



Community Paper on Climate Extremes

Challenges in Estimating and Understanding Recent Changes in the Frequency and Intensity of Extreme Climate and Weather Events

World Climate Research Programme Open Science Conference

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About PCIC

The Pacific Climate Impacts Consortium is a regional climate service centre at the University of Victoria that provides practical information on the physical impacts of climate variability and change in the Pacific and Yukon Region of Canada. PCIC operates in collaboration with climate researchers and regional stakeholders on projects driven by user needs. For more information see <http://pacificclimate.org/>.

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Preface

This *Community Paper on Extremes* is part of a collection of community discussion papers developed for the World Climate Research Programme (WCRP) and its Open Science Conference in Denver, Colorado October 24-28, 2011. It is a work-in-progress, intended to stimulate discussion and further refinements leading to a set of recommendations for the advancement of research on climate and weather extremes into the future. This discussion, and parallel discussions taking place in the development of other community papers in the collection, will assist in planning the future direction of the WCRP beyond 2015.

As a regional climate service provider the Pacific Climate Impacts Consortium has a keen interest in these discussions and is highly supportive of the community-driven process involved, which promises to not only advance research on climate extremes, but also the continued refinement of climate-related services to stakeholders. While the future of the WCRP is not yet known, the planning that will take place over the next two years in the wake of the Open Science Conference will have an enormous impact on climate science internationally for many years to come.

Your comments on this paper would be very much appreciated. Please direct any comments to Francis Zwiers using the online contact form provided at <http://pacificclimate.org/contact/Francis%20Zwiers>.

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19 1. Introduction

20 This paper briefly reviews some aspects of the current status of research on changes in climate
21 extremes, identifying gaps and issues that warrant additional work. This paper focuses primarily on
22 the historical instrumental record, giving a sense of the nature of the results that have been
23 obtained, challenges that arise from observational, methodological and climate modelling
24 uncertainties and discussing the extent to which detection and attribution research has been able
25 to link observed changes to external forcing of the climate system. It also very briefly discusses
26 projections for the 21st century. Extremes are not discussed on paleo time scales, in the context of
27 the present (i.e., short term forecasting), or in the context of climate surprises (extreme tipping
28 points). These choices reflect our desire not to attempt too broad a review of the topic due to
29 space constraints of this short paper, as well as a view that very high priority should be given to
30 reducing uncertainty in our understanding of historical changes in extremes over the instrumental
31 period as a prerequisite to confidently predicting changes over the next century. This includes the
32 development of improved and comprehensive observational records, improvement in our ability to
33 confidently detect changes in observations through the development of better physical models,
34 forcing data sets and more powerful statistical techniques, the development and refinement of our
35 understanding of the physical processes that produce extremes, and continued improvement in our
36 ability to attribute causes to those changes. This does not imply that research on extremes on paleo
37 timescales or on the projection of future changes in extremes is of lesser importance, but rather
38 that overall progress on understanding implications of ongoing and future changes in extremes will
39 be strongly dependent upon our ability to document and understand changes in extremes during
40 the period of history that has been (and continues to be) most comprehensively and directly
41 observed.

42 Considerable confusion results from the various definitions of extremes that are used in climate
43 science. Part of this confusion occurs because the word *extreme* can be used to describe either a
44 characteristic of a climate variable or that of an impact. In the case of a climate variable, such as
45 surface air temperature or precipitation, the notion of an extreme is reasonably well defined and
46 refers to values in the tails of the variable's distribution that would be expected to occur relatively
47 infrequently. However, even in this case, there can be ambiguity concerning the definition of
48 extremes. For example, a great deal of climate research on "extremes" deals with indicators of the
49 frequency or intensity of events that, in fact, describe parts of the distribution that are not very
50 extreme, such as warm events that occur beyond the 90th percentile of daily maximum
51 temperature. Such events lie well within the samples of observations that are collected each
52 season, and they are typically studied by determining whether there are trends in their rates of
53 occurrence. They are often referred to as "moderate extremes" in the literature (and we will also
54 use that term occasionally below), but this term is not one that is used in statistical science to
55 describe the upper part of a distribution (statisticians would rather refer to things like the 90th
56 percentile, which they would not consider to be extreme). Nor does the term accurately describe
57 the collection of ETCCDI indices (Klein Tank, et al, 2009) since they characterize various points in the
58 distributions of daily temperature and precipitation observations.

59 In addition to the literature on indices, or “moderate extremes” of climate variables, there is also
60 another body of work that deals with rare values of climate variables that occur only infrequently
61 and are generally not expected to recur each year. In this case the concept corresponds well to that
62 used in the statistical sciences, and thus powerful statistical tools based on extreme value theory
63 are available to aid in the analysis of historical and future extremes (e.g., Coles, 2001; Katz et al,
64 2002). Such tools were originally developed to make statements about what might happen outside
65 the range of the observed sample, such as the problem of estimating the 100 year return value on
66 the basis of a 30- or 40-year sample. Hence, the notion of “extremes” is defined here as very high
67 quantiles, such as the 95th, 99th or 99.9th percentiles of annual maximum values. An important
68 aspect of this theory is to quantify the uncertainty of such extrapolations through the computation
69 of suitably constructed confidence intervals. Increasingly, these tools are being used in the
70 evaluation extreme events simulated in climate models (e.g., Kharin et al, 2007; Wehner et al,
71 2010). These tools are being further developed in the statistical sciences, and there is currently a
72 very high level of interaction between that community and the climate sciences community on the
73 development and application of methods that can be used in the climate sciences, such as the
74 ExtREmes toolkit (see <http://www.assessment.ucar.edu/toolkit/>).

75 In the case of extremes defined by their impacts, the concept of what constitutes an extreme may
76 be less well defined, and this may affect the approaches that are available for analysis. For example,
77 all tropical cyclones that are classified as Category 1-5 storms on the Saffir-Simpson scale are
78 generally considered to be extreme because of their high potential to cause damage from high
79 winds, rainfall, and/or storm surge flooding. Nevertheless, they are a natural component of the
80 climate system and occur in more or less constant numbers (globally) each year. They are more
81 difficult to characterize statistically than, for example, extreme temperature events, because the
82 numbers of tropical cyclones within a region are not constant, the regions affected vary with time,
83 and historical data concerning characteristics that might be used to identify tropical cyclones with
84 characteristics in the tails of the corresponding distributions are subject to substantial
85 inhomogeneities due to the evolution of our observing systems.

86 For the purpose of this article we define “extreme events” as well-defined weather or climate
87 events (including tropical cyclones) that are rare within the current climate. With the term “well-
88 defined” we understand that these events may be defined in terms of measurable physical
89 quantities such as temperature, precipitation, wind speed, runoff levels or similar; and the term
90 “rare” is used to refer to values in the tails of the variable’s distribution as discussed above.

91 It is important to note that the linkage between extreme events and extreme impacts (i.e. natural
92 disasters) is not straightforward. An extreme event does not necessarily imply any damages. Rather
93 the implied damages also depend upon the distribution of values, population density, emergency
94 response measures, etc. Similarly, not all damages from a weather or climate events are related to
95 extreme events as defined above. For instance, poor building practices may allow a “normal” or
96 moderate event to generate extreme damages. This issue is very familiar to the re-insurance
97 industry that uses damage models to link extreme events to impacts (e.g. Klawa and Ulbrich 2003,
98 Watson and Johnson 2004)

99 The structure of the remainder of this paper is as follows. The paper begins in Section 2 with a
100 discussion of the status of research on simple indices that are derived from daily (or occasionally
101 more frequent) observations that are collected primarily at operational meteorological stations.
102 The main focus here is on temperature and precipitation extremes, but wind extremes derived
103 from station data are also discussed. Section 3 discusses storms (extra-tropical cyclones, tropical
104 cyclones and tornadoes). This is followed by a discussion of hydrological extremes (droughts and
105 floods) in Section 4, and extreme sea-levels (e.g., storm surge events) in Section 5. A summary and
106 recommendations are presented in Section 6. Amongst other sources, the paper draws upon the
107 IPCC 4th Assessment Report (IPCC 2007a, IPCC 2007b), the US Global Change Program Special
108 Assessment Product on extremes (i.e., CCSP 3.3, Karl et al, 2008), the recent WMO assessment on
109 tropical cyclones (Knutson et al, 2010), and on a very recently completed review of research on
110 indices by Zhang et al (2011).

111 2. Simple indices derived from daily data

112

113 a. Introduction

114 The indices that are discussed in this section are generally derived from daily observations of
115 individual meteorological variables, such as temperature or precipitation. Indices calculated from
116 daily data have appeal for a number of reasons, including the fact that they are relatively easy to
117 calculate and that they summarize information on changes in variability compactly, and in a way
118 that is accessible to a broad range of users.

119 Indices have been designed to characterize different parts of the distribution of a given variable.
120 The indices that are of interest here are those that characterize aspects of the tails of the
121 distribution (the “extremes”) since these tend to be more relevant to society and natural systems
122 than indices that characterize aspects of the distribution that occur more frequently. This is
123 because the more extreme an event, the more likely it is to cause societal or environmental
124 damage. However, analyses of changes in the frequency or intensity of extremes that are further
125 out in the tail of the distribution are inherently more uncertain because less data are available to
126 identify and characterize possible changes (Frei and Schär 2001). Moreover, extreme impacts may
127 also occur following moderate events, e.g. when these are compounded with other climate events
128 (see discussion in Hegerl et al, 2011) or with increased vulnerability or exposure of society or
129 ecosystems to such events. Conversely, statistically very rare events may not necessarily lead to
130 impacts if there is either no exposure or no vulnerability to the particular event. Thus the impact of
131 an extreme event may depend on its season, its duration, and co-occurrence of further extremes,
132 such as drought conditions with heat waves.

133 Most indices of extremes tend to represent only “moderate extremes,” i.e. those that typically
134 occur at least once a year. In some cases, changes in the tails, as indicated by changes in the
135 indices, are essentially similar to those in other parts of the distribution (Figure 1). However, even
136 for temperature, changes may be seen that are not consistent between means and extremes,

137 minimum and maximum, and upper and lower tail (e.g. Hegerl et al., 2004; Kharin et al., 2007) due
 138 to alterations in feedback processes that may affect different parts of the distribution differently
 139 (e.g. Hirschi et al, 2011). Some indices for climate extremes can also be used for secondary
 140 inference; for example, statistical extreme value theory can be used to estimate long return period
 141 precipitation amounts from long time series of annual maximum daily precipitation amounts (Klein-
 142 Tank et al, 2009). It should be noted that the estimation of return levels is often based on the
 143 assumption of spatial and/or temporal independence among sites or grid points (either on the raw
 144 data or conditionally on their distributional parameters). Consequently, uncertainties can be
 145 underestimated or these assumptions can be challenged.

