0.16 to 0.33 mm yr^{-1} .
268 Paleoclimate evidence (Johnson et al., 2014) for the early Holocene (a period of seasonal
269 regional warming about 2K above pre-industrial) has revealed mass loss from the Pine Island
270 Glacier at a rate comparable to present-day loss, but no collapse. Simulations for the adjacent
271 Thwaites glacier, also in the Amundsen Sea embayment, indicate future mass losses are
272 moderate (less than 0.25 mm yr^{-1}) over the 21st century but generally increase thereafter
273 (Joughin et al., 2014). The likely time scale for collapse, based on various imposed ice shelf
274 basal melt rates, is the time required for 100-200 km of grounding line retreat in the Thwaites
275 Glacier system plus 200-1000 years for an actual collapse event (Joughin et al., 2014). Except
276 possibly for the lowest-melt scenario used in the simulations, the results indicate that early-
277 stage irreversible collapse has already begun (Joughin et al., 2014). One model includes the
278 process of hydrofracture for Antarctic ice shelves (Pollard et al., 2015; DeConto and Pollard,

279 2016), associated with surface melt water forcing open crevasses, leading to ice shelf
280 disintegration, and marine ice cliff instability. In this idealised simulation, hydrofracture
281 caused a rapid deglaciation of WAIS on a timescale of only about 100 years (from the
282 beginning of major retreat on the Antarctic Peninsula through to peak rate of sea-level rise
283 around year 2140 – see their Figure 4c). The ice sheet collapse projected by Deconto and
284 Pollard (2016) does not occur if the strong mitigation scenario of RCP2.6 is followed. Ice
285 loss from the EAIS Wilkes basin may become substantial over timescales beyond a century
286 (up to 3-4m sea-level rise after several millennia), with the loss irreversible above a threshold
287 of regional ice loss (Mengel and Levermann, 2014).

288

289 Some new understanding of the paleo record has emerged. For Antarctica as a whole, there is
290 evidence (Weber et al., 2014) for periods of relatively abrupt Antarctic mass loss following
291 the Last Glacial Maximum (26-19 thousand years ago), possibly associated with a positive
292 feedback involving ocean heat transport. It is likely that WAIS collapse occurred in the last
293 interglacial (125 thousand years ago), when the Southern Ocean temperature anomaly
294 exceeded 2-3°C (Sutter et al., 2016). However, the solar insolation is sufficiently different in
295 this interglacial, that a similar spatial pattern of warming cannot be achieved through present-
296 day increases in CO₂. One study (Levy et al., 2016) combined a range of regional and global
297 paleoproxy information to further constrain the response of Antarctica during the early to
298 mid-Miocene (23-14 million years ago), when CO₂ levels fluctuated between 280 and 500
299 ppm (equivalent to pre-industrial and a value that will be reached in the next few decades).
300 They identify a peak warming period (16 million years ago) showing a consistent picture of
301 global and regional warming, Antarctic ice sheet retreat, and a corresponding sea level rise of
302 10 to 20 m.

303

304

305

306 The Southern Ocean as a whole has not warmed significantly over the last decades (Armour
307 et al., 2016). Instead, local warming has occurred, as deeper warm waters have been forced,
308 by increased circumpolar winds, onto the Amundsen Sea continental shelf. Increasing winds
309 are a consequence of global warming, the depletion of stratospheric ozone, or natural
310 variability. Regardless of the cause of the increased windspeed, warm water has reached the
311 continental shelf of the Bellingshausen and Amundsen seas. As a consequence it has been
312 suggested that a critical threshold for grounding line retreat has already been passed for
313 glaciers in the Amundsen Sea sector (Rignot et al., 2014). High ice shelf thinning rates for
314 this and the Bellingshausen Sea sector of West Antarctica over the last two decades (Paolo et
315 al., 2015) combined with the dramatic shift in mass imbalance of the Southern Antarctic
316 Peninsula (Wouters et al., 2015) also point to a widespread shift in behaviour for this region.

317

318 It cannot be ruled out that the observed ice shelf thinning is a natural fluctuation rather than a
319 consequence of anthropogenic forcing. Thus the likelihood of (partial) collapse of the WAIS
320 has not yet been quantified, and requires improved modelling through ice sheet models fully
321 coupled within global atmosphere-ocean climate models. Some progress has been made along
322 these lines, with realistic ice shelf cavities now represented in ocean models (Beckmann et
323 al., 1999; Dinniman et al., 2007; Losch, 2008; Mathiot et al., 2017), and the idealised
324 simulations of MISOMIP (Asay-Davis et al., 2016).

325

326

327 **4 Potential consequences**

328

329 If a collapse of the WAIS were to occur, it would lead to a global sea level rise of up to 3.3 m
330 (Bamber et al., 2009) on timescales (from the onset of collapse) of 100 years (Deconto and
331 Pollard, 2016) to 400 years (Cornford et al., 2015; Golledge et al., 2015). This inference is
332 supported by records of past sea level rise. Under the low emissions scenario, RCP2.6, sea
333 level contributions remain small, and a collapse of WAIS does not occur in simulations
334 (Golledge et al., 2015; DeConto and Pollard, 2016). For Antarctica as a whole, paleoproxy
335 evidence from the Miocene (Levy et al., 2016) suggests potential sea-level rise of the order of
336 10-20m, for CO₂ levels near 500ppm.

337

338 Surface melt from the Greenland Ice Sheet may influence local ocean circulation, through
339 stratification reducing convection in the Labrador Sea (Yang et al., 2016), and consequently
340 local sea level change, perhaps by 5 cm, in the North-West Atlantic (Swingedouw et al.,
341 2013; Howard et al., 2014).

342

343 On centennial to millennial time scales, Antarctic Ice Sheet melt can moderate warming in
344 the Southern Hemisphere, by up to 10°C regionally, in a 4 x CO₂ scenario (Swingedouw et
345 al., 2008). This behaviour stems from the formation of a cold halocline in the Southern
346 Ocean, which limits sea-ice cover retreat under global warming and increases surface albedo,
347 reducing local surface warming. In addition, Antarctic ice sheet melt, by decreasing Antarctic
348 Bottom Water formation, restrains the weakening of the Atlantic meridional overturning
349 circulation, which is an effect of the bi-polar oceanic seesaw (Pedro et al., 2011).

350 Consequently, it appears that Antarctic ice sheet melting strongly interacts with climate and
351 ocean circulation globally. It is therefore necessary to account for this coupling in future
352 climate and sea-level rise scenarios.

353

354 **5 Cautions (uncertainties).**

355

356 While substantial progress in understanding has been made, it is still unclear what the recent
357 observed changes imply for long-term future ice-sheet loss (Wouters et al., 2013), due to
358 regional natural variability. Some observations suggest that there may be a natural cycle of
359 increase and decrease in the rates of mass loss from coastal glaciers (Murray et al., 2010), so
360 short-term trends should not necessarily be extrapolated into the future (Wouters et al., 2013).
361 Indeed many Greenland glaciers, which accelerated in the early 2000s have since slowed
362 (Moon et al., 2012; Enderlin et al., 2014). There is a possibility that solid earth movement, in
363 response to ice loss, may influence the bedrock slopes, and so reduce further ice loss from the
364 West Antarctic Ice Sheet (Konrad et al., 2015), delaying WAIS collapse by as much as 5000
365 years.

366

367 **6 Comparison with AR5**

368

369 Of the key findings summarised in Table 1, the main new points since AR5 are: observational
370 evidence (Enderlin et al. 2014) that, from Greenland, the proportion of loss from surface melt
371 has increased, becoming more consistent with long term model projections; evidence that
372 some degree of irreversible loss from the WAIS may have begun (Favier et al. 2014, Joughin
373 et al. 2014, Rignot et al. 2014, Wouters et al. 2015); and indications that the East Antarctic
374 Ice Sheet (Miles et al. 2013) and northeast Greenland (Khan et al. 2014) may be more
375 sensitive to climate change than previously expected.

376

377

378

379 **III AMOC**

380

381 **1 Introduction**

382

383 The Atlantic Meridional Overturning Circulation (AMOC) transports large amounts of heat
384 northwards in the Atlantic Ocean, resulting in a milder climate in northwest Europe and the
385 North Atlantic than would otherwise be experienced (for recent reviews of AMOC behaviour
386 and observations see Srokosz et al., 2012; Srokosz and Bryden, 2015; Buckley and Marshall,
387 2016). The IPCC AR5 report (Collins et al., 2013) concludes that it is very likely that the
388 AMOC will weaken over the 21st century, although there is a large spread in the predicted
389 weakening among climate models. A large or rapid (over a decadal time scale) reduction in
390 the AMOC would likely have substantial impacts on global climate, although a collapse
391 (rapid shutdown) of the AMOC by 2100, however, was judged as very unlikely (Collins et
392 al., 2013). These assessments have not changed since the previous IPCC assessment.

393

394 **2 Observed recent changes**

395

396 The RAPID-MOCHA array has been observing the AMOC at 26°N since 2004 and now has
397 acquired over a decade of data (Rayner et al., 2011; McCarthy et al., 2015b). This dataset has
398 revealed large variability on timescales from daily to interannual (see Figure 2). This
399 included a large (30%), temporary decrease in AMOC strength over 2009-2010 (McCarthy et
400 al., 2012; Bryden et al., 2014), which resulted in cooling in the upper North Atlantic Ocean in
401 2010 north of the latitude of the RAPID array and warming to the south (Cunningham et al.,
402 2013; Bryden et al., 2014). This decrease began with a strengthening of the upper mid-ocean

403 recirculation in early 2009 and was compounded by a slowdown in the northward Ekman
404 transport and Gulf Stream flow in late 2009 and early 2010 (accounting for 61%, 27% and
405 12% of the slowdown, respectively; Bryden et al., 2014). This decrease was well outside the
406 range predicted for interannual AMOC variability in coupled ocean-atmosphere models
407 (McCarthy et al., 2012; Roberts et al., 2014); note that model resolution may be an issue here.
408 Roberts et al. (2013b) reproduced this AMOC decrease using an initial condition ensemble of
409 ocean simulations driven by observed surface forcing (albeit with too weak an AMOC),
410 suggesting that the atmosphere may have had a dominant role in the temporary AMOC
411 decrease. However, the origin of, and complete explanation for, the 2009-10 event remains
412 uncertain. To-date no explanations have fully accounted for the changes in Lower North
413 Atlantic Deep Water (LNADW at 3000 to 5000m depth) and the lack of change in the Upper
414 North Atlantic Deep Water (UNADW between 1000 and 3000m depth) described by
415 McCarthy et al. (2012).

