

1 **Atlantic Multidecadal Oscillation and Northern Hemisphere’s climate**
2 **variability**

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12 **Abstract** Proxy and instrumental records reflect a quasi-cyclic 50-to-80-year climate
13 signal across the Northern Hemisphere, with particular presence in the North Atlantic.
14 Modeling studies rationalize this variability in terms of intrinsic dynamics of the Atlantic
15 Meridional Overturning Circulation influencing distribution of sea-surface-temperature
16 anomalies in the Atlantic Ocean; hence the name Atlantic Multidecadal Oscillation
17 (AMO). By analyzing a lagged covariance structure of a network of climate indices, this
18 study details the AMO-signal propagation throughout the Northern Hemisphere via a
19 sequence of atmospheric and lagged oceanic teleconnections, which the authors term the
20 “stadium wave”. Initial changes in the North Atlantic temperature anomaly associated
21 with AMO culminate in an oppositely signed hemispheric signal about 30 years later.
22 Furthermore, shorter-term, interannual-to-interdecadal climate variability alters character
23 according to polarity of the stadium-wave-induced prevailing hemispheric climate
24 regime. Ongoing research suggests mutual interaction between shorter-term variability
25 and the stadium wave, with indication of ensuing modifications of multidecadal
26 variability within the Atlantic sector.

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27 Results presented here support the hypothesis that AMO plays a significant role in
28 hemispheric and, by inference, global climate variability, with implications for climate-
29 change attribution and prediction.

30 **Keywords** Climate indices · Teleconnections · AMO · NAO · ENSO · PDO

31

32 **1 Introduction**

33 A pressing challenge for climate science in the era of global warming is to better
34 distinguish between climate signals that are anthropogenically forced and those that are
35 naturally occurring at decadal and longer timescales. Identifying the latter is essential for
36 assessing relative contributions of each component to overall climate variability.

37 Multi-century proxy records reflect the signature of such a low-frequency climate
38 signal, paced at a typical timescale of 50 to 80 years, with pronounced presence in the
39 North Atlantic (Stocker and Mysak 1992; Mann et al. 1995; Black et al. 1999, Shabalova
40 and Weber 1999; Delworth and Mann 2000; Gray et al. 2004). A similar signature has
41 been detected in the instrumental record (Folland et al. 1986; Kushnir 1994; Schlesinger
42 and Ramankutty 1994; Mann and Park 1994, 1996; Enfield and Mestas-Nuñez 1999;
43 Delworth and Mann 2000; Zhen-Shan and Xian 2007). Modeling studies, with divergent
44 results dependent upon model design (Delworth et al. 2007; Msadek et al. 2010a, 2010b),
45 qualitatively support this pattern and tempo of variability in terms of the Atlantic
46 Meridional Overturning Circulation's (AMOC) influencing redistribution of sea-surface-
47 temperature (SST) anomalies in the Atlantic Ocean (Delworth and Mann 2000; Dong and
48 Sutton 2002; Latif et al. 2004; Knight et al. 2005). Observed multidecadal SST variability
49 in the North Atlantic Ocean, believed consequent of variability in the AMOC, has been
50 termed the Atlantic Multidecadal Oscillation (AMO: Kerr 2000; Enfield et al. 2001).

51 AMO significantly impacts climate in North America and Europe (Enfield et al.
52 2001, Sutton and Hodson 2005, Knight et al. 2006; Pohlmann et al. 2006). In fact, its
53 influence may have a global character; a suite of studies have addressed bits and pieces of
54 requisite interactions (Dong and Sutton 2002, 2007; Dima and Lohmann 2007; Grosfeld
55 et al. 2008; Sutton and Hodson 2003, 2005, 2007; Sutton et al. 2003; Timmermann et al.
56 2007; Zhang and Delworth 2005, 2007; Zhang et al. 2007; Polyakov et al. 2009).

57 Atlantic-related multidecadal variability reflected in geographically diverse proxy
58 and instrumental records, combined with an abundance of numerical studies that link the
59 hemispheric climate signal to AMO-related North Atlantic variability, guide our
60 hypothesis and motivate our present investigation. We endeavor to put these previous
61 results into perspective through use of a statistical methodology that characterizes
62 Northern Hemispheric multidecadal climate variability and provides rigorous statistical
63 bounds on the associated uncertainties. Emergent is a picture of a climate signal
64 propagating from the North Atlantic throughout the Northern Hemisphere via a sequence
65 of atmospheric and lagged oceanic teleconnections. We term this signal the “stadium
66 wave”. The stadium wave is examined here both in terms of its role in generating a
67 multidecadally varying hemispheric climate signature and in its relationship with
68 interannual-to-interdecadal climate variability.

69 Our philosophical approach, data sets, and analysis methods are described in
70 section 2. Section 3 contains the results of our analyses. Section 4 summarizes our work
71 and discusses potential mechanisms underlying the stadium wave.

72

73

74 **2 Approach, data sets and methods**

75 **2.1 Approach**

76 Our strategy was to evaluate collective behavior within a network of climate indices.
77 Considering index networks rather than raw 3-D climatic fields is a relatively novel
78 approach that has its own advantages — such as potentially increased dynamical
79 interpretability, along with apparent increase of signal-to-noise ratio, and enhanced
80 statistical significance — at the expense of detailed phenomenological completeness. In
81 particular, the climate indices may represent distinct subsets of dynamical processes —
82 tentatively, “climate oscillators” — that might, however, exhibit various degrees of
83 coupling (Tsonis et al. 2007). As these indices also pertain to different geographical
84 regions, we directly address the question of the global multidecadal teleconnections.
85 Using fairly standard multivariate statistical tools, we aim to characterize dominant mode
86 of climate co-variability in the Northern Hemisphere and provide rigorous estimates of its
87 statistical significance, namely: What are the chances that the low-frequency alignment of
88 various regional climatic time series, alluded to in an impressive suite of previous climate
89 studies (see section 1), is, in fact, random? Our methodology allows us to give a
90 quantitative answer to this question. Furthermore, objective filtering of the multidecadal
91 signal provides a clearer picture of higher-frequency climate variability and its possible
92 multi-scale interactions within the climate system.

93

94 **2.2 Data sets**

95 Insights gained from extensive literature review regarding proxy records, instrumental
96 data, and climate-model studies; completeness of data sets throughout the 20th century;

97 and preliminary examination of linearly detrended 30-year (should be 13-year) smoothed
98 index-anomaly time series (not shown), collectively prompted our choice to use two
99 subsets of climate indices to address our hypotheses. A main set of eight indices focused
100 on the role played by the stadium-wave teleconnection sequence in the evolution of the
101 multidecadal hemispheric climate signal. Seven complementary indices were then used to
102 generate a larger fifteen-member network (**Table 1**)³; the goals here were to (i) check the
103 robustness of our results to the choice of index subset; (ii) to paint a more complete
104 picture of multidecadal variability; and (iii) to investigate more fully the role of higher-
105 frequency behavior in relation to multidecadal stadium-wave propagation.

106 The original subset of eight indices contained the Northern Hemisphere area-
107 averaged surface temperature (NHT), the Atlantic Multidecadal Oscillation (AMO),
108 Atmospheric-Mass Transfer anomaly (AT), the North Atlantic Oscillation (NAO), El
109 Niño/Southern Oscillation (NINO3.4), the North Pacific Oscillation (NPO), the Pacific
110 Decadal Oscillation (PDO), and the Aleutian Low Pressure (ALPI) indices. The
111 complementary set of seven indices used to generate our expanded network included: the
112 North Pacific Gyre Oscillation (NPGO), ocean-heat-content anomalies of the upper 700
113 and 300 meters in the North Pacific Ocean (OHC700 and OHC300), the Pacific North
114 American (PNA) pattern, the Western Pacific (WP) pattern, the Pacific Meridional Mode
115 (PMM), and the Atlantic Meridional Mode (AMM) indices. In order to maintain
116 consistency with previous studies to which we compare our results, some of the analyses
117 below will also include another ENSO index, the NINO3 index based on the Kaplan et al.
118 (1998) sea-surface temperature data set. All of these indices have been used extensively

³ The time series of the indices used is later shown in Fig. 7, in terms of their decomposition into multidecadal signal and remaining higher-frequency variability.

119 in previous studies and highlight distinctive subsets of climatic processes; see Table 1 for
120 references and brief description of each index.

121 We focused on boreal winter — the period when seasonally sequestered SST
122 anomalies re-emerge (Alexander and Deser 1995), thereby enhancing conditions for
123 potential oceanic–atmospheric coupling. Allied with this goal, the majority of our indices
124 represent the months DJFM. Exceptions include AMO, AT, and OHC, which are annual;
125 this due to logistics of data acquisition, with expectation of minimal impact on results.

126 Gapless time series of all eight primary indices over the 20th century period
127 (1900–1999) are available⁴. Records are less extensive for the complementary set of
128 seven indices — available for only the second half of the 20th century. To in-fill missing
129 values, we used a covariance-based iterative procedure (Schneider 2001; Beckers and
130 Rixen 2003; Kondrashov and Ghil 2008) described in section 2.3.2. This procedure
131 assumes co-variability-statistics between continuous and gapped indices remain uniform
132 throughout the century. Therefore, some detected relationships among reconstructed
133 indices in the first half of the 20th century may partially reflect statistical co-dependencies
134 of late 20th century indices. We guard against interpreting such relationships in dynamical
135 ways (see section 3.2). Prior to analysis, all raw indices were linearly detrended and
136 normalized to unit variance.

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⁴ This was one of the reasons to consider these as our primary subset.

141 **2.3 Methodology**

142 **2.3.1 Spatiotemporal filtering and climate-signal identification**

143 Compact description of dominant variability implicit in the multi-valued index time
144 series described in section 2.3 was accomplished by disentangling the lagged covariance
145 structure of this data set via Multichannel Singular Spectrum Analysis (M-SSA:
146 Broomhead and King 1986; Elsner and Tsonis 1996; Ghil et al. 2002). M-SSA is a
147 generalization of the more widely used Empirical Orthogonal Function (EOF;
148 Preisendorfer 1988) analysis. M-SSA excels over EOF analysis in its ability to detect
149 lagged relationships characteristic of propagating signals and has been used extensively
150 to examine various climatic time series (Vautard and Ghil 1989; Ghil and Vautard 1991;
151 Vautard et al. 1992; Moron et al. 1998; Ghil et al. 2002 and references therein).

152 M-SSA is, in fact, EOF analysis applied to an extended time series, generated
153 from the original time series, augmented by M lagged (shifted) copies thereof. In M-SSA
154 terminology, each of the multivalued index time series (e.g., AMO, NAO, etc.) is referred
155 to as a channel. Because each multivalued index time series represents a spatial region,
156 eigenfunctions of this extended lagged covariance matrix provide spatiotemporal filters,
157 which define modes that optimally describe lagged co-variability of the original
158 multivariate data set — analogous to EOFs optimally describing zero-lag co-variability.

159 M-SSA modes are represented in the original index space by their reconstructed
160 components (RC). Each RC is effectively the narrow-band filtered version of an original
161 multivariate time series, whose filters are related to coefficients (EOF weights) of M-SSA
162 decomposition of the time series under consideration. In contrast to principal components
163 (PCs) of EOF analysis, RCs are not mutually orthogonal, but their sum across all M-SSA

164 modes is identical to the original time series. We identify leading M-SSA modes with our
165 climate signal and use the sum of their RCs to visualize variability associated with this
166 climate signal.

167

168 **2.3.2 Treatment of missing data**

169 As mentioned in section 2.2, records are complete for the eight indices in the original
170 subset, yet are limited for the seven indices comprised in our complementary set — these
171 are available only since about 1950. M-SSA can be used to infill incomplete data sets;
172 hence, we applied this technique to our extended set of 15 climate indices. The
173 imputation procedure was developed in Beckers and Rixen (2003) and Kondrashov and
174 Ghil (2008); it is a simpler version of the method proposed by Schneider (2001). At the
175 initial stage, missing values of each index were replaced by this index’s time mean,
176 which was computed over available data points. Anomalies associated with this stage’s
177 “climatology” were formed, followed by application of M-SSA to the resulting anomaly
178 field. Time series in channels with missing data were then reconstructed (data at points
179 with actual data were not changed). Only the first N M-SSA RCs were retained, where N
180 was chosen to correspond to the number of modes that account for a certain fraction (we
181 use 90%) of the full data set’s variability. Remaining modes were regarded as noise and
182 did not participate in the reconstruction. At the next iteration, the climatology based on a
183 newly computed index time series was subtracted from the corresponding indices. The
184 procedure was repeated until convergence, i.e., until in-filled data values stop changing
185 within 10% of their standard error. Results of the data infilling were insensitive to the M-
186 SSA window used, for window sizes $M=10, 20, 30,$ and 40 .