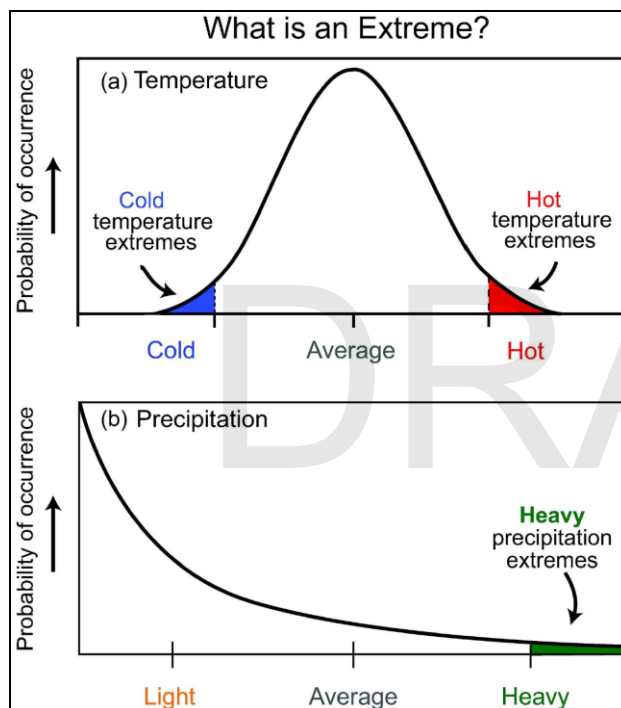


Figure 1: Schematic representations of the probability distributions of daily temperature, which tends to be approximately Gaussian (exceptions can be caused by soil freezing, or by energy balance constraints, e.g. Fischer and Schär 2009), and daily precipitation, which has a skewed distribution. Extremes are denoted by the shaded areas. In the case of temperature, changes in the frequencies of extremes are strongly affected by changes in the mean; a relatively small shift of the distribution to the right would substantially increase warm extremes and decrease cold extremes. In addition, the frequency of extremes can also be affected by changes in the shape of the tails of the temperature distribution, which could become wider or narrower, or could become somewhat skewed rather than being symmetric as depicted. In a skewed distribution such as that of precipitation, a change in the mean of the distribution generally affects its variability or spread, and thus an increase in mean precipitation would also likely imply an increase in heavy precipitation extremes, and vice-versa. In addition, the shape of the right hand tail could also change, affecting extremes. Furthermore, climate change may alter the frequency of precipitation and the duration of dry spells between precipitation events. From Folland et al (1995) and Peterson et al (2008).

146 In addition to indices that summarize various aspects of the tails of the daily variability of individual
 147 meteorological parameters, there have also been a variety of attempts to build indices that
 148 incorporate information from multiple parameters to summarize information related to impacts,
 149 such as fire weather indices that were first developed for operational use in wild fire risk
 150 management (e.g., Van Wagner, 1987) and subsequently used to study the potential impacts of
 151 climate change on wild fire frequency and extent (e.g., Flannigan et al, 2005). Similar types of
 152 development are seen in a variety of indices (another example being heat indices such as that
 153 described by Steadman, 1979; Karl and Knight 1997; Fischer and Schär 2010). Since these types of
 154 indices are impact specific, their construction must ultimately be informed by the characteristics

155 and functioning of the system (ecological, social, or economic) or biological organism that is
156 impacted (health, agriculture). This requires inter- and trans-disciplinary collaboration, and involves
157 a range of potential compound indices far greater than would be required to monitor and
158 understand change in the physical climate system.

159 b. Status

160 i) Temperature and precipitation indices

161 Many indices have been defined (e.g., Frich et al, 2002; Klein-Tank et al, 2009) for the purpose of
162 monitoring changes in the moderately far tails of surface variables such as temperature and
163 precipitation that are routinely observed on a daily, or more frequent, basis. These indices include:
164 (i) absolute quantities such as the annual maximum and minimum temperature and the annual
165 maximum precipitation; (ii) the frequency of exceedance beyond a fixed absolute threshold, such as
166 the annual count of the number of days with precipitation amounts greater than 20 mm; (iii) the
167 frequency of exceedance above or below fixed relative thresholds such as the 90th percentile of
168 daily maximum temperature or the 10th percentile of daily minimum temperature where the
169 threshold is determined from a climatological base period such as 1961-90; and (iv) dimensionless
170 indices, such as the proportion of annual precipitation that is produced by events larger than the
171 95th percentile of daily precipitation amounts, where the threshold is again determined from a fixed
172 base period. These indices are studied because they describe aspects of temperature and
173 precipitation variability that have been linked, in one way or another, to societal or ecological
174 impacts. Their calculation is actively coordinated by the CLIVAR/CCI/JCOMM Joint Expert Team on
175 Climate Change Detection and Indices (ETCCDI). The state of development of these indices has
176 recently been reviewed comprehensively by Zhang et al (2011).

177 The calculation of indices requires high quality, high frequency (daily or better), homogeneous
178 meteorological data. High quality data are available from hydro-meteorological services in many
179 parts of the world, and are often freely available for scientific research at least nationally, if not on
180 a fully open basis internationally, though various limitations to (mostly raw) data access remain an
181 issue (see also point i below). Data availability is generally greater in developed countries than in
182 developing countries, where resources and/or mandate sometimes limit the collection and
183 dissemination of daily meteorological observations, although restricted data access also remains a
184 problem in some developed countries. The ETCCDI has an ongoing program of open source
185 software development and international workshops that are intended to train developing world
186 scientists in the homogenization of data that are collected by their hydro-meteorological services,
187 and in the subsequent calculation of indices (Peterson et al, 2008). The calculated indices are
188 published in the peer-reviewed literature (e.g., Aguilar et al, 2009) and are subsequently
189 contributed to global scale index datasets such as HadEX (Alexander et al, 2006) and the
190 subsequent updates (e.g. Donat and Alexander, 2011; Alexander and Donat, 2011), thereby helping
191 to improve the global coverage of these datasets and consequently enabling more confident global
192 scale monitoring and detection and attribution.

193 While the ETCCDI type of approach is helpful, there are nevertheless ongoing challenges. These
194 include:

- 195 i. Concerns about the reproducibility of the entire chain of index production. Currently the
196 reproducibility of the entire processing sequence cannot be guaranteed because, while
197 methods and codes are freely available, the underlying daily station data are not always
198 openly accessible to the international scientific community because regional data gathering
199 organizations may not have the capacity or mandate to support open data dissemination.
- 200 ii. Lack of access to daily station data also implies a lack of access to metadata describing the
201 history of observing stations. This is important concern because small changes in observing
202 station location or exposure can lead to large artificial changes in extremes. In the absence
203 of station metadata, it is often difficult to determine if such issues have corrupted indices
204 derived from the underlying data.
- 205 iii. Lack of real-time updating, particularly for regions that are unable to contribute to the
206 Global Historical Climate Network (GHCN, see [http://www.ncdc.noaa.gov/oa/climate/ghcn-](http://www.ncdc.noaa.gov/oa/climate/ghcn-daily/)
207 [daily/](http://www.ncdc.noaa.gov/oa/climate/ghcn-daily/)). This is a concern because maintaining and monitoring indices is not always part of
208 the primary mandate of the developing world scientists who participate in the ETCCDI
209 workshops and are involved in index development for their countries or regions. It should
210 be noted however, that the Asia Pacific Network (APN; Manton et al, 2001), which has
211 focussed on a specific region, has been successful in running repeat workshops that have
212 allowed for the updating of indices in that region.
- 213 iv. The potential loss of scientific information that results from providing only a limited
214 number of pieces of information about the distribution of daily temperature and
215 precipitation.
- 216 v. Potential difficulties in characterizing the statistical distributions of many indices, which
217 makes it more difficult to make reliable statistical inferences about things such as the
218 presence or absence of trend in a time series of annual indices.
- 219 vi. Consideration of specific impacts often requires information that relies upon simultaneous
220 values of several climate variables. For instance, health impacts from heat waves depend
221 upon temperature and humidity (and additional factors), information that cannot be
222 recovered from standard indices.

223 A further challenge is that the spatial coverage of index datasets remains far from being truly
224 global, with significant fractions of the globe still under-sampled, for example, in Africa and South
225 America (see Fig. 2a-c). Further challenges in the production of global datasets are also related to
226 the choice of gridding framework in addition to parameter choices that are made within a chosen
227 gridding method (e.g. Donat and Alexander, 2011). This adds additional uncertainty to long term
228 variability measures and trend estimates (see Fig. 2d).

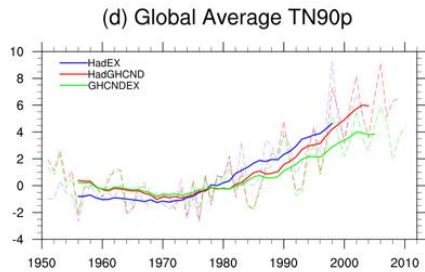
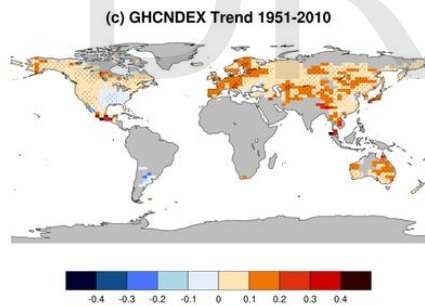
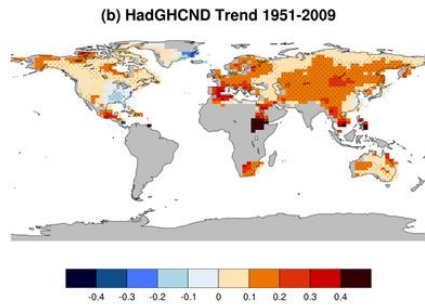
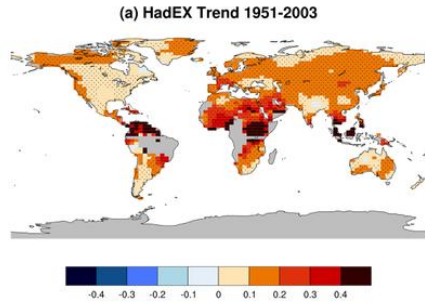


Figure 2: Annual trends in warm nights (TN90p) using different datasets for the periods indicated. The datasets are (a) HadEX (Alexander et al., 2006), (b) HadGHCNDEX (ETCCDI indices calculated from an updated version of HadGHCND (Caesar et al. 2006)) and (c) GHCNDEX (Donat and Alexander, 2011). (d) represents the global average timeseries plots for each of the three datasets with associated 11-year running means.

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230 The index approach also has several scientific limitations. One such limitation, for which a solution
231 has been found, is the possibility that inhomogeneities could be introduced into index time series
232 unintentionally, such as can occur in the case of threshold crossing frequency indices when
233 thresholds representative of the far tails are estimated from a fixed observational base period (e.g.,
234 Zhang et al, 2005). Another limitation, which can also be circumnavigated, is that differences in the
235 recording resolution of observational data can cause non-climatic spatial variations in threshold
236 crossing frequency and trends (e.g., Zhang et al, 2009). A third limitation is that in a changing
237 climate, the number of exceedances of thresholds based on a climatological base climate may
238 saturate, e.g. exceedances may never or almost always occur under strong climate change. Thus,
239 percentage exceedance indices are only useful for characterizing change in the distribution that is
240 not too far from the base period (see e.g. Portmann et al., 2009). A further limitation is that the
241 nature of index data, which typically provides only one value per year, may limit the range of
242 possible approaches that can be used to analyze change in certain types of extremes. For example,
243 long return period extremes (e.g., the intensity of the 20-year extreme daily precipitation event)
244 can be estimated from the annual extremes that are recorded in HadEX, but the analyst can only do
245 so using the so- called block-maximum approach to extreme value analysis, which only considers
246 the most extreme of a series of values observed within a block of a defined length (e.g. the annual
247 maximum). In contrast, it is often argued by statisticians that the so-called peaks-over-threshold
248 approach, by which all values exceeding a given threshold are used in the analysis, may result in
249 more confident estimates of long period return values since it has the potential to utilize the
250 information about extremes that is available in a long time series of daily values more effectively
251 than the block-maximum approach. It should be noted however, that the peaks-over-threshold
252 approach is difficult to apply to large gridded datasets, such as the output from global climate
253 models, because of the challenges associated with finding an automated procedure for reliably
254 determining the appropriate threshold at each location in the grid. A further consideration is that
255 most available index datasets do not currently provide the date (or dates) on which the extreme
256 values were recorded. This creates a limitation when attempting to study the association between
257 the occurrences of extremes in different variables or between climate extremes on the one hand
258 and impacts on the other. It also limits the ability to study changes in the seasonality of extremes,
259 and it impairs process based analyses of the conditions leading to recorded extremes.