416

417 The links between changes in the AMOC, upper ocean heat content and atmospheric response
418 represent an active area of research. For example, the ocean has been implicated in the re-
419 emergence of sea surface temperature anomalies from the winter of 2009-10 during the
420 following early winter season of 2010-11, which contributed to the persistence of the
421 negative winter North Atlantic Oscillation (NAO) and wintry conditions in northern Europe
422 (Taws et al., 2011). Such behaviour may lead to improved predictions of the NAO and winter
423 conditions (Maidens et al., 2013; Scaife et al., 2014).

424

425 The AMOC overturning has also decreased from 2004-2014 (Figure 2; Srokosz and Bryden,
426 2015; Frajka-Williams et al., 2016); the majority of this was due to a weakening of the
427 geostrophic flow (Smeed et al., 2014: who analysed the first eight and a half years of data).

428 This trend has been associated with decreases in subsurface density in the subpolar gyre,
429 similar to those seen in climate models when there is a reduction in the AMOC (Robson et
430 al., 2014). It is unclear at this stage whether the decrease is forced (so part of a longer-term
431 downturn). Some recent work suggests that it may be part of a downturn after a previous
432 increase (Jackson et al., 2016; Frajka-Williams, 2015). Statistical tests on the observations
433 (Smeed et al. 2014) suggested that the AMOC decrease is statistically significant, even if the
434 low AMOC event of 2009-10 is excluded. Roberts et al (2014) found similar trends as part of
435 natural variability in 2 out of 14 global climate models, and in all models considered when
436 corrections are made to include more realistic high frequency variability. They concluded that
437 more than a decade of observations would be required to detect and attribute an
438 anthropogenic weakening of the same trend as observed over the period 2004-2012 (although
439 this rate does not appear to have been maintained since 2012, Figure 2). In an earlier model
440 study, Roberts et al. (2013a) estimated that a minimum of two decades of data would be
441 required to detect an anthropogenic trend in the AMOC, based on multimodel 1% per year
442 CO₂-forced experiments. This means that the existing AMOC observing system would need
443 to make measurements until at least 2024. Another study (Mercier et al., 2015) analysed
444 repeat hydrographic data along a Greenland to Portugal section from 1993 to 2010, finding an
445 overall decline in the AMOC over that period. Send et al. (2011) observed a decreasing trend
446 in the transport of the deep western boundary current at 16°N (one component of the AMOC)
447 over a similar period. In the South Atlantic, based on a combination of satellite altimeter and
448 hydrographic observations, Dong et al. (2015) note that “since 2010 the MOC has exhibited
449 low values when compared to the 1993–2011 mean values.” Linking the observations of the
450 AMOC obtained at different latitudes by different observational means remains a significant
451 challenge (Elipot et al., 2014; Elipot et al., 2017). Landerer et al. (2015) show that satellite
452 observations of ocean bottom pressure may provide a useful method for examining latitudinal

453 coherence of signals, however it is currently restricted to detrended data and regions of steep
454 topography.

455

456 A recent paper (Rahmstorf et al., 2015) suggested that the trend detected at 26°N is part of an
457 ‘exceptional slowdown’ of the AMOC. They find a relationship between sea surface
458 temperatures and the AMOC in a climate model and then use reconstructions of surface
459 temperatures from paleoclimate records to suggest that there has been a weakening that is
460 unprecedented over the last 1000 years. There are, however, inherent uncertainties around
461 both the relationship used and the temperature reconstructions, raising questions over whether
462 the results are robust. In contrast, a recent reconstruction of the AMOC in the South Atlantic
463 since 1870 (Lopez et al., 2017) suggests that it is presently in a stronger than normal phase.
464 Ultimately, all proxies for the AMOC, such as temperature, coastal sea level (Ezer, 2015;
465 McCarthy et al., 2015b; Frajka-Williams, 2015), or gravity measurements (Landerer et al.,
466 2015) need to be tested and verified against direct observations of AMOC strength, over the
467 time scales of interest, if they are to be used to infer robustly its behaviour over longer
468 periods.

469

470 Future observations and research will improve our assessments of past and on-going AMOC
471 changes. In this context note that the Overturning in the Subpolar North Atlantic Program
472 (OSNAP; Lozier et al., 2017) deployed instruments in 2014 along a line from Canada to
473 Greenland to Scotland, to observe the AMOC in the subpolar gyre, complementing the
474 26.5°N observations in the subtropical gyre. Meanwhile in the South Atlantic there are trans-
475 basin observations of the AMOC beginning to be made at 34.5°S (SAMBA – South Atlantic
476 MOC Basin-wide Array; Meinen et al., 2013; Ansorge et al., 2014). Recently, a new
477 component of the AMOC, the so-called East Greenland spill jet, has been identified from a

478 year of mooring observations (von Appen et al., 2014), but its importance in the long-term for
479 the overall AMOC remains to be confirmed.

480

481 **3 Potential for significant change**

482

483 Paleoclimate studies have suggested that some abrupt changes to climate may have been
484 caused by the AMOC switching from an “on” state, where it transports heat northwards in the
485 Atlantic, to an “off” state (Rahmstorf, 2002). Paleoceanographic studies of the AMOC and
486 abrupt climate change over the last glacial cycle have been recently summarised by Lynch-
487 Stieglitz (2017), who found that the evidence for changes in the AMOC associated with the
488 Younger Dryas and many Heinrich events is strong, and there is some evidence for AMOC
489 changes over many Dansgaard-Oeschger events. However, the ultimate causal links between
490 the co-incident changes in the AMOC and climate are less clear. Further, these studies are
491 hard to interpret in terms of future change, as the conditions in which past abrupt changes
492 occurred were very different to the present. It is thought that abrupt changes may be related
493 to the existence of bistability (where both “on” and “off” states of the AMOC can exist for a
494 given forcing) as predicted by theoretical models of the Atlantic (e.g. Stommel, 1961), Earth
495 system models of intermediate complexity (Rahmstorf et al., 2005) and studies with low
496 resolution global circulation models (Hawkins et al., 2011; Manabe and Stouffer, 1988).
497 Statistical properties of the timeseries of AMOC strength may give warning of approaching a
498 threshold, however a new model study finds that centuries of data from a reliable proxy
499 would be required (Boulton et al., 2014).

500

501 There have been many model studies suggesting that the stability of the AMOC might be
502 affected, or even controlled, by whether the AMOC imports or exports fresh water from the

503 Atlantic, since this can indicate the presence of a positive or negative advective feedback. De
504 Vries and Weber (2005) found that the fresh water transport by the AMOC into the Atlantic
505 was an important indicator of stability in their experiments. The relative importance, for
506 AMOC stability, of freshwater export/import by the AMOC itself, is unclear, however.
507 Other factors have subsequently been found to be important in determining AMOC stability.
508 For example, Jackson (2013) found that, while the overturning component of freshwater
509 transport does partially indicate the sign of the advective feedback in a GCM, the transport of
510 fresh water by the gyres can also play a crucial role. Swingedouw et al. (2013) also found
511 that gyre transports can affect the magnitude of AMOC reduction. The presence of eddies is
512 lacking in many models (due to low resolution) but studies with an eddy resolving model
513 show that they can also affect the fresh water transport (den Toom et al., 2014). Mecking et
514 al. (2016) found that the AMOC in an eddy-permitting model was very slow to recover from
515 an input of fresh water. They found that the freshwater transport by the AMOC was important
516 for maintaining the weak AMOC state, and hypothesised that this transport was changed by
517 the eddy-permitting resolution. Understanding these controls on AMOC stability is crucial to
518 constraining the likelihood of AMOC collapse. A recent paper (Liu et al., 2017) has noted
519 that biases in the models may affect the estimated probability of an AMOC collapse.

520

521

522 **4 Potential consequences**

523

524 A collapse in the AMOC would cause a large relative cooling over the North Atlantic, which
525 would have wide-ranging impacts, such as cooling in the northern hemisphere, warming in
526 the southern hemisphere, and a southward shift in the Inter Tropical Convergence Zone,
527 causing substantial changes in tropical precipitation, (Vellinga and Wood, 2008; Jackson et

528 al., 2015).

529

530 The Amazon is one region sensitive to change in the AMOC, but the impacts are uncertain. A
531 recent study by Parsons et al. (2014) found that a reduction in the AMOC caused an increase
532 in vegetation over the Amazon, due to a change in precipitation seasonality (despite a
533 reduction in annual mean precipitation). This contrasts with an earlier study (Bozbiyik et al.,
534 2011), which found that a reduction in the AMOC causes large reductions in Amazon
535 vegetation due to precipitation reductions.

536

537 Other studies have concentrated on impacts over Europe. Jackson et al. (2015) confirmed an
538 earlier study by Woollings et al. (2012), that a reduction in AMOC strength could drive an
539 increase in the number of winter storms across Europe. Jackson et al. (2015) also showed that
540 the increase in winter storms resulted in greater precipitation over western coasts in Northern
541 Europe, despite a general reduction of precipitation over the northern hemisphere from a
542 cooling-induced reduction in evaporation. They also found regional changes in summer
543 precipitation across Europe, similar to those associated with Atlantic sea temperature found
544 by Sutton and Dong (2012). Haarsma et al. (2015) examined the relationship between
545 European atmospheric circulation and the AMOC across the CMIP5 ensemble. They also
546 found an influence of AMOC strength on European summer precipitation and cloud cover.

547

548 One impact of the AMOC suggested recently is its possible role in the so-called global
549 warming “hiatus” (Chen and Tung, 2014), though various other explanations for the hiatus
550 have been proposed. Another recently observed impact is the reduction in uptake of CO₂ by
551 the Atlantic Ocean due to the weakening of the AMOC over the period 1990 to 2006 (Perez
552 et al., 2013).

553

554 Another recent focus of attention has been the role of the AMOC in sea level rise (SLR) on
555 the eastern seaboard of the USA (Ezer, 2015; Goddard et al., 2015; McCarthy et al., 2015a;
556 Yin et al., 2009). In particular, Goddard et al. (2015) demonstrate that the 2009-10 temporary
557 downturn in the AMOC led to an unprecedented 12.8 cm sea level rise along the coast north
558 of New York over the same period. They show that this rise was a 1-in-850 year event.
559 Furthermore, they note that, “Unlike storm surge, this event caused persistent and widespread
560 coastal flooding even without apparent weather processes. In terms of beach erosion, the
561 impact of the 2009–2010 SLR event is almost as significant as some hurricane events.” This
562 observed short-term change provides evidence for what has been previously suggested only
563 by modelling studies (Levermann et al., 2005), that a slowdown or collapse of the AMOC
564 would lead to significant sea level rise on the eastern seaboard of the USA.