187 We stress our caution in interpretation of statistical relationships found between
188 in-filled indices in the first part of the 20th century. These relationships will be considered
189 real only if their counterparts exist in the late 20th century.

190

191 **2.3.3 Assessing statistical significance of stadium-wave-related cross-correlations**

192 We tested statistical significance of M-SSA identified multidecadal modes (section 3.1)
193 and that of cross-correlations between interannual-to-interdecadal climate indices (section
194 3.2) using two different stochastic models. The first model is a red-noise model fitted
195 independently to each index considered. It has the form

$$196 \quad x^{n+1} = ax^n + \sigma w, \quad (1)$$

197 where x^n is the simulated value of a given index at time n ; x^{n+1} is its value at time $n+1$; w
198 is a random number drawn from the standard normal distribution with zero mean and unit
199 variance, while parameters a and σ are computed by linear regression. Model (1)
200 produces surrogate time series characterized by the same lag-1 autocorrelation as the
201 original (one-dimensional) data. However, by construction, true cross-correlations
202 between surrogate time series of different indices are zero at any lag. Nonzero cross-
203 correlations between these time series are artifacts of sampling due to finite length of the
204 time series considered. Observed correlation between any pair of actual time series is
205 considered to be statistically significant only if it lies outside of the envelope of
206 correlation values predicted by multiple surrogate simulations of this pair generated by
207 model (1).

208 The second model is a multivariate, two-level extension of model (1) [Penland
209 1989, 1996; Penland and Ghil 1993; Kravtsov et al. 2005]. Construction of this second

210 model was based on 15-index *anomalies with respect to the multidecadal signal* in order
211 to concentrate on interannual-to-interdecadal, rather than multidecadal, variability. The
212 model produces random realizations of the 15-valued climate-index anomalies with lag 0,
213 1, and 2-yr covariance structure (including cross-correlations between the indices), which
214 is statistically indistinguishable from the observed covariance structure. We used this
215 model to test if cross-correlations between members of our climate network, *computed*
216 *over various sub-periods of the full time series* (see sections 2.3.4 and 3.2.2), exceeded
217 sampling thresholds quantified by this model's surrogate realizations. In other words, we
218 looked into non-stationary relationships between pairs of indices characterized by
219 stronger-than-normal cross-index synchronization, depending on time segment analyzed.
220 Our detailed methodology for this part of analysis is summarized in section 2.3.4.

221

222 **2.3.4 Testing for abnormal short-term synchronizations within the climate network**

223 **Definition of synchronization measure: Network connectivity.** We first
224 computed absolute values of cross-correlations between pairs of observed indices among
225 all 15 members of our climate network over a 7-yr-wide sliding window⁵. This window
226 size was chosen as a compromise between maximizing the number of degrees of freedom
227 in computing cross-correlations, while minimizing the time segment allotted in order to
228 capture the characteristically brief duration of synchronization episodes (Tsonis et al.
229 2007; Swanson and Tsonis 2009). This procedure resulted in 100-yr time series for each
230 of the 105 cross-correlations among all possible pairs of the 15 indices; for each of the
231 100 years, we stored these cross-correlation values in the lower triangular part of the

⁵ The M-SSA-defined multidecadal signal was subtracted from the indices prior to computing cross-correlations.

232 15×15 cross-correlation matrix, with all other elements of this matrix being set to zero.
233 We then defined the *connectivity* of our climate index network for a given year as the
234 leading singular value of the Singular Value Decomposition (SVD; Press et al. 1994) of
235 the above cross-correlation matrix, normalized by the square root of the total number of
236 nonzero cross-correlations in this matrix (which equaled to 105 for the case of the full 15-
237 index set). The connectivity defined in this way described a major fraction (typically
238 larger than 80%; not shown) of the total squared cross-correlation within the network of
239 climate indices, while providing a more robust measure of synchronization than the raw
240 sum of squared correlations (cf. Swanson and Tsonis 2009). The maxima of connectivity
241 time series defined local episodes of the climate network synchronization.

242 **Identification of synchronizing subsets.** Not all members of the climate network
243 contribute equally to various synchronization episodes. To identify index subsets that
244 maximize synchronization measure during such episodes, we employed the following re-
245 sampling technique. We randomly drew a subset of K different indices (K varied from 6
246 to 10) from the full set of 15 indices and computed 100-yr long connectivity time series
247 based on this subset. We repeated this procedure 10,000 times, producing 10,000 100-yr
248 long connectivity time series. We then identified indices and years for which the
249 connectivity value, based on at least one subset including this index, exceeded 95th and
250 99th percentiles of all connectivities that were based on the entire combined set of (100
251 yr) ×(10,000 samples) surrogate connectivity values.

252 **Statistical significance of synchronizations.** To estimate statistical significance
253 of synchronizations, we produced 1000 random realizations of the 15-valued climate-
254 index time series using the second stochastic model described in section 2.3.3. While

255 these surrogates mimic actual observed indices in the overall (climatological) covariance
256 structure, they, by construction, are mathematically unaware of the magnitude and/or
257 phasing of the multidecadal signal that was subtracted from the original data.

258 Connectivity time series for surrogate indices were then computed for each of the
259 1000 realizations in the same way as for the original data. The 95th percentile of surrogate
260 connectivity for each year defines the 5% *a priori* confidence levels. These levels should
261 be used if one tests for some pre-defined years in the climate record being characterized
262 by statistically significant synchronizations.

263 If no such a priori reasons exist, one has to use *a posteriori* confidence levels,
264 which can be defined as follows. For each synthetic realization, we determine the
265 maximum value of connectivity *over the entire 100-yr period*; repetition of this procedure
266 for all stochastic realizations results in 1000 values of maximum surrogate connectivities.
267 The 95th percentile of these maximum surrogate connectivities gives the 5% *a posteriori*
268 confidence level for rejection of the null hypothesis stating that connectivity for each year
269 is statistically the same as its true stationary value based on the correlation matrix
270 computed over the entire 100-yr period.

271

272 **3 Results**

273 **3.1 Multidecadal “stadium wave”**

274 Individual and cumulative variances of M-SSA modes for the main set of eight climate
275 indices are shown in **Figs. 1a, b**. The leading pair of modes is well separated from the
276 others according to the North et al. (1982) criterion. For the latter estimation, we have
277 computed the number of effective temporal degrees of freedom N^* in our data set using

278 the Bretherton et al. (1999) formula $N^* = N(1 - r^2)/(1 + r^2)$, where $N=100$ is the length of
279 the time series, and $r \approx 0.65$ is the maximum lag-1 autocorrelation among our set of eight
280 indices. We will hereafter identify this statistically significant leading M-SSA pair with
281 our climate “signal.” This dominant M-SSA pair *is unlikely to be due to random*
282 *sampling of uncorrelated red-noise processes*, since it accounts for a substantially larger
283 variance [exceeding the 97th percentile – upper dashed line in panel (a)] of the
284 corresponding surrogate spectra generated by model (1).

285 Normalized RCs corresponding to joint M-SSA modes 1 and 2 are shown in **Fig.**
286 **2**; note that the reconstructions of *negative* NHT and AMO indices are displayed. Visual
287 inspection of these RCs indicates they are dominated by a multidecadal climate signal
288 with the time scale of about 50–80 yr. Maximum-entropy spectral analysis (Press et al.
289 1994) of each of these RCs centers on the period of the broadband variability identified at
290 about 64 yr, indicative of long time scales associated with ocean dynamics.

291 Our climate signal accounts for a substantial fraction of variance in several
292 indices: NHT, AMO, PDO, and AT (see the values corresponding to zero low-pass filter
293 time scale in **Fig. 3a**). This latter manifestation (AT) demonstrates a clear atmospheric
294 response to ocean-induced multidecadal climate variations. On the other hand, fractional
295 weakness of the multidecadally varying climate signal exists in NAO, ALPI, NPO, and
296 NINO3.4; these are dominated by shorter-term fluctuations. Despite dominance of higher
297 frequency behavior in the NAO, ALPI, NPO, and NINO3.4 modes, M-SSA still identifies
298 within these indices a statistically significant, coherent multidecadal signal.

299 To demonstrate robustness of these results, we repeated M-SSA analysis using the
300 running-mean boxcar low-pass filtered indices. We used three filters, with sliding

301 window sizes of 10, 20, and 30 years. Ends of the data series are processed using shorter-
302 filter-size running means and one-sided (or asymmetric) filters. Fraction of variance
303 accounted for by the leading M-SSA pair of low-pass filtered data generally increases
304 with scale of low-pass filter, the fractional increase of variance being most dramatic for
305 ALPI, NPO, NAO, and NINO3.4 (Fig. 3a). Normalized RCs of the leading M-SSA mode
306 of low-pass filtered indices are close to their counterparts obtained via M-SSA analysis of
307 raw indices. Fig. 3b shows this behavior in NAO. Note boxcar low-pass filtering is
308 intrinsically characterized by a large phase error. This error is not entirely eliminated by
309 M-SSA time–space covariance-based filtering, explaining shifts between the maxima and
310 minima of filtered and raw RCs in Fig. 3b.

311 Figure 3b also illustrates well the distinction between M-SSA spatiotemporal
312 filtering and time filtering, underscoring the former’s advantage in analyzing climate
313 data. Geophysical time series are typically characterized by presence of noise with
314 substantial power in the low-frequency portion of the spectrum — the so-called red noise.
315 Low-pass time filtering of a synthetic mixture of quasi-periodic time series with a strong
316 red-noise component may erroneously reflect spurious low-frequency variability that has
317 little to do with the shape of the actual quasi-periodic signal used to construct this time
318 series. On the other hand, if the quasi-periodic signal is characterized by a coherent
319 spatial structure that is distinct from noise, and if multivariate time series representing
320 geographically diverse indices possessing this signal are available, separation of signal
321 from noise is more robust (see Ghil et al. 2002 for further details). Note M-SSA
322 reconstruction of low-pass filtered NAO index in Fig. 3b (light solid line) does not follow
323 exactly excursions of the original low-pass filtered NAO index (heavy solid line).

324 Deviations from purely temporal filtering reflect corrections resulting from M-SSA,
325 which relies on cross-index lagged correlations of the dominant multidecadal signal.

326 To confirm a truly multivariate nature of the leading M-SSA signal, we employed
327 the following bootstrap procedure. We randomly generated multiple (~15,000)
328 combinations of the eight primary indices and applied M-SSA to each randomly
329 assembled set, resulting in multiple sets of eight-valued RC time series — in some cases
330 with certain indices being repeated, and others being omitted. For each of the eight
331 indices, this procedure resulted in 10,000 sets of RC time series corresponding to leading
332 M-SSA pairs of random index sets. Any statistical characteristic based on RCs of the
333 original index set can also be computed for surrogate sets. The latter estimates can be
334 used to assess the expected spread of statistical quantities computed. For example, the
335 bootstrap-based estimate of dominant timescale of the leading M-SSA signal produces
336 the mean value of 63 yr, with the standard uncertainty of ± 5 yr and 95% confidence
337 interval close to twice the standard uncertainty. These estimates are thus consistent with
338 results of the original eight-index analysis. *Robustness of bootstrap-based M-SSA*
339 *reconstructions highlights lagged co-variability among the entire set of eight indices; the*
340 *leading M-SSA signal is not “dominated” by the variability in a particular M-SSA*
341 *channel (say, NHT or AMO).*

342 It is apparent from Fig. 2 that the climate-signal indices are correlated at nonzero
343 lags, except for the normalized RCs 6 (NPO) and 7 (PDO), which turned out to be
344 virtually identical. We objectively determined the lead–lag relationships between the RCs
345 using the cross-correlation analysis (not shown), which unambiguously identifies, based
346 on maximum cross-correlation, the order of RCs in the “stadium wave” propagation

347 across the phase space of the reconstructed climate indices. The phase of each RC in the
 348 stadium-wave sequence was computed by least-square fitting two-predictor time series
 349 $\{\sin(2\pi/64), \cos(2\pi/64)\}$, with t changing from 0 to 100, to each of the RCs. The
 350 following phase shifts between the “adjacent” RC pairs resulted:

351

352 $-NHT \rightarrow(4 \text{ yr})\rightarrow -AMO \rightarrow(7 \text{ yr})\rightarrow AT \rightarrow(2 \text{ yr}) \rightarrow NAO \rightarrow(5 \text{ yr})\rightarrow NINO3.4 (3 \text{ yr})\rightarrow$
 353 $NPO/PDO \rightarrow(3 \text{ yr})\rightarrow ALPI \rightarrow(8 \text{ yr})\rightarrow NHT \rightarrow(4\text{yr})\rightarrow AMO \rightarrow (7 \text{ yr})\rightarrow -AT \rightarrow(2 \text{ yr})\rightarrow -NAO$
 354 $\rightarrow(5 \text{ yr})\rightarrow -NINO3.4\rightarrow (3 \text{ yr})\rightarrow -NPO/PDO\rightarrow (3 \text{ yr})\rightarrow -ALPI\rightarrow (8 \text{ yr})\rightarrow -NHT.$

355

356 This “stadium wave” is summarized as a Hoffmuller diagram in **Fig. 4**. The complete
 357 cycle of the stadium wave has naturally a period of 64 yr, and a bootstrap-based standard
 358 uncertainty of ± 5 yr. The bootstrap-based estimates of the above lags and their
 359 uncertainties are listed in **Table 2**. Note that the fractional uncertainties vary considerably
 360 from one transition leg to another, due mainly to differences in mean duration of various
 361 transitions in the presence of fairly uniform absolute uncertainty. This describes the case
 362 for all but the $-NHT\rightarrow-AMO$ and $-AMO \rightarrow +AT$ transitions, which involve indices
 363 strongly dominated by low-frequency anomalies. Uncertainty of lag-time between these
 364 indices is smaller than that associated with index pairs involving noisier raw indices.