260 An additional limitation is that the block over which the annual extreme is calculated is generally
261 not adjusted to match the annual cycle. For example, the annual maximum of daily maximum
262 temperature is reported in HadEX for the calendar year at all locations. This is appropriate for mid-
263 latitude Northern Hemisphere locations, but creates a situation where the extreme temperature
264 recorded during an extremely hot Southern Hemisphere summer could be recorded in different
265 years for adjacent locations, which would have the effect of reducing the apparent spatial
266 dependence between extremes at nearby locations. Also, there is the possibility that some
267 persistent extreme events could be double-counted in the Southern Hemisphere, for example, if a
268 single event spanning the days between two calendar years produced the warmest temperatures of
269 the year in both calendar years. One can imagine similar concerns regarding precipitation.

270 As noted, methods have been developed to prevent inhomogeneities in indices that count
271 exceedances beyond quantile based thresholds and to account for the effects of different data
272 reporting resolutions (Zhang et al, 2005, 2009). Other limitations could be overcome by adding a
273 modest number of additional indices to the “standard” ETCCDI list. For example, one could include
274 within the suite of indices the r most extreme values observed annually for some small number $r > 1$
275 and not just the most extreme value annually, thereby enabling the application of the more
276 efficient “ r -largest” extreme value analysis techniques (e.g., Smith, 1986; Zhang et al, 2004). In
277 addition, it would be appropriate to redefine the ETCCDI indices such that they describe annual
278 extremes and counts that pertain to a year that is defined in a climatologically appropriate manner,
279 where the definition of the year would depend upon location and parameter, taking into account
280 the form of the annual cycle for the specific aspect of the parameter that is relevant for each index.
281 It should be noted that the definition of the year has implications for many types of indices and not
282 just annual extremes as discussed above. As specific example is CDD (consecutive dry days, see
283 Klein-Tank et al, 2009), an index that can show very large changes in climate models under future
284 emissions scenarios (e.g. Tebaldi et al. 2006, Orlowsky and Seneviratne 2011). CDD calculated on
285 the basis of the calendar year has a different interpretation in places where the climatological dry
286 period spans the year boundary as opposed to places where the climatological dry period occurs in
287 the middle of the year; while dry periods may be of comparable length in both types of places, CDD
288 will tend to report them as being substantially shorter in the former. In contrast, a CDD index that
289 was calculated from years that are defined locally in such a way that the climatological dry period
290 occurs everywhere in the middle of the year would have a more uniform interpretation across
291 different locations.

292 There are a number of factors that limit our ability to evaluate how well models simulate indices by
293 comparison against observed indices. These include observational limitations, such as the wide
294 variation in the density and coverage of observing stations, the likelihood that there are few regions
295 in the world where precipitation station density is sufficient to reliably estimate grid box mean
296 precipitation on GCM and RCM scales (see discussion in Zhang et al., 2007), the still relatively
297 limited duration of satellite and other remotely sensed data products with high reporting frequency
298 (e.g., Kossin et al. 2007; Lau et al 2008), and ongoing concerns about the homogeneity (e.g., Elsner
299 et al. 2008) and calibration of remote sensing products. As a consequence, model evaluation often
300 relies on proxies for direct observations, such as reanalysis products. This is a reasonable approach
301 for variables such as surface temperature that are directly constrained by observations in
302 reanalyses, but is more problematic in the case of variables such as precipitation which is generally
303 not observationally constrained in reanalyses (the North American Regional Reanalysis, Messinger
304 et al, 2006, is an exception; it uses precipitation observations to adjust latent heating profiles).
305 Furthermore, the observational data streams assimilated in reanalysis data products are not
306 consistent over time, e.g. because of the relatively short length of satellite data, which may affect
307 their use for the assessment of climatic trends (e.g. Bengtsson et al. 2004, Grant et al. 2008). Taking
308 these various limitations into account, models are found to simulate the climatology of surface
309 temperature extremes with reasonable fidelity (Kharin et al., 2007) on global and regional scales
310 when compared against reanalyses, although there are uncertainties associated with, for example,

311 the representation of land-atmosphere feedback processes in models (Seneviratne et al, 2006). In
312 contrast, intercomparisons between models, reanalyses, and large scale observational precipitation
313 products such as CMAP (Xie et al, 2003) suggest large uncertainties in all three types of
314 precipitation products; particularly in the tropics (e.g., see Figure 6 in Kharin et al, 2007)

315 Scaling issues (e.g., differences between intensity, variability and representativeness of point
316 observations from rain gauges or gridded observed precipitation versus the grid box mean
317 quantities simulated by climate models; Klein-Tank et al, 2009; Chen and Knutson 2008),
318 uncertainties in observational gridded products (Donat and Alexander 2011), and incomplete
319 process understanding continue to limit the extent to which direct quantitative comparison can be
320 made between station observations and models (Mannshardt-Shamseldin et al, 2010). It should be
321 noted, however, that models of sufficiently high resolution may be capable of simulating
322 precipitation extremes of comparable intensity to observed extremes. For example, Wehner et al
323 (2010) show the global model that they study produces precipitation extremes comparable to
324 observed extremes at a horizontal resolution of approximately 60 km. However, most global
325 models continue to operate at substantially lower resolutions, leading to ambiguities in the
326 interpretation of projected changes in extremes. Nevertheless, precipitation change at large scales
327 is determined primarily by changes in the global hydrological cycle that are reflected in changes in
328 evaporation, atmospheric moisture content, circulation (which affects moisture transport and
329 convergence), and energy and moisture budgets, providing a fundamental basis for the qualitative
330 (in terms of the direction of change and its large scale features), if not quantitative (in terms of the
331 absolute values of the changes and their detailed geographic patterns), interpretation of modelled
332 precipitation changes. The scaling issue can sometimes be partially overcome transforming
333 observed and simulated precipitation to dimensionless scales that can more readily be
334 intercompared, such as has been done by Min et al (2011). A disadvantage of such transformations,
335 however, is that the translation of extremes onto a probability or other type of relative scale may
336 impede the physical interpretation of trends and variability.

337 ii) Wind indices

338 To date, temperature and precipitation indices have been studied most intensively. Indices of wind
339 extremes, while of enormous importance in engineering applications, have received less attention,
340 in part because of the greater difficulty in obtaining homogeneous high-frequency wind data. Wind
341 records are often affected by non-climatic influences, such as development in the vicinity of an
342 observing station that alters surface roughness over time. It has also been postulated by Vautard et
343 al (2010) that large scale changes in vegetative cover over many land areas has altered surface
344 roughness and that this may be an important contributor to the apparent stilling (reduction) of
345 surface wind speeds in many mid-latitude regions (e.g., Zwiers, 1987; Roderick et al, 2007).

346 An alternative to using direct anemometer observations of wind speeds is to consider a proxy that
347 is based on pressure readings that are usually more homogeneous than wind speed observations.
348 Several storm proxies currently being used are derived from pressure readings at single stations,

349 such as the statistics of 24-hourly local pressure changes or of the frequency of low pressure
350 readings. These single station proxies relate to synoptic experience and reflect storminess indirectly
351 as they seek to detect atmospheric disturbances (e.g. Schmith et al, 1998; Hanna et al, 2008; Allan
352 et al, 2009; Barring and von Storch, 2004; Barring and Fortuniak, 2009). Another approach to
353 explore past storminess is to make use of the statistics of geostrophic wind speeds. Geostrophic
354 wind speeds can be derived by considering mean sea-level pressure gradients in networks of
355 reliable surface pressure records over homogenous mid-latitude domains, such as the north-east
356 Atlantic and western Europe (e.g., Schmidt and von Storch, 1993; Alexandersson et al, 1998;). These
357 records, which continue to be developed in the North Atlantic and European region (e.g., Wang et
358 al, 2011) and are also being developed in the Australian regions (e.g., Alexander et al, 2011), are
359 available for much longer periods of record than the more limited anemometer network. For the
360 North Atlantic region for which they have been most extensively developed, they show
361 predominately the effects of natural low frequency variability in atmospheric circulation on
362 variations in storminess and extreme geostrophic wind speeds.

363 Recently Krueger and von Storch (2011a) used a regional climate model to evaluate the underlying
364 assumption that the extremes of geostrophic wind speed are indeed representative of surface wind
365 speed extremes. They also considered the sensitivity of the proxy to the density of stations in the
366 network, concluding that higher density networks should give more reliable estimates of wind
367 speed extremes. Work is currently underway to evaluate the robustness of such proxies to
368 instrumental error in pressure readings and to inhomogeneity in one or more of the surface
369 pressure records that are used to derive the geostrophic winds. Further, a study that evaluates how
370 well a number of single-station pressure proxies represent storminess has recently been completed
371 (Krueger and von Storch, 2011b) and concludes that all single-station pressure proxies considered
372 were linearly related to storm activity, with absolute pressure tendency being most strongly
373 correlated.

374 Another possibility for the construction of wind speed and storminess indices is provided by
375 reanalyses, such as the NCEP (Kistler et al, 2001), ERA-40 (Uppala et al, 2005), or the 20th Century
376 (20CR) reanalysis of Compo et al (2011), which is based only on surface observations and covers the
377 period 1871-2008. In contrast with wind speed observations and recent extreme wind speed
378 reconstructions from surface pressure readings (e.g., Wang et al., 2011), all reanalyses appear to
379 show an increase in European storm indicators during the last few decades of the 20th century
380 (Smits et al, 2005; Donat et al, 2011). For tropical cyclones, the intensities of the storms (i.e.,
381 maximum near-surface sustained one-minute wind speeds) can also be estimated globally using
382 satellite data, at least since the early 1980s (Kossin et al. 2007; Elsner et al. 2008).

383 c. Role of external influences

384 i) Temperature extremes

385 Considerable progress has been made in the detection and attribution of externally forced change
386 in surface temperature extremes since the feasibility of such studies was first demonstrated by
387 Hegerl et al (2004). Studies that detect human influence on surface temperature extremes are
388 available on the global and regional scale, use a range of indices that probe different aspects of the
389 tails of the surface temperature distribution including the frequency (e.g., Morak et al, 2011; Figure
390 3, which also shows that human influence can be detected in the frequency of warm nights in most
391 regions) and magnitude (e.g., Christidis et al, 2005, 2011; Zwiers et al, 2011) of extreme surface
392 temperature events; results are robust across a range of methods and across both types of indices.
393 Some studies use methods that rely on extreme value theory (e.g., Christidis et al, 2011; Zwiers et
394 al, 2011), and are therefore best suited for studying change in the far tails of the temperature
395 distribution, whereas other studies that consider less extreme parts of the distribution (Christidis et
396 al 2005; Morak et al., 2011) appropriately use standard fingerprinting approaches (e.g., Hegerl et al,
397 2007). Some studies (e.g., Christidis et al, 2011) are also able to separate and quantify the
398 responses to anthropogenic and natural external forcing from observed changes in surface
399 temperature extremes, thereby increasing confidence in the attribution of a substantial part of the
400 observed changes to external forcing on global scales.

