



HAL
open science

Decadal climate variability in the tropical Pacific: Characteristics, causes, predictability, and prospects

Scott Power, Matthieu Lengaigne, Antonietta Capotondi, Myriam Khodri,
Jérôme Vialard, Beyrem Jebri, Éric Guilyardi, Shayne Mcgregor, Jong-Seong
Kug, Matthew Newman, et al.

► **To cite this version:**

Scott Power, Matthieu Lengaigne, Antonietta Capotondi, Myriam Khodri, Jérôme Vialard, et al.. Decadal climate variability in the tropical Pacific: Characteristics, causes, predictability, and prospects. *Science*, 2021, 374 (6563), 10.1126/science.aay9165 . hal-03431253

HAL Id: hal-03431253

<https://hal.science/hal-03431253v1>

Submitted on 6 Jan 2025

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

TITLE

Decadal climate variability in the tropical Pacific: Characteristics, causes, predictability, and prospects

AUTHORS

Power, S; Lengaigne, M; Capotondi, A; et al.

JOURNAL

Science

DEPOSITED IN ORE

05 October 2021

This version available at

<http://hdl.handle.net/10871/127350>

COPYRIGHT AND REUSE

Open Research Exeter makes this work available in accordance with publisher policies.

A NOTE ON VERSIONS

The version presented here may differ from the published version. If citing, you are advised to consult the published version for pagination, volume/issue and date of publication

Decadal climate variability in the tropical Pacific: characteristics, causes, predictability and prospects

Scott Power^{1,2,3}, Matthieu Lengaigne⁴, Antonietta Capotondi^{5,6}, Myriam Khodri⁷, Jérôme Vialard⁷, Beyrem Jebri⁷, Eric Guilyardi^{7,8}, Shayne McGregor², Jong-Seon Kug⁹, Matthew Newman^{5,6}, Michael J. McPhaden¹⁰, Gerald Meehl¹¹, Doug Smith¹², Julia Cole¹³, Julien Emile-Geay¹⁴, Daniel Vimont¹⁵, Andrew T. Wittenberg¹⁶, Mat Collins¹⁷, Geon-Il Kim⁹, Wenju Cai^{18,19,20}, Yuko Okumura²¹, Christine Chung²², Kim M. Cobb²³, François Delage²², Yann Y. Planton¹⁰, Aaron Levine¹⁰, Feng Zhu²¹, Janet Sprintall²⁴, Emanuele Di Lorenzo²⁵, Xuebin Zhang¹⁸, Jing-Jia Luo²⁶, Xiaopei Lin^{19,20}, Magdalena Balmaseda²⁷, Guojian Wang¹⁸, Benjamin J. Henley^{2,3}

July 27, 2021

Corresponding Author:

Professor Scott B. Power, Dip. Ed.

Director, Centre for Applied Climate Sciences, University of Southern Queensland

Email: scott.power@usq.edu.au; Phone: +61 400 650 48; ORCID ID: 0000-0002-9596-4368.

¹ Centre for Applied Climate Sciences, University of Southern Queensland, QLD, Australia

² School of Earth, Atmosphere, and Environment, Monash University, VIC, Australia

³ ARC Centre of Excellence for Climate Extremes, Monash University, VIC, Australia

⁴ MARBEC, University of Montpellier, CNRS, IFREMER, IRD Sète, France

⁵ Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, CO, USA

⁶ Physical Sciences Laboratory, NOAA, Boulder, CO, USA

⁷ LOCEAN, Sorbonne Universités/UPMC/CNRS/IRD, Paris, France

⁸ National Centre of Atmospheric Science, University of Reading, Reading, UK

⁹ Division of Environmental Science and Engineering, Pohang University of Science and Technology (POSTECH), Pohang, South Korea

¹⁰ NOAA/Pacific Marine Environmental Laboratory, Seattle, WA, USA

¹¹ National Center for Atmospheric Research, Boulder, CO, USA

¹² Met Office Hadley Centre, Exeter, UK

¹³ Department of Earth and Environmental Sciences, University of Michigan, Ann Arbor, USA

¹⁴ Department of Earth Sciences, University of Southern California, Los Angeles, USA

¹⁵ Atmospheric and Oceanic Science, University of Wisconsin–Madison, Madison, Wisconsin

¹⁶ NOAA Geophysical Fluid Dynamics Laboratory, Princeton, NJ, USA

¹⁷ College of Engineering, Mathematics and Physical Sciences, University of Exeter, Exeter, EX4 4QE, UK

¹⁸ Centre for Southern Hemisphere Oceans Research, CSIRO Oceans and Atmosphere, Hobart 7001, TAS, Australia

¹⁹ Frontier Science Center for Deep Ocean Multispheres and Earth System and Laboratory of Physical Oceanography, Ocean University of China

²⁰ Qingdao National Laboratory for Marine Science and Technology, Qingdao 266003, China

²¹ Institute for Geophysics, Jackson School of Geosciences, The University of Texas at Austin, Austin, Texas

²² Bureau of Meteorology, Docklands, VIC, Australia

²³ School of Earth and Atmospheric Sciences, Georgia Institute of Technology, Atlanta, USA

²⁴ Scripps Institution of Oceanography, University of California San Diego, La Jolla, California, USA

²⁵ Program in Ocean Science & Engineering, Georgia Institute of Technology, Atlanta, Georgia, USA

²⁶ Institute for Climate and Application Research (ICAR)/CICFEM/KLME/ILCEC, Nanjing University of Information Science and Technology, Nanjing, China

²⁷ European Centre for Medium-Range Weather Forecasts, Reading, United Kingdom

19 **Structured Abstract**

20 **BACKGROUND**

21 Tropical Pacific Decadal climate Variability and change (TPDV) affects the global climate system,
22 extreme weather events, agricultural production, streamflow, marine and terrestrial ecosystems, and
23 biodiversity. While major international efforts are underway to provide decadal climate predictions,
24 there is still a great deal of uncertainty about the characteristics and causes of TPDV, and the
25 accuracy to which it can be simulated and predicted. Here we critically synthesize what is currently
26 known and what is not known, and provide recommendations to improve our understanding of
27 TPDV and our ability to predict it.

28

29 **ADVANCES**

30 TPDV is evident in instrumental records, paleoclimate records over past millennia, and climate
31 models. TPDV can occur spontaneously as "internal" variability, as is largely the case in the central
32 equatorial Pacific, or in response to "external" forcing. While internal TPDV arises to a large extent
33 as a residual of independent El Niño-Southern Oscillation events, it can also result from oceanic
34 processes occurring at decadal timescales involving the upper-ocean overturning circulation
35 known as subtropical-tropical cells, and in response to internal atmospheric variability in the extra-
36 tropical Pacific and changes in sea surface temperature in other ocean basins. "Externally-forced"
37 TPDV, in the form of mean-state changes that unfold on decadal timescales or forced decadal
38 variability, can be driven by anthropogenic (e.g., greenhouse gas (GHG) increases, sulphate
39 aerosols changes) and natural processes (e.g., volcanic eruptions). External forcing can also affect
40 the behavior and characteristics of internal TPDV.

41

42 In the western tropical Pacific, GHG-forced warming has reached levels that are unprecedented in
43 the historical record. Further greenhouse warming in the equatorial Pacific will ensure that record-
44 setting high temperatures will be experienced for decades to come. Increases in equatorial
45 precipitation and in precipitation variability in parts of the tropical Pacific, and a southward
46 expansion of the southern hemisphere Hadley Cell, are projected by climate models with some
47 confidence. Yet projected changes in eastern equatorial Pacific surface temperature, and changes
48 in the strength of the Walker Circulation and trade winds, remain very uncertain.

49
50 Skill in decadal predictions of temperature in the western Pacific is apparent, though it appears to
51 be largely underpinned by GHG warming. There are also indications of multiyear skill in
52 predicting some biogeochemical quantities important for fisheries and the global carbon budget.

53
54 The limited length of the instrumental records, the scarcity of paleoclimate data, and TPDV
55 representation biases in climate models have so far prevented a complete characterization and
56 understanding of TPDV and have limited our ability to predict TPDV.

57

58 **OUTLOOK**

59 While several mechanisms have been proposed to explain TPDV, their relative importance as
60 sources of decadal prediction remains unclear. Issues in need of greater understanding include the
61 role played by the upper ocean overturning circulation in controlling tropical Pacific sea surface
62 temperatures at decadal timescales, the impact of external forcing on the Walker circulation and
63 characteristics of internally-generated TPDV, and the extent to which sea surface temperature
64 variability in other basins drives TPDV. A better understanding of the origin and spatial pattern of

65 current predictive skill is also needed. Improving predictions and projections requires
66 improvements in the quality, quantity, and length of instrumental and paleoclimate records, in the
67 performance of climate models and data assimilation methods used to make predictions.

68 **Introduction**

69 Climate variability in the tropical Pacific affects global climate on a wide range of timescales. On
70 interannual timescales, the tropical Pacific is home to the El Niño-Southern Oscillation (ENSO),
71 the most energetic and influential climate phenomenon in the world (1). Less well known is that
72 decadal variations and changes in the tropical Pacific, referred to here collectively as “Tropical
73 Pacific Decadal Variability” (TPDV), also profoundly affects the climate system. In the following,
74 we will use TPDV to refer to any form of decadal climate variability or change that occurs in the
75 atmosphere, the ocean and over land within the tropical Pacific. “Decadal” is used here in a broad
76 sense to encompass multiyear through multidecadal timescales, including variability about the
77 mean–state on decadal timescales, externally forced mean-state changes that unfold on decadal
78 timescales, and decadal variations in the behavior of higher-frequency modes like ENSO.

79
80 Naturally occurring, spontaneously generated TPDV can arise in the absence of any change to
81 external forcing (e.g., greenhouse gas (GHG) increases or volcanic eruptions). Climate scientists
82 refer to such variability as “internally-generated” or “internal” variability, and will be referred to
83 here as “internal TPDV” (2). Internal TPDV affected the rate at which globally-averaged surface
84 air temperature rose over the past century. This was dramatically illustrated by the recent and highly
85 publicized “global warming slow-down”, when decadal surface cooling in the eastern equatorial
86 Pacific (shading in Fig. 1A) associated with a major redistribution of heat in the subsurface ocean
87 offset the anthropogenic global warming trend at the turn of the 21st century (3, 4) (bottom curve
88 on Fig. 1D). Trade winds intensification associated with this cooling also contributed to rapid sea-
89 level rise in the western Pacific during recent decades (5) (contours in Fig. 1A). More generally,
90 internal TPDV has further been reported to modulate drought, wildfire, floods, extreme weather,

91 polar sea-ice extent (6, 7), decadal variations in the impact that ENSO has on rainfall, river flow
92 and agricultural production, and the skill with which ENSO impacts can be predicted, as
93 demonstrated for Australia (8). Uncertainty in the magnitude of internal TPDV simulated in global
94 climate models may also be linked to uncertainty in simulated climate sensitivity (9) – a measure
95 of the degree of global warming that occurs in response to anthropogenic increases in atmospheric
96 GHG concentrations (10).