565

566 **5 Cautions**

567

568 There are large inter-model differences in projections of future AMOC decline amongst
569 models used for the IPCC AR5 report. Reintges et al. (2016) found that uncertainties in
570 AMOC projections were dominated by uncertainties in fresh water changes amongst the
571 models, with contributions from uncertainties in both changes in surface fresh water fluxes
572 and ocean fresh water transports.

573

574 Several studies have shown that many GCMs have biases in the fresh water transport of the
575 AMOC (importing instead of exporting fresh water), and that this might affect the simulated
576 stability of the AMOC. The source of this bias is unclear. Jackson (2013) attributed the bias
577 to an over-evaporative Atlantic in the model and notes the difference from observations in

578 salinity profiles in the South Atlantic. Liu et al. (2014) suggested that the presence of a
579 double Atlantic ITCZ (a common GCM bias) results in a tropical salinity bias that stabilises
580 the AMOC. Another source of uncertainty is the transport of saline water from the Indian
581 Ocean to the Atlantic by eddies that are shed from the Agulhas current. Current GCMs do not
582 resolve the scales required to correctly represent these eddies, but a recent study by Biastoch
583 and Böning (2013) used a high resolution nested model to resolve this region. They found
584 that a southwards shift of the southern hemisphere westerlies (as is expected to occur under
585 anthropogenic climate change) results in a decrease in salinity transport into the Atlantic,
586 however this change in salinity is small and has little impact on the AMOC. The lack of
587 eddy-resolving resolutions in current GCMs might also have an impact on the transient
588 response of the AMOC to increased freshwater input (Weijer et al., 2012; Mecking et al.,
589 2016).

590

591 There is also substantial uncertainty about the future inputs of freshwater into the Atlantic,
592 particularly since the climate models lack dynamic ice sheet models which could
593 substantially speed up the input of freshwater from the Greenland ice sheet. Separate studies
594 including additional freshwater inputs from the Greenland ice sheet find that projected
595 changes do not have major impacts on the AMOC, although there is uncertainty about future
596 changes in freshwater fluxes from Greenland (Bamber et al., 2012). Böning et al. (2016)
597 concluded that meltwater from the Greenland ice sheet has resulted in a gradual freshening of
598 Labrador Sea, but that this has had no significant impact on the AMOC yet. A recent study
599 found that the MOC became less sensitive to fresh water inputs when CO₂ levels were high,
600 because of increases in stratification caused by warming and changes in the wind-driven
601 circulation (Swingedouw et al., 2015). Another study suggests that future increases in
602 precipitation over the Arctic, leading to increased freshwater flux into the North Atlantic

603 could also affect the AMOC (Bintanja and Selten, 2014: see Methods). The most recent GCM
604 study that accounted for Greenland melting (Bakker et al., 2016) concluded that Greenland
605 Ice Sheet "melting affects AMOC projections, even though it is of secondary importance."
606

607 **6 Comparison with AR5**

608 The main development since the publication of AR5 has been the updated observations of
609 overturning from the RAPID-MOCHA array (Smeed et al., 2014; Srokosz and Bryden,
610 2015), which shows a decline over the period 2004-2014. Studies suggest that this was
611 related to decadal variability. This does not preclude the presence of a longer-term decline,
612 but the time series is too short to make definitive statements. Continuous observations from
613 the existing AMOC observing system until at least around 2024, combined with further
614 understanding of the past record from multiple proxy information, and more model studies,
615 will be required to isolate a forced decline in the AMOC. Another key finding is the
616 unprecedented rise in US east coast sea level associated with the 2009-10 downturn in the
617 AMOC (both of which subsequently recovered). Although this is a change on a shorter time
618 scale than the 100 year time scale associated with climate change, it shows that changes in
619 the AMOC may have impacts on multiple time scales. Finally, the inclusion of Greenland
620 melting in GCMs has been found to affect AMOC projections, but appears to be of secondary
621 importance.

622

623

624

625

626

627 **IV Tropical forests (Amazon focus)**

628

629 **1 Introduction**

630

631 Tropical forests regulate and supply to society a range of services, which bring benefits at
632 global to local scales. As well as sustaining high biodiversity they influence climate through
633 biogeochemical (carbon cycle) and biophysical (water and energy) mechanisms. Over the
634 period 1990-2007, intact tropical forests took up carbon at the rate of $1.2 \pm 0.4 \text{ Pg C year}^{-1}$
635 (corresponding to about half the global land carbon sink), compared with $0.50 \pm 0.08 \text{ Pg C}$
636 year^{-1} by the boreal forests (Pan et al., 2011), and around $1.1 \pm 0.8 \text{ Pg C year}^{-1}$ losses of
637 forest carbon stocks to the atmosphere through land use change over 2000-2009 (Settele et
638 al., 2014). However, large droughts can cause elevated mortality rates, especially for larger
639 trees (Phillips et al., 2010; Nepstad et al., 2007; da Costa et al., 2010; McDowell and Allen,
640 2015) and temporary shifts from ecosystem carbon sink to carbon source (Phillips et al.,
641 2010; Lewis et al., 2011; Gatti et al., 2014). Estimates of the impact of the 2005 and 2010
642 Amazon droughts (mostly through increases in tree mortality during and lagging the
643 droughts) stand at 1.6 and 2.2 Pg C, respectively (Phillips et al., 2009; Lewis et al., 2011).

644

645 Tropical forests are subject to interacting effects from atmospheric CO₂, climate and land-use
646 change (e.g. Coe et al., 2013). Land-use change effects include direct deforestation, and
647 accidental 'leakage' fires (where intentional fires spread accidentally over a wider forest area).
648 Forest fragmentation (an important by-product of deforestation) lengthens the forest edge.
649 Since most forest fires occur at the forest edge, because of greater human activity,
650 fragmentation accelerates the rate of forest erosion by fire. Deforestation increases albedo

651 and reduces evapotranspiration, altering climate both locally and downwind; aerosols from
652 deforestation fires may also reduce rainfall (Marengo et al., 2011). Climate change could alter
653 vegetation productivity and mortality, both directly, and indirectly by modifying fire
654 behaviour. Increased atmospheric CO₂ may increase tree growth (where nutrients are not
655 limiting), but also increase tree mortality from lianas (vines). The full vegetation response to
656 CO₂ and climate changes may take decades to be completely realised (Jones et al., 2009), and
657 the subsequent carbon release even longer.

658

659 The AR5 finds that large-scale dieback due to climate change alone is unlikely by the end of
660 this century (medium confidence). However, it states with medium confidence that “severe
661 drought episodes, land use, and fire interact synergistically to drive the transition of mature
662 Amazon forests to low-biomass, low-statured fire-adapted woody vegetation” (Settele et al.,
663 2014). New research has largely, but not exclusively, focused on the Amazon: due in part to
664 early climate model projections of climate-driven Amazon dieback (Cox et al., 2000). Severe
665 Amazonian droughts in the last decade have provided insights on forest responses to extreme
666 dry conditions. In addition to forest and climate monitoring, throughfall exclusion and
667 prescribed-burn experiments have allowed in-situ study of the effects of longer-term drought
668 and fire. Numerical studies have also increased in number and progress has been made in
669 putting the early results into context.

670

671 **2 Observations**

672

673 New studies have given greater confidence that the Amazon represents a long-term net
674 carbon sink (Brienen et al., 2015; Espirito-Santo et al., 2015; Gatti et al., 2014), but also
675 suggest (Brienen et al., 2015) that its strength has weakened progressively as tree mortality

676 rates increase (Figure 3). Potential drivers for the mortality increase include more frequent or
677 more severe droughts, and feedbacks of faster growth on mortality, resulting in shortened tree
678 longevity (Bugmann and Bigler, 2011).

679

680 The response of trees to elevated CO₂ remains uncertain. Some recent longer-term studies of
681 tropical tree rings (van der Sleen et al., 2015; Battipaglia et al., 2015; Groenendijk et al.,
682 2015) have found no evidence for sustained increases in tree growth or carbon uptake, but as
683 Brien et al. (2012) point out, tree-ring studies are subject to biases which preclude robust
684 statements about ecosystem-level changes. So far, multi-decadal plot data been used
685 systematically to probe recent growth trends at continental scale only in Amazonia (Brien
686 et al., 2015). Here they indicate a long-term increase in growth rates since the 1980's, as well
687 as a lagging increase in mortality rates, consistent with a long-term growth stimulation, such
688 as by CO₂.

689

690 There is greater confidence that Amazon forests are adversely affected by drought. There has
691 been new work on the response to the 2010 drought, and also the 1997 and 2005 events
692 (Tomasella et al., 2013). A new attribution study (Shiogama et al., 2013) of the 2010 drought
693 showed that, while sea surface temperature anomalies in the tropical Pacific and Atlantic
694 likely increased the probability of drought (in addition to biomass burning; Marengo et al.,
695 2011), unforced atmospheric variability probably also played a large role. Atmospheric
696 measurements (Gatti et al., 2014) confirmed earlier plot-based findings (Phillips et al., 2010;
697 Lewis et al., 2011) that the Amazon switched from a temporarily from a net carbon sink to a
698 source during the 2010 drought. Compared to these short-term natural droughts, the impact
699 was seen to be much stronger in the long-term persistent experimental droughts induced by a
700 forest throughfall exclusion experiment in eastern Amazonia (da Costa et al., 2014), and by

701 2014, 13 years of 50% throughfall exclusion at Caxiuana had caused a cumulative biomass
702 loss of $45.0 \pm 2.7\%$ (Rowland et al., 2015). Consistent with previous suggestions that effects
703 of a single drought persist for several years (Saatchi et al., 2013; Phillips et al., 2010), even
704 during the anomalously wet year of 2011, the Amazon was still estimated to be carbon
705 neutral overall (Gatti et al. 2014; possibly due to lagged effects of the 2010 drought).
706
707 Drought mortality, especially in larger trees, is a major pathway for carbon release (Phillips et
708 al., 2010; Nepstad et al., 2007; da Costa et al., 2010), but underlying mechanisms are not well
709 understood (Meir et al., 2015), and poorly represented in current vegetation models (Powell
710 et al., 2013). However, hydraulic failure is suggested as the primary cause from the Caxiuana
711 drought experiment (Rowland et al., 2015). A study of detailed plot-level responses to the
712 2010 drought in several sites, compared to other years (Doughty et al., 2015) suggested that
713 trees may prioritise growth in response to reduced photosynthesis from short-term drought,
714 leaving some trees more vulnerable to mortality. In contrast, the long-term (> 12 years)
715 response to persistent experimental rainfall exclusion (Rowland et al., 2015), shows no
716 decline in photosynthetic capacity (although photosynthesis may have declined if mean
717 stomatal conductance declined), but an increase in leaf dark respiration in tree taxa vulnerable
718 to drought mortality (possibly a sign of drought stress). It has been suggested that early
719 warning of drought mortality events may be plausible based on observations of tree
720 properties (Camarero et al., 2015). Overall, the AR5 viewpoint of persistent drought causing
721 a shift towards lower statured, low-biomass forest is retained (Rowland et al., 2015).
722
723 Drought can also cause abrupt increases in fire-induced tree mortality over sub-seasonal
724 timescales (Brando et al., 2014), driving a lagged increase in carbon emissions over
725 subsequent years. More than 85,500 km² of the southern Amazon was burnt by understory

726 fires during 1999-2010, with evidence for a strong climate control on fire (Morton et al.,
727 2013).