365 M-SSA results for the extended data set support those of the original stadium-
 366 wave analysis. Leading M-SSA modes 1 and 2 describe the multidecadal stadium wave,
 367 with no modification of phase relationships between the original eight indices (Fig. 4 and
 368 **Fig. 5**). While all 15 indices reflect the dominant multidecadal signature, a stronger high-
 369 frequency component characterizes most of the seven complementary indices, rendering
 370 their fractional variance of the stadium-wave signal weak. Only those with strong

371 fractional variance — PNA and OHC300/700 — are displayed in the Hoffmuller diagram
372 in Fig. 4. OHC700 exhibits the strongest fractional variance among the trio from the
373 complementary indices, with a strong multidecadal component nearly in phase with that
374 of the AT index. OHC300, on the other hand, exhibits a much lower fractional variance
375 and its phasing is close to the NAO and NINO3.4 multidecadal signals. These
376 observations invite speculation that ocean-heat-content variability at various depths
377 within the subsurface North Pacific is governed by different dynamical processes related
378 to indices within the stadium-wave network (see also discussion in section 4.2).

379

380 **3.2 Interannual-to-interdecadal variability**

381 **3.2.1 Variability unrelated to the stadium wave**

382 To extract a residual signal from our 15 indices, we subtract the multidecadal signal (**Fig.**
383 **6**, blue lines) associated with the stadium wave. This forms a network of climate indices
384 (**Fig. 6**, red lines) that emphasizes higher-frequency, interannual-to-interdecadal climate
385 variability. To study possible connections within this network, we first identify
386 statistically significant (lagged) cross-correlations between index anomalies so computed
387 (see **Table 3**). *A posteriori* significance levels were determined for each index pair using
388 surrogate indices simulated by the red-noise model (1); see section 2.3.3. We computed
389 maximum lagged correlations (using the entire 100-yr time series and over a range of all
390 possible lags) for 100 sets of surrogate index time series, followed by sorting these
391 correlations in ascending order, taking the 99th sorted values as the 1% significance level.

392 Analysis results of high-frequency variability of the residual signal point to a
393 Pacific-centered subset of indices — PNA, PDO, ALPI, NINO3.4, WP, NPO — whose

394 members exhibit behavior similar to one another (that is, they are correlated at zero or
395 one-year lags). Guided by this observation, we applied M-SSA to this select group of
396 indices as a way to verify suspected relationships among them and to more graphically
397 describe the cross-correlations between them. Relationships identified in cross-
398 correlation analysis are corroborated by M-SSA results (**Figs. 7 and 8**). In particular, we
399 find NINO 3.4 and NPO correlated at zero lag; although the low fractional variance of
400 NPO suggests caution regarding its degree of uncertainty. The same is true for WP (see
401 section 4.3). NINO3.4 and NPO lead others within this subset by one year, with the
402 typical time scale of variability being on the order of three to four years. Based on time
403 scale and spatial confinement, behavior of the Pacific-centered set has strong association
404 with ENSO. These results are not surprising or novel. However, they do demonstrate that
405 the network approach we took is sound, as it corroborates previous findings by other
406 researchers on ENSO teleconnections.

407 Another statistically significant cross-index relationship identified includes that
408 between OHC700 and OHC300, which is, of course, not unexpected given the similar
409 nature of the two indices. There is also the anti-correlation between the members of
410 PMM/NPGO pair. Both of the latter indices are dominated by interdecadal anomalies
411 (see Fig. 6). There is also a statistically significant positive simultaneous correlation
412 between the interdecadal anomalies of AMO and NHT indices, demonstrating that the
413 North Atlantic decadal-to-interdecadal SST anomalies directly explain a significant
414 fraction of the NHT decadal-to-interdecadal variability. Finally, the cross-correlation
415 analysis identifies substantial co-variability within the Atlantic-based index trio: AMM,

416 AMO, and NAO, thus suggesting interesting dynamical connections among various
417 components of the coupled AMOC system.

418

419 **3.2.2 Effects of the stadium wave on interannual-to-interdecadal climate anomalies**

420 Next we wish to see if collective behavior of higher-frequency variability of the residual
421 signal has a relationship with the stadium-wave multidecadal signal. This line of inquiry
422 was motivated by work of Tsonis et al. (2007) and Swanson and Tsonis (2009). These
423 authors identified five intervals within the 20th century during which certain climate
424 indices abnormally synchronized — matching with strong statistical significance in both
425 rhythm and phase of their interannual variability; the five intervals were centered at the
426 years 1916, 1923, 1940, 1957, and 1976. Three of five synchronizations (1916, 1940,
427 1976) coincided with hemispheric climate-regime shifts, which were characterized by a
428 switch between distinct atmospheric and oceanic circulation patterns, a reversal of NHT
429 trend, and by an altered character of ENSO variability (intensity and dominant timescale
430 of variability (Federov and Philander 2000; Dong et al. 2006; Kravtsov 2010, among
431 others). As the authors had expected, based on synchronized chaos theory, these
432 “successful” synchronizations (i.e., the ones preceding climate shifts) were accompanied
433 by increased network coupling defined via the climate index network’s phase-prediction
434 measure. Two synchronizations *not* leading a climate-regime shift were *not* accompanied
435 by increased network coupling. We observe here that the timing of the regime shifts has a
436 multidecadal pace, consistent with the stadium wave; we are thus motivated to further
437 explore synchronizations in our extended (compared to Tsonis et al.) climate network.

438 We first determine the years of abnormal synchronizations and synchronizing
439 subsets within the main 15-index set using the methodology described in section 2.3.4.
440 Results presented in **Fig. 9** identify five time bands characterized by high connectivity
441 values among the subsets of the full 15-index set; these bands are centered at years 1916,
442 1923, 1940, 1957 and 1976 — thus consistent with studies by Tsonis et al. Synchronizing
443 subsets comprise the following indices (based on the 99th percentile of re-sampled sub-
444 networks; see section 2.2.4): AMM, NAO, NPGO, PNA, WP, PDO for the 1916 subset;
445 AMO, AMM, NAO, NPGO, PNA, ALPI for the 1923 subset; AMO, AT, NAO, PNA,
446 PMM, ENSO, PDO, ALPI for the 1940 subset; PNA, ENSO, NPO, PDO, ALPI, NHT for
447 the 1957 subset; and AMO, AT, NAO, OHC700, OHC300, PNA, WP, PDO ALPI, NHT
448 for the 1976 subset. Connectivity time series computed for each subset exhibit
449 statistically significant maxima during corresponding time bands (**Fig. 10**). While the
450 statistical significance of connectivity maxima drops substantially if one accounts for
451 degrees of freedom associated with possibility of drawing multiple index subsets of a
452 given size from the full set of 15 indices, all of the connectivity maxima in Fig. 10 still
453 exceed the 5% *a priori* levels in this case (not shown). The use of *a priori* levels may, in
454 fact, be more appropriate for successful synchronization episodes, given the evidence for
455 regime shifts and predictions of synchronized chaos theory that rationalize expected
456 synchronizations preceding these shifts (Tsonis et al. 2007; Swanson and Tsonis 2009).

457 Our results here provide a more detailed picture of the 1916, 1940 and 1976
458 (“successful”) synchronizations and expose differences between these three “successful”
459 and two “unsuccessful” episodes (i.e., 1923 and 1957) in terms of index membership of
460 corresponding synchronizing subsets. In particular, the 1957 episode is confined mostly

461 within the Pacific; while the Atlantic reflects a greater influence during the 1923
462 synchronization. Successful synchronizations tend toward a more symmetrical
463 contribution from both the Atlantic and Pacific sectors and PNA participates in all
464 synchronizations. Note that timing between successful synchronizations, shifts between
465 alternating periods of enhanced and diminished interannual variance of ENSO and NAO
466 (**Fig. 11**), and the stadium-wave tempo share similar pace, thus suggesting possible
467 stadium-wave influence on synchronizations within the climate network.

468 In summary, while our research on the interrelationships between the
469 multidecadal signal and higher-frequency variability continues, the above results suggest
470 that the major synchronization episodes and regime shifts of interannual-to-decadal
471 climate variability all over the world are paced by the multidecadal climate variations
472 originating in the North Atlantic (see section 4).

473

474 **4 Summary and discussion**

475 **4.1 Summary**

476 We considered the Northern Hemisphere's climate variability in a network of well-
477 known indices describing climatic phenomena all over the Northern Hemisphere (NH).
478 Our network approach — via data compression to a subspace of dynamically and
479 geographically distinct indices — provides means to establish rigorous estimates of
480 uncertainty associated with multidecadal variability observed in the instrumental climate
481 records and to address the question of how likely these observed multidecadal
482 teleconnections are to be due merely to random sampling of uncorrelated red-noise time
483 series. Multi-channel Singular Spectrum Analysis (M-SSA) of this network (Figs. 1–5)

484 identifies the dominant signal with a time scale of 50–80 yr, which propagates through
485 the phase space of the indices considered as the “stadium wave” (Fig. 4, Table 2). In
486 section 4.2 below, we interpret this stadium wave in terms of the sequence of atmospheric
487 and multi-year-lagged oceanic teleconnections originating from the Atlantic Multidecadal
488 Oscillation (AMO) — an extensively studied intrinsic oceanic mode associated with the
489 variability of Meridional Overturning Circulation (MOC). The stadium-wave propagation
490 is reflected in the NH area-averaged surface temperature signal, which explains, by
491 inference, at least a large fraction of the multidecadal non-uniformity of the observed
492 global surface temperature warming in the 20th century.

493 In section 3.2, we decomposed the climate indices into the stadium-wave
494 multidecadal and shorter-term residual variability (Fig. 6). The dominant residual signal
495 is the Pacific-centered suite of interannual ENSO teleconnections (Table 3 and Figs. 7
496 and 8). Results summarized in Table 3 also show that a major portion of the shorter-term
497 (with respect to the stadium wave time scale), decadal-to-interdecadal variations of NH
498 surface temperature (NHT) is directly related to the corresponding variability of the
499 North Atlantic sea-surface temperature (SST). In combination with the inferred
500 AMO/NHT stadium-wave connections, these findings imply that the North Atlantic SSTs
501 exert a strong influence on the hemispheric climate across the entire range of time scales
502 considered.

503 One of the most intriguing findings that came out of our analysis is the apparent
504 connection between the hemispheric climate regime induced by the stadium-wave
505 dynamics and the character of interannual climate variability. In particular, the pacing of
506 the stadium wave is consistent with the timing between the episodes of abnormal

507 synchronization of Atlantic/Pacific teleconnections (Figs. 9–11). The termination of these
508 synchronization episodes coincided with hemispheric climate-regime shifts characterized
509 by the switch of the climate regime from the one characterized by the high ENSO
510 variance (pre-1939), to the one with low ENSO variance (1939–1976), and back to the
511 high ENSO variance regime (post-1976); see Fig. 11. The NAO interannual variance
512 exhibits similar multidecadal variability, as shown in Fig. 11, suggesting Pacific–Atlantic
513 feedback.

514 Thus, results presented in this note suggest that AMO teleconnections, as captured
515 by our stadium-wave, have implications for decadal-scale climate-signal attribution and
516 prediction — both significant to the developing field of climate research.