97
98 The tropical Pacific also changes in response to external forcing, including GHG increases, volcanic
99 eruptions and anthropogenic aerosols. This component of TPDV will be referred to as “external
100 TPDV”. The observed low-frequency sea-surface temperature (SST) evolution over the western
101 tropical Pacific warm-pool is dominated by a long-term warming trend similar to the global
102 anthropogenic warming signal (bottom curves of Fig. 1D), that has been linked to a drying trend in
103 the East Asian monsoon (11). Further warming in the region is also expected to reduce coastal fish
104 populations, shift tuna distribution eastward, cause record-breaking high temperatures to occur
105 more often (12) and fundamentally alter coral reefs, with major impacts on biodiversity, Pacific
106 Island communities, and livelihoods (12, 13).

107
108 Major international efforts are underway to provide decadal climate predictions that are intended to
109 help decision makers plan for coming years and decades (14) that take both internally generated
110 and externally-forced TPDV into account, as they will both influence future climate. The enormous
111 challenges currently faced by groups producing decadal predictions demand a better understanding
112 of the mechanisms of TPDV. To that end, here we synthesize our current understanding of TPDV,
113 its spatial and temporal characteristics, its many proposed mechanisms – both natural and

114 anthropogenic and the interactions between them, and the current ability of state-of-the-art
115 modeling and prediction systems to simulate and predict TPDV. A wide and diverse array of
116 evidence is used, from historical records, instrumental and paleoclimate observations, mathematical
117 models of Earth's climate, and decadal prediction systems, to assess the degree of confidence we
118 have in proposed mechanisms and the extent to which those processes provide a degree of
119 predictability (2, 14, 15).

120

121 **Advances**

122 *Observed TPDV*

123 Decadal SST fluctuations peak in the equatorial central/eastern Pacific (contours in Fig. 1B),
124 alternating between decadal periods of anomalously warm and cold phases (top panel of Fig. 1D).
125 This evolution broadly matches the positive and negative phases of the Interdecadal Pacific
126 Oscillation (δ) (represented as vertical shading in Fig. 1D), characterized by opposite SST and sea-
127 level signals in the eastern and western tropical Pacific (Fig. 1A). While important on decadal
128 timecales, this variability only modestly contributes to total SST variations in the equatorial eastern
129 Pacific (shading in Fig. 1B), which are largely dominated by ENSO-related interannual SST
130 fluctuations. The relative contribution of TPDV is considerably larger in the western Pacific, where
131 the low-frequency SST signal is dominated by a long-term warming trend similar to the global
132 anthropogenic warming signal (bottom curves of Fig. 1D). Consequently, internal TPDV, estimated
133 here in the 8-40 years range, dominates in the central Pacific and in off-equatorial bands in the
134 eastern part of the basin, especially in the northern Hemisphere. Internal TPDV has a weak signature
135 in the western tropical Pacific (Fig. 1C), where the longer timescales of external TPDV prevail
136 instead.

137
138 Confidently characterizing TPDV is complicated by the short historical record. Indeed, historical
139 observations of the tropical Pacific are sparse before the mid-20th century, which creates
140 uncertainty in tropical Pacific SST records prior to 1950. Paleoclimate records offer key
141 complementary information extending further into the past (Fig. 1D). A recent synthesis of
142 dozens of monthly- to annually-resolved Pacific coral records exhibit strong decadal variability
143 over the last four centuries (*16*). In particular, isotopic measurements from corals in the central
144 and southwest tropical Pacific (*17, 18*) (white dot and star in Fig. 1A) display a monotonic trend
145 toward warmer and wetter conditions over the twentieth century (blue curves in Fig. 1D),
146 supporting the central Pacific rainfall increase in response to anthropogenic forcing found in
147 models. On shorter timescales, these timeseries display opposite signals, capturing the contrast
148 between warm/wet and cold/dry regions related to the internal TPDV pattern (Fig. 1A). Recent
149 advances in paleoclimate data assimilation have enabled the construction of gridded tropical
150 Pacific SST fields extending through the last millennium (*19*), that match qualitatively well the
151 observed SST evolution over the instrumental period (Fig. 1D). These reconstructions, however,
152 exhibit increasing uncertainty as fewer records become available back in time (Fig. 1D, gray
153 envelopes).

154

155 **Internal TPDV**

156 The dominant internal TPDV SST signal can be characterized by the leading empirical orthogonal
157 function and related Principal Component of 8-40 year variability of tropical Pacific SST (Fig. 2,
158 A and B), which accounts for about half of the tropical Pacific SST variability in the 8-40 year
159 timescale (contours in Fig. 1C). The related SST pattern throughout the entire Pacific basin can then

160 be obtained through linear regression upon this principal component. This internal TPDV SST
161 pattern (Fig. 2A; shading) strongly resembles the ENSO SST pattern in the tropics (Fig. 2E), but
162 has a generally broader latitudinal extent (20, 21). This pattern is also associated with a zonal seesaw
163 in tropical Pacific mean sea-level pressure, as described by the Southern Oscillation index (Fig.
164 2B), a measure of the difference in sea level pressure between Tahiti and Darwin, leading to the
165 anti-correlation between this index and the SST-based internal TPDV index at decadal timescales
166 (Fig. 2B). The internal TPDV SST pattern (Fig. 2A) is consistent with the patterns of variability
167 associated with the Interdecadal Pacific Oscillation (8) over the Pacific basin and the Pacific
168 Decadal Oscillation in the North Pacific (22), highlighting the important role of the tropical Pacific
169 in forcing and synchronizing decadal variability in both hemispheres (22).

170
171 The simplest explanation for internal TPDV - the *null hypothesis* - is that it arises as a residual of
172 largely independent ENSO events (21, 23, 24). We hence first establish the extent to which observed
173 TPDV might result from decadal averages of ENSO events that are randomly distributed, without
174 modification by independent decadal dynamics or decadal clustering through nonlinear interactions.
175 In this view, the random occurrence of decadal epochs with a larger number of El Niño (La Niña)
176 can be expected to result in an El Niño (La Niña)-like residual SST anomaly (21). The null hypothesis
177 is supported by the good correspondence between the time series of the leading pattern of internal
178 TPDV and the relative number of El Niño and La Niña events during partially-overlapping 8-year
179 periods (Fig. 2B). In addition, each ENSO event is not a static pattern, but undergoes an evolution
180 from a precursor phase (Fig. 2D), through a mature phase (Fig. 2E), to a decay phase (Fig. 2F), so
181 that the average over this seasonal evolution across multiple events can result in a latitudinally
182 broader pattern (Fig. 2C) very similar to the pattern in Fig. 2A (24).

183
184 This picture is further modified by ENSO asymmetries. For example: the strongest El Niño events
185 are larger than the strongest La Niña events; La Niña events tend to last longer than El Niño events;
186 and strong El Niño events tend to occur further east than strong La Niña events, impacting the
187 details of TPDV (25). Differences in spatial patterns, with some events having greatest amplitude
188 in the central or eastern tropical Pacific (1, 26) may result in mean pattern differences during
189 different decadal epochs (25), which may themselves occur purely by chance (25, 27).

190
191 Whether decadal changes in the background state, even if randomly forced, feedback and modulate
192 ENSO characteristics is a focus of current research (28). To go beyond the null hypothesis,
193 therefore, we would need evidence that slow oceanic processes provide sources of decadal
194 predictability beyond the ENSO timescale. For example, the ocean integration of ENSO-related
195 surface fluxes may result in low-frequency SST variations with enhanced decadal predictability, as
196 illustrated in a modeling context (21).

197
198 In addition, changes in the strength of the wind-driven upper-ocean overturning circulation, known
199 as the Subtropical-Tropical Cells (STCs), which connect the subtropical and equatorial regions,
200 have been related to decadal variations of equatorial SSTs in both observations (29) and models
201 (30, 31). As schematically depicted in Fig. 3A, the STCs include subsurface equatorward flow,
202 equatorial upwelling, and poleward flow in the surface Ekman layer, so that a strengthening
203 (weakening) of the STCs results in enhanced (reduced) equatorial upwelling of cold subsurface
204 waters, impacting equatorial SSTs. The adjustment of the STCs to changes in surface wind forcing
205 is accomplished through the propagation of oceanic Rossby waves (30) that travel to the western

206 ocean boundary, and then along the boundary to the equator as coastal-trapped waves, where they
207 alter the depth of the equatorial thermocline and influence SST anomalies (Fig. 3A) (32, 33). Since
208 Rossby waves are more efficiently excited by wind anomalies with larger spatial scales and longer
209 timescales, they can dynamically “filter” the wind forcing, and contribute to an enhancement of
210 low-frequency power (34).

211
212 Other modeling studies suggest that temperature anomalies subducted in the subtropics can reach
213 the equatorial thermocline and influence equatorial SSTs (35), especially when they are density-
214 compensated by associated salinity anomalies (so-called “spiciness” anomalies) and can then
215 propagate toward the equator along mean density surfaces with minimal dissipation (36). While
216 the ability of spiciness anomalies to reach the equator from their source regions has been recently
217 demonstrated in a modeling context (36), their impact on equatorial SSTs in nature remains to be
218 determined.

219
220 The above ocean processes occurring on decadal timescales are mostly wind-forced. In particular,
221 modeling studies (31) indicate that subtropical winds play a key role in altering the strength of the
222 STCs, but the origin of these anomalous winds is still unclear. Extra-tropical influences could be
223 a source of sub-tropical wind variations. For example, internal atmospheric variability in the
224 northern midlatitudes during winter can create subtropical SST and wind anomalies that persist
225 through summer due to strong air-sea coupling (37), developing into a SST pattern that extends
226 from the coast of California toward the central-western equatorial Pacific (Fig. 2D), known as the
227 “North Pacific Meridional Mode” (38). Similarly, an anomalous SST pattern, known as the South
228 Pacific Meridional Mode (39), can develop along the coast of South America (Fig. 2D). These

229 Meridional Modes are considered ENSO precursors, but their associated winds could also provide
230 anomalous forcing for the slow tropical oceanic processes described above (40). As ENSO
231 precursors, they are also part of a seasonal progression from the extra-equatorial ENSO precursor
232 stage, to ENSO development, to extra-tropical ENSO teleconnections that can act as a filter of
233 decadal variance in the Pacific basin (41). Climate model sensitivity experiments, where the North
234 and South Pacific Meridional Modes were alternatively suppressed (42), suggest a potentially more
235 important influence of the South Pacific on internal TPDV, consistent with other model-based
236 studies (43, 44). SST anomalies in the Atlantic and Indian Oceans could also influence these
237 tropical Pacific winds (45, 46), as discussed below.