401 There is the potential to further develop techniques in order to be able to conduct the analysis
402 more fully within the framework of extreme value theory and more confidently separate signals by
403 utilizing recent developments in the spatial modelling of extremes via so-called max-stable
404 processes (e.g., Smith, 1990; Schlatter, 2002; Vannitsem and Naveau, 2007; Blanchet and Davison,
405 2011). By working within the framework of extreme value theory, as has already been done in the
406 recent studies of Christidis et al (2011) and Zwiers et al (2011), it should become possible to all
407 attribute changes in the *likelihood* of extreme events to external causes, thereby contributing to the
408 scientific underpinnings that will be required for event attribution (see the Community Paper lead
409 by Stott et al). For example, Zwiers et al (2011) provide rough estimates of circa 1990s expected
410 waiting times for events that nominally had a 20-year expected waiting time in the 1960s, showing
411 that cool temperature extremes have become substantially less frequent globally, whereas warm
412 temperature extremes have become modestly more frequent. A further area where statistics can
413 make important contributions is in accounting for spatial dependence between extremes. Most
414 work described above considers grid points or stations independently of each other. However,
415 statistical space-time modelling can account for spatial dependence between parameters of
416 extreme value distributions, for example, by setting prior expectations of spatial dependence that
417 are updated with data. These methods can account for complex space-time structure of extremes
418 and make use of information in data more completely (e.g., Sang and Gelfand, 2009, 2010; Heaton
419 et al, 2010)

420 A limitation of the studies that have been conducted to date is that they have been confined to the
421 20th century, in part due to the design of the CMIP3 experiment which ended the historical
422 simulations and the single forcing runs at 1999 or 2000, but more importantly, because suitable
423 observational datasets providing broad coverage of annual temperature extremes have not been
424 updated to the more recent decade (e.g., Alexander et al, 2006). Also, modelling groups

425 participating in CMIP3 generally were not able to make available large volumes of high frequency
 426 (daily or higher) output or ensembles of historical single forcing runs (e.g., runs with historical
 427 greenhouse gases or aerosol forcing only). Consequently, currently available studies that separate
 428 signals have only been performed with single climate models rather than with multi-model
 429 ensembles. All of these problems should be alleviated at least to some extent in the near future
 430 with the advent of updated research quality datasets, such as HadEX2 (Alexander and Donat, 2011),
 431 and the growing availability of CMIP5 simulations (Taylor et al, 2009) that are currently being
 432 completed by the climate modelling community and will make available high frequency output
 433 more broadly than their predecessors in CMIP3, enabling a more thorough exploration of model
 434 uncertainties.

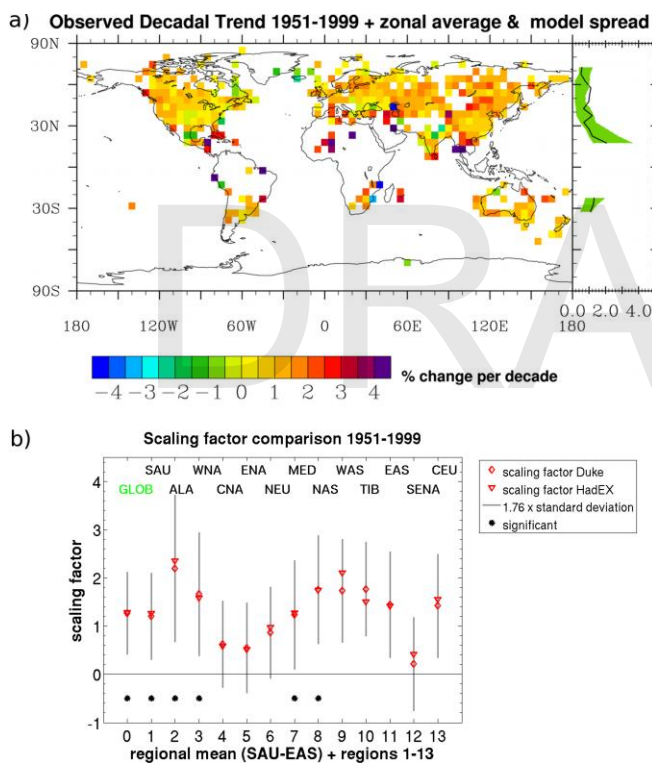


Figure 3: (a) 1951–1999 observed decadal trend of TN90 (in % change per decade) based on a combination of HadEX (Alexander et al, 2006) and additional index data from Kenyon and Hegerl (2007). The zonal average of the observations (black line) and the spread of trends in an ensemble of CMIP3⁹ “ALL” forcings model simulated trends for the same period (green shaded area) is shown on the side of the plot. (b) The scaling factors (red markers) of observed changes projected onto the multi-model mean fingerprint for the period 1951–1999. The “diamonds” indicate scaling-factors based the Kenyon and Hegerl (2007) dataset (labelled Duke in the legend), and the “triangles” indicate scaling-factors based on HadEX. Grey bars indicate 5–95% uncertainty ranges. Regions in which results are detectable at the 5% significance level and where model simulated internal variability is consistent with regression residuals are indicated with an asterisk. Results indicate broad increases in the frequency of warm nights, and the detection of anthropogenic influence in the pattern of observed increases globally and in most regions. From Morak et al (2011).

435

436 The studies available to date use only a limited number of models. Across many of these studies
 437 results suggest that the climate model simulated pattern of the warming response to historical
 438 anthropogenic forcing in cold extremes fits observations best when its amplitude is scaled by a

⁹ Coupled Model Intercomparison Project Phase 3, see http://www-pcmdi.llnl.gov/ipcc/about_ipcc.php

439 factor greater than one (i.e., when the simulated warming signal is scaled up). Conversely, the
440 expected warming signal in extremes of warm daily maximum temperature extremes generally
441 needs to be scaled down, and in fact, has only recently been detected in observations through the
442 use of more sophisticated statistical techniques (Christidis et al, 2011; Zwiers et al, 2011). These
443 results point to the possibility that the forcing and/or response mechanisms, including the
444 possibility of feedbacks that operate differently during the warm and cold seasons and during
445 different parts of the diurnal cycle (day versus night), may not be fully understood (e.g. Portmann
446 et al, 2009) or accurately modelled. Recent examples include work by Seneviratne et al (2006,
447 2010) and Nicholls and Larsen (2011) concerning the role of land-atmosphere feedbacks in the
448 development of temperature extremes, and by Sillmann et al (2011) on the role of blocking in the
449 development of cold temperature extremes in winter over Europe.

450 It should be noted that not all extremes change in the same direction as the mean. One example for
451 temperature is freeze-thaw cycles that occur when night-time low temperatures are below freezing
452 but day-time highs are above freezing. The number of freeze-thaw cycles can have a considerable
453 impact on infrastructure. One can imagine locations currently below a cold threshold such that
454 freeze-thaw cycles currently never occur, but as climate warms such cycles start to occur and then
455 become more frequent. Similarly, locations with a current climate just cool enough for freeze-thaw
456 cycles to be common now could see large decreases with warming. Other extremes that could
457 change in opposing directions include events that depend on both temperature and precipitation.
458 Rain on frozen ground, for example, could actually become more common at sufficiently cold
459 locations despite warming if more winter precipitation occurs, as is projected at mid-latitudes.
460 Conversely locations that are just cool enough for rain on frozen ground to occur at present could
461 see reductions in the frequency of occurrence of rain on frozen ground because rising temperatures
462 would shorten or eliminate the annual period of frozen ground. Rain on snow and rapid snow melt
463 could change in either direction for similar reasons. McCabe et al (2007) find decreasing trends in
464 rain on snow throughout most of the Western US but increases at many of the coldest locations.
465 Similarly, Ye et al (2008) find increasing rain on snow trends in European Russia where trends are
466 driven by increased precipitation at the coldest locations but inhibited by reductions in snowfall
467 days at warmer locations. Hamlet et al (2005) showed that even mean snowpack, which depends
468 on both temperature and precipitation, can change in opposite directions at nearby locations:
469 Western US precipitation increases tend to drive increasing historical snowpack trends at the
470 coldest locations while temperature increases have driven decreasing snowpack trends at relatively
471 warmer locations.

472 ii) Precipitation extremes

473 As is also the case with change in the mean state, in comparison with surface air temperature only
474 limited progress has been made in determining the extent to which external influences on the
475 climate system have influenced changes in the intensity or frequency of heavy or extreme
476 precipitation. Various studies have found that extreme precipitation can have heavy tailed
477 behaviour (with a shape parameter of around 0.2 when annual maxima of daily precipitation are

478 fitted with a generalized extreme value distribution). While climate models simulate substantial
479 precipitation extremes, it is not clear that they simulate daily intensities that are as heavy-tailed as
480 observed. For example, Kharin and Zwiers (2005) do not find strong evidence for heavy tailed
481 behaviour in the model that they studied, estimating shape parameters that are positive, but near
482 zero. Averaging in space and time smoothes the tail behaviour recorded at weather stations but
483 this reduces the applicability for impact studies. In addition, it is a real challenge to detect and
484 attribute changes whenever the variable of interest has a positive shape parameter. In such cases,
485 uncertainties grow rapidly with a slight change in the shape parameter and consequently very long
486 time series are necessary. Thus there are substantial statistical challenges associated with the
487 detection and attribution of the precipitation response to external forcing.

488 Nevertheless, there is a modest body of literature that has investigated whether there is evidence
489 that natural or anthropogenic forcing has affected global mean precipitation over land (e.g., Gillett
490 et al, 2004; Lambert et al, 2005), the zonal distribution of precipitation over land (e.g., Zhang et al,
491 2007) and the quantity of precipitation received at high northern latitudes (Min et al, 2008). Since
492 the variability of precipitation is related to the mean (there is greater short term precipitation
493 variability in regions that receive more precipitation), the detection of human influence on the
494 mean climatological distribution of precipitation should imply that there has also been an influence
495 on precipitation variability, and thus extremes. Hegerl et al (2004) found in a model-study that
496 changes in moderately extreme precipitations may be more robustly detectable than changes in
497 mean precipitation since models robustly expect extreme precipitation to increase across a large
498 part of the globe.