728

729 Forest responses to warming remain uncertain, and more forest warming field experiments
730 are needed (Cavaleri et al., 2015). In one such experiment (Slot et al., 2014), although
731 respiration increased with warming, thermal acclimation did occur. A new meta study
732 integrating experimental and observational results (Vanderwel et al., 2015) suggests that
733 acclimation could potentially half increases in leaf dark respiration over the century,
734 compared with null model expectations that ignore acclimation. On the other hand, a global-
735 scale analysis of interannual variability has suggested (Anderegg et al., 2015) that nighttime
736 respiration in tropical forests may be highly sensitive to warming.

737

738 Some new observational studies have found substantial reductions in evapotranspiration in
739 some (Oliveira et al., 2014; da Silva et al., 2015; Panday et al., 2015), but not all (Rodriguez
740 et al., 2010) deforested regions. The full effects of deforestation over the Xingu river basin (a
741 southeast tributary of the Amazon) may have been masked by climate variability (Panday et
742 al., 2015).

743

744 **3 Potential for significant change**

745

746 A recent review of wider sources of evidence (Coe et al., 2013) identified South/South-East
747 Amazonia as particularly vulnerable: due to high deforestation rates locally and in the upwind
748 savanna region; its susceptibility to small climate shifts (being in a transitional climate zone
749 between forest and savanna); and greater climate model agreement on future rainfall
750 reductions in this region (compared to the West Amazon). A review for African rainforests

751 (Malhi et al., 2013) highlighted similar points: while deforestation rates have historically
752 been relatively low over Africa, there is potential for significant future increases; and African
753 forests have climate close to the limit of rainforest sustainability. Models tend to predict
754 rainfall reductions(increases) over western(central) equatorial Africa (James et al., 2014).
755 Rainfall decreases over west equatorial Africa can be large in some models (James et al.,
756 2014), although the forest response is hard to predict.

757

758 The observations and field experiments summarised above have given greater confidence that
759 forests are significantly affected by drought, emphasising the importance of extreme climate
760 events in causing extensive tree losses (through drought and heat mortality and increased
761 fire). On the other hand, some acclimation of trees to warming has been demonstrated.
762 These vegetation responses are not well understood or represented in current dynamic
763 vegetation models (e.g. Powell et al., 2013).

764

765 A recent study using three terrestrial biosphere models (Zhang et al., 2015) found the
766 direction and severity of precipitation change to be critical. A greater model consensus for a
767 projected lengthening and deepening of the dry season in Amazonia was found in CMIP5
768 compared with CMIP3 (Joetzjer et al., 2013), although it is unclear whether this represents a
769 statistically significant improvement in model performance. A new observationally
770 constrained model study (Boisier et al., 2015) found a greater lengthening of dry seasons over
771 the Amazon than projected by unconstrained models (as found, with a different method, by
772 Shiogama et al., 2011). A key uncertainty in terms of impacts is the extent to which forest
773 whole-ecosystem responses to climate change might be protected by the wide functional
774 diversity in many tropical forests. The work of Fauset et al. (2012) from Ghana suggests that
775 by not accounting for biological diversity, most vegetation models may underestimate forest

776 resilience.

777

778 In terms of land-use, it has become clearer that as well as direct deforestation, the indirect
779 effects of deforestation on forest fragmentation, and on climate locally and downwind, must
780 be considered in regulatory policies (e.g. Harper et al., 2014; Lawrence and Vandecar, 2015;
781 Brinck et al., 2017; Wu et al., 2017). While a 70% decline has been reported in deforestation
782 in the Brazilian Amazon between 2005 and 2013 (Nepstad et al., 2014), maintaining low
783 levels of deforestation in a sustainable manner remains a challenge (Nepstad et al., 2014).
784 Despite the reduction in deforestation since 2004, around half of the area burnt during 1999-
785 2010 over the southern Amazon occurred during 2007 and 2010, when deforestation activity
786 was relatively low, suggesting that fire-free land use needs to be encouraged as well as
787 reducing direct deforestation (Morton et al., 2013). Achieving similar reductions in
788 deforestation in other countries may be challenging due to issues with governance and
789 monitoring capability (DeFries et al., 2013), but for Indonesia, accounting for spatial
790 variation in costs and benefits of avoided deforestation does reveal low cost options (Graham
791 et al., 2017).

792

793 Various positive feedbacks (fire-vegetation and climate-vegetation; eg. Hirota et al., 2011;
794 Staver et al., 2011; Hoffmann et al., 2012) exist that could lead to abrupt reductions in forest
795 cover, for relatively small change in external forcings, and inhibit reversibility, but the
796 processes are poorly characterised. The timescale of any abrupt change depends on the
797 processes and the spatial scale considered, and may be strongly dependent on stochastic
798 climate variability. Very locally, loss of tree cover from fire or drought mortality can occur
799 over seasonal timescales in the event of severe drought. However, one model study (Higgins
800 and Scheiter, 2012) found that, while transitions between vegetation states may be abrupt

801 locally, over continental and larger scales the effect on the carbon cycle is much more gradual
802 (because the timing of transitions varies with location). It has been suggested (Verbesselt et
803 al., 2016) that temporal autocorrelation in satellite data provides evidence for threshold
804 behaviour in forests – and potential for monitoring forest resilience. Evidence for alternative
805 stable states has recently been reported in vegetation height (Xu et al., 2016) as well as in tree
806 fractional cover (Staver et al., 2011; Pausas and Dantas, 2017; Hirota et al., 2011). However,
807 the spatial scale over which abrupt or irreversible change might extend depends on the
808 strength of these positive feedbacks compared to environmental control on vegetation cover,
809 and demonstrating whether alternative stable states exist over large scales is challenging
810 (Good et al., 2016; Staal et al., 2016). Spatial interaction between forest and savanna can
811 reduce the area over which alternative stable states exist (a clear exploration is provided by
812 Staal et al., 2016). Indeed, recent observational work has challenged the notion that savanna
813 and forest represent ‘alternative stable states’ over large areas of the tropics (Veenendaal et
814 al., 2015; Wuyts et al., 2017). Local fire-vegetation feedbacks are seen in prescribed burning
815 experiments (Silverio et al., 2013), but over large scales, only 10% of the locations burnt in
816 the 2005 drought showed repeated burning by 2010 (Morton et al., 2013). A model study
817 (Moncrieff et al., 2014) found that the area over which alternative stable states are possible
818 could be large in present-day conditions, but declined substantially with future CO₂ increases.
819 Hoffmann et al. (2012) noted that forest-fire feedbacks themselves can be sensitive to tree
820 growth-rates – and hence to climate change.

821

822 **4 Potential consequences**

823

824 The observations summarised above give greater confidence that the Amazon represents a net
825 carbon sink, but this appears to have been declining at least for a decade, and the long-term

826 future of this sink is uncertain. Persistent drought would be likely to cause a transition to
827 lower statured, lower biomass forest, from mortality of larger trees (Rowland et al., 2015),
828 and severely threaten biodiversity (Esquivel-Muelbert et al., 2017). Extreme events over the
829 Amazon could have a large impact on the global carbon cycle and offset or counteract
830 potential regional increases in biomass (Reichstein et al., 2013).

831

832 While it is accepted that tropical deforestation tends to reduce evapotranspiration locally,
833 consequent changes in rainfall are complex and depend on the scale and pattern of
834 deforestation (Lawrence and Vandecar, 2015). Including deforestation feedback on climate
835 (via precipitation) is key in assessing river runoff change (Stickler et al., 2013; Lima et al.,
836 2014). Stickler et al. (2013) estimate that when feedbacks on climate are included, the sign
837 of change in hydropower generation potential for the plants under construction on the
838 Amazonian Xingu River is reversed, declining to 25% of maximum plant output by 2050
839 under business-as-usual land-use projections (with 40% deforestation by 2050). The net
840 runoff response in the Amazon is basin-dependent (Lima et al., 2014) and is sensitive to the
841 scale and pattern of deforestation (Lawrence and Vandecar, 2015). Deforestation may reduce
842 the length of the wet season, such that large-scale expansion of agriculture in Amazonia may
843 be unsustainable (Oliveira et al., 2013; Arvor et al., 2014). Land-use-driven stream warming
844 of at least 3-4K (in mean daily maximum temperature) in southeastern Amazonian has also
845 been observed (Macedo et al., 2013) - well above the ~1K threshold for changes in fish
846 physiology, growth and behaviour. Overall, multiple ecosystem services need to be taken
847 into account when considering optimal management (Donoso et al., 2014).

848

849

850 **5 Cautions (uncertainties)**

851

852 Accurate projections are partly limited by the availability of observations. Inaccessibility of
853 tropical forests increases reliance on remote sensing data, but also makes verifying remote
854 sensing data (notably, precipitation, biomass, and vegetation productivity data) challenging.
855 New studies have shown that great caution is required in interpreting satellite retrievals of
856 variability in greenness (Morton et al., 2014). The tropical forest biome constitutes one of the
857 largest terrestrial carbon sinks, but it is also associated with relatively large uncertainties (Pan
858 et al., 2011), because of its great ecological complexity, huge scale, and multiple
859 anthropogenic processes affecting it (Lewis et al., 2015).

860

861 There is substantial uncertainty in the CMIP5 projections of future precipitation in tropical
862 forest regions, (Collins et al., 2013), although there is greater degree of inter-model
863 agreement in some seasonal changes, such as a lengthening and a deepening of the dry season
864 in Amazonia (Joetzjer et al., 2013; Boisier et al., 2015). However, the representation of
865 present-day Amazon precipitation still contains large biases. Large uncertainties are also
866 associated with the modelled response of vegetation to temperature (Galbraith et al., 2010;
867 Huntingford et al., 2013) and to CO₂ (Rammig et al., 2010). Processes of direct mortality
868 from fire and drought (and effects of fire on aerosol) are often either unrealistic or absent
869 from models (e.g. Powell et al., 2013), and the range of plant functional types is extremely
870 limited in relation to the large biodiversity and hence range of potential tree-level responses
871 in most tropical forests.

