

Regional climate impacts of future changes in the mid-latitude atmospheric circulation: a storylines view

Article

Accepted Version

Zappa, G. (2019) Regional climate impacts of future changes in the mid-latitude atmospheric circulation: a storylines view. *Current Climate Change Reports*, 5 (4). pp. 358-371. ISSN 2198-6061 doi: <https://doi.org/10.1007/s40641-019-00146-7> Available at <https://centaur.reading.ac.uk/86475/>

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To link to this article DOI: <http://dx.doi.org/10.1007/s40641-019-00146-7>

Publisher: Springer

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1 **Regional climate impacts of future changes in the mid–latitude**
2 **atmospheric circulation: a storylines view**

3 **Giuseppe Zappa**

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5 Received: date / Accepted: date

6 **Abstract**

7 **Purpose of review.** Atmospheric circulation exerts a strong control on regional climate and
8 extremes. However, projections of future circulation change remain uncertain, thus affecting the as-
9 sessment of regional climate change. The purpose of this review is to describe some key cases where
10 regional precipitation and windiness strongly depend on the mid–latitude atmospheric circulation
11 response to warming, and summarise this into alternative plausible storylines of regional climate
12 change.

13 **Recent findings.** Recent research has enabled to better quantify the importance of dynamical
14 aspects of climate change in shaping regional climate. The cold season precipitation response in
15 Mediterranean–like regions is identified as one of the most susceptible impact–relevant aspects
16 of regional climate driven by mid–latitude circulation changes. A circulation–forced drying might
17 already be emerging in the actual Mediterranean, Chile and southwestern Australia. Increasing

G. Zappa

Department of Meteorology, University of Reading, Reading, UK

Tel.: +44 (0)118 378 6016

E-mail: g.zappa@reading.ac.uk

18 evidence indicates that distinct regional changes in atmospheric circulation and European windi-
19 ness might unfold depending on the interplay of different climate drivers, such as surface warming
20 patterns, sea ice loss and stratospheric changes.

21 **Summary.** The multi-model mean circulation response to warming tends to show washed-out sig-
22 nals due to the lack of robustness in the model projections, with implications for regional changes.
23 To better communicate the information contained within these projections, it is useful to discuss
24 regional climate change conditionally on alternative plausible storylines of atmospheric circulation
25 change. As progress continues in understanding the factors driving the response of circulation to
26 global warming, developing such storylines will provide end-to-end and physically self-consistent
27 descriptions of plausible future unfoldings of regional climate change.

28

29 **Keywords** regional climate change · atmospheric circulation · CMIP5 · Mediterranean climates ·
30 storylines · precipitation projections

31 1 Introduction

32 In the last decade consensus has started to grow on how atmospheric circulation will respond to
33 global warming [1]. On average, climate projections from multi-model ensembles indicate an over-
34 all poleward shift of the mid-latitude westerlies [2], associated with a poleward expansion of the
35 Hadley circulation [3], and a reduction in the number of extratropical cyclones [4,5]. Changes con-
36 sistent with the models' projections are starting to be observed in the real world but, due to the
37 large year to year internal variability in the climate system, not even the observed trends in the
38 zonal-mean aspects of atmospheric circulation have yet been unequivocally attributed to warm-
39 ing [6]. Nonetheless, if greenhouse gas emissions are not mitigated and climate model projections
40 are realised, future changes in the atmospheric circulation will not pass unobserved. In the mid-
41 latitudes, atmospheric circulation determines the trajectory of weather systems and their associated
42 precipitation and wind speed extremes [7]; it stirs the transport of moisture from dry to wet regions
43 [8]; it drives hot extremes in summer and cold extremes in winter through the establishment of
44 persistent anticyclones [9]. As such, atmospheric circulation change can have a diverse range of
45 societal impacts.

46 Despite its potential to drive climate impacts, projections of circulation change have not yet
47 translated into high confidence statements on regional climate change [10]. This lack of confidence
48 depends on multiple causes. Zonal-mean aspects of circulation change, such as the Hadley cell
49 expansion, are not sufficient to constrain the response of regional climate over land [11,12]. At the
50 regional and seasonal scale the uncertainty in how atmospheric circulation responds to warming
51 remains large, to the extent that different models can even show opposite forced responses [13].
52 It would be tempting to treat these uncertainties probabilistically and to take the multi-model
53 mean as a best projection, but such an approach is not supported on firm theoretical grounds [14].
54 While the multi-model mean usually outperforms individual models in global metrics of climate,

55 this is not typically the case for regional aspects of atmospheric circulation, which can be better
56 represented in some individual climate models than in the multi-model mean [15]. Different features
57 of regional circulation change tend to be averaged out, leading to an overly smooth and possibly
58 too weak signal.

59 Given this uncertainty in the response of regional climate to global warming, the development of
60 storylines, or narratives, of climate change has been proposed as an informative way to characterise
61 and communicate future climate projections to stakeholders and policy makers [16, 17]. By storyline
62 is meant a possible and physically self-consistent future unfolding of global and regional climate
63 events. In a storyline approach, multiple storylines are identified in order to span the uncertainty in
64 the future projections from multi-model ensembles. However, the focus is not placed on attributing
65 a probability to the different storylines but on understanding the driving physical factors, the chain
66 of mechanisms involved and the implications at the regional level.

67 This report aims to review the future impacts that might unfold from atmospheric circulation
68 change. Thinking in terms of storylines therefore becomes particularly useful and it naturally leads
69 to frame the problem in terms of the following question: what additional information could be gained
70 at the regional level if the response in the large-scale atmospheric circulation were known? After
71 reviewing some methodological aspects (section 2), this question will be discussed for a selection
72 of regional climate impacts associated with changes in precipitation (section 3) and windiness
73 (section 4). For each case, based on published literature, different plausible storylines of atmospheric
74 circulation change will be analysed. Conclusions are presented in section 5.