238
239 Finally, changes in subtropical-tropical winds may arise as a response to the equatorial SST
240 anomalies themselves, as shown in some modeling studies (32, 47). In this view, atmospheric
241 teleconnections triggered by the tropical SST anomalies at decadal timescales alter the extra-
242 equatorial atmospheric circulation and produce wind anomalies of the opposite sign that force a
243 phase reversal of the decadal cycle. This view of internal TPDV as arising from low-frequency
244 processes that are independent of ENSO, with important implications for decadal predictability,
245 remains very challenging to demonstrate observationally, due to the insufficient duration of the
246 instrumental record in the presence of climate noise that may obscure the various deterministic
247 links.

248
249 ***Representation of internal TPDV in climate models***
250 In practice, evaluating internal TPDV simulated by climate models is challenging because: (i) the
251 instrumental record is relatively short (e.g., Fig. 1D); (ii) relatively few multi-century paleoclimate

252 records exist for the core regions of internal TPDV; (iii) internal TPDV in climate model
253 simulations exhibits large changes from one century to the next (48); and (iv) the characteristics
254 of internal TPDV vary markedly from model to model (Fig. 4).

255

256 As illustrated in Fig. 4, climate models still display major deficiencies in simulating key aspects
257 of internal TPDV (49, 50). For example, most models capture to first order the observed SST
258 pattern but the equatorial Pacific warming extends too far to the west (compare shading in Fig. 4A
259 and Fig. 4B; see also (49)), as do simulated ENSO SST patterns (51). Models also markedly
260 underestimate sea-level signals in the tropical western Pacific and extra-tropical central Pacific
261 associated with internal TPDV (compare contours on Fig. 4A and Fig. 4B). Similarly, the
262 magnitude of simulated internal TPDV varies considerably from one model to another and is
263 underestimated by a majority of models not only in terms of SST (Fig. 4C), but also in terms of
264 trade wind strength (Figs. 4A vs. 4B) and associated mean-sea-level pressure gradients in the
265 tropical Pacific atmosphere (4, 52). This underestimation partly arises because ENSO simulations
266 in most models tend to be too quasi-biennial and not persistent enough (53), impeding the ability
267 of models to generate decadal anomalies through the null hypothesis (52). As a consequence, while
268 all models exhibit some link between ENSO decadal variability and internal TPDV, they strongly
269 underestimate the strength of this relationship (Fig. 4D).

270

271 *Sources of externally-forced TPDV*

272 As mentioned above, external radiative forcing from natural (e.g., volcanic eruptions and solar
273 variability) and anthropogenic (e.g., GHGs, ozone and sulfate aerosols) sources also contribute to
274 TPDV. The resulting external TPDV is directly related to decadal variability in the forcing (e.g.,

275 intermittent volcanic eruptions, slowly-increasing GHGs, varying anthropogenic aerosols
276 emission), with possible contributions from the slow adjustment timescale of the ocean. Here, we
277 examine the expected tropical Pacific responses from both anthropogenic and natural external
278 forcing, which are represented schematically in Figure 3B.

279
280 GHGs such as carbon dioxide are the major source of anthropogenic climate warming and have
281 been increasing steadily over past decades. Despite the spatially uniform nature of well-mixed
282 GHGs, the warming of the ocean surface simulated by climate models exhibits substantial spatial
283 variations (54). Most models project an enhanced warming in the equatorial Pacific (Fig. 3B),
284 giving rise to tropical rainfall changes (54) altering global teleconnection patterns, increasing the
285 frequency of extreme ENSO events (55) and regulating the magnitude of climate sensitivity (56).
286 One study concluded that a GHG-forced enhancement of oceanic stratification leads to increasing
287 Rossby wave speed, which decreases the amplitude and shortens the period of internal TPDV (57),
288 whereas another study using a single model found GHG enhanced the amplitude of internal
289 decadal variability (56).

290
291 Unlike GHGs, anthropogenic tropospheric aerosols display large spatio-temporal variations
292 because of localised emission sources, and act to cool global surface temperature by reflecting
293 sunlight. Models suggest that they induce SST and rainfall changes that are similar in pattern but
294 opposite in sign to those of GHGs, especially in the tropical Pacific (59), hence weakening the
295 GHG-induced warming.

296

297 Large volcanic eruptions can also contribute to external TPDV by injecting aerosols into the
298 stratosphere. This cools the troposphere for a year or more (2) and the ocean for up to a decade -
299 thereby temporarily reducing the rate of global thermosteric sea-level rise (60). While the impact
300 of volcanic eruptions on global temperature is evident, their contribution to external TPDV is less
301 clear (61, 62). Volcanic eruptions have been suggested to (i) influence ENSO (63) and, by
302 inference, TPDV, and (ii) to contribute to cooling the western Pacific warm pool on decadal scales
303 (64). Models tend to simulate enhanced, long-term cooling in the eastern equatorial Pacific, but
304 observations are still too sparse to adequately test these model results (65).

305

306 *Confidence in the attribution of observed changes and future projections*

307

308 Although GHG forcing generally dominates external TPDV at multi-decadal and longer
309 timescales, anthropogenic aerosols and volcanic eruptions may have significantly contributed to
310 regional tropical SST variations over recent decades (61, 62). Relatively small decadal changes in
311 top-of-atmosphere solar irradiance have presumably a smaller influence than GHG, although the
312 11-yr solar cycle, amplified by coupled atmosphere-ocean processes, has been proposed to
313 modulate the Walker Circulation on decadal timescales (66). Timescales involved in internal and
314 external TPDV overlap, which makes them difficult to distinguish from one another, especially
315 when considering the relatively short climate record and potential errors in models. As a result,
316 there are varying degrees of confidence in the attribution of some of the observed trends to either
317 internal or external forcing.

318

319 In the following section, we discuss key aspects of future model projections in the tropical Pacific,
320 comparing them with the observational record.

321

322 • ***Western Pacific warming***

323 The western Pacific exhibits a prominent warming trend since the 1950s (Fig. 1D; Fig. 5A), which
324 dominates the evolution of SST in this region (Fig. 1B). This warming trend is accurately captured
325 by historical simulations (compare Fig. 5A and Fig. 5B) and clearly stands out against the weak
326 background natural variability in this region (Fig. 5C), reflecting the fact that the signal-to-noise
327 ratio for this projected warming is among the highest in the world (12). Coral-based SST estimates
328 indicate that such a warming period is likely unprecedented in the western Pacific region
329 throughout the last 1,250 years (67). The warming trend for air temperatures over land in west
330 Pacific island countries is so large that every year since the early 1990s has been warmer than all
331 years prior to 1970 (68). The resulting increase of the western Pacific warm-pool size has
332 confidently been attributed to GHG forcing arising from human activity (68, 69) although remote
333 influence from the natural multidecadal climate variability in the Atlantic (70) and major volcanic
334 eruptions (64) may also have modulated the SST warming rate there.

335

336 • ***Hadley Cell***

337 While recent observational datasets significantly differ before the 1950's, they consistently report
338 a southward expansion of the southern edge of the Southern Hemisphere Hadley Cell since 1979
339 (Fig. 5D; (71)). Although internal climate variability also contributed, this widening over the last
340 40 years can confidently be attributed to the combined effect of ozone depletion and rising GHGs
341 (Fig. 3B; Fig. 5D; (71)). The mechanism behind this widening is still subject to debate but likely

342 reflects how subtropical atmospheric baroclinic eddies respond to tropospheric (GHGs) and
343 stratospheric (ozone) changes in the atmospheric background state (71). This southward expansion
344 is associated with a lower rate of warming (Fig. 5AB) and ocean acidification (54, 72) in the
345 southeastern tropical Pacific than in the rest of the tropics, probably driven by an intensification of
346 the southeastern Pacific trade winds, which strengthen the Peru-Chile upwelling system near the
347 coast, increase heat loss through air-sea fluxes and modulate the oceanic mixed layer offshore (68).
348 Models also project a widening of the Northern Hemisphere Hadley Cell that is currently not yet
349 detectable in observations due to a larger influence of internal climate variability (71, 73).

350

351 • *The Walker Circulation and equatorial SST gradients*

352 As illustrated in Fig. 5B, E and F, most state-of-the-art models project a weakening of the
353 equatorial trade winds and Walker Circulation and a faster warming rate at the equator, in
354 particular in the eastern equatorial Pacific (55). In agreement with instrumental observations and
355 historical simulations (51, 74) central tropical Pacific corals also point to a wet trend over the 20th
356 century (Fig. 1D), accompanied by even wetter periods during positive phases of the Interdecadal
357 Pacific Oscillation (18). A leading explanation for the Walker Circulation weakening is that
358 rainfall increases less in models than predicted by the Clausius-Clapeyron relation, implying
359 increased atmospheric stability and a reduced mass-flux between the boundary layer and free
360 atmosphere, resulting in a weakened Walker Circulation (75). The enhanced equatorial eastern
361 Pacific warming has been explained by a feedback loop between the weaker evaporative cooling
362 in the cold tongue (54) and reduced trade winds, and a limitation of the SST increase by cloud
363 feedbacks over the West Pacific (55). Recent studies also suggest that the subtropical
364 anthropogenic warming also contributes to the enhanced equatorial warming by slowly making its

365 way to the equatorial thermocline through the oceanic STCs (76, 77). There is, however, no
366 consensus to date on the dominant mechanism responsible for the projected equatorial Pacific SST
367 gradient changes.

368
369 A key uncertainty of external TPDV is that simulated changes do not match recent observed
370 historical trends over, e.g., 1981-2012 (4, 52), which are characterized by a marked strengthening
371 of the Walker Circulation over this period. Such signals are typical of internal TPDV (Fig. 4AB).
372 Indeed, recent studies attribute a large part of this recent observed evolution in the central
373 equatorial Pacific to internal TPDV (4, 52, 78), which is a strong contributor to SST variations in
374 this region (Fig. 1C). This is illustrated by the relatively large model ensemble spread displayed in
375 Figures 5 E and F, which largely encompass the observed SST and surface wind evolution.