517

518 **4.2 Discussion: Stadium-wave’s hemispheric propagation**

519 Statistical results developed above describe only co-variability within the climate-index
520 network, not causality. We interpret these results based on a wide variety of observational
521 and modeling studies, which rationalize the existence of the multidecadal climate signal
522 — Atlantic Multidecadal Oscillation (AMO) — that originates dynamically in the North
523 Atlantic Ocean and propagates throughout the Northern Hemisphere via a suite of
524 atmospheric and oceanic processes, or the stadium wave.

525 Inconsistent modeling results foil simple narrative, with dependency of the
526 simulated low-frequency climate variability upon model configuration, initialization, and
527 spatial resolution (Metzger and Hurlburt 2001; Yualeva et al. 2001; Kushnir et al. 2002;
528 Kelly and Dong 2004; Minobe et al. 2008). Instrumental records, on the other hand, are
529 short, and while they have been shown to reflect a multidecadal signature (Folland et al.

530 1986; Kushnir 1994; Schlesinger and Ramankutty 1994; Delworth and Mann 2000), their
531 direct interpretation is difficult. These collective challenges are mollified somewhat by
532 observations from proxy records (Mann et al. 1995, 1998; Black et al. 1999; Shabalova
533 and Weber 1999; Gray et al. 2004), which do indicate the tendency of a diverse collection
534 of paleo-climate indices to exhibit climate variability on a shared multidecadal cadence –
535 a finding similar to ours.

536 Despite caveats of model studies, their results are instructive in context of our
537 proposed stadium wave, describing potential mechanisms for climate-signal generation
538 and its subsequent hemispheric propagation. Hypotheses for AMO origin often find
539 common ground in the suspicion that AMOC plays a nontrivial role in generating and
540 setting the pace of the AMO. Many models produce intrinsic interdecadal and
541 multidecadal variability (Delworth et al. 1993; Delworth et al. 1997; Timmermann et al.
542 1998; Cubasch and Voss 2000; Delworth and Greatbatch 2000; Hilmer and Jung 2000;
543 Eden and Jung 2001; Shindell et al. 2001; Bryden et al. 2005; Getzlaff et al. 2005;
544 Delworth and Dixon 2006; Latif et al. 2004, 2006; Dima and Lohmann 2007; Vellinga
545 and Wu 2004; Delworth et al. 2007; Msadek et al. 2010), although mechanisms behind
546 the observed variations are still a matter of debate. Models with periodicities and boreal-
547 winter atmospheric projections closest to observation seem to require a deep or
548 interactive ocean (Knight et al. 2005; Msadek et al. 2010b), underscoring the
549 hypothesized AMO link to AMOC-related ocean dynamics.

550 The Atlantic SST Dipole — an index based on interhemispheric Atlantic SST
551 anomalies — is thought to reflect AMOC-forced SST fluctuations related to the AMO
552 (Black et al. 1999; Keenlyside et al. 2008). Motivated by this observation, we added the

553 dipole to our MSSA mix of indices. What we found was the SST-dipole-index RC
554 identical in phasing to that of the NHT RC. NHT leads AMO by about four years — and
555 so does AMOC, according to the SST Dipole proxy. We also experimented with the
556 addition of a paleo-proxy for the SST Dipole (*G.bulloides*; see Black et al. 1999) to our
557 MSSA collection of indices and found, once again, the phasing identical to the NHT RC.
558 Considering the collective evidence — the dipole phasing, paleo-proxy phasing, and
559 model studies, we suggest the AMOC indeed plays a significant role in generating the
560 AMO.

561 **The AMOC generated multidecadal SST signal and its relationship with the**
562 **overlying atmosphere:** Bjerknes (1964) argued that long-term top-of-the-atmosphere
563 radiative balance remains fairly constant and so is the total poleward heat transport
564 accomplished by ocean and atmosphere. Therefore, when one vehicle of poleward heat
565 transport weakens, the other strengthens to compensate, the phenomenon known as
566 “Bjerknes Compensation.” Shaffrey and Sutton (2006) and Van der Swaluw et al. (2007)
567 showed via modeling studies that Bjerknes compensation is operative in the northern
568 North Atlantic Ocean, with maximum expression between 60°N and 80°N, on decadal
569 and longer time scales. Note that the AMO SST index reflects the ocean surface signature
570 of AMOC multidecadal variability and is thus not representative of the oceanic heat
571 transport, which may exhibit decadal delays with respect to SST due to oceanic
572 dynamical inertia (see, for example, Marshall et al. 2001a,b; Kravtsov et al. 2008, and
573 references therein). Consistent with this notion, our results find that the –AMO index
574 leads the AT index (related to atmospheric mass and heat transport) and NAO index
575 (related to atmospheric mass) by 7 and 9 yr, respectively. This implies that

576 oceanic/atmospheric heat transport has a minimum/maximum a few years after peak
577 negative SST anomalies in the North Atlantic, with atmospheric transport causing NAO
578 to peak shortly after. These timings are consistent with Msadek et al. (2010b).

579 **Relationship of the multidecadal SST signal transmitted with the overlying**
580 **atmosphere:** Mechanisms behind the North Atlantic SST anomalies' influence on the
581 overlying atmosphere on decadal-to-multidecadal time scales, as suggested by the above
582 energy-balance considerations, remain the subject of extensive research. For example,
583 Kelly and Dong (2004) and Dong and Kelly (2004) found that strong basin-scale
584 westerlies in both the North Atlantic and North Pacific co-occur with anomalously high
585 ocean heat content and strong negative heat flux (out of the ocean) in the western-
586 boundary currents and their extensions (see also Hansen and Bezdek 1996; Sutton and
587 Allen 1997, McCartney 1997; Marshall et al. 2001a). These results argue that there exists
588 a positive, reinforcing feedback on SST anomalies through oceanic modification of
589 overlying atmospheric circulation (Palmer and Sun 1985; Latif and Barnett 1994, 1996;
590 Kushnir and Held 1996; Rodwell et al. 1999; Latif et al. 2000; Mehta et al. 2000;
591 Robertson et al. 2000; Czaja and Frankignoul 2002; Wu and Gordon 2002; Sutton and
592 Hodson 2003; Wu and Rodwell 2004; Xie 2004; Xie and Carton 2004). Note, however,
593 that details of this response are sensitive to the location of the heat source with respect to
594 the mid-latitude storm track (Peng et al. 1997; Peng and Whittaker 1999; Peng and
595 Robinson 2001; Czaja and Marshall 2001; Peng et al. 2002; Nakamura et al. 2004; Xie et
596 al. 2004; Minobe et al. 2008), with substantial inconsistencies among the models — see
597 Kushnir et al. (2002) for a review and Msadek et al. 2010b.

598 **Generation of the hemispheric response: Atlantic–Pacific teleconnections: A**
599 variety of models with prescribed AMO-related SSTs show consistent hemispheric
600 response (Delworth et al. 1993; Kushnir 1994; Hakkinen 1999; Delworth and Mann
601 2000; Sutton and Hodson 2003; Sutton and Hodson 2005; Knight et al. 2006; Grosfeld et
602 al. 2007; Sutton and Hodson 2007). Longitudinal and latitudinal migrations of
603 atmospheric centers-of-action (COA) appear to govern circumpolar communication of
604 regionally generated climate signals (Kirov and Georgieva 2002; Polonsky et al. 2004;
605 Grosfeld et al. 2006; Dima and Lohmann 2007; Msadek et al. 2010b). Dominant direction
606 of wind flow across the entire range of longitudes (AT) and inter-basin connectivity
607 modify accordingly: Wang et al. (2007) find that COA migrations generate intervals
608 when climate patterns over the North Pacific and over the Eurasian continent upstream
609 are linked, as are regions downstream, via an enhanced PNA and an eastwardly extended
610 jet stream. During these time segments of enhanced circum-hemispheric stationary wave
611 patterns, ENSO influence on ALPI becomes secondary to mid-latitude dynamics (Wang
612 et al. 2007). Latitudinal shifts in atmospheric COAs — strongly influenced by NPO/WP
613 patterns (Sugimoto and Hanawa 2009; Frankignoul et al. 2011) — have also been shown
614 to influence interdecadal-scale migrations of western-boundary-current extensions
615 (oceanic-gyre frontal boundaries), with impact on western-boundary dynamics and air-
616 sea interaction (Kwon et al. 2010; Frankignoul et al. 2011). These sets of processes, after
617 accounting for interannual delays due to oceanic dynamical adjustment, rationalize the
618 sequence of the hemispheric and Pacific centered indices in our stadium wave.

619 Other proposed mechanisms for conveying an Atlantic-born climate signal to the
620 North Pacific include stability changes within the tropical thermocline (Dong and Sutton

621 2005; Dong et al. 2006; Zhang and Delworth 2005, 2007; Zhang et al. 2007;
622 Timmermann et al. 2007) and latitudinal shifts of Intertropical Convergence Zones
623 (ITCZ) (Vellinga and Wood 2002; Vellinga and Wu 2004; Vimont and Kossin 2007).
624 Tropical Pacific multidecadal changes may further modify the North Pacific response to
625 AMO via an atmospheric bridge (Lau and Nath 1994; Deser and Blackmon 1995; Zhang
626 et al. 1996). All the while, the AMOC continues its quasi-oscillatory behavior; the related
627 AMO teleconnection sequence evolves accordingly. Possibly the Pacific communicates
628 back to the Atlantic via tropical connections; this hypothesis will be addressed in a future
629 study.

630

631 **4.3 Discussion: Interannual-to-interdecadal variability**

632 Our results support the notion that higher-frequency variability, especially in the Pacific,
633 is modified according to polarity of the multidecadal signal. A Pacific subset — PNA,
634 PDO, ALPI, NINO3.4, WP, and NPO — thus becomes a focus for this multidecadally
635 alternating interannual-to-interdecadal behavior (see Table 3 and Fig. 8). Perhaps striking
636 is the comparatively low fractional variance displayed by the NPO and its upper-
637 atmosphere counterpart — WP — both of which show significant correlation with
638 NINO3.4. Potentially relevant to this conundrum is the observation of a non-stationary
639 relationship between the NPO and ENSO (Wang et al. 2007), with ENSO-related signal
640 in the ALPI strengthening and weakening at a multidecadal pacing and affecting the
641 hemispheric interconnectedness of mid-latitude Pacific-centered indices, including NPO
642 and WP.

643 Absent from our Pacific grouping are NPGO and PMM, which other studies have
644 linked to ENSO by way of NPO. Our results suggest a strong negative relationship
645 between higher-frequency behavior of NPGO and PMM, but not a significant relationship
646 between NPGO and either NINO3.4 or NPO, in contrast to some other studies (Di
647 Lorenzo et al. 2009; Alexander et al. 2010). This discrepancy may be due to differences
648 underpinning methods and data used. For example, different authors used different
649 versions of the NPO index (Di Lorenzo et al. 2008; Chhak et al. 2008; Ceballos et al.
650 2009; Di Lorenzo et al. 2009; and Di Lorenzo et al. 2010), with possibly large ensuing
651 differences in the NPO–NPGO correlation. The same is true for various versions of
652 ENSO related indices. Further potential explanations for our finding no significant
653 correlations between NPO and PMM, as well as ENSO and PMM, include the seasonal
654 dependence of these cross-correlations missed in our boreal-winter index analysis, and
655 likely nonlinearity of the ENSO–PMM connection (Vimont et al. 2001, 2003 a, b; Chiang
656 and Vimont 2004; Chang et al. 2007), which is not optimally described by linear
657 correlation measures.

658 In conclusion, dynamics presented above — potential mechanisms underlying
659 stadium-wave dynamics and related dynamics on interdecadal time scales — are topics of
660 active and controversial research, reliant upon technological leaps in data retrieval and
661 computer modeling to advance them toward consensus. We suggest momentum on these
662 investigations is building and that our statistical analysis is consistent with these
663 emerging hypotheses.

664

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1133 95: 115-121

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1144 **Table Captions**

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1146 **Table 1.** Observed climate-index network.

1147 **Table 2.** The stadium-wave lags and their uncertainties (yr) estimated using a bootstrap
1148 procedure (see text). The time scale T of the observed stadium-wave cycle is
1149 given by the twice-the-sum of first-row values, while its standard uncertainty —
1150 by the square root of twice-the-sum of the second row's squared values. These
1151 calculations result in the estimate of $T=63\pm 5$ yr, consistent with the 64-yr scale
1152 obtained in the 8-index (and 15-index) M-SSA analyses.