499 Min et al (2011) recently investigated this possibility, finding evidence for a detectable human
500 influence in observed changes in precipitation extremes during the latter half of the 20th century.
501 This was accomplished by transforming the tails of observed and simulated distributions of annual
502 maximum daily precipitation amounts into a probability based index (PI) before applying an optimal
503 detection formalism, thereby partly circumnavigating the scaling issues that are associated with
504 precipitation. It should be noted however, that some strong assumptions are implicit in such
505 transformations that are not necessarily verifiable. For example, it is implicitly assumed that forced
506 changes in precipitation extremes result in comparable changes in PI at different scales, even
507 though the mechanisms that generate extreme precipitation locally may be quite different from
508 those that determine extreme events on climate model grid box scales and larger. Even with the
509 transformation, it was found that a best fit with observations required that the magnitude of the
510 large-scale climate model simulated responses to external forcing be increased by a considerable
511 factor, with a greater increase in magnitude being required in the case of historical simulations that
512 take into account a combination of anthropogenic and natural forcing (ALL forcing), than for
513 simulations accounting only for the former (ANT forcing; see Figure 4). The discrepancy between
514 scaling factors for ALL and ANT forcing is understandable given that the anthropogenically forced
515 signal is still small, and that natural forcing (from changes in solar and volcanic activity) would have
516 offset some of the response to ANT forcing, thereby weakening the ALL signal during the latter part
517 of the 20th century. This leads to smaller expected changes in the ALL fingerprint, which are more

518 strong affected by noise and thus more difficult to detect, than the ‘cleaner’ signal from ANT
 519 forcing. The on-line supplementary information accompanying Min et al (2011) includes an
 520 extensive set of sensitivity analyses that consider a broad range of uncertainties affecting their
 521 results.

522 The cause of the
 523 discrepancies between
 524 observed and simulated
 525 changes in both mean and
 526 extreme precipitation
 527 remains to be fully
 528 understood. Explanations
 529 could include uncertainties
 530 in observations, forcing, or
 531 the representation of moist
 532 processes in models. The
 533 observations used in
 534 detection studies to date
 535 have been limited to the
 536 20th century, and have been
 537 based exclusively on station
 538 data. Thus coverage is
 539 limited to land areas only
 540 and in many regions, is
 541 inadequate due to
 542 limitations in observing
 543 network density, access to

544 existing observations for the purposes of scientific research, or lack of capacity or mandate to
 545 facilitate the dissemination of observations. Remote sensing products may eventually solve these
 546 problems, but they have not been used in detection and attribution studies due to homogeneity
 547 concerns and lack of sufficiently long records, although they have been used in some cases for
 548 model evaluation (e.g., Kharin et al, 2007). Without broader coverage it is difficult to assess, for
 549 example, whether discrepancies in changes between models and observations are a global
 550 phenomenon or whether they are regional in nature, reflecting, for example, differences in
 551 moisture transport between models and the observed world. Topography, land-atmosphere
 552 coupling, and the representation of teleconnected patterns of variability all affect precipitation and
 553 are subject to uncertainty due to limited resolution in climate models or lack of complete process
 554 knowledge. In addition, wide uncertainty also remains in aerosol forcing (e.g., Forster et al, 2007),
 555 aerosol transport, the effect of aerosols upon the production of precipitation, and so on, which may
 556 affect both temperature extremes and precipitation extremes. Further, there are differences in the
 557 mechanisms of response to long- and short-wave forcing (e.g., Mitchell et al, 1987; Allen and

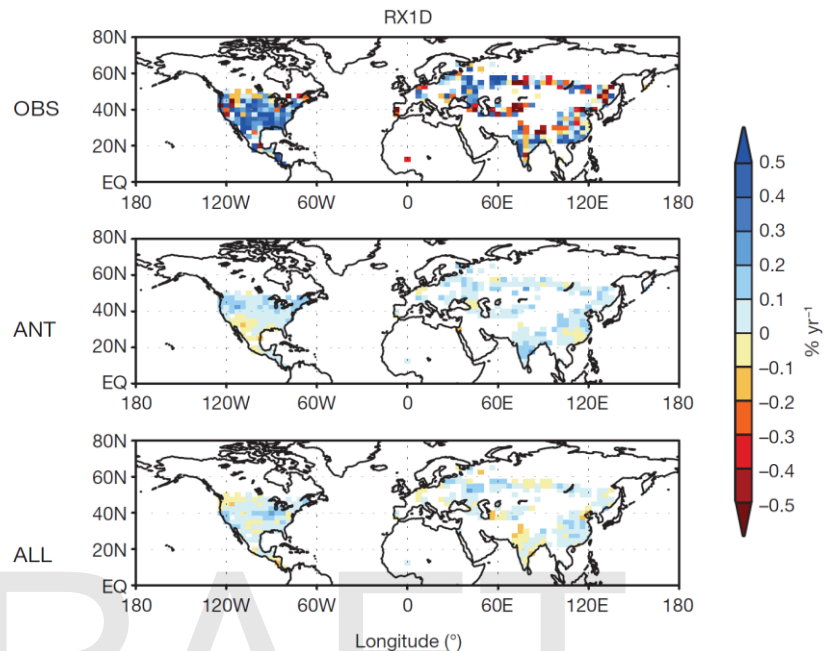


Figure 4: Geographical distribution of trends of extreme precipitation indices (PI) for annual maximum daily precipitation amounts (RX1D) during 1951–99. Observations (OBS); model simulations with anthropogenic (ANT) forcing; model simulations with anthropogenic plus natural (ALL) forcing. For models, ensemble means of trends from individual simulations are displayed. Units: per cent probability per year. From Min et al (2011; see paper for details).

558 Ingram, 2002) and thus the possibility that models may over- or under-simulate the response to
559 one or the other type of forcing.

560 3. Storms

561
562 High energy cyclonic phenomena driven by latent heat release occur in the atmosphere on a
563 number of scales, ranging from individual tornadoes to mesoscale convective complexes to tropical
564 and extra-tropical cyclones. They often cause extensive damage directly by high wind speeds
565 and/or heavy precipitation, and this may be compounded by the effects of flying debris, drifting
566 snow, storm surges and high waves, and wind driven ice movements and other associated events.

567 a. Extra-tropical cyclones

568
569 Extratropical cyclones (synoptic-scale low pressure systems) exist throughout the mid-latitudes and
570 are associated with extreme winds, sea levels, waves and precipitation. Climate models project
571 changes in the large scale flow and reduced meridional temperature gradients as a consequence of
572 greenhouse gas forcing, both of which affect extra-tropical cyclone development, and consequently
573 changes in their number distribution (Lambert and Fyfe, 2006) and in the positioning of extra-
574 tropical storm tracks (Bengtsson et al, 2006).

575 Climate models represent the general structure of the storm track pattern reasonably well
576 (Bengtsson et al., 2006; Greeves et al., 2007; Ulbrich et al., 2008; Catto et al., 2010) although
577 models tend to have excessively zonal storm tracks (Randall et al., 2007). Detecting changes in
578 extra-tropical cyclone numbers, intensity, and activity based on reanalysis remains challenging due
579 to concerns about inhomogeneity that is introduced through changes over time in the observing
580 system, particularly in the southern hemisphere (Hodges et al., 2003; Wang et al., 2006). Even
581 though different reanalyses correspond well in the Northern Hemisphere (Hodges et al., 2003;
582 Hanson et al., 2004), the observing system may also have influenced cyclone characteristics there
583 as well (Bengtsson et al., 2004).

584 Numerous studies using reanalyses suggest that the main northern and southern hemisphere storm
585 tracks have shifted polewards during the last 50 years (e.g., Trenberth et al, 2007). Idealized studies
586 (e.g., Brayshaw et al., 2008; Butler et al., 2011) suggest that greenhouse gas forcing from increases
587 in well mixed greenhouse gases and decreases in stratospheric ozone may have played a role in
588 these shifts. However, for the moment, studies of pressure-based indices (see above) (e.g., Wang et
589 al., 2011 for the European/North Atlantic region, see Figure 5; Alexander et al, 2011 for south-
590 eastern Australia) are not able to provide corroborating evidence of a poleward shift in the principal
591 storm track locations, since in both hemispheres, the domain over which pressure triangles needed
592 to produce these indices are available is rather limited. Ongoing work with single station pressure
593 proxies may help to alleviate this situation in the future. For example, a regional study over Canada
594 that considered changes in observed cyclone deepening rates based on pressure tendencies at

595 stations (Wang et al, 2006) found qualitative agreement between reanalyses and station data
 596 suggesting a northward shift of the winter storm track over Canada.

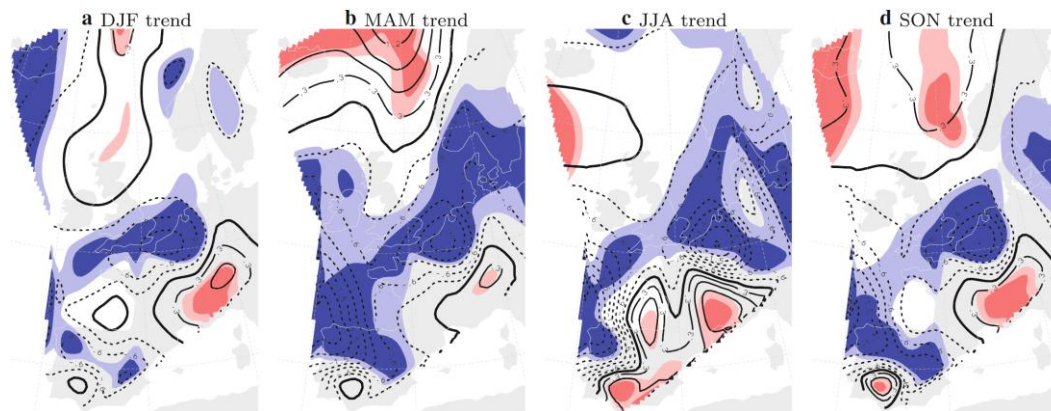


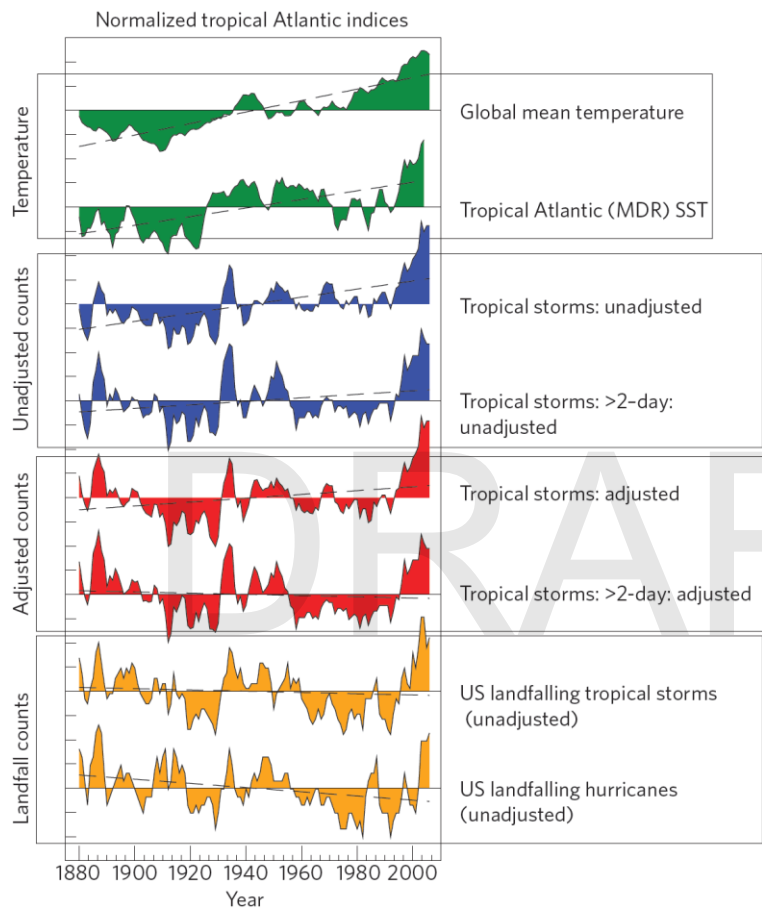
Figure 5: Example of an analysis is trends in seasonal storm indices derived from long surface pressure records. This figure shows contour maps of Kendall's linear trend estimates (in unit per century) in seasonal storm indices defined as the 99th percentile of sub-daily geostrophic wind speed estimated from pressure triangles for the period 1902–2007 in a domain the covers western Europe and the eastern North Atlantic. The contour interval is 0.3. The zero contours are shown in bold. Positive trends are shown in thin solid contours, and reddish shadings indicate at least 20% significance; and negative trends in dashed contours and bluish shadings. The darker shadings indicate areas with trends that are significant at the 5% level or lower. From Wang et al (2011).