376
377 On the other hand, many recent studies suggest plausible mechanisms by which external forcings
378 might also have contributed to the recent strengthening. For example, model results indicate that
379 the reduction in tropospheric sulfate aerosol emissions from North America and Europe and the
380 concurrent increase in China - perhaps augmented by changes driven by volcanic eruptions (62,
381 79) – might have contributed to the recent tropical Pacific cooling (61). Other modeling studies
382 suggest that the observed faster warming in the Indian and/or Atlantic relative to the Pacific Ocean
383 are conducive to enhanced trades in the Pacific and reinforced the recent tropical Pacific cooling
384 (46). Increasing GHGs likely contributed to this observed Indian-Pacific differential warming, but
385 their contribution to the enhanced Atlantic warming is unclear (80). Finally, some models
386 reproduce the observed Walker Circulation strengthening and equatorial cooling (81), with a
387 plausible mechanism related to the poleward export of the added equatorial Pacific heat by the

388 sustained meridional divergence of the near-equatorial upper-ocean currents (82, 83). Model-based
389 studies further suggest that the fast equatorial cooling related to this oceanic thermostat mechanism
390 will be followed by a slower transition to an enhanced equatorial warming and Walker Circulation
391 weakening, in response to subtropical warm anomalies advected into the equatorial thermocline
392 by the STCs (76, 77).

393
394 It is thus unclear from current literature if the recent observed Walker Circulation and cold tongue
395 strengthening is a response to external forcing that is only reproduced by a few models, or if it
396 simply arises from internal variability hiding a subtle opposite secular trend (78). This results in a
397 rather low confidence in the projected weakening of the Walker Circulation and related enhanced
398 equatorial eastern Pacific warming in climate models. Several studies indeed argue that the
399 enhanced equatorial warming in most climate projections may arise from common present-day
400 climate model biases within the tropical Pacific (84) or from an underestimation of interbasin
401 interactions (45, 46, 84). Confidence in these projections is further reduced by the large
402 uncertainties on the impact of aerosols on radiation, cloud microphysics and SST (85). These
403 caveats imply that it is currently not possible to conclude with confidence whether GHG forcing
404 has weakened, strengthened, or had no effect on the Walker Circulation and equatorial upwelling.

405
406 • *Changes in ENSO*

407 Improving the reliability of these projections is key, partly because projected changes in the
408 equatorial zonal SST gradient strongly influence ENSO in climate models (55). The projected
409 warming pattern in the equatorial Pacific in most climate models indeed increases ENSO-driven
410 and decadal precipitation anomalies in part of the tropical Pacific (51, 55, 74), and is tied to an

411 increase in the amplitude of ENSO anomalies (55). Recent paleo-climatic evidence suggests that
412 the increase in ENSO variability since the 1950s stands out in the context of the past millennia
413 (86), lending support to the inter-model agreement on increased ENSO-driven precipitation
414 variability under greenhouse warming. These findings have significant implications, given the
415 large societal impacts of projected changes in ENSO, and the fact that any increase in ENSO-
416 driven precipitation variability (51) or the frequency of extreme ENSO states (55) may energize
417 internal TPDV through the various forms of our null-hypothesis (21, 24, 25).

418

419 **Outlook**

420 Predicting the climate of the tropical Pacific over the next decade and beyond, including
421 precipitation, temperature, sea-level, and biogeochemistry, would have far-reaching societal and
422 environmental benefits. However, because of the partially chaotic nature of the climate system,
423 decadal predictions can, at best, provide an outlook of annual to multi-year average conditions or
424 risks, rather than a more detailed picture of daily or seasonal conditions (2). A decadal prediction
425 would be typically expressed in terms of probabilities, such as the probability that temperature in
426 the tropical Pacific averaged over the next five years will exceed the temperature in the tropical
427 Pacific averaged over the past 30 years. While the changes in average conditions may be small,
428 they can produce marked differences in the probability of extremes (12).

429

430 Experimental prediction systems have been developed (2, 14) to exploit any predictability arising
431 from the mechanisms discussed in the previous sections. Results from an ensemble prediction
432 system suggests that initialisation with observations in much of the tropical Pacific tends to
433 contribute towards predictive skill for surface temperature, for forecast lead times only up to

434 approximately two years (Fig. 6A) and is mostly associated with predictability arising from ENSO
435 (87), though another study concluded that trans-basin climate variability connected with TPDV
436 can be predicted up to three years ahead (46). It might also be that climate models underestimate
437 the degree of skill that actually exists in the real world (88). At longer lead times, skill arises mainly
438 from external forcing (2, 89) (Fig. 6A).

439
440 While predictive skill of decadal average SST is found in most of the tropical Pacific (15, 87) (Fig.
441 6B), it is not evident everywhere. In particular, there is limited skill in the central tropical Pacific
442 north of the equator, extending to the northeast Pacific (Fig. 6B). This is an important region
443 because SST variability there can impact climate in many parts of the world. This low skill may
444 be because the intrinsic predictability of internal variability beyond two years is genuinely low
445 there and any predictable forced response is weak compared with unpredictable internal variability.
446 Alternatively, the combined impact of internal variability and the externally-forced signal may be
447 predictable but the models might miss or misrepresent key mechanisms underpinning the
448 predictability. If this is the case, then the impact of TPDV on ENSO behaviour might also be
449 currently underestimated.

450
451 A significant advancement identified in this review is that skill in decadal predictions of SST in
452 the western Pacific is apparent in the last two generations of dynamical decadal prediction systems
453 (2) (Fig. 6B). While it is likely that this primarily arises from anthropogenic warming, climate
454 models also simulate substantial externally-forced decadal variability in this region about the long-
455 term warming trend (Fig. 5C). This suggests that other types of external forcing have also
456 contributed to TPDV in west Pacific SSTs. Whether this enhances predictability in West Pacific
457 SSTs or not is still unclear.

458

459 There are indications that in the tropical Pacific, multi-year variability in some biogeochemical
460 quantities important for fisheries and the global carbon budget such as net primary production and
461 carbon dioxide uptake can be predicted with greater skill than SST (90, 91). This may be because
462 the biogeochemical quantities are more influenced by subsurface and spatially integrated
463 quantities, which tend to exhibit greater predictability than does SST (21). Limited evidence also
464 suggests that there may be some skill in predicting atmospheric sea-level pressure and sea-surface
465 height (92), changes in the phase of the Interdecadal Pacific Oscillation (e.g.; (93, 94), related
466 precipitation averaged over the Asian-Australian monsoon, Australia more broadly, and western
467 North America (95), and soil moisture – with implications for drought and wildfire – over parts of
468 the southwestern U.S. (6).

469

470 In summary, our review of TPDV predictability finds that although responses to anthropogenic
471 GHG increases offer predictability in some variables (e.g., Fig. 5C; Fig. 6A and B), confidence in
472 the modeled response in the tropical Pacific is generally low; predictability from tropospheric
473 aerosols is still debated; volcanic eruptions likely provide predictability immediately after the
474 eruption has occurred (63); changes in anthropogenic aerosols (e.g., due to industrial growth and
475 pollution aerosols) provide longer-timescale forcing; and TPDV arising from solar forcing likely
476 exists but is small in models compared with other sources of external forcing and unlikely to be a
477 significant source of predictability.

478

479 **The Way Forward**

480 This review has highlighted some important advances in our understanding of the tropical Pacific
481 climate variability and change at decadal and longer timescales. It has also highlighted the
482 complexity of the interactions between variations that occur naturally and those that are forced by
483 external factors of both natural and anthropogenic origin, and the knowledge gaps and
484 uncertainties associated with both components and their interactions. While several plausible
485 mechanisms for both internal and externally-forced TPDV have been proposed, their relative
486 importance and relevance to predictability needs to be further clarified. Specific open science
487 questions include:

- 488 1. How important are oceanic processes involving the STCs in driving predictable decadal
489 climate variations? Do the mechanisms involving STC variability and the associated wind
490 forcing arise independently of ENSO? How large is the predictability associated with these
491 oceanic processes?
- 492 2. How robust are climate model projections in the tropical Pacific and what are the dominant
493 processes driving these changes? In particular, how will the Walker Circulation, equatorial
494 SST and internal variability respond to future greenhouse gas increases?
- 495 3. Why do forecast systems appear to offer predictive skill in the western and southern
496 tropical Pacific, but not in the north-eastern tropical Pacific?

497 Improvements in the quality, quantity, and length of observational records available for
498 characterizing decadal variability are critical to address these science questions, and to initialize
499 and verify decadal prediction systems. This will require sustaining and enhancing the ocean and
500 climate observing systems, data rescue efforts to recover historical observations from data-sparse
501 regions, and the development of new monthly- to annually-resolved paleoclimate records from

502 TPDV centers of action, with a focus on obtaining multiple records in those regions to enhance
503 signal-to-noise ratios. Continued advances in paleoclimate data assimilation (19) will also be
504 critical for the integration of paleoclimate and instrumental observations with models to obtain
505 more complete and reliable fields.

506 While substantial model improvements have been made in recent decades for some features of the
507 climate system (10), models are still limited in their ability to accurately represent observed TPDV
508 and there are large model-to-model differences in the magnitude of simulated TPDV. As noted in
509 the Introduction, there is evidence suggesting that there may be a link between these differences
510 and model-to-model differences in global climate sensitivity. Improving the simulation of TPDV
511 might therefore yield a narrower range in climate sensitivities and greater clarity on our climatic
512 future.

513 Despite their shortcomings, climate models are essential tools for advancing our ability to
514 understand and predict future change in the tropical Pacific. The underlying causes of the
515 shortcomings are still elusive and dedicated efforts using novel approaches are required to identify
516 the major sources of errors in both local and remote feedbacks. Enhanced efforts on the specific
517 role of the STCs in driving TPDV in models may facilitate improved understanding of the
518 mechanisms involving variability of the STCs and their associated wind forcing. In the longer
519 term, improving climate models will be essential for achieving more realistic simulations, as well
520 as more reliable predictions and projections, of TPDV. Advances are expected from improvements
521 in: the representation of subgridscale processes; data assimilation into forecast systems; and
522 computing technology enabling higher spatial resolution, less reliance on parameterizations,
523 longer model runs, and larger ensemble sizes.