1153 **Table 3.** Maximum lagged correlations (with corresponding lags) between climate
1154 indices computed over the full 100-yr 20th century record, as well as over its
1155 second half (1950–1999); the latter correlation values are shown in parentheses.
1156 Only the correlations that are statistically significant at 5% *a posteriori* level for
1157 *both* of these periods — relative to the null hypothesis of zero correlation at any
1158 lag — are shown. Correlations equal or exceeding 0.5 (upon rounding up) are in
1159 bold font. The negative lags correspond to the column-member of an index pair
1160 leading its corresponding row member*. For example, the maximum lagged
1161 correlation between NINO3.4 and PNA is equal to 0.47 (0.54 during the second
1162 half of the 20th century), with NINO3.4 leading by 1 yr.

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* Note: “column-members” arranged in the leftmost column; “row-members” arranged in the top row.

1166 **Figure Captions**

1167 **Fig. 1.** M-SSA spectrum of the network of eight climate indices (see text): (a) Individual
1168 variances (%); (b) cumulative variance (% of the total). The M-SSA embedding
1169 dimension (window size) $M=20$. The errorbars in (a) are based on North et al.
1170 (1982) criterion, with the number of degrees of freedom set to 40, based on the
1171 decorrelation time scale of ~ 2.5 yr. The +-symbols and dashed lines in panel (a)
1172 represent the 95% spread of M-SSA eigenvalues base on 100 simulations of the
1173 eight-valued red-noise model (1), which assumes zero true correlations between
1174 the members of the eight-index set.

1175 **Fig. 2.** Normalized reconstructed components (RCs) of the eight-index network. The RC
1176 time series shown have been normalized to have a unit variance. The indices are
1177 synchronized at, generally, non-zero lags, except for the indices NPO and PDO
1178 indices, whose rescaled RCs are virtually identical. Note that the reconstructions
1179 of *negative* NHT and AMO indices are displayed.

1180 **Fig. 3.** (a) Fraction of each index variance (%) accounted for jointly by M-SSA modes 1
1181 and 2, as a function of the low-pass filter time scale (yr). Boxcar low-pass
1182 filtering has been applied to each index prior to the M-SSA analysis; the value of
1183 0 corresponds to no prior low-pass filtering (the normalized RCs for this case are
1184 shown in Fig. 2), the values 10, 20, and 30 indicate 10-yr, 20-yr, and 30-yr
1185 running-mean boxcar filtering, with shorter-filter-size running means and one-
1186 sided (or asymmetric) filters used to process ends of the plotted curves. (b)
1187 Filtered NAO time series: heavy line — 30-yr running-mean boxcar filtered time
1188 series; light line — reconstruction based on M-SSA modes 1 and 2 of the 30-yr

1189 low-pass filtered 8-index set; dashed line — reconstruction based on M-SSA
1190 modes 1 and 2 of the raw 8-index set (the same as in Fig. 2). All time series have
1191 been normalized to have unit variance.

1192 **Fig. 4.** Hoffmuller diagram of the “stadium-wave” propagation in the phase space of
1193 normalized RCs (for M-SSA modes 1 and 2 combined) of the extended set of 15
1194 multidecadal indices (eight of the indices were described above, while the other
1195 seven indices are NPGO, OHC700, OHC300, PNA, WP, PMM, and AMM: see
1196 **Figs. 5, 6** and text). For the additional indices, only the ones with substantial
1197 fraction of variance within leading pair of RCs (see Fig. 5) are shown. The
1198 horizontal cross-section in locations marked by index names shown on the left for
1199 the original eight indices, and on the right for the additional indices would
1200 represent time series of the corresponding index as plotted in Figs. 2 and 5. The
1201 indices are sorted from bottom to top of the figure in the order determined by
1202 cross-correlation analysis, while the vertical “distance” between each index equals
1203 the lead time of maximum cross-correlation between adjacent indices. Negative
1204 indices are also included to show one complete cycle (with the estimated period
1205 of 64 yr) of the stadium-wave propagation.

1206 **Fig. 5.** The M-SSA results based on the analysis of the full 15-index data set. The
1207 quantities shown describe the statistics of the seven climatic indices considered in
1208 addition to the original 8-index set (see Figs. 2 and 3). (a) Fraction of raw-index
1209 variance accounted for jointly by M-SSA modes 1 and 2 (Fig. 3a shows this
1210 quantity for the original eight indices). (b) The same as in Fig. 2, but for the

1211 additional seven climate indices [see panel (a) and figure legend for abbreviated
1212 index names].

1213 **Fig. 6.** The network of 15 climate indices: (i) M-SSA modes 1 and 2 RCs (blue); and (ii)
1214 anomaly with respect to the corresponding RC (red). The blue and red lines give
1215 the raw climate index (detrended and normalized to have a unit variance). The
1216 abbreviated index names are given in panel captions.

1217 **Fig. 7.** M-SSA spectrum of the Pacific-centered multi-index set consisting of interdecadal
1218 anomalies (Fig. 6, red lines) of PNA, WP, NINO3.4, NPO, PDO, ALPI indices.
1219 The bottom panel shows raw time series of the NINO3.4 index and its RC based
1220 on M-SSA modes 1 and 2.

1221 **Fig. 8.** The M-SSA results for the Pacific-centered index subset (see Fig. 7). See panel
1222 legend for abbreviated index names. Same conventions as in Fig. 5.

1223 **Fig. 9.** Identification of index sub-networks. For a given year (on the vertical axis), bright
1224 yellow and orange cells indicate that there exists at least one six-index network
1225 that includes the index this cell represents (the indices are arranged along the
1226 horizontal axis), which is characterized by the network connectivity exceeding the
1227 95th percentile of the connectivity for *all possible* 6-index networks and for *all*
1228 *years*; the yellow cells are characterized by the connectivity that exceeds the 99th
1229 percentile. See caption to Fig. 10 and text for the definition of network
1230 connectivity. The results for the subset sizes in the range from 4 to 8 (not shown)
1231 are analogous.

1232 **Fig. 10.** The time series of index connectivity for the sub-networks identified in Fig. 9.
1233 The indices comprising each subset are listed in the caption of each panel. The

1234 measure of connectivity used here is the leading singular value of the lower
1235 triangular part of the index cross-correlation matrices computed over the 7-yr
1236 sliding window. The horizontal dashed lines indicate 5% a priori and a posteriori
1237 confidence levels based on the linear stochastic model that reproduces the
1238 *climatological* cross-correlations between the indices.

1239 **Fig. 11.** Anomalies (with respect to the mean value of 1) of NAO (light black line) and
1240 Nino3 (light red line) standard deviations, along with the M-SSA mode's 1–2 RC
1241 of AT index (heavy blue line; this line is the same as that in the middle panel of
1242 Fig. 7, but multiplied here by a factor of 0.1). The Nino3 standard-deviation
1243 anomaly has been scaled by a factor of 0.5 for easier visual analysis. The standard
1244 deviations for each index shown were computed over the 31-yr-wide sliding
1245 window, with shorter-filter-size running means and one-sided (or asymmetric)
1246 filters used to process ends of the plotted curves.

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1267 **Table 1.** Observed climate-index network.
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<u>Index/Acronym</u>	<u>Reference/Data source</u>	<u>Description/General Information</u>
Aleutian Low Pressure Index <u>AL or ALPI</u>	Beamish et al. (1997) http://www.pac.dfo-mpo.gc.ca/sci/sa-mfpd/downloads/indices/alpi.txt	Relative intensity of SLP in N.Pacific (~50°N) in winter (DJFM). Calculated as mean area (km ²) w/ SLP<100.5kPa. Expressed as anomaly relative to the 1950–1997 mean.
Atlantic Multidecadal Oscillation Index <u>AMO</u>	Kerr (2000);Enfield et al. (2001); Sutton and Hodson (2003) http://www.esrl.noaa.gov/psd/data/timeseries/AMO/	North Atlantic SSTA averaged across 0–60°N, 75–7.5°W; 1871–2003 Monopolar SSTA pattern N. Atlantic
Atlantic Meridional Mode Index <u>AMM</u>	Chiang and Vimont (2004) http://www.cdc.noaa.gov/Timeseries/Monthly/AMM/ www.aos.wisc.edu/~dvimont/Research/MMode/Data/AMM.txt	SST projected onto 10-m wind field 21°S to 32°N; 74°W to 15°E;; SST gradient across Intertropical Convergence Zone (ITCZ) in Atlantic. (proxy for position of Atlantic ITCZ) (Observations available from 1949)
Atmospheric-Mass Transfer Anomalies <u>AT</u>	Vangenheim (1940); Girs (1971) http://alexeylyubushin.narod.ru/lyubushin@yandex.ru	Dominance of air transfer direction 30°–80°N; 45–75°E; Proxy for atmospheric-heat transfer; reflects longitudinal/latitudinal shifts in position of atmospheric centers-of-action, w/ implications for hemispheric communication of climate signal.
North Atlantic Oscillation Index <u>NAO</u>	Hurrell (1995) http://www.cgd.ucar.edu/cas/jhurrell/indices.data.html#npanom	Normalized SLP difference Azores and Iceland ~ 1 st PC (Principal Component) of SLP in the North Atlantic; re-distribution of atmospheric mass b/n subpolar and subtropical latitudes, reflecting jet-stream variations.
Northern Hemisphere Temperature <u>NHT</u>	Jones & Moberg (2003); Rayner et al., (2006) ftp://ftp.cru.uea.ac.uk/data http://www.cru.uea.ac.uk/cru/data/temperature/hadcrut3nh.txt	Average surface land and sea-surface temperatures of Northern Hemisphere 1850–2003 NHT characterizes overall climate variability, with strongest changes occurring at high northern latitudes.
El Niño/ Southern Oscillation region 3.4 <u>NINO3.4</u>	Calculated from the HadSST1 http://www.cdc.noaa.gov/gcos_wgsp/Timeseries/Data/nino34.long.data	Average SSTA 5°N–5°S; 170°W –120°W Proxy for ENSO behavior
North Pacific Gyre Oscillation Index <u>NPGO</u>	Di Lorenzo et al. (2008) http://www.o3d.org/npgo/data/NPGO.txt http://eros.eas.gatech.edu/npgo/	2-nd PC of SSH in NE Pacific. Measures changes in North Pacific gyre circulation (observations available since 1950)
North Pacific Oscillation Index <u>NPO</u>	Rogers (1981); Walker and Bliss (1982); Wang et al. (2007) wanglin@mail.iap.ac.cn cw@post.iap.ac.cn	Meridional dipole in SLP over North Pacific; pressure variations b/n Hawaii and Alaska/Alberta, reflective of meridional re-distribution of atmospheric mass. We used that of Wang et al. 2007: 2 nd EOF SLP _{PA} 100°E - 120°W, 0°-90°N.

<p>Indices of Ocean-Heat-Content Anomalies</p> <p><u>OHCA300</u> <u>OHCA700</u></p>	<p>Willis et al. (2004) jwillis@pacific.jpl.nasa.gov http://www.nodc.noaa.gov/OC5/DATA_ANALYSIS/basin_data.html</p>	<p>Measured in North Pacific for 0 to 300m & 0 to 700m upper layers May serve as proxies for ocean-dynamic-induced SST variations (since these are presumably not confined to the oceanic mixed layer, while those caused by atmosphere-ocean heat flux are).</p> <p>(Observations available since 1955)</p>
<p>Pacific Decadal Oscillation Index</p> <p><u>PDO</u></p>	<p>Mantua et al. (1997); Minobe (1997, 1999) ftp://ftp.atmos.washington.edu/mantua/pnw_imfacts/INDICES/PDO.latest</p>	<p>Leading PC of SSTA north of 20°N in North Pacific, with century-scale globally averaged SSTA removed; Strongly related to intensity and location of Aleutian Low</p>
<p>Pacific Meridional Mode Index</p> <p><u>PMM</u></p>	<p>Chiang and Vimont (2004) http://www.cdc.noaa.gov/Timeseries/Monthly/PMM www.aos.wisc.edu/~dvimont/Research/MMode/Data/PMM.txt</p>	<p>Anomalous meridional SST gradient across the Intertropical Convergence Zone (ITCZ) in the Pacific; proxy for Pacific ITCZ position; MCA of SST and 10-m winds 32°N to 21°S, 175°E to 95°W.</p> <p>(observations available since 1949)</p>
<p>Pacific North American Pattern Index</p> <p><u>PNA</u></p>	<p>Wallace and Gutzler (1981); Bell and Halpert (1995); Overland et al. (1999) http://www.jisao.washington.edu/data/pna/ http://www.cpc.ncep.noaa.gov/products/precip/CWlink/pna/norm.pna.monthly.b5001.current.ascii</p>	<p>Time series related to combination of standardized 500-hPa geopotential height values (Z) in four centers (Wallace and Gutzler 1981); upper tropospheric pattern associated with PDO; associated with intensity variations and longitudinal shifts of the Aleutian Low and with strength and eastward extent of the East Asian jet.</p> <p>$PNA = 0.25 * [Z(20N,160W) - Z(45N,165W) + Z(55N,115W) - Z(30N,85W)].$</p> <p>(Observations available since 1951)</p>
<p>West Pacific Pattern Index</p> <p><u>WP</u></p>	<p>Wallace and Gutzler (1981) http://www.beringclimate.noaa.gov/data/</p>	<p>Difference in the normalized 500-hPa height anomalies between two points (60°N, 155°E and 30°N, 155°E); upper tropospheric pattern associated with NPO. Involves latitudinal shifts of AL, with resulting influence on oceanic-gyre frontal boundary and western-boundary dynamics – a potential “hotspot” of oceanic-atmospheric coupling (Sugimoto and Hanawa 2009)</p> <p>(Observations available since 1950)</p>