75 **2 How to identify regional impacts of mid-latitude circulation change**

76 Developing physically self-consistent storylines of atmospheric circulation change relies on having
77 a causal understanding of the chain of mechanisms involved. Achieving this understanding requires

78 tackling two separate problems. At a global level, the challenge lies in understanding what climate
79 aspects, e.g. sea surface temperature patterns and sea ice, drive the uncertainty in the regional
80 response of atmospheric circulation. Identifying such drivers ultimately requires numerical exper-
81 imentation [18]. Furthermore, at the regional level, an additional challenge lies in understanding
82 the impacts of the response of atmospheric circulation for regional climate change. This requires
83 separating the other aspects of regional climate change that directly result from energy imbalance
84 and surface warming, often called thermodynamic aspects [19]. A clean separation is generally
85 not possible. The different methods either attempt to directly quantify the regional changes due
86 to circulation, or, conversely, to quantify the thermodynamic response expected for no change in
87 circulation, and then define the dynamical part as a residual. Some of these approaches are now
88 discussed.

89 2.1 Internal variability analogs

90 The response of atmospheric circulation to global warming can resemble, or even project on, present-
91 day modes of internal atmospheric variability [20]. In this case, the impacts that future changes
92 in the atmospheric circulation might have on the regional climate can be directly estimated by
93 identifying analogs of the projected circulation change in the present-day observational record. By
94 referring back to the present-day climate, any thermodynamic influence is by construction excluded.

95 This approach has been implemented in several ways. The most direct technique relies on lin-
96 early regressing the circulation response on the dominant modes of variability in the atmospheric
97 circulation, such as obtained via EOF analysis [21–23]. Alternatively, the linearity assumption can
98 be relaxed by describing the circulation response as a change in the frequency of occurrence of
99 present-day weather regimes obtained by clustering algorithms such as k-means or self-organising
100 maps [24, 9, 25–27]. A limit of these approaches is that accurate present-day analogs may not al-

ways be available. For example, in the Mediterranean area in winter, the response of atmospheric circulation to greenhouse forcing does not project on individual present-day modes of internal atmospheric variability [28,23]. Using clustering algorithms does not necessarily address this issue, as global warming can force a change in the structure of weather regimes between the present and the future climate simulations [29,26]. In these cases, variability analogs might only isolate part of the signal associated with future circulation changes. Alternative more flexible strategies have shown potential to address these issues. The “constructed circulation analogs” technique aims to optimally reproduce the circulation response pattern through linear combinations of several analogs extracted from large ensembles of climate simulations or atmospheric reanalyses [30,31]. Techniques based on partial least-squares regression are effective at identifying the atmospheric circulation patterns that exert the largest impacts on a climate aspect of interest [32]. These approaches have shown good skill at capturing the influence of circulation variability and change on surface temperature, and some promising results are also emerging for precipitation [33,34].

2.2 Budget equations

A useful complementary approach to internal variability analogs consists in the inspection of atmospheric budget equations. Two most notable applications have been the use of the moisture budget [35] and of the energy budget [36] to understand variability and change in regional hydro-climate.

The moisture budget equation directly informs on the change in the balance between precipitation and evaporation (P-E) as, in steady state, P-E depends on the transport of moisture from other regions. Part of the impact of the circulation response to warming on P-E can be estimated – assuming linearity – as the change in the moisture transport due to the response in the time-mean winds ($\delta\bar{\mathbf{v}}$) acting on the present-day climatology of moisture ($\overline{Q_p}$), i.e. $\int -\nabla \cdot (\delta\bar{\mathbf{v}} \overline{Q_p}) dz$. This decomposition is particularly informative where the mean circulation dominates the transport of

124 moisture, e.g. in the tropics [37]. In the mid-latitudes, it can also be important to account for
125 variations in the transport of moisture due to transient atmospheric eddies, such as those populat-
126 ing mid-latitude storm tracks. The transport by transient eddies depends on the daily covariance
127 between wind and moisture anomalies, which makes decomposing this term into a dynamical and
128 thermodynamic contribution substantially more challenging [38]. The same linear decomposition is
129 used to estimate the impact of atmospheric circulation change on precipitation using the energy
130 budget equation [39]. However, the relationship between circulation and hydro-climate is less direct
131 than for the moisture budget, as a change in the transport of dry static energy can not only be
132 balanced by changes in condensational latent heat release, i.e. precipitation, but also by changes in
133 the surface sensible and radiative heat fluxes.

134 2.3 Regional climate models

135 Finally, experimental approaches can be used to isolate the relative impacts of atmospheric cir-
136 culation changes and warming on regional climate. Regional climate models (RCMs) have been
137 particularly useful for this purpose. One such example consists in the so-called “pseudo global
138 warming” experiments [40], in which a warming signal is added to the boundary conditions driving
139 a present-day RCM simulation. By construction, the approach isolates the response of regional
140 climate to warming in the absence of changes in the large-scale atmospheric circulation. This en-
141 ables to ask how specific past high-impact weather events might have evolved in a warmer climate
142 [41]. In a similar way, the boundary conditions of a RCM can be modified to incorporate only
143 the projected changes in the atmospheric circulation [42]. Further decompositions of the boundary
144 conditions have been proposed in order to isolate the impact of changes in the mean circulation, in
145 the mean static stability and, as a residual, in the transient eddies [43, 44]. A possible limitation is
146 that changes in surface warming, static stability and circulation are physically connected, so that

147 decomposing the boundary conditions in a way that retains meaningful physical balances requires
148 particular care. Nonetheless, the method offers a unique opportunity to directly test how different
149 changes in the atmospheric circulation and warming may determine the response in regional aspects
150 of climate change.

151 In summary, different methods have different strengths and limitations. No single approach is
152 able to globally and unambiguously define the impact of future circulation changes on regional
153 climate, but confidence can be built by comparing results from different approaches.