872

873 **6 Comparison with AR5**

874

875 The new literature has not altered the broad, general view given in AR5. Probably the

876 greatest advances lie in increased confidence that, at least over the Amazon, drought
877 adversely affects the forest carbon balance – and improved understanding of how this occurs.
878 Many uncertainties remain, and estimating the likelihood of basin-scale forest dieback
879 remains challenging.
880

881

882 **V Responses to Ocean Acidification**

883

884 **1 Introduction**

885 Increased concentrations of atmospheric CO₂ reduce seawater pH, increase the solubility of
886 calcium carbonate (reducing saturation state), and cause other chemical changes, together
887 known as ocean acidification. The biogeochemical, ecological and societal implications of
888 ocean acidification have received greatly increased research attention during the past decade
889 (Riebesell and Gattuso, 2015; Mathis et al., 2015). Ocean acidification risks and impacts
890 were included as a component of climate change in the IPCC's Fourth Assessment Report,
891 with more detailed analyses in the Fifth Assessment Report, particularly by Working Group
892 II (Portner et al., 2014).

893 Analyses of geological ocean acidification events and modelling studies show that physico-
894 chemical recovery from perturbations in ocean carbonate chemistry of similar magnitude to
895 projected changes takes many thousands of years (Zeebe and Ridgwell, 2011), due to slow
896 rates of deep ocean mixing and of chemical equilibration with seafloor sediments. The rate
897 of CO₂ increase today is estimated to be around 10 times faster than any natural ocean
898 acidification event during the past 66 million years (Honisch et al., 2012; Zeebe et al., 2016).
899 The longterm hysteresis effects are inherent in the response of global ocean chemistry to
900 atmospheric CO₂ forcing, and there is only very limited capacity to accelerate future recovery
901 by actively removing CO₂ from the atmosphere (Mathesius et al., 2015). Species' extinctions
902 are necessary irreversible.

903 Many different thresholds for ocean acidification impacts can be considered under conditions
904 of steadily increasing atmospheric CO₂ levels; the focus here is on increased solubility of

905 calcium carbonate (in particular, the saturation state for aragonite, the form of carbonate in
906 the shells and structures of many marine organisms) and the risk of rapid loss of tropical
907 corals.

908 **2 Observed recent changes**

909 The IPCC Fifth Assessment Report (Rhein et al., 2013) provided decadal measurements of
910 ocean carbonate chemistry in near-surface waters at three oceanic monitoring sites; and other
911 datasets are also now available (WMO, 2014; Bates, 2017). All these observations
912 unequivocally show decreasing pH in the upper ocean at rates (-0.0011 to -0.0024 yr^{-1})
913 closely matching those expected from rising atmospheric CO_2 . Both physical and biological
914 factors are responsible for the spatial and temporal variability in these datasets; whilst
915 seasonality is usually smoothed-out for trend analyses (WMO, 2014), it is of high ecological
916 importance, determining the conditions experienced by marine organisms (Sasse et al., 2015).
917 There is much less temporal variability of pH in the ocean mid-waters and at greater depth;
918 however, there are also fewer longterm measurements. Atlantic observations (Woosley et al.,
919 2016) confirm an anthropogenically-driven decrease in surface pH of ~ 0.0021 yr^{-1} with
920 greatest changes in the top $\sim 1000\text{m}$; however, some decrease also occurs at greater depths.
921 Such changes are superimposed on a natural decrease of pH with depth, with North Atlantic
922 seafloor values generally being in the range $7.70 - 7.75$ (Vazquez-Rodriguez et al., 2012)
923 compared to a global mean surface value of ~ 8.1 , and typical seasonal ranges of $7.9 - 8.3$.

924

925 Correlations between observed ocean acidification and biological or ecosystem changes are
926 not necessarily causal, since other environmental factors are also likely to be involved. The
927 strongest observational evidence relates to ocean acidification effects on pteropods
928 (planktonic snails) in the Southern Ocean and northeast Pacific (Bednarsek et al., 2014a;

929 Bednarsek et al., 2012; Bednarsek et al., 2017); on cultivated oysters (Barton et al., 2015); on
930 warm-water corals, and at natural CO₂ vents (discussed below).

931

932 Longterm reductions of up to ~30% in the natural calcification and growth rates of tropical
933 corals have been reported in several studies (e.g. Silverman et al., 2014). Linkage to ocean
934 acidification has been demonstrated by in situ treatments of a natural coral community in the
935 Great Barrier Reef (Albright et al., 2016). When water chemistry was restored to pre-
936 industrial conditions by short-term alkalinity enrichment, coral growth rates increased by
937 ~7%.

938 Observations at natural, shallow-water CO₂ vents consistently show marked decreases in
939 overall biodiversity as pH declines (Hall-Spencer et al., 2008; Fabricius et al., 2011; Gambi et
940 al., 2016). Microbes in sediment are also affected (Raulf et al., 2015). Non-calcifying
941 seaweeds and sea grasses out-compete calcifying organisms under such high CO₂, low pH,
942 conditions, although some genetic adaptation of the latter can occur (Garilli et al., 2015).

943 **3 Potential for significant change**

944 Experimental studies have shown that many marine species are likely to be negatively
945 affected from future ocean acidification if high CO₂ emissions continue, with risk of
946 ecosystem alterations at the global scale (CBD, 2014; Gattuso et al., 2015; Nagelkerken and
947 Connell, 2015). Taxonomic variability in biotic responses to ocean acidification is, however,
948 high. Furthermore, many interactions occur with temperature, food availability and other
949 stressors (Wittmann and Portner, 2013; Ramajo et al., 2016; Kroeker et al., 2013; Kroeker et
950 al., 2017); responses may be sex-specific (Ellis et al., 2017); and impacts on behaviour,
951 competition and predator-prey relationships are complex (Nagelkerken and Munday, 2016;
952 Nagelkerken et al., 2017). Whilst the potential for evolutionary adaptation is largely

953 unknown (Sunday et al., 2014) , the sensitivity of populations could be shaped by regional
954 adaptation to local conditions causing differences between geographically separated
955 populations of the same species (Calosi et al., 2017).

956

957 Marine ecosystems are susceptible to non-linear changes occurring over just a few years
958 (regime shifts; Mollmann et al., 2015) that cannot be easily reversed once thresholds, that
959 may be of different kinds, are exceeded (Mumby et al., 2011; Hughes et al., 2013; Plaganyi et
960 al., 2014). Two such ocean acidification-related thresholds were projected (Steinacher et al.,
961 2013) in the context of allowable carbon emissions: aragonite undersaturation in the Southern
962 Ocean, and the carbonate chemistry conditions necessary for warm-water coral reef survival.

963

964 Hauri et al. (2016) used a multi-model ensemble to determine changes in aragonite saturation
965 state (Ω) around Antarctica and southern South America in an unabated CO₂ emissions
966 scenario (RCP 8.5). The monthly occurrence of aragonite undersaturation ($\Omega < 1.0$) at the
967 surface and at 100m water depth increased rapidly in most of these areas (Figure 4),
968 particularly between 2040 - 2070 when atmospheric CO₂ levels are projected to be 500 - 650
969 ppm.

970

971 Similar effects are projected for the Arctic Ocean, where all surface waters north of 66° are
972 projected to be unsaturated for aragonite by 2100 under RCP 8.5 (Popova et al., 2014; Qi et
973 al., 2017). Regional differences are, however, greater - with surface undersaturation expected
974 to have already occurred in the Siberian shelves and Canadian Arctic Archipelago (i.e. with
975 current atmospheric CO₂ values of ~400 ppm), but not until the 2080s in the Barents and
976 Norwegian seas (at ~ 900 ppm). The ecological significance of aragonite unsaturation is that

977 such conditions are chemically corrosive to unprotected shells made of that form of
978 carbonate, e.g. those of pteropods (Bednarsek et al., 2014b; Bednarsek et al., 2017).

979

980 Coral exoskeletons are also made of aragonite: the depth distribution of coldwater corals is
981 closely correlated with the aragonite saturation horizon (Guinotte et al., 2006; Jackson et al.,
982 2014), whilst the calcification rate of both coldwater and tropical corals is sensitive to
983 saturation state, responding semi-linearly over a wide range of values (McCulloch et al.,
984 2012; Comeau et al., 2013). The dead unprotected reef-like structures of coldwater corals are
985 especially susceptible to dissolution (Hennige et al., 2015).

986

987 Most tropical coral reefs occur in waters where $\Omega > 3.0$ (Manzello et al., 2014; Mongin et al.,
988 2016), and that value has been used as a threshold for modelling climate change impacts
989 (Steinacher et al., 2013). Whilst tropical coral growth can continue where $\Omega < 3.0$ (Comeau
990 et al., 2013; Shamberger et al., 2014), growth rates need to exceed bioerosion (Andersson and
991 Gledhill, 2013) and to be sufficiently rapid to allow reef recovery between temperature-
992 induced bleaching events (Frieler et al., 2013). In theory, tropical corals could avoid the risk
993 of bleaching by colonizing new sites where water temperatures have previously been too cool
994 (Couce et al., 2013). However, the rate of current change may be too rapid for that to occur –
995 and there are many geological precedents for ‘coral reef crises’, involving mass extinctions
996 during geological warming and/or ocean acidification events (Kiessling and Simpson, 2011).
997 Based on these considerations, many coral researchers consider atmospheric levels of ~350
998 ppm CO₂ to be the ‘safe’ limit to ensure coral reef survival (ISRS, 2015).

999 **4 Potential consequences**

1000 The potential consequences of future ocean acidification are extremely wide-ranging,

1001 particularly for high emission scenarios. They include physico-chemical impacts (reduction
1002 in seawater capacity to absorb further CO₂); species-specific physiological and behavioural
1003 changes; perturbations in marine community processes, ecosystem functions and
1004 biogeochemical feedbacks; and changes in ocean ecosystem services, with societal effects on
1005 food security, coastal protection and climate regulation. The scale of the biological and
1006 socio-economic changes is, however, uncertain.

1007

1008 An overall reduction in marine diversity and abundances is expected to occur in a high CO₂
1009 world (Nagelkerken and Connell, 2015); nevertheless, not all species will be negatively
1010 affected. Some marine species that may be favoured also provide societal benefits, e.g. sea-
1011 grasses (Garrard and Beaumont, 2014), but not all. Thus ‘nuisance’ species, such as jellyfish,
1012 seem generally tolerant of ocean acidification (Hall-Spencer and Allen, 2015).

