Pacific decadal climate variability: indices, patterns and tropical-extratropical interactions

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Highlights:

- Review of patterns and indices of PDV using observations and palaeoclimate data
- IPO, PDO and SPDO related to, affected by, but distinct from ENSO in extratropics
- New stochastic tool provides probabilistic estimate of recent PDV state
- Bi-hemispheric tropical-extratropical interactions might explain coherence IPO/PDO
- Intense research focus needed on PDV observations, palaeoclimate and modelling
Abstract

Pacific decadal variability (PDV) plays a critical role in the climate system. Here I present a
review of indices and patterns of decadal climate variability in the Pacific from observations
and palaeoclimate reconstructions. I examine the spatial characteristics of Pacific sea surface
temperature variability and the metrics used to track observations of PDV. I find
commonalities between the PDV patterns, the Interdecadal Pacific Oscillation (IPO) and its
North and South Pacific counterparts, the Pacific Decadal and South Pacific Decadal
Oscillations (PDO and SPDO). I present a tool to provide probabilistic quantification of the
recent state of the IPO, and use the tool to provide reliable estimates of IPO state up to 2 years
prior to the present. The tool indicates a probability of 80-90% that the IPO remained in its
negative state until 2014-2015. I review palaeoclimate reconstructions of the IPO and PDO,
and outline advances and challenges in our pre-instrumental understanding of PDV. I draw
attention to a Pacific-wide tropical-extratropical mechanism that suggests that the cool and
warm phases of PDV are not driven by tropical or extratropical variability alone, but are
instead the result of continuous tropical-extratropical interactions on decadal timescales. I
conclude by noting the persistent uncertainties and emphasising the need to better
understand decadal variability through continual improvements in observations, an
expansion of palaeoclimate exploration and data collection and renewed efforts in model
development.

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1. Introduction

Variability on decadal timescales is critically important for planning. It is the intermediate timescale nestled between the better understood and more predictable interannual climate variability such as that associated with the El Niño Southern Oscillation (ENSO) and North Atlantic Oscillation (NAO) and the centennial and longer timescale changes due to external factors such as rising greenhouse gas emissions. Decadal-scale changes in the mean climate state are near enough to pose tangible risks to existing critical infrastructure, but long enough to be a significant challenge to climate prediction (Meehl et al. 2009). These challenges are due primarily to the limitations of an under-observed climate and its chaotic natural variability sensitive to initial conditions.

Decadal variations in the risk of extreme episodes such as floods, droughts and heatwaves on decadal timescales have implications for risk assessment, with direct relevance to the insurance industry, the value of public and private assets and the safety of engineered materials and structures which have typically been designed under the assumption of risk stationarity. Agricultural and water supply systems are typically designed to withstand seasonal to interannual variability. But hydrological persistence on decadal timescales can pose additional unplanned threats to human and natural systems. Near term decadal prediction is now an intrinsic part of climate change projection and has a prominent chapter in the regular Assessment Reports (e.g. AR5) of the Intergovernmental Panel on Climate Change (IPCC, Kirtman et al. 2013).

Natural internal patterns of Pacific Decadal Variability (PDV), particularly the Interdecadal Pacific Oscillation (IPO, Folland et al. 1999; Power et al. 1999), the closely-related Pacific Decadal Oscillation (PDO, Mantua et al. 1997) and the Atlantic Multidecadal Oscillation (AMO, Knight et al. 2005) have been implicated in periods of acceleration and slowdown in the rate of global warming at the surface in the post-industrial period (England et al. 2014; Maher et al. 2014; Kosaka & Xie 2013; Kosaka & Xie 2016; Fyfe et al. 2016; Crowley et al. 2014). These patterns are also responsible for coherent and persistent hydroclimatic impacts globally. The uncertainty in the trajectory of global mean surface temperature in the coming 1-2 decades due to internal climate variability is higher than the uncertainty from emissions scenarios (Hawkins & Sutton 2009). This places decadal variability and decadal prediction firmly in the spotlight.