525 **References**

526

- 527 1. A. Timmermann, S.-I. An, J.-S. Kug, F.-F. Jin, W. Cai, A. Capotondi, K. Cobb, M.
528 Lengaigne, M. J. McPhaden, M. F. Stuecker, K. Stein, A. T. Wittenberg, K.-S. Yun, T.
529 Bayr, H.-C. Chen, Y. Chikamoto, B. Dewitte, D. Dommenges, P. Grothe, E. Guilyardi, Y.-
530 G. Ham, M. Hayashi, S. Ineson, D. Kang, S. Kim, W. M. Kim, J.-Y. Lee, T. Li, J.-J. Luo,
531 S. McGregor, Y. Planon, S. Power, H. Rashid, H.-L. Ren, A. Santoso, K. Takahashi, A.
532 Todd, G. Wang, G. Wang, R. Xie, W.-H. Yang, S.-W. Yeh, J. Yoon, E. Zeller, X. Zhang,
533 El Niño–Southern Oscillation complexity. *Nature*. **559**, 535–545 (2018).
- 534 2. B. Kirtman, S. B. Power, A. J. Adedoyin, G. J. Boer, R. Bojariu, C. Ines, F. Doblas-Reyes,
535 A. M. Fiore, M. Kimoto, G. Meehl, M. Prather, S. Abdoulaye, C. Schär, R. Sutton, G. J.
536 van Oldenborgh, G. Vecchi, H.-J. Wang, Near-term Climate Change: Projections and
537 Predictability. *Fifth Assess. Rep. IPCC*, 953–1028 (2013).
- 538 3. Y. Kosaka, S. Xie, Recent global-warming hiatus tied to equatorial Pacific surface
539 cooling. *Nature*. **501**, 403–407 (2013).
- 540 4. M. H. England, S. McGregor, P. Spence, G. A. Meehl, A. Timmermann, W. Cai, A. Sen
541 Gupta, M. J. Mcphaden, A. Purich, A. Santoso, Recent intensification of wind-driven
542 circulation in the Pacific and the ongoing warming hiatus. **4**, 1–6 (2014).
- 543 5. X. Zhang, J. A. Church, Sea level trends, interannual and decadal variability in the Pacific
544 Ocean. *Geophys. Res. Lett.* **39**, 1–8 (2012).
- 545 6. Y. Chikamoto, A. Timmermann, M. J. Widlansky, M. A. Balmaseda, L. Stott, Multi-year
546 predictability of climate, drought, and wildfire in southwestern North America. *Sci. Rep.*
547 **7**, 6568 (2017).
- 548 7. G. A. Meehl, J. M. Arblaster, C. M. Bitz, C. T. Y. Chung, H. Teng, Antarctic sea-ice
549 expansion between 2000 and 2014 driven by tropical Pacific decadal climate variability.
550 *Nat. Geosci.* **9**, 590–595 (2016).
- 551 8. S. Power, T. Casey, C. Folland, A. Colman, V. Mehta, Inter-decadal modulation of the
552 impact of ENSO on Australia. *Clim. Dyn.* **15**, 319–324 (1999).
- 553 9. R. Colman, S. B. Power, What can decadal variability tell us about climate feedbacks and
554 sensitivity? *Clim. Dyn.* **51**, 3815–3828 (2018).
- 555 10. IPCC, *Climate Change 2013: The Physical Science Basis. Contribution of Working Group
556 I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*
557 (Cambridge University Press, Cambridge, 2013).
- 558 11. L. Wang, G. Huang, W. Chen, W. Zhou, W. Wang, Wet-to-dry shift over Southwest
559 China in 1994 tied to the warming of tropical warm pool. *Clim. Dyn.* **51**, 3111–3123
560 (2018).
- 561 12. S. B. Power, F. P. D. Delage, Setting and smashing extreme temperature records over the
562 coming century. *Nat. Clim. Chang.* **9**, 529–534 (2019).
- 563 13. J. E. Johnson, V. Allain, B. Basel, J. D. Bell, A. Chin, L. X. C. Dutra, E. Hooper, D.
564 Loubser, J. Lough, B. R. Moore, S. Nicol, in *Springer Climate* (2020), pp. 359–402.
- 565 14. Y. Kushnir, A. A. Scaife, R. Arritt, G. Balsamo, G. Boer, F. Doblas-Reyes, E. Hawkins,
566 M. Kimoto, R. K. R. K. Kolli, A. Kumar, D. Matei, K. Matthes, W. A. W. A. Müller, T.
567 O’Kane, J. Perlwitz, S. Power, M. Raphael, A. Shimpou, D. Smith, M. Tuma, B. B. Wu,
568 Towards operational predictions of the near-term climate. *Nat. Clim. Chang.* **9**, 94–101

- (2019).
- 569
570 15. G. J. Boer, G. J. Kharin, W. J. Merryfield, Differences in potential and actual skill in a
571 decadal prediction experiment. *Clim. Dyn.*, 868–874 (2019).
- 572 16. J. E. Tierney, N. J. Abram, K. J. Anchukaitis, M. N. Evans, C. Giry, K. H. Kilbourne, C.
573 P. Saenger, H. C. Wu, J. Zinke, Tropical Corals 400 yrs reconstructed from coral archives.
574 *Paleoceanography*. **30**, 226–252 (2015).
- 575 17. E. P. Dassié, A. Hasson, M. Khodri, N. Lebas, B. K. Linsley, Spatiotemporal Variability
576 of the South Pacific Convergence Zone Fresh Pool Eastern Front from Coral-Derived
577 Surface Salinity Data. *J. Clim.* **31**, 3265–3288 (2017).
- 578 18. S. C. Sanchez, N. Westphal, G. H. Haug, H. Cheng, R. L. Edwards, T. Schneider, K. M.
579 Cobb, C. D. Charles, A Continuous Record of Central Tropical Pacific Climate Since the
580 Midnineteenth Century Reconstructed From Fanning and Palmyra Island Corals: A Case
581 Study in Coral Data Reanalysis. *Paleoceanogr. Paleoclimatology*. **35**, 1–15 (2020).
- 582 19. R. Tardif, G. J. Hakim, W. A. Perkins, K. A. Horlick, M. P. Erb, J. Emile-Geay, D. M.
583 Anderson, E. J. Steig, D. Noone, Last Millennium Reanalysis with an expanded proxy
584 database and seasonal proxy modeling. *Clim. Past*. **15**, 1251–1273 (2019).
- 585 20. Y. Zhang, J. M. Wallace, D. S. Battisti, ENSO-like interdecadal variability: 1900-93. *J.*
586 *Clim.* **10**, 1004–1020 (1997).
- 587 21. S. Power, R. Colman, Multi-year predictability in a coupled general circulation model.
588 *Clim. Dyn.* **26**, 247–272 (2006).
- 589 22. M. Newman, M. A. Alexander, T. R. Ault, K. M. Cobb, C. Deser, E. Di Lorenzo, N. J.
590 Mantua, A. J. Miller, S. Minobe, H. Nakamura, N. Schneider, D. J. Vimont, A. S. Phillips,
591 J. D. Scott, C. A. Smith, The Pacific decadal oscillation, revisited. *J. Clim.* **29**, 4399–4427
592 (2016).
- 593 23. T. R. Ault, C. Deser, M. Newman, J. Emile-Geay, Characterizing decadal to centennial
594 variability in the equatorial Pacific during the last millennium. *Geophys. Res. Lett.* **40**,
595 3450–3456 (2013).
- 596 24. D. J. Vimont, The Contribution of the Interannual ENSO Cycle to the Spatial Pattern of
597 Decadal ENSO-Like Variability. *J. Clim.* **18**, 2080–2092 (2005).
- 598 25. G. Il Kim, J. S. Kug, Tropical pacific decadal variability induced by nonlinear rectification
599 of El Niño–Southern Oscillation. *J. Clim.* **33**, 7289–7302 (2020).
- 600 26. A. Capotondi, A. T. Wittenberg, J. Kug, K. Takahashi, M. J. McPhaden, in *El Niño*
601 *Southern Oscillation in a Changing Climate*, M. J. McPhaden, A. Santoso, W. Cai, Eds.
602 (John Wiley & Sons, Inc, First., 2021), pp. 65–86.
- 603 27. M. Newman, S.-I. Shin, M. A. Alexander, Natural variation in ENSO flavors. *Geophys.*
604 *Res. Lett.* **38** (2011), doi:10.1029/2011GL047658.
- 605 28. T. Sun, Y. M. Okumura, Impact of ENSO-Like Tropical Pacific Decadal Variability on
606 the Relative Frequency of El Niño and La Niña Events. *Geophys. Res. Lett.* **47**, 1–10
607 (2020).
- 608 29. M. J. McPhaden, D. Zhang, Slowdown of the meridional overturning circulation in the
609 upper Pacific Ocean. *Nature*. **415**, 603–608 (2002).
- 610 30. A. Capotondi, M. A. Alexander, C. Deser, M. J. McPhaden, Anatomy and Decadal
611 Evolution of the Pacific Subtropical–Tropical Cells (STCs). *J. Clim.* **18**, 3739–3758
612 (2005).
- 613 31. G. Graffino, R. Farneti, F. Kucharski, F. Molteni, The effect of wind stress anomalies and
614 location in driving Pacific subtropical cells and tropical climate. *J. Clim.* **32**, 1641–1660