154 **3 Impacts of circulation change on regional hydro–climate**

155 What more could be learnt on regional hydro–climate change if the response of atmospheric circu-
156 lation were known? On the one hand, soil moisture drought risk is directly increased by warming
157 through a thermodynamic increase in evapotranspiration [45], although partly balanced by the
158 counter–acting effect of enhanced CO₂ on stomatal closure [46]. On the other hand, mid–latitude
159 precipitation, together with river runoff, are strongly controlled by storms and circulation [47]. These
160 aspects will be reviewed in this section, by considering different storylines of circulation change rel-
161 evant for the precipitation response to warming in three regions with a Mediterranean–like climate:
162 the Mediterranean proper, California and Chile. A comprehensive analysis of hydro–climate vari-
163 ability and change in the Mediterranean–like climates from a multi–model mean perspective is given
164 in reference [48].

165 **3.1 Winter Mediterranean circulation change**

166 The Mediterranean area has long been identified as a “hot-spot” of climate change [49], due to a
167 large projected decline in precipitation, which is of the order of 6% per degree of global warming in
168 the mean of the CMIP5 model projections [1]. Furthermore, a reduction in Mediterranean precipi-

169 tation since 1900 is also revealed by reconstructions from rain gauges. This has led many authors to
170 conclude that the projected precipitation decline and increase in meteorological droughts is already
171 happening [50–53], a finding reported with “medium confidence” in the IPCC 1.5 degrees report
172 [54]. The observed precipitation reduction is largest in the Southern and Eastern Mediterranean in
173 winter. In these areas, the observed precipitation trends largely exceed those projected by the mean
174 of the CMIP5 models [52], possibly because of the influence from internal variability in the atmo-
175 spheric circulation [34]. Crucially, these observed trends might already have led to serious societal
176 impacts, such as the case of the 2006–2009 Syrian drought and civil war [55], thus highlighting the
177 vulnerability of the Mediterranean region to hydro–climate variability and change.

178 Despite this evidence, the real magnitude of the Mediterranean climate change signal is still
179 poorly understood, due to the influence from multi–decadal climate variability [56,34], the ob-
180 servational uncertainty in the precipitation reconstructions [50] and the large uncertainty in the
181 magnitude of the projected precipitation decline. However, up to 80% of the inter–model variance
182 in the precipitation projections is linked to uncertainties in the atmospheric circulation [57,58], and
183 this dependence can be used to define distinct storylines of regional climate change. Based on ref-
184 erence [57], Fig. 1a–b compares future projections (2060–2099) of cold season Euro–Mediterranean
185 precipitation change per degree of global warming evaluated for two subsets of CMIP5 models hav-
186 ing, respectively, the least and most negative change in the 850 hPa zonal wind in North Africa, i.e. a
187 simple index for Mediterranean circulation change. A notable anticyclonic circulation anomaly and
188 a larger and more extensive precipitation reduction, i.e. a high–impact storyline for Mediterranean
189 hydro–climate change, is projected in the subset of models responding with a large long–term circu-
190 lation change (Fig. 1b), while these responses are nearly absent, apart from the East Mediterranean,
191 in the opposite case (Fig. 1a). These differences cannot just be explained by internal variability [57],
192 so that the two storylines reflect different ways in which the atmospheric circulation may respond
193 to warming.

194 The large precipitation reduction in the high-impact storyline can be confidently attributed
195 to the change in the atmospheric circulation [50,59,60]. Associated with the anticyclonic anomaly,
196 climate models project increased atmospheric subsidence and low-level divergence. According to
197 moisture budgets, this mean circulation change is the dominant factor leading to the reduced
198 fresh water availability over land, via increased evaporation and an export of moisture out of the
199 region [59]. At the same time, climate models also project the Mediterranean storm track to become
200 weaker and the number of Mediterranean cyclones to decrease [61,62]. This reduced synoptic cyclone
201 activity causes a reduction in the number of rainy days, which only in the Northern Mediterranean
202 region is compensated by a thermodynamically-driven increase in the precipitation generated by
203 each storm [62]. Moisture budgets and storm tracking approaches hence provide complementary
204 views on the dependence of Mediterranean drying on circulation change.

205 A complete dynamical explanation of this localised anticyclonic response and of its driving
206 factors is yet unavailable. The response only weakly projects on the first two modes of Euro-Atlantic
207 atmospheric variability [23], and changes in the structure of the global stationary waves appear
208 to be involved [15,48]. Nonetheless, a large circulation response, as in the high-impact storyline
209 (Fig. 1b), tends to be favoured by an amplified warming of the tropical upper troposphere and
210 by a strengthening of the NH stratospheric vortex [63]. This interpretation is at least qualitatively
211 supported by experiments with atmospheric general circulation models in which the tropical SSTs
212 [50] and the stratospheric vortex [64] are perturbed in a controlled manner. Amplified tropical
213 warming is found to be particularly linked to East Mediterranean rainfall [63], possibly by inducing
214 a weakening of the Mediterranean storm track, while the stratospheric vortex is mostly linked
215 with Western Mediterranean rainfall, via changes in the position of the North Atlantic storm track
216 [64]. Interestingly, the projected reduction in Mediterranean precipitation per degree of warming is
217 larger in the mean of the CMIP3 models compared to the more recent CMIP5 models (Fig. 12.41 in

218 reference [1]). Understanding such differences across model generations would be important to test
219 the impact of remote climate responses and circulation biases on Mediterranean climate change.

220 3.2 The summer NAO and European rainfall

221 If circulation plays an undisputed role on Mediterranean hydro-climate change in the cold season,
222 its role in the warm season is more subtle, and it is here discussed for comparison.