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 1282 estimate of $T=63\pm 5$ yr, consistent with the 64-yr scale obtained in the 8-index (and 15-
 1283 index) M-SSA analyses.
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	-NHT → -AMO	AMO → AT	AT → NAO	NAO → NINO3.4	NINO3.4 → NPO/PDO	NPO/PDO → ALPI	ALPI → NHT
Mean	4.3	6.8	2.4	4.8	2.4	2.9	8.2
STD	0.5	0.4	1.3	1.6	1.1	1.3	1.2
½ 95%	0.5	0.5	2.5	3.0	2.0	2.5	2.0

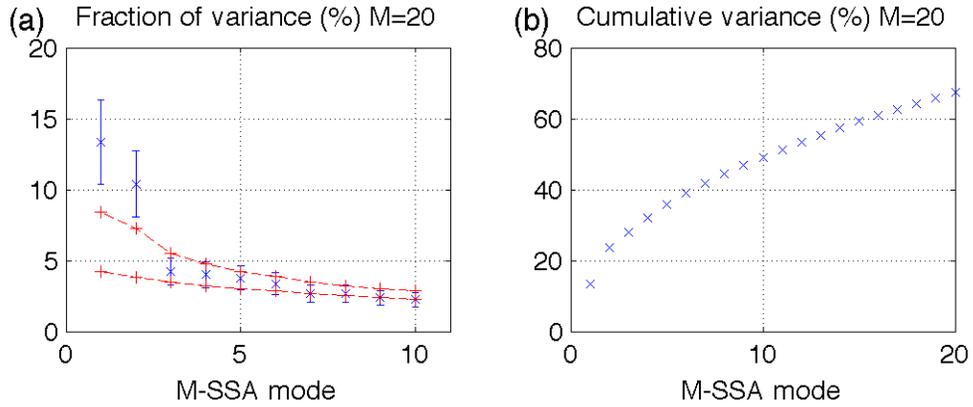
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Table 3. Maximum lagged correlations (with corresponding lags) between climate indices computed over the full 100-yr 20th century record, as well as over its second half (1950–1999); the latter correlation values are shown in parentheses. Only the correlations that are statistically significant at 5% *a posteriori* level for both of these periods — relative to the null hypothesis of zero correlation at any lag — are shown. Correlations equal or exceeding 0.5 (upon rounding up) are in bold font. The negative lags correspond to the column-member of an index pair leading its corresponding row member*. For example, the maximum lagged correlation between NINO3.4 and PNA is equal to 0.47 (0.54 during the second half of the 20th century), with NINO3.4 leading by 1 yr.

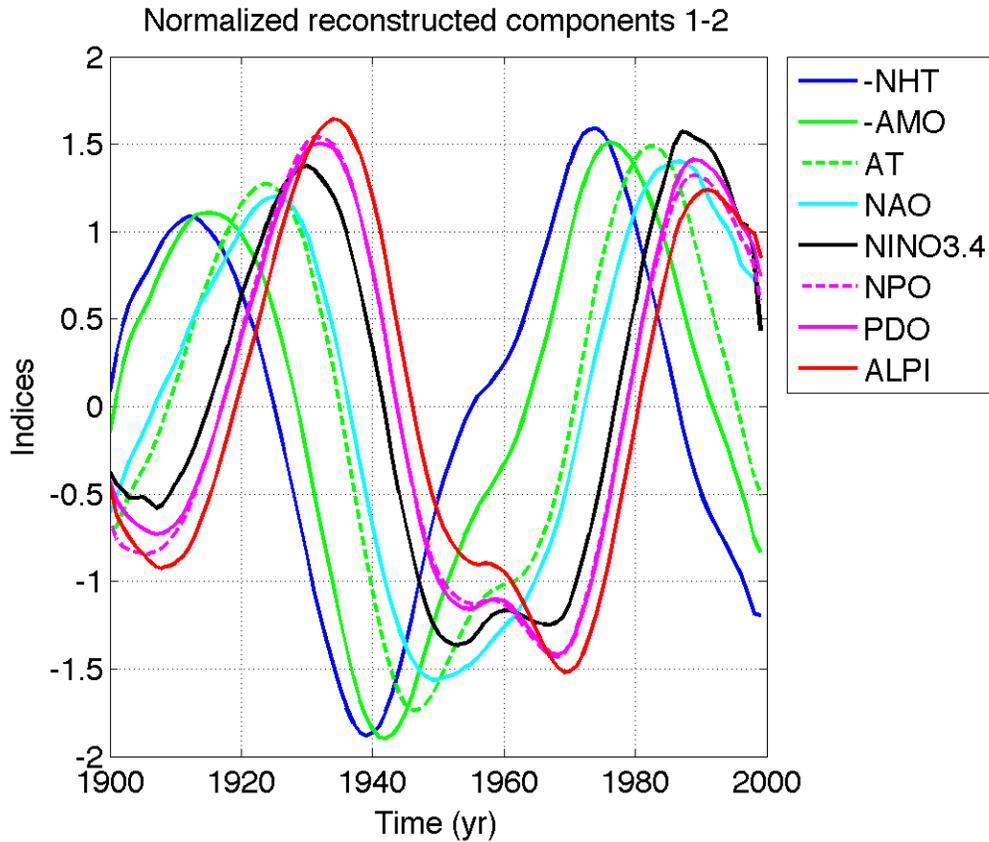
	NPGO	OHC 7	PNA	WP	AMO	NAO	NINO 3.4	PDO
OHC3		0.9 (0.9), 0						
PMM	-0.54 (-0.52), 0							
NINO3.4			0.47 (0.54), -1	0.52 (0.49), -1				
NPO				0.45 (0.46), -1			0.36 (0.48), 0	
PDO			0.75 (0.76), 0				0.41 (0.52), 1	
ALPI			0.77 (0.86), 0					0.55 (0.64), 0
AMM					0.54 (0.6), 1	-0.57 (-0.52), 0		
NHT			0.42 (0.44), 0		0.66 (0.74) 0			

* Note: “column-members” arranged in the leftmost column; “row-members” arranged in the top row.



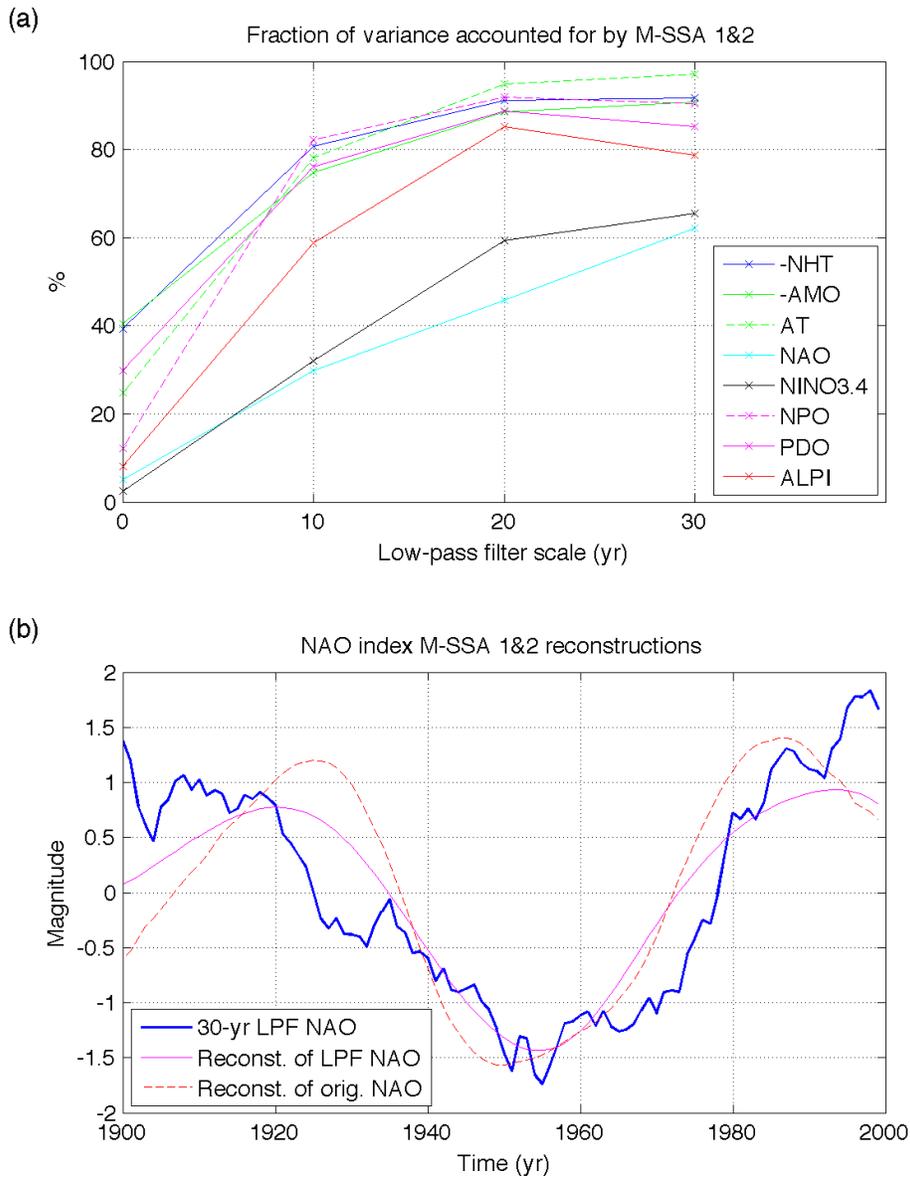
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Fig. 1. M-SSA spectrum of the network of eight climate indices (see text): (a) Individual variances (%); (b) cumulative variance (% of the total). The M-SSA embedding dimension (window size) $M=20$. The errorbars in (a) are based on North et al. (1982) criterion, with the number of degrees of freedom set to 40, based on the decorrelation time scale of ~ 2.5 yr. The +-symbols and dashed lines in panel (a) represent the 95% spread of M-SSA eigenvalues base on 100 simulations of the eight-valued red-noise model (1), which assumes zero true correlations between the members of the eight-index set.