- 615 (2019).
- 616 32. G. A. Meehl, A. Hu, Megadroughts in the Indian Monsoon Region and Southwest North
617 America and a Mechanism for Associated Multidecadal Pacific Sea Surface Temperature
618 Anomalies. *J. Clim.* **19**, 1605–1623 (2006).
- 619 33. A. Capotondi, M. A. Alexander, Rossby waves in the tropical North Pacific and their role
620 in decadal thermocline variability. *J. Phys. Oceanogr.* **31**, 3496–3515 (2001).
- 621 34. A. Capotondi, M. A. Alexander, C. Deser, Why Are There Rossby Wave Maxima in the
622 Pacific at 10°S and 13°N? *J. Phys. Oceanogr.* **33**, 1549–1563 (2003).
- 623 35. J. Luo, S. Masson, S. Behera, P. Delecluse, S. Gualdi, A. Navarra, T. Yamagata, South
624 Pacific origin of the decadal ENSO-like variation as simulated by a coupled GCM. **30**, 4–
625 7 (2003).
- 626 36. M. Zeller, S. McGregor, E. van Sebille, A. Capotondi, P. Spence, Subtropical-tropical
627 pathways of spiciness anomalies and their impact on equatorial Pacific temperature. *Clim.*
628 *Dyn.* (2020), doi:10.1007/s00382-020-05524-8.
- 629 37. D. J. Vimont, D. S. Battisti, A. C. Hirst, Footprinting: A seasonal connection between the
630 tropics and mid-latitudes. *Geophys. Res. Lett.* **28**, 3923–3926 (2001).
- 631 38. J. C. H. Chiang, D. J. Vimont, Analogous Pacific and Atlantic Meridional Modes of
632 Tropical Atmosphere–Ocean Variability. *J. Clim.* **17**, 4143–4158 (2004).
- 633 39. H. Zhang, A. Clement, P. Di Nezio, The south pacific meridional mode: A mechanism for
634 ENSO-like variability. *J. Clim.* **27**, 769–783 (2014).
- 635 40. E. Di Lorenzo, G. Liguori, N. Schneider, J. C. Furtado, B. T. Anderson, M. A. Alexander,
636 ENSO and meridional modes: A null hypothesis for Pacific climate variability. *Geophys.*
637 *Res. Lett.* **42**, 9440–9448 (2015).
- 638 41. Y. Zhao, E. Di Lorenzo, The impacts of Extra-tropical ENSO Precursors on Tropical
639 Pacific Decadal-scale Variability. *Sci. Rep.* **10**, 1–12 (2020).
- 640 42. G. Liguori, E. Di Lorenzo, Separating the North and South Pacific Meridional Modes
641 Contributions to ENSO and Tropical Decadal Variability. *Geophys. Res. Lett.* (2019),
642 doi:10.1029/2018GL080320.
- 643 43. Y. M. Okumura, Origins of tropical pacific decadal variability: Role of stochastic
644 atmospheric forcing from the South Pacific. *J. Clim.* **26**, 9791–9796 (2013).
- 645 44. C. T. Y. Chung, S. B. Power, A. Sullivan, F. Delage, *Sci. Rep.*, in press,
646 doi:10.1038/s41598-019-52805-2.
- 647 45. S. McGregor, M. F. Stuecker, J. B. Kajtar, M. H. England, M. Collins, Model tropical
648 Atlantic biases underpin diminished Pacific decadal variability. *Nat. Clim. Chang.* **8**, 493–
649 498 (2018).
- 650 46. W. Cai, L. Wu, M. Lengaigne, T. Li, S. McGregor, J.-S. Kug, J.-Y. Yu, M. F. Stuecker, A.
651 Santoso, X. Li, Y.-G. Ham, Y. Chikamoto, B. Ng, M. J. McPhaden, Y. Du, D.
652 Dommenges, F. Jia, J. B. Kajtar, N. Keenlyside, X. Lin, J.-J. Luo, M. Martín-Rey, Y.
653 Ruprich-Robert, G. Wang, S.-P. Xie, Y. Yang, S. M. Kang, J.-Y. Choi, B. Gan, G.-I. Kim,
654 C.-E. Kim, S. Kim, J.-H. Kim, P. Chang, Pantropical climate interactions. *Science (80-.)*.
655 **363**, eaav4236 (2019).
- 656 47. R. Farneti, F. Molteni, F. Kucharski, Pacific interdecadal variability driven by tropical-
657 extratropical interactions. *Clim. Dyn.* **42**, 3337–3355 (2014).
- 658 48. Y. Peng, C. Shen, H. Cheng, Y. Xu, Simulation of the Interdecadal Pacific Oscillation and
659 its impacts on the climate over eastern China during the last millennium. *J. Geophys. Res.*
660 *Atmos.* **120**, 7573–7585 (2015).

- 661 49. K. Lyu, X. Zhang, J. A. Church, J. Hu, Evaluation of the interdecadal variability of sea
662 surface temperature and sea level in the Pacific in CMIP3 and CMIP5 models. *Int. J.*
663 *Climatol.* **36**, 3723–3740 (2016).
- 664 50. B. J. Henley, G. Meehl, S. B. Power, C. K. Folland, A. D. King, J. N. Brown, D. J.
665 Karoly, F. Delage, A. J. E. Gallant, M. Freund, R. Neukom, Spatial and temporal
666 agreement in climate model simulations of the Interdecadal Pacific Oscillation. *Environ.*
667 *Res. Lett.* **12**, 44011 (2017).
- 668 51. S. Power, F. Delage, C. Chung, G. Kociuba, K. Keay, Robust twenty-first-century
669 projections of El Niño and related precipitation variability. *Nature.* **502**, 541–5 (2013).
- 670 52. G. Kociuba, S. B. Power, Inability of CMIP5 models to simulate recent strengthening of
671 the walker circulation: Implications for projections. *J. Clim.* **28**, 20–35 (2015).
- 672 53. H. Bellenger, E. Guilyardi, J. Leloup, M. Lengaigne, J. Vialard, ENSO representation in
673 climate models : from CMIP3 to CMIP5. *Clim. Dyn.* **42**, 1999–2018 (2014).
- 674 54. S.-P. Xie, C. Deser, G. A. Vecchi, J. Ma, H. Teng, A. T. Wittenberg, Global Warming
675 Pattern Formation: Sea Surface Temperature and Rainfall. *J. Clim.* **23**, 966–986 (2010).
- 676 55. W. Cai, A. Santoso, G. Wang, L. Wu, M. Collins, M. Lengaigne, S. Power, A.
677 Timmermann, in *El Niño Southern Oscillation in a Changing Climate*, M. J. McPhaden,
678 A. Santoso, W. Cai, Eds. (John Wiley & Sons, Inc., First Edit., 2021;
679 <https://doi.org/10.1002/9781119548164.ch13>), pp. 289–307.
- 680 56. T. Andrews, J. M. Gregory, M. J. Webb, The dependence of radiative forcing and
681 feedback on evolving patterns of surface temperature change in climate models. *J. Clim.*
682 **28**, 1630–1648 (2015).
- 683 57. G. Liguori, E. Di Lorenzo, Meridional Modes and Increasing Pacific Decadal Variability
684 Under Anthropogenic Forcing. *Geophys. Res. Lett.* **45**, 983–991 (2018).
- 685 58. S. Li, L. Wu, Y. Yang, T. Geng, W. Cai, B. Gan, Z. Chen, Z. Jing, G. Wang, X. Ma, The
686 Pacific Decadal Oscillation less predictable under greenhouse warming. *Nat. Clim. Chang.*
687 **10**, 30–34 (2020).
- 688 59. S.-P. Xie, B. Lu, B. Xiang, Similar spatial patterns of climate responses to aerosol and
689 greenhouse gas changes. *Nat. Geosci.* **6**, 828 (2013).
- 690 60. M. A. Balmaseda, K. E. Trenberth, E. Källén, Distinctive climate signals in reanalysis of
691 global ocean heat content. *Geophys. Res. Lett.* **40**, 1754–1759 (2013).
- 692 61. D. M. Smith, B. B. Booth, N. J. Dunstone, R. Eade, L. Hermanson, G. S. Jones, A. A.
693 Scaife, K. L. Sheen, V. Thompson, Role of volcanic and anthropogenic aerosols in the
694 recent global surface warming slowdown. *Nat. Clim. Chang.* **6**, 936 (2016).
- 695 62. W. Hua, A. Dai, M. Qin, Contributions of Internal Variability and External Forcing to the
696 Recent Pacific Decadal Variations. *Geophys. Res. Lett.* **45**, 7084–7092 (2018).
- 697 63. S. McGregor, M. Khodri, N. Maher, M. Ohba, F. S. R. Pausata, S. Stevenson, in *El Nino*
698 *Southern Oscillation in a changing climate*, M. McPhaden, A. Santoso, W. Cai, Eds.
699 (John Wile & Sons, Inc., First Edit., 2021), vol. 253, pp. 267–287.
- 700 64. V. M. Mehta, H. Wang, K. Mendoza, Simulations of three natural decadal climate
701 variability phenomena in CMIP5 experiments with the UKMO HadCM3, GFDL-CM2.1,
702 NCAR-CCSM4, and MIROC5 global earth system models. *Clim. Dyn.* **51**, 1559–1584
703 (2018).
- 704 65. N. Maher, A. Sen Gupta, M. H. England, Drivers of decadal hiatus periods in the 20th and
705 21st centuries. *Geophys. Res. Lett.* **41**, 5978–5986 (2014).
- 706 66. G. A. Meehl, J. M. Arblaster, K. Matthes, F. Sassi, H. van Loon, Amplifying the Pacific

- 707 Climate System Response to a Small 11-Year Solar Cycle Forcing. *Science* (80-.). **325**
708 (2009), doi:10.1126/science.1172872.
- 709 67. T. Chen, K. M. Cobb, G. Roff, J. Zhao, H. Yang, M. Hu, K. Zhao, Coral-Derived Western
710 Pacific Tropical Sea Surface Temperatures During the Last Millennium. *Geophys. Res.*
711 *Lett.* **45**, 3542–3549 (2018).
- 712 68. G. Wang, S. B. Power, S. Mcgree, Unambiguous warming in the western tropical Pacific
713 primarily caused by anthropogenic forcing. *Int. J. Climatol.* **36**, 933–944 (2016).
- 714 69. E. Weller, S. K. Min, W. Cai, F. W. Zwiers, Y. H. Kim, D. Lee, Human-caused Indo-
715 Pacific warm pool expansion. *Sci. Adv.* **2**, 1–8 (2016).
- 716 70. C. Sun, F. Kucharski, J. Li, F. F. Jin, I. S. Kang, R. Ding, Western tropical Pacific
717 multidecadal variability forced by the Atlantic multidecadal oscillation. *Nat. Commun.* **8**,
718 1–10 (2017).
- 719 71. K. M. Grise, S. M. Davis, I. R. Simpson, D. W. Waugh, Q. Fu, R. J. Allen, K. H.
720 Rosenlof, C. C. Ummenhofer, K. B. Karnauskas, A. C. Maycock, X. W. Quan, T. Birner,
721 P. W. Staten, Recent tropical expansion: Natural variability or forced response? *J. Clim.*
722 **32**, 1551–1571 (2019).
- 723 72. G. Beaugrand, A. Conversi, A. Atkinson, J. Cloern, S. Chiba, S. Fonda-Umani, R. R.
724 Kirby, C. H. Greene, E. Goberville, S. A. Otto, P. C. Reid, L. Stemann, M. Edwards,
725 Prediction of unprecedented biological shifts in the global ocean. *Nat. Clim. Chang.* **9**,
726 237–243 (2019).
- 727 73. B. Jebri, M. Khodri, V. Echevin, G. Gastineau, S. Thiria, J. Vialard, N. Lebas,
728 Contributions of internal variability and external forcing to the recent trends in the
729 southeastern pacific and peru-chile upwelling system. *J. Clim.* **33**, 10555–10578 (2020).
- 730 74. A. G. Pendergrass, R. Knutti, F. Lehner, C. Deser, B. M. Sanderson, Precipitation
731 variability increases in a warmer climate. *Sci. Rep.* **7**, 1–9 (2017).
- 732 75. G. A. Vecchi, B. J. Soden, A. T. Wittenberg, I. M. Held, A. Leetmaa, M. J. Harrison,
733 Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. **441**
734 (2006), doi:10.1038/nature04744.
- 735 76. U. K. Heede, A. V. Fedorov, N. J. Burls, Time Scales and Mechanisms for the Tropical
736 Pacific Response to Global Warming: A Tug of War between the Ocean Thermostat and
737 Weaker Walker. *J. Clim.* **33**, 6101–6118 (2020).
- 738 77. M. F. Stuecker, A. Timmermann, F. F. Jin, C. Proistosescu, S. M. Kang, D. Kim, K. S.
739 Yun, E. S. Chung, J. E. Chu, C. M. Bitz, K. C. Armour, M. Hayashi, Strong remote
740 control of future equatorial warming by off-equatorial forcing. *Nat. Clim. Chang.* **10**, 124–
741 129 (2020).
- 742 78. E. S. Chung, A. Timmermann, B. J. Soden, K. J. Ha, L. Shi, V. O. John, Reconciling
743 opposing Walker circulation trends in observations and model projections. *Nat. Clim.*
744 *Chang.* **9**, 405–412 (2019).
- 745 79. C. Takahashi, M. Watanabe, Pacific trade winds accelerated by aerosol forcing over the
746 past two decades. *Nat. Clim. Chang.* **6**, 768–772 (2016).
- 747 80. L. Dong, M. J. McPhaden, Why has the relationship between Indian and Pacific Ocean
748 decadal variability changed in recent decades? *J. Clim.* **30**, 1971–1983 (2017).
- 749 81. T. Kohyama, D. L. Hartmann, D. S. Battisti, La Niña-like mean-state response to global
750 warming and potential oceanic roles. *J. Clim.* **30**, 4207–4225 (2017).
- 751 82. A. C. Clement, R. Seager, M. A. Cane, S. E. Zebiak, An ocean dynamical thermostat. *J.*
752 *Clim.* **9**, 2190–2196 (1996).