223 In the warm season, P-E and precipitation are both projected to decline, particularly in Western
224 and Southern Europe [65,59]. At the same time, the North Atlantic jet is robustly projected to shift
225 poleward as part of a positive trend in the summer North Atlantic Oscillation (SNAO) [65,21,66,67].
226 Analyses of moisture budgets identify the change in the mean atmospheric circulation – particularly
227 the northerly flow linked to the positive SNAO trend – as the dominant contributor to the decline of
228 P-E over most of Europe [59]. However, internal variability analogs [65,21], and a RCM experiment
229 [42], suggest that only in Northwestern Europe (including UK, Northern France and Northern
230 Germany) more than 50% of the mean response and inter-model spread in the precipitation change
231 can be attributed to SNAO. Consistent with these findings, the projected precipitation reduction
232 in Southern Europe tends to be comparable in the two sets of CMIP5 climate models featuring the
233 smallest and largest poleward shift in the North Atlantic jet (Fig. 1c-d).

234 These apparently contrasting results can be reconciled in light of the additional warming-
235 mediated processes that contribute to the response of precipitation in the warm season, particularly
236 in Southern Europe [44]. As the land warms, the soil is projected to become drier, leading to a
237 reduction in evapotranspiration and in the surface relative humidity. These local changes conse-
238 quently lead to a reduction in clouds and precipitation, which may further enhance the aridity of the
239 soil through an increase in surface shortwave radiation [42,68]. Model differences in the representa-
240 tion of moisture-feedback and cloud-temperature interactions are responsible for the uncertainty

241 in the magnitude of the precipitation change forced via this mechanism [68]. However, the induced
242 response resembles a suppression of the local hydrological cycle so that, whilst being important
243 for precipitation, it could have only a negligible impact on P-E. This supports the view that the
244 interaction between circulation, clouds and soil moisture would deserve more investigation [69]. For
245 the purpose of developing storylines, dynamic and thermodynamic driving factors would both need
246 to be accounted for to describe possible future changes in European summer hydro-climate.

247 The projected positive SNAO trend is partly linked to the projected weakening of the Atlantic
248 Meridional Overturning Circulation (AMOC) [67], and it has not yet emerged in the observations.
249 On the contrary, the observed SNAO trend has been largely negative since the 1990s [70]. Natural
250 decadal variability in the SSTs associated to the Atlantic Multi-decadal Variability (AMV) could
251 in part explain this mismatch [71,72,22], but it has also been speculated that the negative SNAO
252 trend could be a forced response to sea-ice loss not captured by climate models [70]. As the AMV
253 is now entering a phase reversal [73], new observations will help to evaluate the respective roles
254 played by SST variability and sea-ice loss.

255 3.3 The Pacific jet and California

256 California is the only Mediterranean-like climate where the mean of the CMIP5 model projections
257 indicates a slight wetting rather than a large drying [74,48]. However, the severe multi-year drought
258 of 2011–2016 raised the question of whether the event might have been made more likely by climate
259 change. California precipitation mainly results from winter North Pacific storms tracking eastward
260 toward the US coast [75]. In the drought period, a series of atmospheric ridges formed at the end
261 of the North Pacific storm track, diverting the storms towards higher latitudes [76]. Most studies
262 agree that while warming is likely to have amplified the soil moisture drought by increasing the
263 evaporative demand [77,78], the anomalous ridge was the result of internal climate variability,

264 although partly forced by an enhanced zonal SST gradient in the tropical Pacific ocean [79–81].

265 But could a future less rainy California be entirely excluded?

266 The uncertainty in the precipitation response to climate change in California has been explored
267 in relation to different aspects of atmospheric circulation: the North Pacific subtropical jet [82, 83],
268 the North Pacific storm track [84, 75], the location of the subtropical highs [85] and the stationary
269 waves [15]. These different studies have revealed a coherent picture of how different aspects of
270 circulation interact to generate either drier or wetter conditions in the model projections. Inspired by
271 reference [15], two possible storylines are summarised in Fig. 1e-f, where projections of precipitation
272 change per degree of global warming have been conditioned on the long-term response in the
273 meridional wind at 300 hPa at the western coast of North America. One possibility – a best case
274 storyline for California drought (Fig. 1e) – is that the circulation response to warming will be
275 manifest in a strengthening and eastward extension of the subtropical jet towards North America
276 [82]. The strengthening of the subtropical jet would imply more favourable conditions for low-
277 latitude storm development and hence a downstream southward shift in the storm track [84]. It
278 would also induce a lengthening of the stationary wave pattern, which is associated with a shift of
279 the Aleutian low and a stronger southerly flow on the western coast of North America [15]. In this
280 scenario, the shift in the storm-track activity [84] as well as the increased precipitation generated
281 by each storm [75] can be expected to make California more rainy under climate change (Fig. 1e).
282 The alternative storyline (Fig. 1f) is characterised by a poleward shift of the subtropical highs in
283 both the east and the west North Pacific [85]. In this case, a slight ridge would develop on the
284 western coast of North America, leading to a northerly flow anomaly and a drier California. As for
285 the Mediterranean, changes in stationary waves, rather than the zonal-mean Hadley cell expansion,
286 appear to be important for the hydro-climate response in this region [48].

287 Different hypotheses have been raised on what processes might control these different projec-
288 tions. The extension of the subtropical jet and the trough in the east Pacific resemble the circulation

289 response to El Niño, which would point to the tropical Pacific as a key driver [86]. Consistent with
290 a tropical driving, the southward shift of the jet in the Northeast Pacific only occurs within the
291 slow response to greenhouse forcing, which includes the development of El Niño-like tropical SST
292 anomalies [87]. However, apart from one previous CMIP3 study [88], more recent analyses suggest
293 that the spread in the Northeast Pacific circulation change is linked to uncertainties in the SSTs
294 in the Northwest Pacific, rather than in the tropics [83,86]. These extratropical SST anomalies
295 resemble those characterising the Pacific Decadal Oscillation, and their sign is consistent with a
296 surface forcing of atmospheric circulation via a modulation of the subtropical vertical wind shear
297 [85]. Alternatively, it may be that the tropical Pacific is an important driver, but its influence
298 in the inter-model spread is obscured by confounding factors arising from the different models’
299 basic states. For example, the influence of biases in the climatology of stationary waves [15] and
300 in the teleconnection of ENSO [89] have been suggested to play a role, although these analyses
301 have reached opposite conclusions on whether an East Pacific trough or ridge is more likely under
302 climate change. Reconciling these two results would be important to increase confidence in future
303 projections of North American hydro-climate.