1013

1014 With regard to the carbonate undersaturation threshold identified above, the loss of pteropods
1015 from polar oceans would have wider consequences for food-webs, also affecting higher
1016 predators (fish, seabirds and sea mammals) of high commercial or conservation value, even if
1017 those groups are not directly affected by ocean acidification. Increasing acidification in the
1018 Southern Ocean represents a risk to another key pelagic species, Antarctic krill. The hatch-
1019 rate for krill eggs decreases markedly at pCO₂ values > 1000 µatm (Kawaguchi et al., 2013),
1020 and major reduction in their abundance could also jeopardise the entire ecosystem.

1021

1022 The potential loss of tropical coral reefs would have major consequences for coastal
1023 protection, tourism and fisheries, with the global economic value of those ecosystem services
1024 estimated to be up to ~ \$1000 billion per year (Brander et al., 2012). However uncertainties

1025 in economic costs are high, and many other factors, in addition to ocean acidification, are
1026 affecting the future health and survival of coral reefs.

1027 **5 Cautions (uncertainties).**

1028 Many uncertainties remain regarding ocean acidification impacts in the context of specific
1029 thresholds (Pandolfi, 2015) and interactions with other stressors (CBD, 2014; Gattuso et al.,
1030 2015). The scaling-up of impacts from organisms to communities, food webs, ecosystems
1031 and economic impacts is challenging (Andersson et al., 2015; Ekstrom et al., 2015; Turley,
1032 2017) – particularly since ocean acidification impacts do not act on their own, but co-occur
1033 with other stressors, both climate-related (warming, de-oxygenation and sea-level rise)
1034 (Gattuso et al., 2015; Howes et al., 2015; Kroeker et al., 2017) and non-climate-related
1035 (pollution, over-fishing and habitat loss) (Breitburg et al., 2015). Furthermore, coastal
1036 ecosystems seem likely to be at greatest risk from ocean acidification, but these are inherently
1037 complex and difficult to simulate in models because of interactions with sediment processes
1038 and riverine inputs (Artioli et al., 2014), and other factors causing local variability in
1039 carbonate chemistry(Chan et al., 2017).

1040 **6 Comparison with AR5**

1041 Since IPCC AR5, many ocean acidification studies have demonstrated variability in
1042 environmental conditions and biological responses, and the complexity of multi-stressor
1043 interactions. Such research therefore may seem to have increased, rather than reduced
1044 uncertainty. Nevertheless, understanding of ocean acidification and its impacts has
1045 significantly improved: observations have greater geographical coverage, integrating
1046 chemical and biological measurements, whilst new meta-analyses and assessments have
1047 confirmed previously-identified patterns and have also provided additional insights.
1048 Furthermore, greater attention has been given to important topics such as palaeo- ocean

1049 acidification events; socio-economic modelling; acclimatization and adaptation; and the
1050 vulnerability of cold-water corals.

1051 Many of those more recent studies relate to the thresholds outlined here. In particular, there
1052 is now greater confidence that extensive aragonite undersaturation, with major ecological
1053 consequences, would occur throughout the water column in high latitudes within a few years
1054 of atmospheric CO₂ exceeding 450-500 ppm, and that warming will need to be well below
1055 2K to avoid damaging interactions between ocean acidification and temperature for tropical
1056 coral reefs.

1057

1058

1059

1060 **VI Conclusions**

1061

1062 This report reviews the major new advances in understanding of four systems with potential
1063 for climate thresholds, focussing on progress since IPCC AR5. Advances are reported in the
1064 context of observed recent changes, the potential for significant change, and the associated
1065 consequences. The key findings are summarised in Table 1. Overall, compared to AR5, a
1066 large number of studies have added further detail to our understanding of these systems, but
1067 the broad headline summaries of AR5 have not greatly changed.

1068

1069 Declines have been observed (in the Greenland and West Antarctic ice sheets, the AMOC,
1070 and ocean corals) that could be partly driven by anthropogenic activity, although the role of
1071 natural variability is uncertain. For the West Antarctic Ice Sheet, some degree of irreversible
1072 collapse may already have begun. For tropical forests the picture is more mixed, with some
1073 long-term increases in carbon storage, but also evidence of a more recent weakening in the
1074 Amazon carbon sink.

1075

1076 For various reasons, long term maintenance of detailed observing systems is critical. Early
1077 warning of approaching thresholds may be possible, as well as attribution of change to
1078 anthropogenic or natural drivers. Further observations are also needed to improve the
1079 models. Current numerical models have improved, but still suffer from biases, and lack key
1080 processes or sufficient spatial resolution. Detailed process-based observations are needed, to
1081 separate different drivers, and mechanisms of response, and forced change from internal
1082 variability. In each of the four systems, there are a range of drivers of change (e.g. CO₂,
1083 atmospheric temperature, regional ocean temperatures - affected by ocean circulation as well

1084 as large scale warming, surface winds, precipitation, fire and atmospheric composition).
1085 Further, the systems can have different mechanisms of response (e.g. dynamical thinning of
1086 outlet glaciers versus surface mass balance for Greenland; or, for tropical forests, productivity
1087 versus mortality, and also changes in allocation of new carbon and inter-species competition).
1088 For tropical forests and ocean biological organisms, the potential for evolutionary adaptation
1089 is a key unknown. Field experiments have provided key information for tropical forests and
1090 ocean acidification, and more are required.

1091
1092 For these systems, there is only limited quantitative information about the difference, in
1093 likelihood of crossing a threshold, between futures reaching 1.5 and 2K global-mean
1094 warming above pre-industrial levels. For ice-sheets and the effects of ocean acidification
1095 (combined with warming) on marine ecosystems, it is reasonable to assume that the
1096 likelihood of crossing a critical threshold is higher for a 2K world than a 1.5K world. For
1097 Greenland, rates of mass loss and sea level rise are a non-linear function of the temperature
1098 increase because of the combined effect of dynamic thinning at the margins and the
1099 temperature-elevation feedback (Applegate et al., 2015). A simplified model study of this ice
1100 sheet suggested that the global-mean warming threshold for irreversible loss could be only
1101 0.8–3.2K (best estimate 1.6°C) above pre-industrial (Robinson et al., 2012); while one long-
1102 term coupled model simulation found the threshold of zero surface mass balance may be
1103 crossed somewhere between 2 and 3K above pre-industrial levels (Vizcaino et al., 2015). For
1104 ocean acidification, there is now greater confidence that extensive aragonite undersaturation
1105 (with major ecological consequences) will occur in high latitudes if atmospheric CO₂ exceeds
1106 450-500 ppm, and that warming will need to be well below 2K to avoid risk of damaging
1107 interactions between ocean acidification and temperature for tropical coral reefs.

1108

1109 For the ice sheets, AMOC and tropical forests, the potential consequences of crossing a
1110 threshold (section 4 for each system - e.g. sea-level rise from decline in ice sheets) are in
1111 general better constrained than the likelihood (or timing) of crossing the threshold under
1112 particular forcing scenarios. Given this, we suggest that the risk of collapse in such systems
1113 could be managed by ongoing detailed monitoring, including of variables that might give
1114 early warning of collapse; and by assessment of the potential timescales and impacts of
1115 collapse using theory and models (however, for ocean acidification, while there is a real risk
1116 of crossing thresholds in ecosystems this century, the potential impacts are complex and
1117 poorly understood, due to possible interactions amongst different species). Ongoing model
1118 development and analysis will help target observations and will improve our understanding of
1119 the likelihood of collapse. These recommendations are similar to those of NRC (2013), with
1120 the additional focus on timescales and impacts of collapse.

1121

1122

1123 **Acknowledgements**

1124

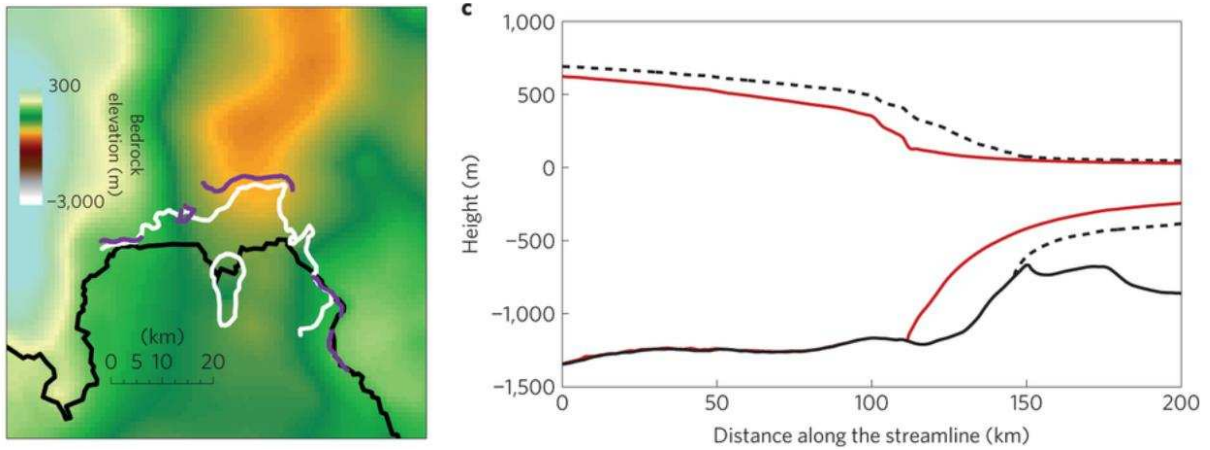
1125 This work was supported by the Joint UK DECC/Defra Met Office Hadley Centre Climate
1126 Programme (GA01101). MAS was funded by the NERC RAPID-AMOC programme. OLP
1127 was supported by the European Research Council Advanced Grant, ‘Tropical Forests in the
1128 Changing Earth System’. PW and CT were supported by the UK Ocean Acidification
1129 research programme funded by NERC, Defra and DECC. BK, KH and GK were funded by
1130 European Union’s Seventh Framework Programme AMAZALERT project (282664). AH
1131 was funded under NERC grant NE/K016016/1: Process-based Emergent Constraints on
1132 Global Physical and Biogeochemical Feedbacks and by the Joint Weather and Climate
1133 Research Programme (JWCRP)’.