597 Detection and attribution studies examining whether human influence has played a role in changes
 598 in cyclone number, intensity or distribution have not yet been conducted. However, human
 599 influence has been detected in the global sea level pressure (Gillett et al, 2005; Gillett and Stott,
 600 2009) and in one study, in geostrophic wind energy derived from sea level pressure records (Wang
 601 et al, 2009). Gillett and Stott (2009) show that observed patterns of trends, which indicate
 602 decreases in high latitude sea level pressure and increases elsewhere, is robust when calculated
 603 from data for 1949–2009. Observed changes were consistent with expectations based on the model
 604 (HadGEM1) used in that study, suggesting that anthropogenic influence has contributed to both
 605 pressure decreases at high latitudes and increases at low latitudes. The mechanism for the latter is
 606 not well understood. Using an approach that would not formally be considered to a detection and
 607 attribution method, Fogt et al (2009) find that both coupled climate model simulated trends and
 608 observed trends in the Southern Annular Mode (SAM) lie outside the range of internal climate
 609 variability during the austral summer, suggesting that human influence has contributed to the
 610 observed SAM trends.

611 b. Tropical cyclones

612 About 90 tropical cyclones have been observed annually since the introduction of geostationary
 613 satellites. The global frequency has remained more or less constant over this period, albeit with
 614 substantial variability in the frequency of tropical cyclones and locations of their tracks within
 615 individual ocean basins (e.g., Webster et al., 2005; Kossin et al., 2010).

616 Tropical cyclones are typically classified in terms of their intensity according to the Saffir-Simpson
 617 scale as indicated by near-surface wind speed or central pressure. Long-term records of the
 618 strongest storms are potentially less reliable than those of their weaker counterparts (Landsea et
 619 al, 2006). In addition to intensity, other impact-relevant characteristics of tropical cyclones include
 620 frequency, duration, track, precipitation, and the structure and areal extent of the wind field in
 621 tropical cyclones, the latter of which can be very important for damage through storm surge as well
 622 as the direct wind-related damage.



645 Figure 6: Tropical Atlantic indices. Green-shaded curves depict global
 646 mean temperature (HadCRUT3 data set) and August–October main
 647 development region (MDR; 10° N–20° N, 80° W–20° W) SST anomalies
 648 (HadISST data set). Blue-shaded curves represent unadjusted tropical
 649 storm counts. Red-shaded curves include time-dependent
 650 adjustments for missing storms based on ship-track density. The
 651 curve labelled '>2-day' depicts storms with a duration greater than
 652 2.0 days. Orange shaded curves depict US landfalling tropical storms
 653 and hurricanes (no adjustments). Solid black lines are five-year means
 654 (1878–2008); dashed black lines are linear trends. Vertical axis ticks
 655 represent one standard deviation. Series normalized to unit standard
 656 deviation. Only the top three series have statistically significant linear
 trends ($p = 0.05$). Figure and caption are from Knutson et al (2010)
 and are based on Vecchi and Knutson (2008) and Landsea et al (2009).

Forming robust physical links between changes in tropical cyclone characteristics and natural or human-induced climate changes is a major challenge. Historical tropical cyclone records are known to be heterogeneous due to changing observing technology and reporting protocols (e.g., Landsea et al, 2004) and because data quality and reporting protocols vary substantially between regions (Knapp and Kruk, 2010). The homogeneity of the global record of tropical cyclone intensity derived from satellite data has been improved (Knapp and Kossin, 2007; Kossin et al, 2007), but these records represent only the past 30-40 years. Statistically significant trends have not been observed in records of the global annual frequency of tropical cyclones (e.g., Webster et al, 2005). Trends in frequency have been identified in the North Atlantic, but are contested (see below). Frequency trends have not been identified in other basins. Power dissipation appears to have increased in the North Atlantic

657 and more weakly in the western North Pacific over the past 25 years (Emanuel, 2007), but the
658 interpretation of longer-term trends is constrained by data quality concerns. A similar metric, the
659 globally accumulated tropical cyclone energy, has recently shown very large variations; it reached a
660 high point in 2005, and subsequently declined to a 40-year low point (Maue, 2009). It remains
661 difficult to robustly place tropical cyclone metrics for recent decades into a longer historical context
662 (Knutson et al, 2010) because pre-satellite records are incomplete and therefore require the use of
663 methods to estimate storm undercounts other biases; these methods have provided mixed
664 conclusions to date (e.g., for the North Atlantic basin, see Holland and Webster, 2007; Landsea,
665 2007; Mann et al, 2007; ; Vecchi and Knutson 2008; Landsea et al. 2009; Knutson et al, 2010; see
666 also Figure 6).

667 Our understanding of the factors that affect tropical cyclone metrics and their variation is
668 improving but remains incomplete. Anthropogenic forcing has been identified as a cause of SST
669 warming in tropical cyclogenesis regions (e.g., Santer et al, 2006; Gillett et al, 2008). Potential
670 intensity theory (Bister and Emanuel, 1998) links changes in the mean thermodynamic state of the
671 tropics to cyclone potential intensity and implies that a greenhouse warming could induce a shift
672 towards greater intensities. This has received some support from dynamical hurricane model
673 simulations (summarized in Knutson et al. 2010, Table S2). These results suggest that human
674 influence could have altered tropical cyclone intensities over the 20th century. However, as noted
675 above, the available evidence concerning historical trends and detectable anthropogenic influence
676 on tropical cyclone characteristics is mixed. A global analysis of trends in satellite-based tropical
677 cyclone intensities has identified an increasing trend that is largest in the upper quantiles of the
678 distribution (Elsner et al, 2008), and also most pronounced in the Atlantic basin. However, this
679 record extends back only to 1981 which is regarded as too short to distinguish a long-term trend
680 from pronounced multi-decadal variability in the Atlantic basin. Historical data show that tropical
681 cyclone power dissipation is related to sea surface temperatures (SSTs), near-tropopause
682 temperatures and vertical wind shear (Emanuel, 2007), but it has been suggested that the spatial
683 pattern of SST variation in the tropics may exert an even stronger influence on Atlantic hurricane
684 activity than absolute local SSTs (Swanson, 2008; Vecchi and Soden, 2007; Ramsay and Sobel,
685 2011). This would have important implications for the interpretation of climate model projections
686 (Vecchi et al, 2008). Related to this, a growing body of evidence suggests that the SST threshold for
687 tropical cyclogenesis (currently about 26°C) would increase at about the same rate as the tropical
688 SST increase due to greenhouse gas forcing (e.g., Ryan et al, 1992; Knutson et al, 2008; Johnston
689 and Xie, 2010). This means, for example, that the areas of simulated tropical cyclogenesis would not
690 expand along with the 26°C isotherm in climate model projections The most recent assessment by
691 the World Meteorological Organization (WMO) Expert Team on Climate Change Impacts on Tropical
692 Cyclones (Knutson et al., 2010) concluded that it remains uncertain whether past changes in any
693 measure of tropical cyclone activity (frequency, intensity, rainfall) exceeds the variability expected
694 through natural causes, after accounting for changes in observing capabilities over time.

695 Based on a variety of model simulations, it is expected that global tropical cyclone frequency will
696 either decrease or display little change as a consequence of greenhouse warming, but that there

697 will be an increase in mean wind speed intensity and in tropical cyclone rainfall rates over the 21st
698 century (Meehl et al., 2007; Knutson et al., 2010). Projected changes for individual basins are more
699 uncertain than global mean projections, as they show large variations between different modelling
700 studies. Studies which have compared tropical cyclone projections downscaled from different
701 climate models using a single downscaling framework (e.g., Zhao et al. 2009; Sugi et al. 2009)
702 suggest that at the regional scale, the uncertainties in tropical cyclone projections due to
703 differences in projected SST patterns is substantial. Concerning detection and attribution of
704 tropical cyclone changes, in addition to the substantial uncertainty in historical records, a further
705 challenge for identifying such predicted changes in observations is that the projected changes are
706 typically small compared to estimated observed natural variability. Modelling studies (e.g. Knutson
707 and Tuleya, 2004; Bender et al, 2010) suggest, on the basis of idealized simulations, that
708 unambiguous detection of the effect of greenhouse gas forcing on tropical cyclone characteristics
709 may still be decades off. Detection of such an anthropogenic influence through the use of tropical
710 cyclone damage statistics could require an even longer period of record (Crompton et al. 2011).

711 c. Tornadoes and other types of small scale severe weather

712 Tornadoes typically occur during severe thunderstorms in which rapid vertical motion and the
713 resulting convergence of angular momentum produces the potential for very high local vorticity.
714 While our understanding of tornadoes has increased in recent years (e.g., Trapp et al, 2005), the
715 body of research that is available globally on changes in tornado frequency and intensity remains
716 limited. This is in part because the available data are inhomogeneous in time (e.g., Brooks, 2004)
717 due to changes in reporting practices as well as changes in population and public awareness, and
718 the introduction of technology such as Doppler radar, all of which undoubtedly affect detection
719 rates. The assessments of Trenberth et al (2007) and Karl et al (2008) contain brief sections
720 summarizing available research on tornadoes and other types of small scale severe weather. The
721 scale of these phenomena implies that using current generation climate models any change in their
722 likelihood of occurrence can only be inferred indirectly from climate models by considering changes
723 in atmospheric conditions such as stability and vertical shear that affect their occurrence. For this
724 reason, as well as the inadequacy of the observational record, detection and attribution studies
725 have not been attempted. Projections of future changes in the incidence and intensity of tornadoes
726 due to greenhouse warming and other climate forcings also remain uncertain, partly because
727 competing influences on tornado occurrence and intensity might change in different ways. Thus, on
728 the one hand, greenhouse gas induced warming may lead to greater atmospheric instability due to
729 increases in temperature and moisture content, suggesting a possible increase in severe weather,
730 but on the other hand vertical shear may decrease due to reduced pole-to-equator temperature
731 gradients (Diffenbaugh et al., 2008).

732 4. Hydrological Extremes

733 We discuss here floods and droughts, which are complex phenomena with large impacts that affect
734 large numbers of people each year. Space and time scales can be large, particularly in the case of

735 droughts which can occur on sub-continental to continental scales and have extended durations of
736 years or longer. In contrast, some types of flooding can be localized and of short duration, although
737 flooding may also occur in large basins over an extended period of time (months). While floods and
738 droughts generally represent opposite ends of the spectrum of variability in a region's hydrological
739 balance, it should be noted that the two phenomena are not completely mutually independent. For
740 example, extreme precipitation events, with the possibility of local flash flooding, can occur during
741 drought (e.g., Carmichael et al, 2010).