304 As a note of caution, the El Niño-like tropical Pacific SST response projected by the models
305 is not yet backed up by the observational record. In contrast, trends in SST reconstructions show
306 muted warming in the eastern tropical Pacific Ocean, thus leading to an enhancement of the zonal
307 tropical Pacific SST gradient [90]. This could be just an expression of internal variability, but some
308 authors note that the El Niño-like response is a “majority decision” in an area where climate
309 models might not represent all relevant processes [91,90]. If models were systematically wrong, and
310 the forced response was that of an increased zonal SST gradient, California rainfall might follow a
311 different storyline to what current climate models project [76].

312 3.4 The SH jet shift and Chilean drought

313 In the Southern Hemisphere (SH), the mid-latitude atmospheric circulation response to climate
314 change is to a large extent described as a shift towards the positive phase of the Southern Annular
315 Mode (SAM) [92]. The positive trend in the SAM is observed in the reanalyses in all seasons, with
316 a largest trend in the austral summer due to the additional forcing from ozone depletion in the SH
317 polar stratosphere [93,94].

318 The climate impacts due to this forced circulation response can be estimated by considering
319 analogs associated with the observed SAM variability on intra-seasonal and inter-annual timescales.
320 The positive SAM is associated with a poleward shift of the storm track, so that precipitation
321 increases at high-latitudes ($\sim 60\text{S}$ – 70S) and decreases in the mid-latitudes ($\sim 40\text{S}$ – 50S) in all seasons
322 [95]. Its impact in the subtropics ($\sim 30\text{S}$) has instead a strong seasonality [96]. In the austral
323 winter (JJA), via the shift in the storm track, the positive SAM is associated with subtropical
324 precipitation decrease, a process that has contributed to the observed negative trend in winter
325 rainfall in Southwestern Australia [97]. In the austral summer (DJF), the positive SAM leads
326 instead to precipitation increase in various subtropical land regions [93,94], via a dynamically-
327 induced shift of the descending branch of the Hadley cell [95]. For example, the wetting trend of
328 Eastern South America in the last decades of the 20th century has been in part attributed to the
329 positive trend in the SAM [98].

330 A particularly large response from mid-latitude circulation changes is expected to occur in
331 Chile, where the positive SAM leads to year-round dry anomalies shifting from Central (30S – 38S)
332 to Southern (38S – 47S) Chile with the seasonal cycle [99]. It is estimated that from 1960 to 2016
333 rain gauges in Central and Southern Chile have recorded a precipitation reduction of about 2% and
334 5% per decade, respectively [99]. Despite observational uncertainties are substantial, all datasets
335 show a negative precipitation trend in Central Chile over the past century [48]. The size of these

336 precipitation trends are influenced by a recent multi-year drought, but they are largely congruent
337 with the precipitation response expected from the positive trend in the SAM [100]. ENSO and the
338 Pacific Decadal Oscillation also affect precipitation variability in Chile, but they have only played
339 a minor role on the observed precipitation trend compared to the SAM-related shift in the storm
340 track [99].

341 Based on these results, and since changes in the stationary waves are less important than in the
342 NH [48], developing storylines of regional climate change will require accounting for the response in
343 the latitude and strength of the SH jet and storm tracks. For example, Fig. 1g,h compares future
344 projections of annual-mean circulation and precipitation change in South America per degree of
345 global warming for the CMIP5 climate models featuring the smallest and largest long-term poleward
346 shift of the SH jet. Consistent with the expectations from internal variability and moisture budgets
347 [48], the projected Chilean precipitation reduction per degree of warming is larger in the mean of the
348 climate models featuring a large poleward shift in the westerlies (Fig. 1h). This exploratory analysis
349 suggests that quantifying the impact on precipitation due to the uncertainty in the SH atmospheric
350 circulation response is important to develop storylines of future changes in the frequency of Chilean
351 droughts [101]. The response in the SH jet could itself be linked to different remote climate responses.
352 In particular, the uncertainty in the jet latitude depends on the magnitude of the polar stratospheric
353 cooling [102, 103] and of the tropical warming, for example via the response in the cloud cover [104],
354 while the uncertainty in the jet strength has also been linked to the loss of Antarctic sea ice [105].
355 However, particularly in austral winter, the models with an equatorward bias in the latitude of
356 the SH jet tend to project a larger poleward shift in response to climate change. This relationship
357 had been suggested as a way to narrow the uncertainty in the future projections, but the argument
358 previously proposed to explain this dependence has been shown not to hold [106]. Nonetheless, these
359 findings indicate that it is important to consider ensembles of models with a realistic present-day
360 simulation of the SH jet to generate plausible storylines of regional climate change in the SH.

361 **4 Impacts on windiness: the European case**

362 The response of the large-scale atmospheric circulation also has implications for other impact-
363 relevant aspects of extratropical storm tracks. One such aspect is surface windiness associated with
364 intense extratropical cyclones. The mean response of extratropical cyclones to climate change is
365 reviewed in Catto et al. 2019 [107], within this same section of Current Climate Change Reports.
366 Here, the focus is instead placed on discussing the uncertainty in the response, using future changes
367 in European windiness as a case study.

368 Central Europe, including the UK and Northern Germany, is vulnerable to wind-storm damage
369 due to intense North Atlantic extratropical cyclones [108]. As extratropical cyclones grow on the
370 baroclinicity associated with the mid-latitude jet stream [109], intense cyclones are favoured by a
371 strong and zonally-extended North Atlantic jet, as found for positive values of the North Atlantic
372 Oscillation (NAO) [110,111]. Severe European wind-storm damage particularly took place in the
373 1990s, a decade of persistent positive NAO [112], and the possibility of a longer-term GHG-induced
374 upward trend remains open but debated [108,113]. In the future projections, climate models indi-
375 cate a strengthening of the mean westerlies in Central Europe [114]. Several previous studies also
376 reported a future increase in the frequency of moderate to extreme European wind speeds [115-
377 117], with consequences for both wind-storm damage [118] and wind-energy production [119,120],
378 but most analyses of the CMIP5 climate models tend to suggest that these changes are non-robust
379 and often small compared to internal variability [121-124].