- 753 83. R. Seager, M. Cane, N. Henderson, D. E. Lee, R. Abernathey, H. Zhang, Strengthening
754 tropical Pacific zonal sea surface temperature gradient consistent with rising greenhouse
755 gases. *Nat. Clim. Chang.* **9**, 517–522 (2019).
- 756 84. J. J. Luo, G. Wang, D. Dommenges, May common model biases reduce CMIP5’s ability
757 to simulate the recent Pacific La Niña-like cooling? *Clim. Dyn.* **50**, 1335–1351 (2018).
- 758 85. N. Bellouin, J. Quaas, E. Gryspeerdt, S. Kinne, P. Stier, D. Watson-Parris, O. Boucher, K.
759 S. Carslaw, M. Christensen, A. L. Daniau, J. L. Dufresne, G. Feingold, S. Fiedler, P.
760 Forster, A. Gettelman, J. M. Haywood, U. Lohmann, F. Malavelle, T. Mauritsen, D. T.
761 McCoy, G. Myhre, J. Mülmenstädt, D. Neubauer, A. Possner, M. Rugenstein, Y. Sato, M.
762 Schulz, S. E. Schwartz, O. Sourdeval, T. Storelvmo, V. Toll, D. Winker, B. Stevens,
763 Bounding Global Aerosol Radiative Forcing of Climate Change. *Rev. Geophys.* **58**, 1–45
764 (2020).
- 765 86. J. Emile-Geay, K. M. Cobb, J. E. Cole, M. Elliot, F. Zhu, in *El Niño Southern Oscillation*
766 *in a Changing Climate*, M. J. McPhaden, A. Santoso, W. Cai, Eds. (John Wiley and Sons,
767 Inc., First., 2021), vol. 253, pp. 87–118.
- 768 87. P. N. DiNezio, C. Deser, A. Karspeck, S. Yeager, Y. Okumura, G. Danabasoglu, N.
769 Rosenbloom, J. Caron, G. A. Meehl, P. N. Di Nezio, C. Deser, A. Karspeck, S. Yeager, Y.
770 Okumura, G. Danabasoglu, N. Rosenbloom, J. Caron, G. A. Meehl, *Geophys. Res. Lett.*,
771 in press, doi:10.1002/2017GL074904.
- 772 88. A. A. Scaife, D. Smith, A signal-to-noise paradox in climate science. *npj Clim. Atmos. Sci.*
773 **1** (2018), doi:10.1038/s41612-018-0038-4.
- 774 89. D. M. Smith, R. Eade, A. A. Scaife, L.-P. Caron, G. Danabasoglu, T. M. DelSole, T.
775 Delworth, F. J. Doblas-Reyes, N. J. Dunstone, L. Hermanson, V. Kharin, M. Kimoto, W.
776 J. Merryfield, T. Mochizuki, W. A. Müller, H. Pohlmann, S. Yeager, X. Yang, Robust
777 skill of decadal climate predictions. *npj Clim. Atmos. Sci.* **2**, 1–10 (2019).
- 778 90. R. Sférian, L. Bopp, M. Gehlen, D. Swingedouw, J. Mignot, E. Guilyardi, J. Servonnat,
779 *Proc. Natl. Acad. Sci.*, in press, doi:10.1073/pnas.1315855111.
- 780 91. N. S. Lovenduski, S. G. Yeager, K. Lindsay, M. C. Long, Predicting near-term variability
781 in ocean carbon uptake. *Earth Syst. Dynam.* **10**, 45–57 (2019).
- 782 92. Y. Chikamoto, T. Mochizuki, A. Timmermann, M. Kimoto, M. Watanabe, Potential
783 tropical Atlantic impacts on Pacific decadal climate trends. *Geophys. Res. Lett.* **43**, 7143–
784 7151 (2016).
- 785 93. G. A. Meehl, H. Teng, CMIP5 multi-model hindcasts for the mid-1970s shift and early
786 2000s hiatus and predictions for 2016–2035. *Geophys. Res. Lett.* **41**, 1711–1716 (2014).
- 787 94. M. Thoma, R. J. Greatbatch, C. Kadow, R. Gerdes, Decadal hindcasts initialized using
788 observed surface wind stress: Evaluation and prediction out to 2024. *Geophys. Res. Lett.*
789 **42**, 6454–6461 (2015).
- 790 95. G. A. Meehl, A. Hu, H. Teng, Initialized decadal prediction for transition to positive phase
791 of the Interdecadal Pacific Oscillation. *Nat. Commun.* **7**, 1–7 (2016).
- 792 96. N. A. Rayner, D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell,
793 E. C. Kent, A. Kaplan, Global analyses of SST, sea ice, and night marine air temperature
794 since the late nineteenth century. *J. Geophys. Res.* **108**, 4407 (2003).
- 795 97. M. A. Balmaseda, K. Mogensen, A. T. Weaver, Evaluation of the ECMWF ocean
796 reanalysis system ORAS4. *Q. J. R. Meteorol. Soc.* **139**, 1132–1161 (2013).
- 797 98. E. de Boissésou, M. A. Balmaseda, M. Mayer, Ocean heat content variability in an
798 ensemble of twentieth century ocean reanalyses. *Clim. Dyn.* **50**, 3783–3798 (2018).

- 799 99. G. P. Compo, J. S. Whitaker, P. D. Sardeshmukh, N. Matsui, R. J. Allan, X. Yin, B. E.
800 Gleason, R. S. Vose, G. Rutledge, P. Bessemoulin, S. Brönnimann, M. Brunet, R. I.
801 Crouthamel, A. N. Grant, P. Y. Groisman, P. D. Jones, M. C. Kruk, A. C. Kruger, G. J.
802 Marshall, M. Maugeri, H. Y. Mok, Ø. Nordli, T. F. Ross, R. M. Trigo, X. L. Wang, S. D.
803 Woodruff, S. J. Worley, The Twentieth Century Reanalysis Project. *Q. J. R. Meteorol.*
804 *Soc.* **137**, 1–28 (2011).
- 805 100. P. Poli, H. Hersbach, D. P. Dee, P. Berrisford, A. J. Simmons, F. Vitart, P. Laloyaux, D.
806 G. H. Tan, C. Peubey, J.-N. Thépaut, Y. Trémolet, E. V Hólm, M. Bonavita, L. Isaksen,
807 M. Fisher, ERA-20C: An Atmospheric Reanalysis of the Twentieth Century. *J. Clim.* **29**,
808 4083–4097 (2016).

809

810 ACKNOWLEDGEMENTS

811 We wish to thank Jing Li at WCRP for helping to organise workshops to help advance this
812 review, the first at Centro Nacional de Acuicultura e Investigaciones Marinas (CENAIM) and
813 Escuela Superior Politecnica del Litoral (ESPOL) in Ecuador, the second at LOCEAN-IPSL,
814 Sorbonne University in Paris, both Viatcheslav Kharin and George Boer for providing the data
815 used in Fig. 5A, Giovanni Liguoro and Julien Cretat, for reviewing earlier drafts, Pascale
816 Bracannot for discussions regarding paleoclimate, Roland Sférian for discussions regarding
817 biogeochemistry, and Lea Crosswell for finalizing the Summary figure. Thanks also to reviewers
818 for their very helpful and constructive comments. **Funding:** S.P., W.C., F.D., and C.C. were
819 partially supported by the Earth System and Climate Change Hub of the Australian National
820 Environmental Science Programme. W.C., W.G. and X.Z. were supported by CSHOR, a joint
821 research Centre for Southern Hemisphere Oceans Research between QNLM and CSIRO. M.L.,
822 J.V. and E.G. were supported by the Agence Nationale de la Recherche ARISE project, under
823 grant ANR-18-CE01-0012, and the Belmont project GOTHAM, under grant ANR-15-JCLI-
824 0004-01. A.C. was supported by the NOAA Climate Program Office’s Climate Variability and
825 Predictability (CVP) and Modeling, Analysis, Predictions and Projections (MAPP) Programs.
826 M.C. was supported by NE/N018486/1 and NE/N005783/1. This is PMEL contribution no. 5031.
827 G.M. was partially supported by the Regional and Global Model Analysis (RGMA) component

828 of the Earth and Environmental System Modeling Program of the U.S. Department of Energy's
829 Office of Biological & Environmental Research via National Science Foundation IA 1844590,
830 and by NCAR, which is a major facility sponsored by the National Science Foundation under
831 Cooperative Agreement No. 1852977. S.McG. was supported by the Australian Research
832 Council through grant FT160100162 and DP20. Y.O. was supported by the NOAA Climate
833 Program Office's Modeling, Analysis, Predictions, and Projections Program
834 (NA17OAR4310149) and the NSF Physical Oceanography Program (OCE-1756883). B.H. was
835 supported by an Australian Research Council Linkage Project (LP150100062). M.N. was
836 partially supported by US Department of Energy Grant #0000238382. J.J.Ll was supported by
837 National Natural Science Foundation of China (Grants 42088101 and 42030605). D.S. was
838 supported by the Met Office Hadley Centre Climate Programme funded by BEIS and DEFRA.