742 a. Floods

743 Floods are complex phenomena that are affected by various characteristics of precipitation. For
744 example, freshet flooding is driven by meteorological and synoptic characteristics that control the
745 timing and magnitude of energy fluxes into the snowpack, possibly confounded by the occurrence
746 of rainfall. The frequency and intensity of floods can be altered by natural and human engineered
747 and non-engineered land use effects on drainage basins, which makes the detection of climatic
748 influences difficult. Human engineering induced effects include the possibility that impoundment of
749 water may alter the local precipitation climatology (Hossain, et al, 2009). Storm surge events can
750 cause coastal flooding, which may be exacerbated in estuaries if a storm surge event coincides with
751 heavy discharge. Sea level rise (section 5) can also interact with storm surge events to increase the
752 risk of coastal flooding (Abeyisirigunawardena et al, 2011).

753
754 The IPCC AR4 (Rosenzweig et al, 2007) and the IPCC Technical Paper VI based on the AR4 (Bates et
755 al, 2008) concluded that documented trends in floods show no evidence for a globally widespread
756 change in flooding (see also, for example, Kundzewicz et al, 2005), although there was abundant
757 evidence for earlier spring peak flows and increases in winter base flows in basins characterized by
758 snow storage. They also noted that there was some evidence of a reduction in ice-jam floods in
759 Europe (Svensson et al, 2006). Subsequent research, which continues to be hampered by a limited
760 availability and coverage of river gauge data, provides mixed results. Some studies suggest that
761 there has been an increase in flooding over time in some basins (e.g., some basins in south-east
762 Asia, Delgado et al., 2009; Jiang et al., 2008; and South America, Barros et al., 2004). However,
763 many other studies suggest no climate-driven change (e.g., in northern Asia, Shiklomanov et al.,
764 2007; North America, Cunderlik and Ouarda, 2009; Villarini et al., 2009) or provide regionally
765 inconsistent findings (e.g., in Europe, Allamano et al., 2009; Hannaford and Marsh, 2008; Mudelsee
766 et al., 2003; and Africa, Di Baldassarre et al., 2010), or a change in the characteristics of flooding
767 such as might be expected when a snowmelt driven flood regime switches, with warming, to a
768 mixed snowmelt-rainfall regime (e.g., Cunderlik and Ouarda, 2009).

769
770 River discharge simulation under a changing climate scenario is generally undertaken by driving a
771 hydrological model with downscaled, bias-corrected climate model outputs. However, bias-
772 correction and statistical downscaling tend to ignore the energy closure of the climate system,
773 which could be a non-negligible source of uncertainty in hydrological projections (Milly and Dunne,
774 2011). Most hydrological models must first be tuned on a basin-by-basin basis to account for sub-

775 grid-scale characteristics such as basin hypsometry, the degree of watercourse meander and other
776 channel characteristics. Hydrologic modelling is therefore subject to a cascade of uncertainties from
777 climate forcing, climate models, downscaling approach, tuning, and hydrological model uncertainty;
778 these uncertainties remain difficult to quantify comprehensively.

779
780 Recently, several studies have detected the influence of anthropogenically-induced climate change
781 in variables that may affect floods. These include Zhang et al (2007) who detected human influence
782 in observed changes in zonally averaged land precipitation, Min et al (2008), who detected human
783 influence in northern high-latitude precipitation and Min et al (2011), who detected human
784 influence in observed global scale change in precipitation extremes. Nevertheless, the extent to
785 which such changes in extreme precipitation may lead to changes in flooding depend on the
786 regional climate characteristics of the respective river catchments, as well as on changes in other
787 climate variables such as soil moisture content and mean precipitation. While human influence has
788 not yet been detected in the magnitude/frequency of floods, at least two studies using detection
789 and attribution methodologies that incorporated output from hydrologic models driven with
790 downscaled climate model output have suggested that human influences have had a discernable
791 effect on the hydrology of the regions that they studied. Barnett et al (2008) detected
792 anthropogenic influence in western US snowpack and the timing of peak-flow (see also Hidalgo et
793 al, 2009), and Pall et al (2011), estimated that human influence on the climate system increased the
794 likelihood of a fall 2000 flooding event that occurred in the southern part of the UK.

795
796 Uncertainty is still large in the projected changes in the magnitude and frequency of floods. The
797 largest source of uncertainties in hydrological projections is from differences between the driving
798 climate models, but the choice of future emission scenarios, downscaling method, and hydrologic
799 model also contribute uncertainty (e.g., Kay et al., 2009; Prudhomme and Davies, 2009; Shrestha et
800 al., 2011, Taye et al., 2011). The relative importance of downscaling, bias-correction and the choice
801 of hydrological models as sources of uncertainty may depend on the selected region/catchment,
802 the selected downscaling and bias-correction methods, and the selected hydrological models
803 (Wilby et al, 2008). Chen et al (2011) demonstrated considerable uncertainty was caused by
804 differing results between the several downscaling methods used to make hydrological projections
805 for a snowmelt-dominated Canadian catchment. Downscaling and bias-correction are also a major
806 source of uncertainty in rain-dominated catchments (van Pelt et al, 2009).

807 b. Droughts

808 Drought is a complex phenomenon that is affected by multiple climate variables on multiple times
809 scales, , including atmospheric circulation, precipitation, temperature, wind speed, solar radiation,
810 and antecedent soil moisture and land surface conditions. It can feed back upon the atmosphere via
811 land-atmosphere interactions, potentially affecting the extremes of temperature, precipitation and
812 other variables (e.g., Seneviratne et al, 2010; Nichols and Larsen, 2011). It can take multiple forms
813 including meteorological drought (lack of precipitation), agricultural (or soil moisture) drought and
814 hydrological drought (runoff or streamflow). There are few direct observations of drought-related

815 variables (e.g., Trenberth et al, 2007), including soil moisture, and hence drought proxies such as
 816 the Palmer Drought Severity Index (PDSI – Palmer, 1965; Dai et al, 2004; Heim, 2002) and the
 817 Standardized Precipitation Index (SPI – McKee et al, 1993; Heim, 2002) are often used to monitor
 818 and study changes in drought conditions. However, these indirect indices imply substantial
 819 uncertainties in respective analyses, and in particular the PDSI has several limitations. Hydrologic
 820 drought can be observed/analysed via statistical analysis of discharge records (see e.g., Flieg et al,
 821 2006).

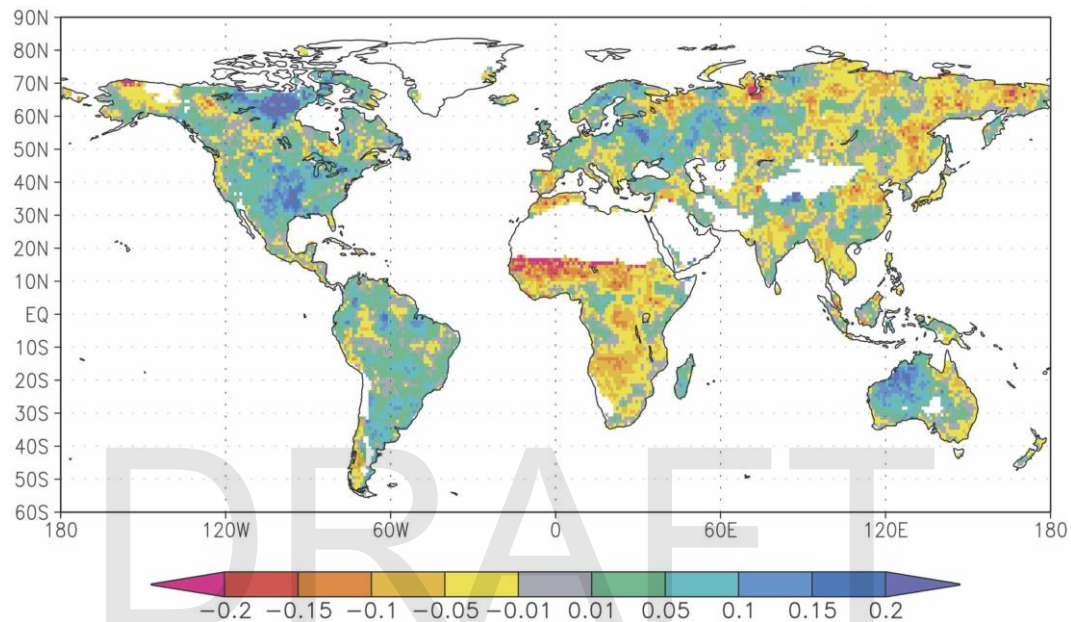


Figure 7: Global distribution of linear trends in annual mean volumetric soil moisture for 1950-2000 obtained from the Variable Infiltration Capacity (VIC) hydrologic model when driven with observationally based forcing. The trends are calculated using the Mann–Kendall nonparametric trend test. Regions with mean annual precipitation less than 0.5 mm day^{-1} have been masked out because the VIC model simulates small drying trends in desert regions that, despite being essentially zero are identified by the nonparametric test. From Sheffield and Wood (2008; Figure 1).

822 Assessments of changes in drought globally remain uncertain. Trenberth et al (2007), using the
 823 dataset of Dai et al (2004), found large increases in dry areas as indicated by the PDSI. However, it
 824 has been noted that the PDSI may not be comparable between diverse climatological regions (e.g.,
 825 Karl, 1983; Alley, 1894). The self-calibrating (sc-) PDSI introduced by Wells et al (2004) attempts to
 826 alleviate this problem by replacing fixed empirical constants with values based on the local climate.
 827 Using the sc-PDSI van der Schrier et al (2006) show that 20th century soil moisture trends in Europe
 828 are not statistically significant Using a more comprehensive land surface model than that implicit in
 829 either the PDSI or sc-PDSI, together with observation-based forcing, Sheffield and Wood (2008)
 830 inferred that during 1950-2000 decreasing trends in drought duration, intensity and severity were
 831 prevalent globally (Figure 7). However, they also noted strong regional variation and increases in
 832 drought indicators in some regions, consistent with some regional studies. For example, Andreadis
 833 and Lettenmaier (2006), using a similar approach, found increasing trends in soil moisture and
 834 runoff in much of US in the latter half of 20th century. On the other hand, Dai (2011) found a global

835 tendency for increases in drought
 836 based on various versions of the
 837 PDSI including the sc-PDSI and soil
 838 moisture from a land surface
 839 model driven observation-based
 840 forcing (Figure 8). Patterns of
 841 change obtained with those
 842 different techniques were largely
 843 consistent, with substantial spatial
 844 variability being a dominant
 845 characteristic. Nevertheless,
 846 inconsistencies between studies
 847 and indicators demonstrate that
 848 there remain large uncertainties
 849 with respect to global assessments
 850 of past changes in droughts,
 851 making it difficult to confidently
 852 attribute observed changes to
 853 external forcing on the climate
 854 system.

855 Characterising hydrologic (i.e.
 856 runoff and streamflow) drought
 857 globally and regionally is also
 858 challenging due to difficulties in
 859 establishing robust and/or
 860 standardised quantitative drought
 861 descriptions over varied hydrologic
 862 regimes (e.g., Flieg et al, 2006).
 863 Some recent examples regarding
 864 analysis of streamflow records for
 865 detection of possible trends in low
 866 flow include work in Europe (Stahl
 867 et al, 2010), Canada (Ehsanzadeh
 868 and Adamowksi, 2007) and the UK
 869 (Hannaford and Marsh, 2006).