380 Since different remote, regional and mesoscale processes can affect the response of intense North
381 Atlantic extratropical cyclones to climate change, thinking in terms of storylines can be particularly
382 suitable to portray different possible scenarios. On the large-scale, the response of mid-latitude
383 storm tracks is affected by the opposite projected changes in the upper- and lower-tropospheric
384 meridional temperature gradients [18]. In the Northeast Atlantic, the impact of the increase in

the upper-tropospheric temperature gradient tends to win over of the reduction of the lower-tropospheric temperature gradient, leading to a net strengthening of the storm track in the multi-model mean [125]. By modulating these temperature gradients, different storylines for the North Atlantic storm track can be considered in relation to the relative magnitude of the tropical upper-tropospheric warming, Arctic warming and polar stratospheric cooling – a measure of stratospheric vortex strength [126,127]. In particular, the response in the strength of the stratospheric vortex can drive an NAO-like uncertainty in the North Atlantic atmospheric circulation [64], with possible implications for European windiness [63]. This is summarised in Fig. 2a–b, where future projections of Euro-Atlantic circulation and windiness change are separately presented for the CMIP5 models with a long-term strengthening and weakening of the vortex. Here, windiness is evaluated as the 98 percentile of daily wind speed at 850 hPa. An increase in European windiness characterises most models featuring a strengthening of the stratospheric vortex (up to about 1.5%/K in Fig. 2b), while it is not found for a weakening of the vortex [63]. Additional storylines have been proposed in relation to the relative magnitude of the tropical versus Arctic warming, as “tropically-amplified” models tend to be associated with a more squeezed and eastward extended jet into Europe [127]. It seems possible that this response could also favour enhanced European windiness, but it remains to be quantified. Taken together, the relative amplitude of the tropical versus Arctic warming, and the change in the stratospheric vortex strength, show promise to characterise the uncertainty in key aspects of the North Atlantic jet response to climate change, such as its zonal extension and waviness [127].

The attribution of the storm-track response to the large-scale drivers discussed above, is complicated by the presence of additional changes in surface baroclinicity within the North Atlantic region. In fact, the uncertainty in the North Atlantic storm track response has also been linked to the magnitude of the weakening of the AMOC, which enhances surface baroclinicity at about 50N by suppressing the warming of northern North Atlantic SSTs [128]. Supporting this pathway,

410 the strength of the North Atlantic storm track increases in experiments inducing a “collapse” of
411 the AMOC [128,129] and it covaries with the AMV on multi-decadal timescales [61,130]. However,
412 climate experiments directly modelling the influence of the projected North Atlantic SST warm-
413 ing patterns provide a less consistent picture: while some studies back up a direct influence on
414 circulation from North Atlantic SSTs [131,132], others only identify a small response [133,134].
415 The latter results would suggest that the weakening of the AMOC is largely communicated via a
416 modulation of remote climate responses, such as the ratio between the Arctic and tropical warming
417 [134]. Indeed, tropically-amplified models tend to have a stronger North Atlantic SST warming hole
418 [127], thus implying it is difficult to separate these drivers on a statistical basis. Reference [131]
419 discusses possible reasons behind the different experimental results on the response of atmospheric
420 circulation to the North Atlantic SST warming pattern.

421 Finally, European windiness also depends on additional uncertainties acting on the cyclone scale.
422 For a given atmospheric flow, the thermodynamic increase in the cyclone-associated precipitation
423 is expected to enhance cyclone growth and propagation speed [135], although this pathway is partly
424 balanced by the increase in atmospheric stratification [136]. The impact of enhanced latent heat
425 release on cyclone growth is unlikely to be fully resolved at the spatial resolution of current climate
426 models [137], thus highlighting the value in studies employing high-resolution models to explore
427 the evolution of historical cyclones in a warmer and moister atmosphere [138]. Furthermore, future
428 projections of changes in cyclone-associated wind speeds are systematically more negative at the
429 surface than at 850 hPa, both in the NH and in the SH [139,124]. It has been hypothesised that
430 this might be linked to the vertical profile of the equator-to-pole baroclinicity change [124], but
431 the role of changes in boundary layer processes should perhaps also be quantified [138].

432 The separation of uncertainties arising from the North Atlantic jet, North Atlantic SSTs and
433 cyclone-associated diabatic process might be questionable, since the jet response is itself influenced
434 by the heat and momentum transport associated with the storm track itself [18]. Nonetheless, the

435 importance of remote SST warming for the North Atlantic upper-tropospheric circulation change
436 [133] suggests there is value in pursuing such an approach. For the purpose of risk assessment,
437 it is the plausibility of high-impact scenarios that is most of interest [140]. Based on the above
438 processes, a worst-case scenario for European windiness change might be expected for a tropically-
439 amplified response with a strengthening of the stratospheric vortex, and an enhanced meridional
440 SST gradient in the North Atlantic. Is such a storyline physically self-consistent? and what regional
441 impacts would it exert if realised? Answering these questions could help to place upper bounds on
442 the future change in European wind-storm risk.

443 5 Conclusions

444 Several aspects of regional climate depend on the response of mid-latitude atmospheric circulation
445 to climate change. The hydro-climate response of Mediterranean-like regions, and European windi-
446 ness downstream of the North Atlantic storm track, are high-impact examples that have received
447 recent attention and have been reviewed in this report. While the internal variability in the at-
448 mospheric circulation is a leading uncertainty in extratropical regional climate change [141], these
449 examples have served to highlight cases where the uncertainty in the forced circulation response is
450 sufficiently large that the magnitude, and sometimes even the direction, of these regional climate
451 trends cannot yet be anticipated, even for a specified level of global warming (Fig. 1 and Fig. 2).