1134

1135

1136 **Figures**

1137



1138

1139 Figure 1. Evidence (from Favier et al. 2014) that the Pine Island Glacier's grounding line is

1140 probably engaged in an unstable 40 km retreat, due to the retrograde bedrock slope. Left:

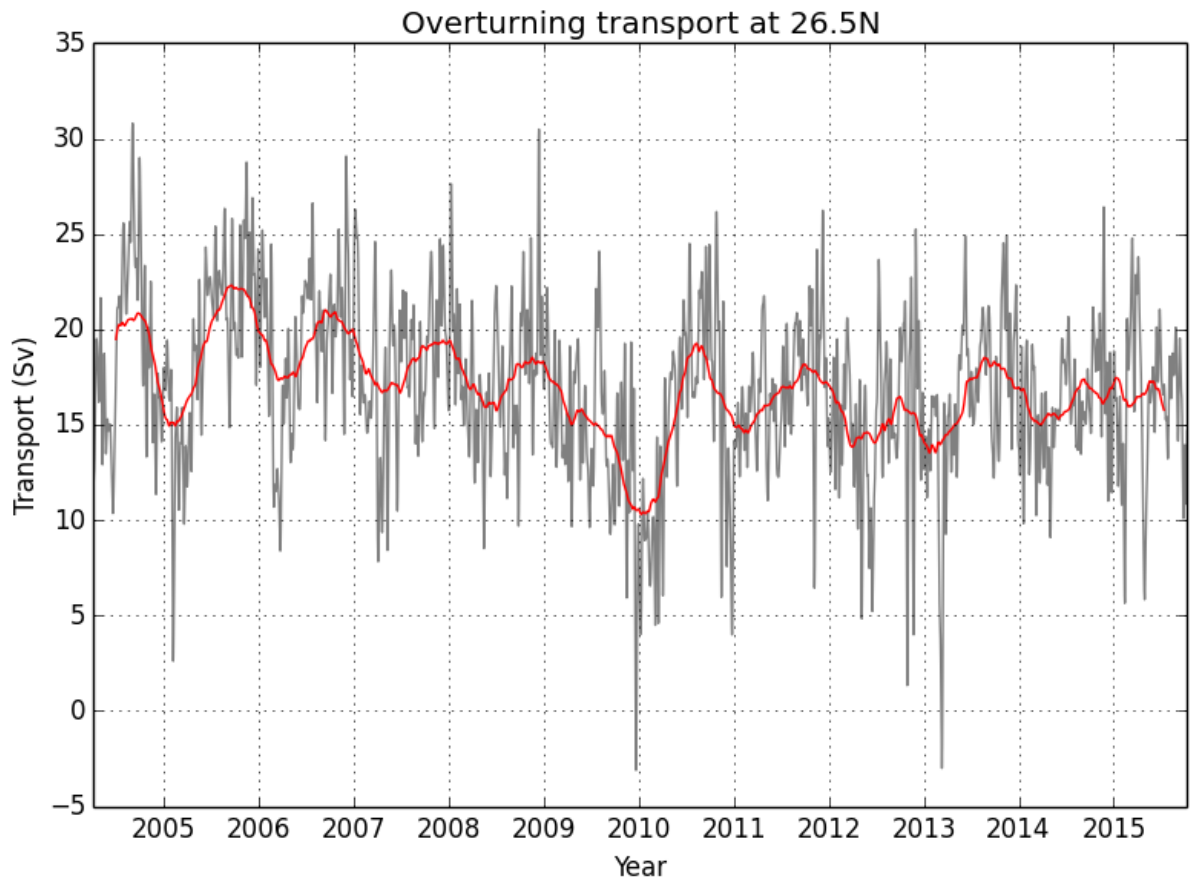
1141 map of bedrock elevation, with the grounding lines for 2009 (white) and 2011 (purple)

1142 shown. Right: bedrock height (solid black line) and geometry of the glacier centreline

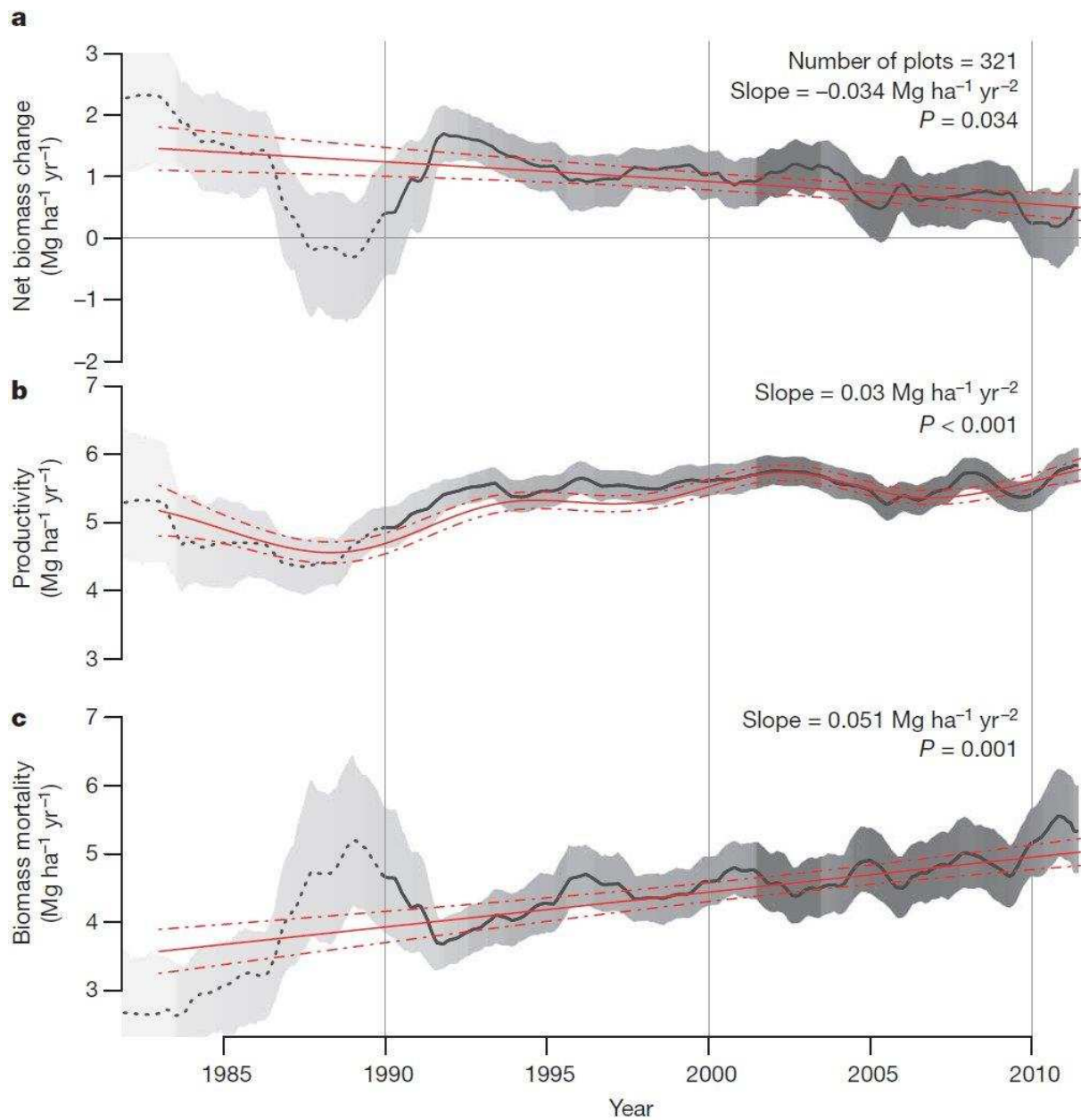
1143 produced by the Elmer/Ice ice-flow model at time (t) = 0 (dotted line) and after 50 years of a

1144 melting scenario (red line).

1145



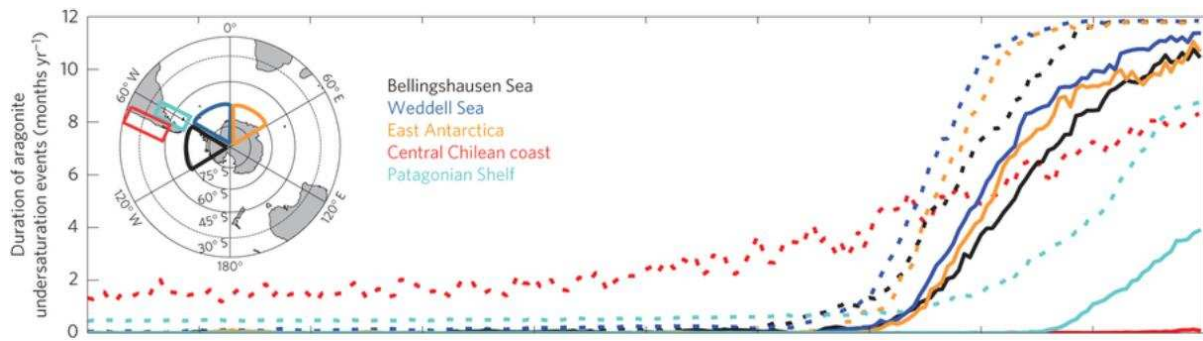
1146
1147 Figure 2. AMOC transport measured at 26.5°N (Smeed et al., 2016). The gray line represents
1148 the 10 day filtered measurements, while the red line was produced using a 180 day running
1149 mean. Clearly visible are the low AMOC event in 2009-10 and the overall decrease in
1150 strength over the measurement period.
1151



1152

1153 Figure 3. Trends in net above-ground biomass change, productivity and mortality rates, for
 1154 321 plots, weighted by plot size (after Brien et al. 2015).

1155



1156

1157 Figure 4. Area-weighted ensemble mean duration (months per year) of aragonite
 1158 undersaturation at the surface (solid lines) and at 100m depth (dashed lines) for three sectors
 1159 of the Southern Ocean (Bellingshausen Sea, Weddell Sea and East Antarctica), the central
 1160 Chilean coast and the Patagonian shelf over the period 1900-2100, with future projections
 1161 based on RCP 8.5. From Hauri et al. (2016).

1162

System	Key findings
Ice sheets	<ul style="list-style-type: none"> • From Greenland, the proportion of loss from surface melt has increased, becoming more consistent with long term model projections. The bedrock topography of the WAIS lends itself to an inherently unstable ice sheet. Some degree of irreversible loss may have begun, although the eventual magnitude and rate of this irreversible loss is uncertain. • There are indications that the East Antarctic Ice Sheet (EAIS) and the northeast Greenland ice stream may be more sensitive to climate change than previously expected. • New paleoclimate evidence for: 1) periods of relatively abrupt Antarctic mass loss following the last glacial maximum; 2) during the early Holocene (sustained warming ~2K above pre-industrial), WAIS mass loss rates comparable to present-day, but no WAIS collapse. • Modelling studies indicate that ice sheet mass loss can be largely avoided under the RCP2.6 scenario. • Significant loss from WAIS will occur on timescales of 100-1000 years.
AMOC	<ul style="list-style-type: none"> • The observed AMOC overturning has decreased from 2004-2014, linked with decreases in subsurface density in the subpolar gyre. It is unclear at this stage whether this AMOC decrease is forced or is internal variability.