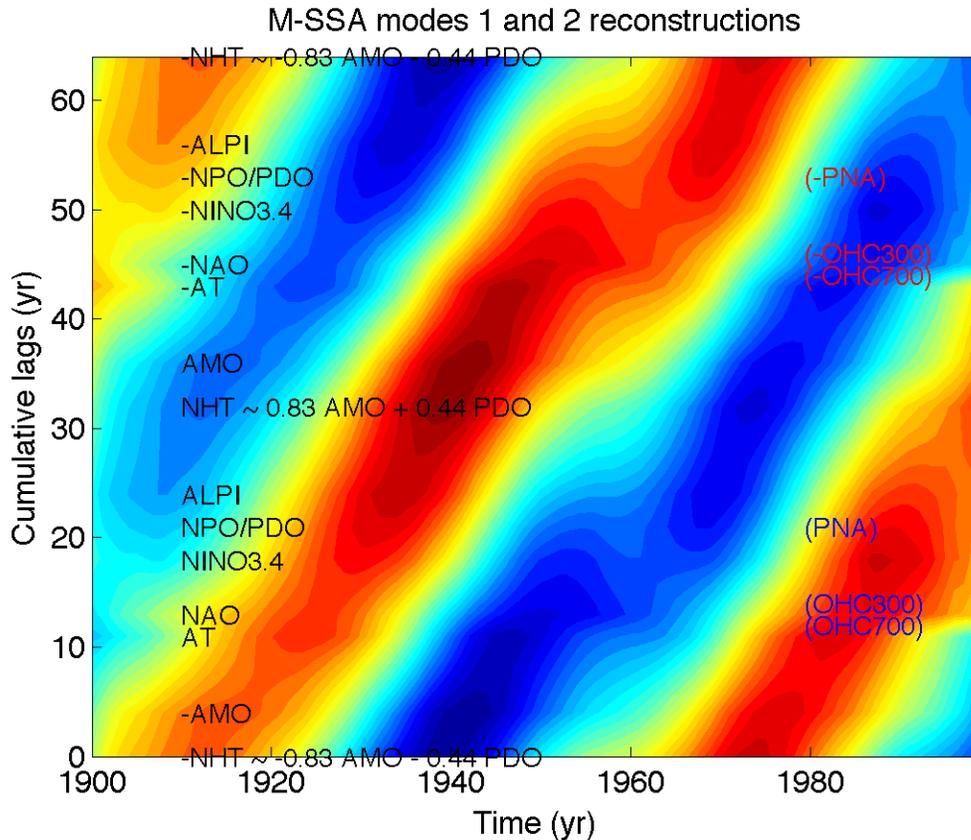


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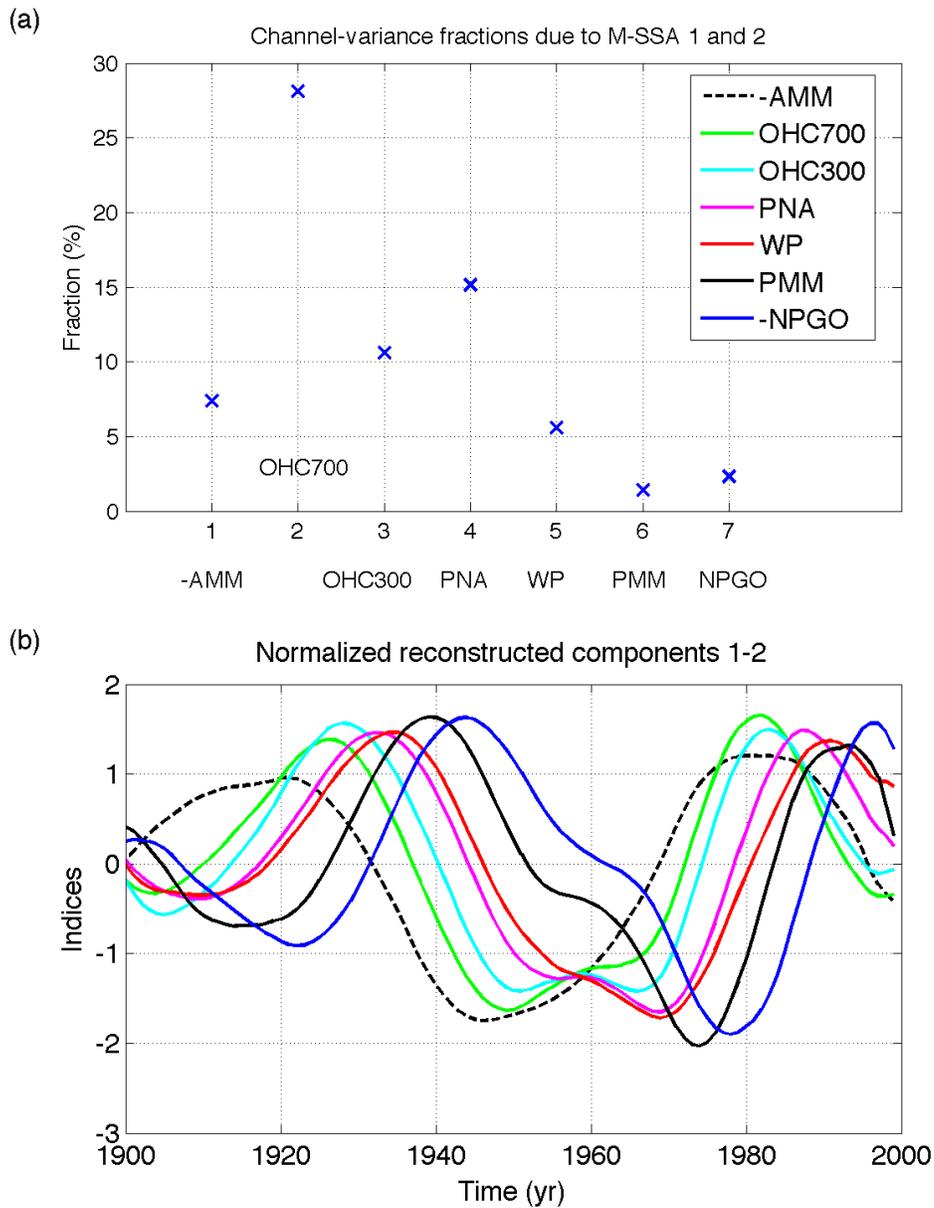


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 1396 and 2, as a function of the low-pass filter time scale (yr). Boxcar low-pass
 1397 filtering has been applied to each index prior to the M-SSA analysis; the value of
 1398 0 corresponds to no prior low-pass filtering (the normalized RCs for this case are
 1399 shown in Fig. 2), the values 10, 20, and 30 indicate 10-yr, 20-yr, and 30-yr
 1400 running-mean boxcar filtering. (b) Filtered NAO time series: heavy line — 30-yr
 1401 running-mean boxcar filtered time series; light line — reconstruction based on M-
 1402 SSA modes 1 and 2 of the 30-yr low-pass filtered 8-index set; dashed line —
 1403 reconstruction based on M-SSA modes 1 and 2 of the raw 8-index set (the same as
 1404 in Fig. 2). All time series have been normalized to have unit variance.
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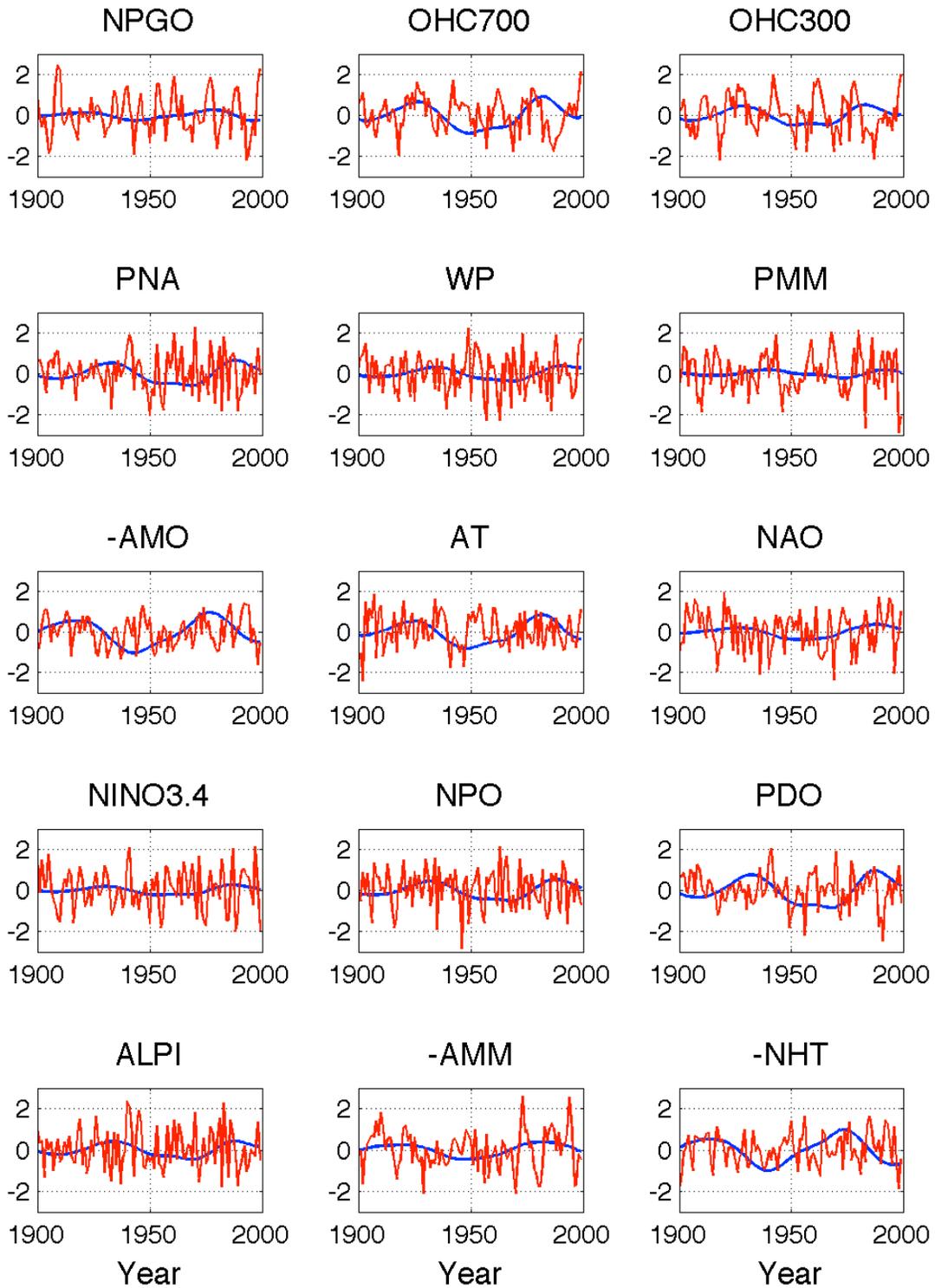
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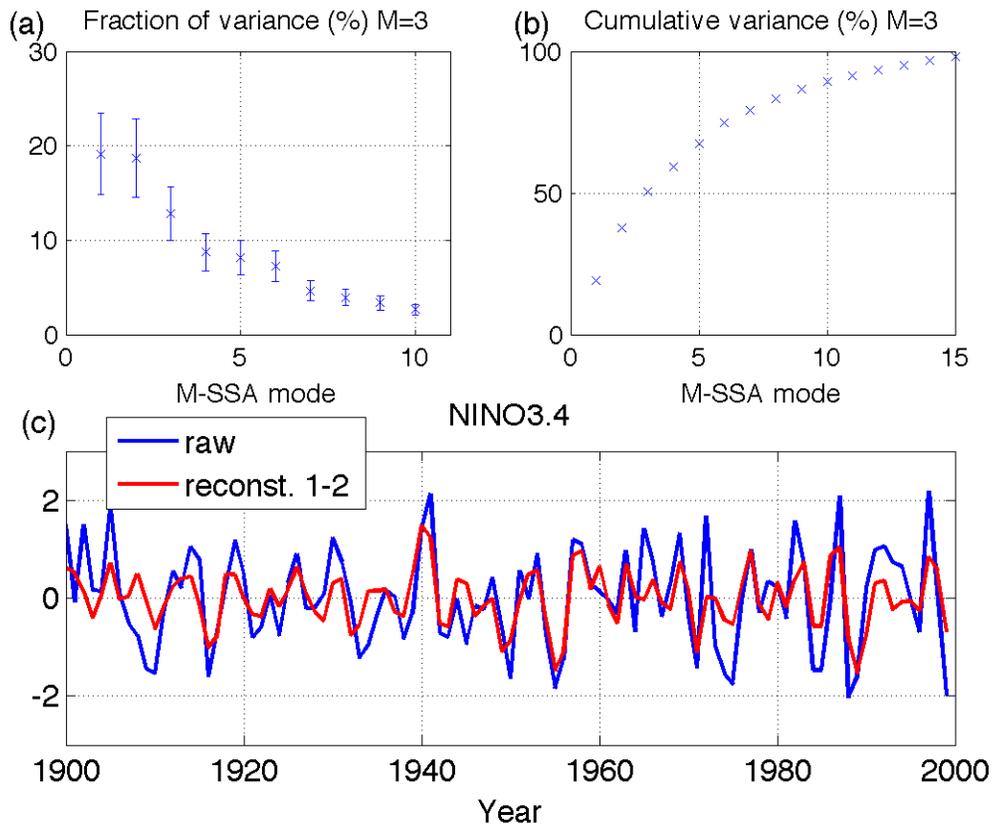
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Fig. 5. The M-SSA results based on the analysis of the full 15-index data set. The quantities shown describe the statistics of the seven climatic indices considered in addition to the original 8-index set (see Figs. 2 and 3). (a) Fraction of raw-index variance accounted for jointly by M-SSA modes 1 and 2 (Fig. 3a shows this quantity for the original eight indices). (b) The same as in Fig. 2, but for the additional seven climate indices [see panel (a) and figure legend for abbreviated index names].