839 **Author contributions:** The manuscript was drafted as a group effort during two specially
840 convened workshops attended by nearly all co-authors. All authors contributed to the manuscript
841 preparation, interpretations, and the discussions that led to the final draft. S.P., A.C. and M.L. led
842 the discussions and coordinated the analysis and writing, with M.L. leading finalisation of
843 Figures 1-5. S.P. drafted Summary Figure 1. S.P., M.L., A. C., M.K. J.V., S.McG. and G.M.
844 coordinated discussions and writing of subsections; J.-S.K., G.-I.K., M. K., D.V., J. B., J. E.-G.,
845 S. McG., C.C., and F.D provided analyses and/or helped with figures. **Competing interests:** The
846 authors declare no competing interests. **Data and materials availability:** All observational and
847 model datasets used here are publicly available or available on request.

848

849

850

851 **Figure captions**

852

853 **Fig. 1. Observed decadal variability in the tropical Pacific.** (A) 1992-2009 linear trend of
854 annual SST (shading, $0.1^{\circ}\text{C decade}^{-1}$) and sea surface height (black contours, 2cm decade^{-1} , dashed
855 contours indicate negative trends, solid contours denote positive trends, and the zero contour is
856 omitted). (B) 8 yr lowpass and (C) 8-40 yr bandpass filtered SST variance (black contours, in 10^2
857 $^{\circ}\text{C}^2$) and ratio of the filtered SST variance to total SST variance (shading). (D) 8 yr lowpass filtered
858 timeseries of SST averaged over the Niño34 region ($5^{\circ}\text{N}-5^{\circ}\text{S}$; $170^{\circ}\text{W}-120^{\circ}\text{W}$), the western
859 tropical Pacific ($10^{\circ}\text{N}-10^{\circ}\text{S}$; $120^{\circ}\text{E}-150^{\circ}\text{E}$) and over the globe from instrumental observations
860 (black lines) and Last Millennium Reconstruction ((19); mean: grey line; interquartile range: light
861 grey shading) and of $\delta^{18}\text{O}$ at Palmyra and Fiji islands (plain and dashed blue lines; positions
862 indicated in (A); (17, 18)). Vertical red and blue bands indicate positive and negative phases of the
863 Interdecadal Pacific Oscillation. SST data: HadISST (96). SSH data: ORAS4 dataset (97).

864

865 **Fig. 2. Internal TPDV - the null hypothesis.** (A) Pacific SST pattern associated with internal
866 TPDV, obtained by regressing the 8-40 year band-pass filtered SST anomalies onto the internal
867 TPDV index. The latter is obtained as the time series (or Principal Component) of the leading EOF
868 of SST anomalies in the 8-40 year band, over the tropical Pacific ($24^{\circ}\text{S}-24^{\circ}\text{N}$; $120^{\circ}\text{E}-80^{\circ}\text{W}$).
869 (B) Timeseries of SST anomalies averaged in the Niño34 region ($5^{\circ}\text{S}-5^{\circ}\text{N}$, $170^{\circ}\text{W}-120^{\circ}\text{W}$; N3.4),
870 a commonly used SST ENSO index; the Southern Oscillation Index (SOI; (18)), a measure of the
871 Walker Circulation strength; the internal TPDV index, and the E-L index, defined as the number

872 of El Niño years minus the number of La Niña years over 8-year running periods. ENSO events
873 are identified using the December Niño3.4 index and an amplitude threshold of 1 standard
874 deviation. Thick black lines in (B) indicate the 8-40year band-pass filtered time series. (C) Average
875 of ENSO-related SST anomalies over the year preceding the peak of an El Niño event (year 0) and
876 the year following the El Niño event (year 1), defined by computing lagged regressions of SST
877 onto the November-December-January averaged Niño3.4 index from lags of -11mo to +12mo, and
878 averaging over all 24 resulting maps. (D), (E), and (F) show individual SST maps from these
879 monthly regressions, illustrating precursor anomalies during the February-March-April (FMA, D)
880 prior to the peak of an event, peak anomalies during October-November-December (OND, E) of
881 the ENSO event, and anomalies during the decay phase in June-July-August (JJA, F) of the years
882 following the peak of an ENSO event. The SST data are from HadISST (96) over the period 1900-
883 2020. Filtering was performed using 5 and 53 point Hanning filter weights.

884

885 **Fig. 3. Mechanisms of internal and external TPDV.** (A) Schematic representation of the ocean
886 processes associated with internal TPDV. The climatological upper ocean overturning circulation
887 (the Subtropical-Tropical Cells, transparent blue arrows) consists of a subtropical subduction
888 component, equatorward subsurface transport, equatorial upwelling, and a poleward surface return
889 flow driven by the equatorial easterly trade winds (large blue arrow), which are the surface
890 component of the Walker Circulation. A positive phase of internal TPDV with warm SST in the
891 tropical Pacific (shading) is associated with a weaker Walker Circulation, reduced equatorial
892 winds, and weaker oceanic overturning circulation. Extra-equatorial wind anomalies may play an
893 important role in driving the changes in the Subtropical-Tropical Cells, whose adjustment is

894 accomplished through the westward propagation of oceanic Rossby waves. After reaching the
895 western boundary, Rossby waves can continue along the boundary to the equator as coastal Kelvin
896 waves and along the equator as equatorial Kelvin waves. The extra-equatorial wind anomalies may
897 be purely stochastic, arise from extra-tropical influences, or as a response to equatorial SST
898 anomalies (see text for details). **(B)** Schematic representation of projected changes associated with
899 external TPDV. The map shows the late 21st century multi-model-mean change in CMIP6 SST,
900 which is dominated by increases in greenhouse gases. High (low) confidence in these projected
901 changes is indicated by solid (dashed) lines. Icons indicate the major external forcings involved in
902 these changes. Greenhouse gas increases and ozone changes induce a robust southward expansion
903 of the Hadley Cell in the southern hemisphere and reduced southern subtropical Pacific warming,
904 in both model projections and observations. The prominent western Pacific warming and the
905 central Pacific rainfall increase detected in models and observations can confidently be attributed
906 greenhouse gas increases. While the projected weakening and enhanced tropical warming is
907 evident in most CMIP6 models, confidence in these projections is low because of inconsistent
908 signals in observations, model biases and the complexity of the mechanisms involved. Volcanic
909 eruptions and changes in solar insolation may also cause decadal variations in the tropical Pacific,
910 though their amplitude is likely small.

911

912 **Fig. 4. Evaluation of internal TPDV in CMIP models.** Maps of the 1st EOF of 8-40yrs bandpass
913 filtered SST over the tropical Pacific (shading), and associated sea-level (contours) and 2m wind
914 (vectors) variability for **(A)** observations (96, 98, 99) and **(B)** a multi-model mean of (10). Box
915 plot showing median, interquartile range, maximum and minimum of CMIP6 historical
916 simulations for **(C)** the standard deviation of the TPDV index, and **(D)** the correlation coefficients

917 between E-L and the internal TPDV index. E-L is a measure of the extent to which El Niño
918 dominates each 8-yr period and is defined as $n(\text{EN}) - n(\text{LN})$, where $n(\text{EN})$ = the number of El Niño
919 years and $n(\text{LN})$ =the number of La Niña years in eight-year blocks. ENSO events are defined using
920 a threshold of 1 STD of Niño3.4 SST. The TPDV index is defined here as the first principal
921 component of the 8-40yrs bandpass filtered SST EOF analysis. Observations are shown as a red
922 star.

923

924 **Fig. 5. Detection and attribution of long-term trends in the tropical Pacific.** (A) Observed (96,
925 98, 99) and (B) multi-model mean (10) maps of 1900-2009 linear trends of SST (shading) and
926 surface winds (vectors) over the Tropical Pacific. Annual time series for CMIP6 historical
927 simulations (grey) and observations (colored) of the SST averaged over (C) over the Niño34 region
928 (5°N - 5°S ; 170°W - 120°W) and (E) the western tropical Pacific (10°N - 10°S ; 120°E - 150°E), the
929 latitude of southern hemisphere Hadley Cell's poleward edge (D) and the strength of equatorial
930 zonal (east-west) winds (F). The latitude of southern hemisphere Hadley Cell's poleward edge the
931 latitudinal anomalies of the latitude where zonal mean precipitation-evaporation is zero while the
932 strength of equatorial zonal (east-west) winds is diagnosed from the 10 m zonal wind anomalies
933 in the Niño3.4 region (positive values indicate a weakening Walker Circulation). CMIP6 results:
934 ensemble mean (black lines); 60% (dark blue shading) and 90% (light blue shading) confidence
935 intervals using a t-distribution. Reanalysis: NOAA-20C (red) (99) and ERA-20C (blue) (100); red
936 lines: annual anomalies (thin lines); and 8-yr running averages (thick lines). SST data: HadISST
937 (96). Notice how the spread of model simulations is larger in the Niño3.4 region than in the western
938 Pacific. The Hadley Cell is calculated over all longitudes, not just the Pacific.

939

940 **Fig. 6. Predicting TPDV. (A)** Actual (solid lines) and potential (dashed lines) correlation skill for
941 the surface air temperature averaged over the tropical Pacific as a function of lead time, for
942 initialized forecasts (red) and for uninitialized simulations (blue), estimated using methods
943 described previously (15). The difference between the initialized and uninitialized simulations is
944 an indication of the potential for forecast improvement (15). **(B)** Correlation skill score using 8-
945 year running mean observations of near-surface air temperature and forecast years 2-9 from
946 initialised multi-model decadal predictions. Skill is measured using the mean of 71 ensemble
947 members from seven modelling systems (89). Darker red indicates higher estimated skill.
948 Hindcasts (2) starting every year from 1960 to 2005, with observations described previously (89).
949 Stippling: outside 95% confidence interval.

950
951
952
953
954
955
956
957
958
959
960
961
962

963

964

965