	<ul style="list-style-type: none"> • There was an unprecedented rise in US east coast sea level associated with the 2009-10 downturn in the AMOC (both of which subsequently recovered).
Tropical forests	<ul style="list-style-type: none"> • Greater confidence that tropical forests are adversely affected by drought. • New climate models continue to suggest that basin-scale Amazon dieback from climate alone (as in an early study) is not typical. However, these studies lack some key processes. • There remains a high level of uncertainty regarding future changes
Ocean acidification	<ul style="list-style-type: none"> • Global trends in ocean acidification driven by increasing CO₂ concentrations are superimposed on a dynamic natural system • Many factors affect variability in biological response; these are now much better understood • Extensive aragonite undersaturation in high latitudes can be expected if atmospheric CO₂ exceeds 450-500 ppm, with effects on key zooplankton and marine food-webs • Tropical coral reefs seem highly vulnerable to the interaction of ocean acidification and warming, with major economic consequences relating to coastal erosion, storm protection, fisheries and tourism.

1164 Table 1. Key new findings, for each system.

1165

1166

1167

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1960

1961 **Figure captions**

1962

1963 Figure 1. Evidence (from Favier et al. 2014) that the Pine Island Glacier's grounding line is
1964 probably engaged in an unstable 40 km retreat, due to the retrograde bedrock slope. Left:
1965 map of bedrock elevation, with the grounding lines for 2009 (white) and 2011 (purple)
1966 shown. Right: bedrock height (solid black line) and geometry of the glacier centreline
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1971 the 10 day filtered measurements, while the red line was produced using a 180 day running
1972 mean. Clearly visible are the low AMOC event in 2009-10 and the overall decrease in
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1975 Figure 3. Trends in net above-ground biomass change, productivity and mortality rates, for
1976 321 plots, weighted by plot size (after Brienen et al. 2015).

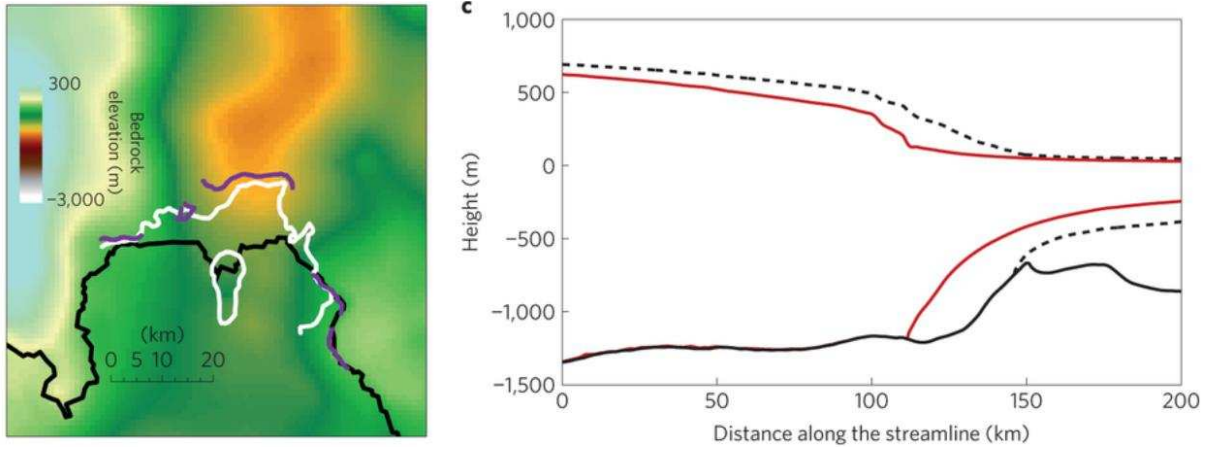
1977

1978 Figure 4. Area-weighted ensemble mean duration (months per year) of aragonite
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1980 of the Southern Ocean (Bellingshausen Sea, Weddell Sea and East Antarctica), the central
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1982 based on RCP 8.5. From Hauri et al. (2016).

1983

1984 **Figures**

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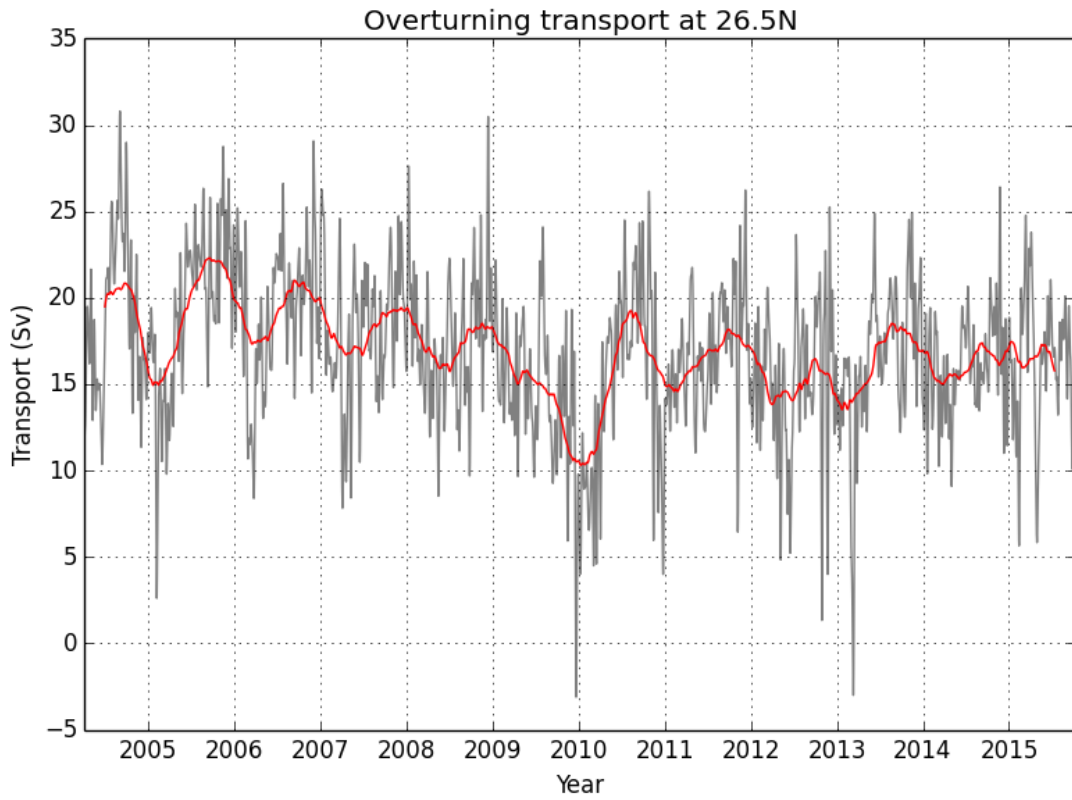
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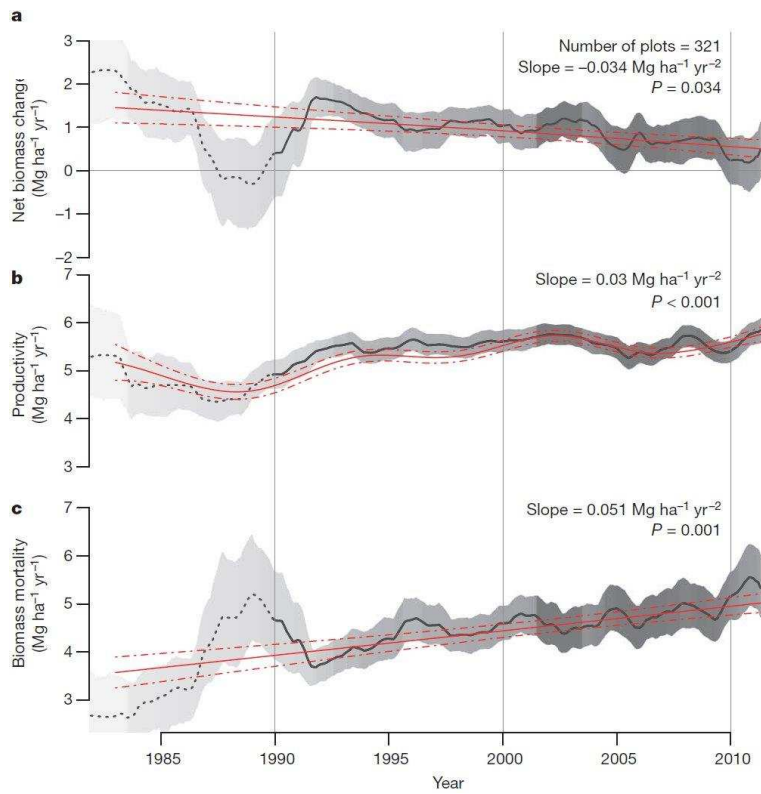
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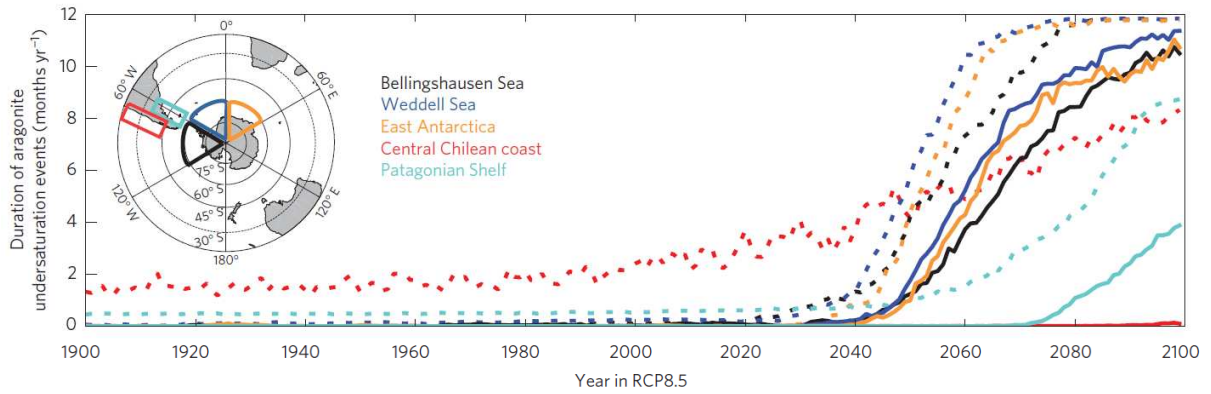
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