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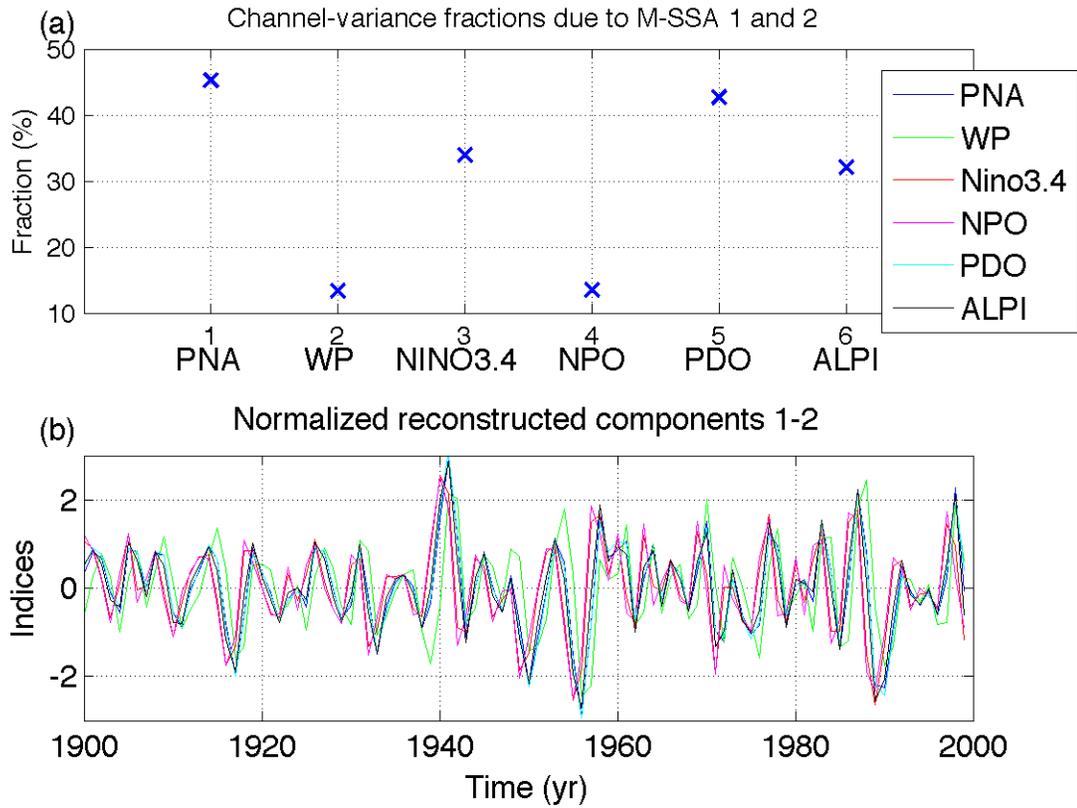
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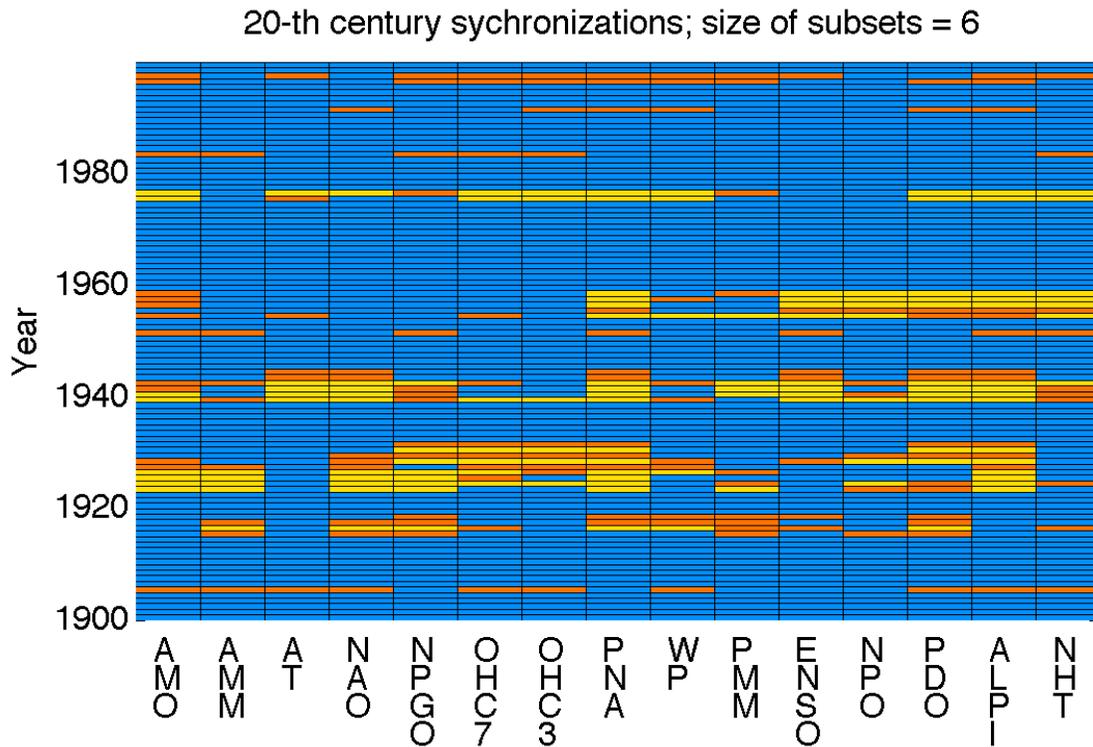
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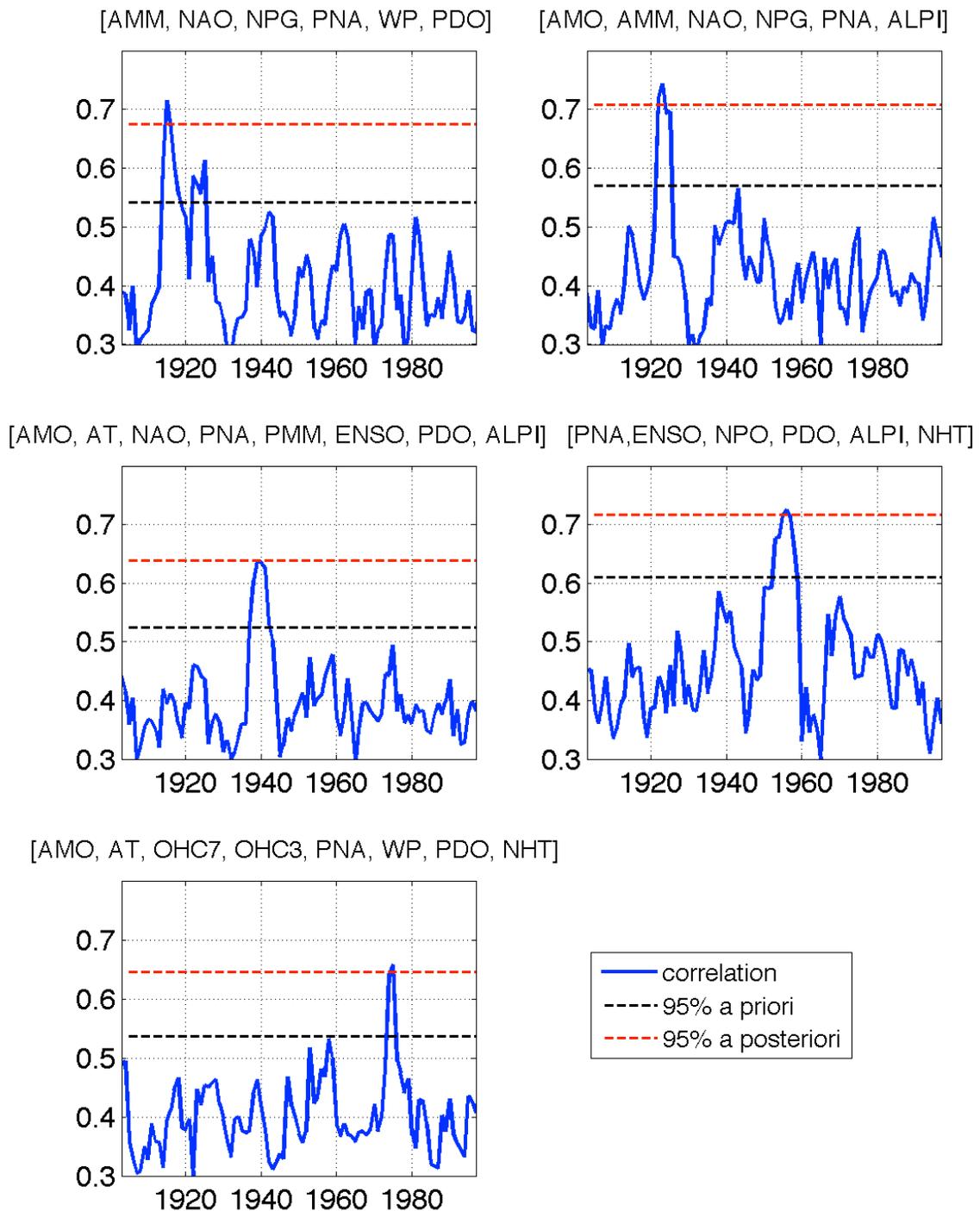
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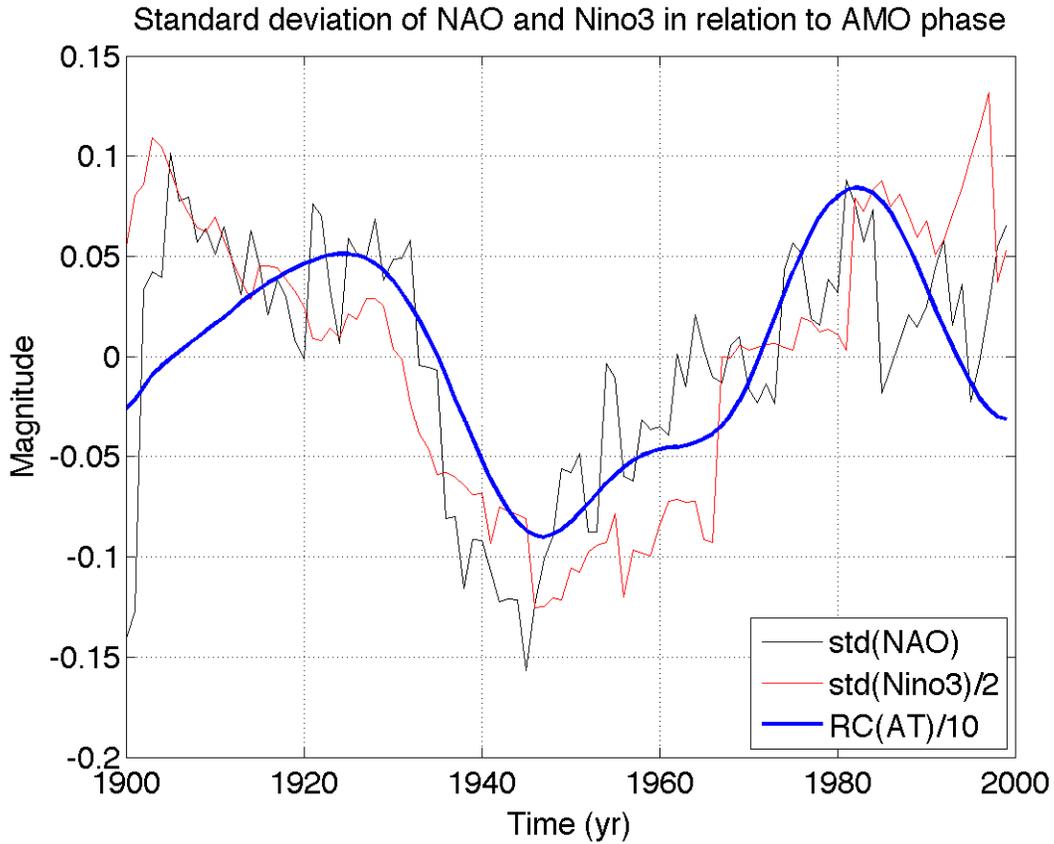
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Fig. 10. The time series of index connectivity for the sub-networks identified in Fig. 9. The indices comprising each subset are listed in the caption of each panel. The measure of connectivity used here is the leading singular value of the lower triangular part of the index cross-correlation matrices computed over the 7-yr sliding window. The horizontal dashed lines indicate 5% a priori and a posteriori confidence levels based on the linear stochastic model that reproduces the climatological cross-correlations between the indices.



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Fig. 11. Anomalies (with respect to the mean value of 1) of NAO (light black line) and Nino3 (light red line) standard deviations, along with the M-SSA mode's 1–2 RC of AT index (heavy blue line; this line is the same as that in the middle panel of Fig. 7, but multiplied here by a factor of 0.1). The Nino3 standard-deviation anomaly has been scaled by a factor of 0.5 for easier visual analysis. The standard deviations for each index shown were computed over the 31-yr-wide sliding window.