870 Despite these uncertainties in global scale studies, there is often more agreement amongst regional
 871 studies of historical and current drought, consistent with the notion that circulation changes should
 872 induce regionally coherent shifts in drought regimes. For example, precipitation is strongly affected
 873 by the El Niño/Southern Oscillation in many parts of the world (Ropelewski and Halpert, 1987)
 874 including extremes (Alexander et al, 2009; Kenyon and Hegerl, 2010; Zhang et al, 2010) and the

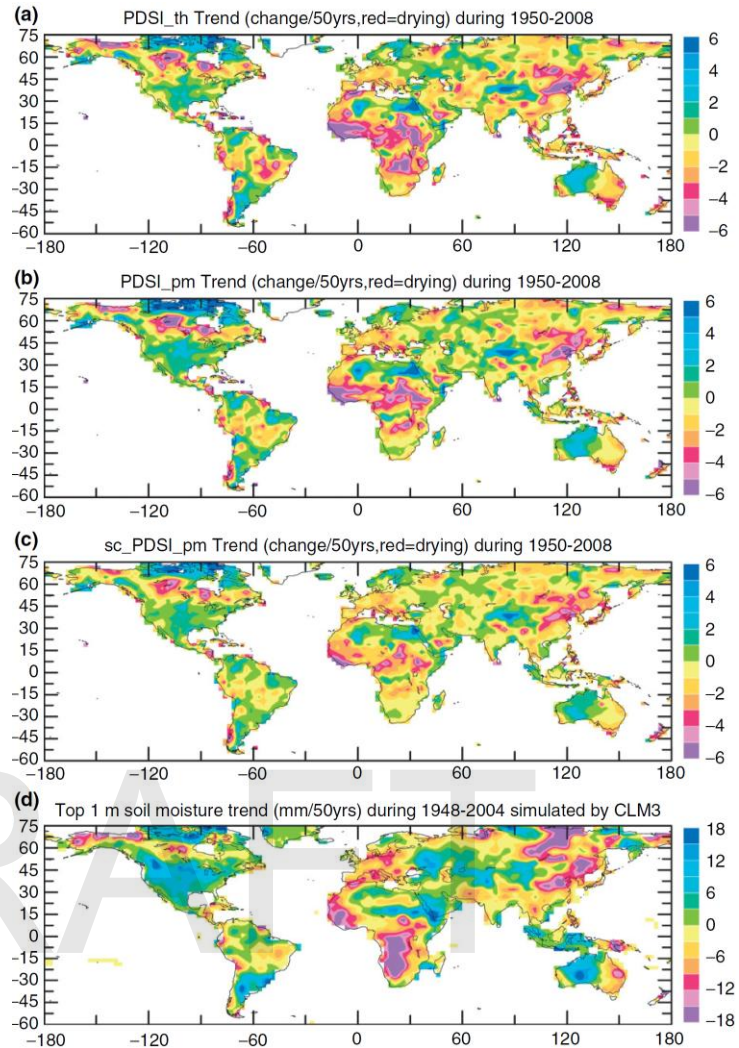


Figure 8: Maps of annual trends (red = drying) from 1950 to 2008 in PDSI (change per 50 years) with potential evapotranspiration (PE) calculated using the (a) Thornthwaite and (b) Penman-Monteith (PM) equation, and (c) annual trends in self-calibrated PDSI with the PM potential evaporation. Also shown (d) is the trend in top-1 m soil moisture content (mm/50 years) from 1948 to 2004 simulated by a land surface model (CLM3) forced by observation-based atmospheric forcing (see Qian et al, 2006, for details). From Dai (2011).

875 resulting teleconnected circulation responses are often linked to the occurrence of precipitation
876 deficits and drought in different regions (e.g., Folland et al, 1986; Hoerling and Kumar, 2003; Held
877 et al, 2005; Hoerling et al, 2006; Giannini et al, 2008, Schubert et al, 2009) although internal
878 atmospheric variability that is not forced by slowly changing boundary conditions can also create
879 drought (e.g., Hoerling et al, 2009). Also, progress is being made in understanding the role of land-
880 atmosphere feedbacks that affect surface conditions (e.g., Koster et al, 2004; Seneviratne et al,
881 2006, 2010; Fischer et al, 2007), although the rate of advance is limited by the availability of
882 observational data.

883 Christensen et al (2007) provide an assessment of regional drought projections based on
884 simulations that were performed for CMIP3, noting consistency across models in projected
885 increases in droughts particularly in subtropical and mid-latitude areas. Uncertainty in drought
886 projections stem from multiple sources. Perhaps the most fundamental of these is the uncertainty
887 in the pattern of sea-surface temperature response to forcing, which is “El Niño like” in many
888 models (Meehl et al, 2007), and which therefore cascades to other aspects of model behaviour
889 through the teleconnected responses to SST change. A second source of uncertainty is associated
890 with the possible alteration of land-atmosphere feedback processes, both as a consequence of
891 change in the physical climate system and change in the terrestrial biosphere. A third source of
892 uncertainty arises because the complexities of drought are at best incompletely represented in
893 commonly used drought indices, leading to potential discrepancies of interpretation. For example,
894 Orlosky and Seneviratne (2011) show, using a more complete ensemble of CMIP3 simulations
895 than was available at the time of Christensen et al (2007), that ensemble projections based on
896 meteorological and agricultural drought indices can be quite different. Also, Burke and Brown
897 (2008), considering several drought indices and two different ensembles of climate model
898 simulations, show little change in the proportion of the land surface that is projected to be in
899 drought based on the SPI, whereas indices that account for change in the atmospheric demand for
900 moisture showed significant increases in the global land area affected by drought. It has been
901 suggested that inferences based on climate model simulated soil moisture may be more robust
902 than those based on other types of drought indicators. This is because model results are often
903 found to be consistent after simple scaling (e.g., Koster et al, 2009; Wang et al 2009).

904 5. Sea level

905 Transient sea level extremes caused by severe weather events such as tropical or extratropical
906 cyclones can produce storm surges and extreme wave heights at the coast. Extreme sea levels may
907 change in the future as a result of both changes in atmospheric storminess and mean sea level rise,
908 neither of which will be spatially uniform across the globe. Sea level change along coast lines may
909 also be affected by some additional factors including glacial isostatic adjustment, coastal
910 engineering, and changes in the Earth’s gravitational field (e.g., Mitrovica et al, 2010) arising from
911 glacial and ice-sheet melting. Global mean sea level rose at an average rate of 1.7 [1.2 to 2.2] mm
912 yr⁻¹ over the 20th century, 1.8 [1.3 to 2.3] mm yr⁻¹ over 1961 to 2003, and at a rate of 3.1 [2.4 to 3.8]
913 mm yr⁻¹ over 1993 to 2003 (Bindoff et al, 2007). Externally induced sea level rise occurs against a

914 backdrop of natural variability in sea level that must be taken into account when attributing causes
 915 to observed changes. For example, natural modes of variability such as the El Niño/Southern
 916 Oscillation (Menendez and Woodworth, 2010), the Pacific Decadal Oscillation
 917 (Abeyasingunawardena and Walker, 2008), the North Atlantic Oscillation (Marcos et al, 2009) and
 918 the position of the South Atlantic high (Fiore et al, 2009) all have transient effects on extreme sea
 919 levels. It is *very likely* that humans contributed to sea level rise during the latter half of the 20th
 920 century (Hegerl et al, 2007), and therefore *more likely than not* that humans contributed to the
 921 trend in extreme high sea levels (Solomon et al, 2007). Both mean and extreme sea level has
 922 continued to rise since the AR4 (Church et al, 2011; Menendez and Woodworth, 2010; Woodworth
 923 et al, 2011; see Figure 7).

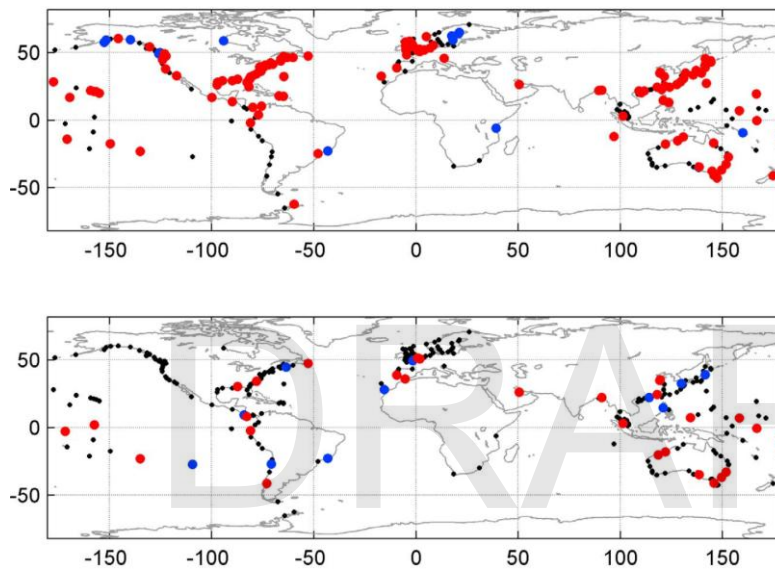


Figure 7: Estimated trends in (upper) annual 99th percentile of sea level based on monthly maxima of hourly tide gauge readings from 1970 onwards, and (lower) 99th percentile after removal of the annual medians of hourly readings. Only trends significant at the 940 level are shown in colour: red for positive trends and blue for negative trends. From Menéndez and Woodworth (2010). The figure show that extreme sea levels have risen broadly, and that the dominate influence on that rise is from the increase in mean sea level.

Meehl et al (2007) projected model based 90% ranges for sea level rise for 2090–2099 relative to 1980–1999 that varied from 18–38 cm in the case of the SRES B1 scenario to 26-59 cm in the case of the A1FI scenario. These estimates accounted for ocean thermal expansion, glaciers and ice caps, and modelled aspects of ice sheets. It was also estimated that an acceleration of the flow of ice from Greenland and Antarctic could increase the upper ends of these ranges by 10–20 cm, and it was noted that insufficient understanding of ice sheet dynamics meant that a larger contribution could not be ruled out. Subsequent studies that

945 extrapolate statistical models linking temperature and sea level have suggested somewhat higher
 946 ranges, for example, 0.75 - 1.90 m (Vermeer and Rahmstorf, 2009, based on SRES B1 to A1FI
 947 scenarios), and 0.90 - 1.30 m (Grinsted et al, 2010, based on the SRES AIB scenario only).

948 Projections of extreme sea level can be produced regionally in several ways. Often, such studies
 949 involve a combination of downscaling and hydrodynamic modelling (e.g., Deberhard and Roed,
 950 2008, who consider the European region and projected both decreases and increases depending
 951 upon location). Such an approach may not be feasible in all locations if the driving climate model
 952 does not simulate the phenomena that are likely to cause storm surge in a given region (e.g.,
 953 tropical cyclones). In such cases it may be possible to construct statistical or idealized models of

954 tropical cyclone characteristics from observations that can then be perturbed to represent future
955 conditions and to drive hydrodynamic models (e.g., McInnes, 2003; Harper et al, 2009; Mousavi et
956 al. 2011). A further approach is to conduct sensitivity analyses to assess the relative impacts on
957 mean sea level rise and wind speed increase (e.g., McInnes et al, 2009).

958 6. Summary and Recommendations

959 **To be completed after the Open Science Conference**

960

DRAFT

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