452 To characterise and communicate this uncertainty, it can be useful to identify different physically
453 self-consistent storylines of how atmospheric circulation and regional climate could respond to
454 warming. Crucially, each storyline needs to be enriched by knowledge of the climate responses that
455 force the respective circulation changes via atmospheric teleconnections [63]. The relative amplitude
456 of tropical and Arctic warming [50, 104, 134], the response of the AMOC [128, 142, 67], the patterns
457 of Pacific SST change [83, 85, 76], and changes in stratospheric vortex strength [64, 103], have here

458 been discussed as possible drivers of the regional climate responses reviewed in this report. Given the
459 uncertainty in these climate responses, the alternative storylines discussed here cannot be discarded
460 and in the present state of knowledge should be considered equally plausible future manifestations
461 of regional climate change.

462 Having confidence in physical storylines requires nonetheless substantial care. In particular, the
463 response of the atmospheric circulation to the remote climate responses, the absence of confounding
464 influences from the models' biases, and the robustness of the storylines across different model
465 generations need to be thoroughly tested. Furthermore, while storylines help to characterise different
466 high-impact future scenarios, they do not immediately enable to reduce the uncertainty in the
467 projections themselves. The analysis of physical relationships between the climate change response
468 and the model biases in the simulation of present-day [143] and past [144,145] climates is needed
469 for making progress, and potentially deem some storylines implausible.

470 While recent works have focused on the response of the mean state to climate change, future
471 research should aim to characterise the dependence of the full range of regional climate variability
472 on remote drivers of atmospheric circulation [146,147]. For example, a recent study based on a
473 single model found that winters with extremely high and low California precipitation could both
474 become more frequent in response to warming due to changes in the amplitude of the year-to-year
475 variability in atmospheric circulation [148]. Making progress in this direction will necessarily require
476 comparing large initial condition ensembles from different climate models. Producing such datasets
477 will be invaluable for advancing research on regional climate change.

478 **6 Appendix: CMIP5 models**

479 Based on their atmospheric circulation response in the RCP8.5 scenario, the following CMIP5
480 climate models have been identified to produce the panels in Fig. 1 and Fig. 2:

- 481 – **Winter Mediterranean circulation change.** Weak anticyclonic response: GISS-E2-H, bcc-
 482 csm1-1-m, CMCC-CESM, MRI-CGCM3. Strong anticyclonic response: GFDL-CM3, IPSL-CM5A-
 483 LR, MIROC-ESM-CHEM, FIO-ESM.
- 484 – **Summer North Atlantic jet shift.** Small poleward shift: GISS-E2-H, GISS-E2-R, IPSL-
 485 CM5A-LR, GFDL-ESM2M. Large poleward shift: ACCESS1-0, GFDL-CM3, IPSL-CM5B-LR,
 486 CSIRO-Mk3-6-0
- 487 – **Northeast Pacific.** Trough: CNRM-CM5, IPSL-CM5A-MR, CSIRO-Mk3-6-0, IPSL-CM5A-
 488 LR. Ridge: GFDL-ESM2M, GISS-E2-R, CMCC-CMS, CMCC-CM
- 489 – **SH jet.** Weak poleward shift: EC-EARTH, CNRM-CM5, CESM1-WACCM, MRI-CGCM3.
 490 Large poleward shift: IPSL-CM5A-MR, IPSL-CM5A-LR, CMCC-CMS, MIROC5
- 491 – **NH stratospheric vortex.** Weakening of the vortex: CMCC-CESM, MRI-CGCM3, CMCC-
 492 CM, CCSM4, IPSL-CM5A-LR, MPI-ESM-LR. Strengthening of the vortex: GFDL-ESM2G,
 493 ACCESS1-3, IPSL-CM5A-MR, MIROC5, GFDL-CM3, MIROC-ESM-CHEM

494 These models have been selected out of a set of 32 CMIP5 models in Fig. 1 and 25 models in Fig. 2,
 495 which requires daily data. As discussed in reference [63], FGOALS-g2 is not considered due to its
 496 much larger bias in the North Atlantic jet latitude, although including it would have no impacts on
 497 the conclusions. Ensemble member r1i1p1 is analysed for all models, apart for EC-EARTH (r2i1p1),
 498 CCSM4 (r6i1p1) and CESM-WACCM (r3i1p1), due to data availability. In Fig. 2, the 98 percentile
 499 of daily wind speed is evaluated on the original models' grids. All data is spatially interpolated on
 500 a regular 2-degree grid for the purpose of averaging the model responses.

501 **Acknowledgements** The author is grateful to Richard Seager and another reviewer for their constructive com-
 502 ments that have helped to improve the manuscript. He also thanks Ted Shepherd for feedback on a draft of this
 503 manuscript and Emanuele Bevacqua and Paulo Ceppi for useful discussions. Finally, he acknowledges the World
 504 Climate Research Programme's Working Group on Coupled Modelling and the climate modelling groups for pro-
 505 ducing and making available the CMIP5 model output. The analysed CMIP5 data is available on the British

506 Atmospheric Data Centre. The work is funded by the ERC advanced grant “Understanding the atmospheric
507 circulation response to climate change (ACRCC, grant number 339390)”.

508 **Conflict of interest statement**

509 The corresponding author states that there is no conflict of interest.

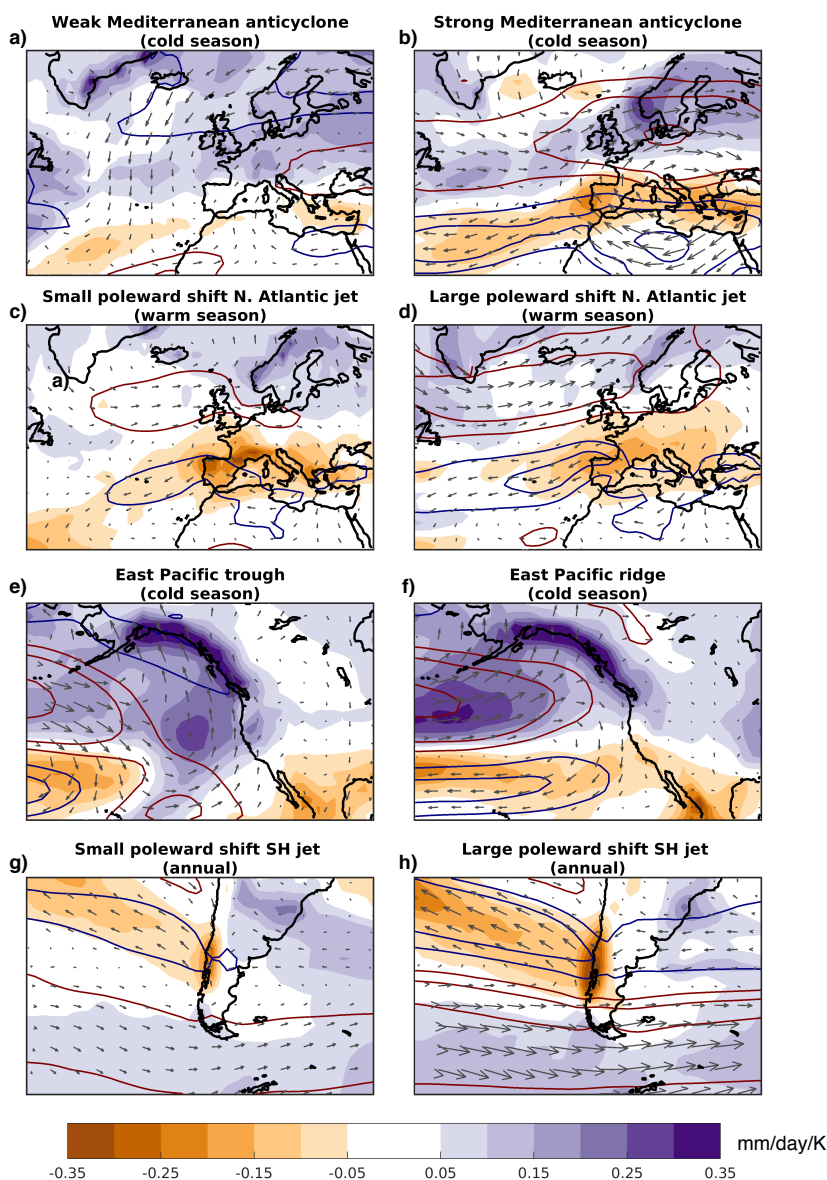


Fig. 1 Different plausible precipitation and wind at 850 hPa responses to climate change per degree of global warming. Each response is obtained by averaging the 4 CMIP5 models with the most and least positive changes in selected indices of atmospheric circulation. a-b) Cold season (NDJFMA) Mediterranean precipitation composites based on the response in the zonal wind at 850 hPa in North Africa (22.5N–33.5N, 10W–37.5E) as defined in [57]. c-d) Warm season (MJJASO) European precipitation composites on the magnitude of the poleward shift of the North Atlantic jet (jet latitude defined as in [87] but evaluated for 60W–30E). e-f) Cold season California precipitation composites on the mean response in the meridional wind velocity at 300 hPa averaged at the western coast of North America (34N–48N, 135W–120W, roughly box WC in [15]). g-h) Annual-mean Chilean precipitation response composites on the magnitude of the poleward shift of the SH jet (jet latitude defined as in [87]). The responses are evaluated for 2060–2099 relative to 1960–1999 in the RCP8.5 scenario, and scaled by the global warming simulated by each model. See the Appendix for the list of models included in each composite. Precipitation change ($\text{mm day}^{-1} \text{K}^{-1}$) is shown as shading. The red (blue) lines indicate isotaches of positive (negative) zonal wind responses at 850 hPa, with c.i. 0.15, 0.3 and 0.6 ($\text{m s}^{-1} \text{K}^{-1}$). The arrows show the mean wind response at the same level.

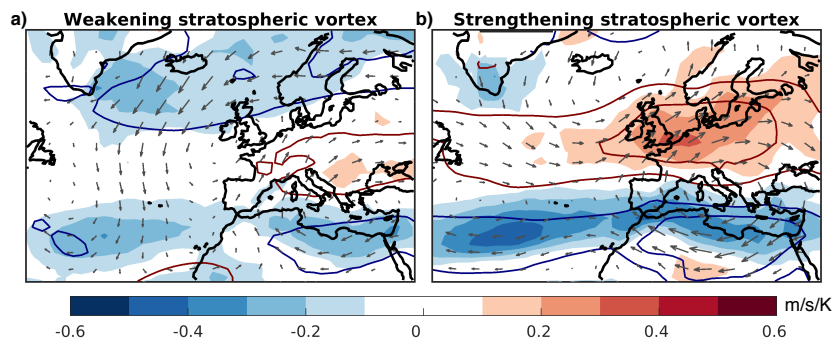


Fig. 2 As in Fig. 1 but for the climate change response in the 98 percentile of the annual distribution of daily wind speed at 850 hPa ($\text{m s}^{-1} \text{K}^{-1}$) conditioned on the response in the strength of the stratospheric vortex. Each panel shows the average response for the 6 CMIP5 models with the most negative (a) and most positive (b) change in the vortex strength. The vortex strength is evaluated as the zonal-mean zonal wind at 20 hPa averaged in 60N–75N in November–April, as defined in [64] apart for the vertical level (20 instead of 10 hPa, due to larger data availability). Similar results (see [63]) are obtained using the stratospheric vortex index from [126], and the tropospheric NAO index from [149] (not shown). Six models are used in each composite to better sample the noise due to internal variability (see the Appendix for details).

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