The El Niño Southern Oscillation (ENSO) is at the heart of the variability in the Earth’s climate system. ENSO is the primary driver of interannual climate variability and is responsible for widespread impacts on global weather (Allan 2000). The Interdecadal Pacific Oscillation (IPO) and the Pacific Decadal Oscillation (PDO) are decadally-varying climate patterns in the Pacific. The terminology of these decadal patterns can be the subject of some confusion. One particular point of confusion is the difference between a pattern and a climate mechanism. The false assumption that the PDO and IPO are known mechanisms leads people to question whether they are “real”. The IPO and PDO are SST patterns, and are therefore
most certainly real. The American Meteorological Society (AMS) maintains an online glossary (glossary.ametsoc.org) which lists both the PDO and the IPO as patterns in SST. The PDO is the leading pattern of decadal to multidecadal variability in North Pacific sea surface temperature North Pacific SST data (Mantua et al. 1997; Zhang et al. 1997; Mantua & Hare 2002). The IPO is a broader SST pattern associated with Pacific-wide SSTs (S Power et al. 1999; Folland et al. 1999; Henley et al. 2015).

The IPO and PDO are closely related to ENSO, and most dynamical studies find that these decadal patterns are likely induced, at least partially, by tropical variability associated with ENSO, along with stochastic atmospheric forcing (Newman et al. 2016). The South Pacific complement of the PDO has been termed the South Pacific Decadal Oscillation (SPDO, Chen & Wallace 2015; Shakun & Shaman 2009). This article refers to the patterns of the IPO, PDO and SPDO collectively as PDV patterns.

A review of the mechanisms of the PDO by Newman et al. (2016) underlined findings that the PDO is an aggregated pattern of a number of physical ocean and atmosphere processes, rather than the fingerprint of a single dynamical mechanism in the North Pacific. Liu (2012) reviewed the dynamics of decadal climate mechanisms in the Pacific and Atlantic, highlighting the importance of stochastic forcing and Rossby wave propagation in the extratropics and calling for further targeted studies using coupled ocean-atmosphere models. The study by Meehl & Hu (2006) provided some of the first comprehensive demonstrations of IPO mechanisms. The review of ENSO by Wang et al. (2012) discussed aspects of Pacific decadal variability and identified uncertainties in the mechanisms and preferred timescales of variability.

In this review, I focus on observational and palaeoclimate indices of PDV and draw specific attention to a plausible mechanism for quasi-symmetry in PDV through tropical-extratropical interactions. I aim for a concise summary of common understanding across studies that focus on the IPO and PDO. Considering the degree of symmetry between North and South Pacific patterns, I consider the possibility that the component of the PDO pattern driven by tropical variability is part of the inter-hemispheric quasi-symmetrical IPO pattern. I review a mechanism driven primarily by ocean-atmosphere interactions between the tropics and extratropics, as described by Farneti et al. (2014). In agreement with Newman et al. (2016), I consider atmospheric forcing and regional ocean currents to influence the North and South Pacific PDO and SPDO patterns largely independently, giving rise to the non-symmetrical aspects of the North and South Pacific Patterns. I note also the identification of alternative mechanisms such as the influence of aerosols on decadal mechanisms (Smith et al. 2016). I find agreement between studies that have focussed on the PDO and IPO and I encourage further momentum towards a focussed global effort on monitoring, modelling and prediction of decadal variability.

This review is organised as follows. In section 2 I outline the major observed patterns of variability in the Pacific, and describe the indices used to track these PDV patterns. In section
I review our understanding of past PDV by examining palaeoclimate reconstructions of the IPO and PDO, comparing their temporal evolution and frequency domain behaviour. In section 4, I explore a Pacific-wide extension of the dynamical process presented by Farneti et al. (2014). I reason that despite significant advances in recent decades, we currently lack sufficient data and dynamical evidence to conclusively reject the major proposed mechanisms, recommending that a number of causal mechanistic arrangements remain plausible. I conclude in section 5 by emphasising the need to better understand decadal variability. This can be achieved through sustained continuous observations, continual improvements in observational networks, renewed efforts in model development and experiments targeting decadal variability, and an expansion of palaeoclimate exploration and data collection, particularly in the Southern Hemisphere.

2. Observations and Indices

2.1. Patterns and timeseries of Pacific decadal variability

As the primary driver of interannual climate variability globally, the El Niño Southern Oscillation (ENSO) has been the subject of focussed observation and modelling for several decades. The El Niño and La Niña phases of ENSO occur irregularly at a period of around 2-7 years. In addition to these interannual variations associated with ENSO, decadal-scale variations in Pacific climate, or PDV, have been identified.

Figure 1 summarises the major patterns of global sea surface temperature (SST) variability (using ERSSTv4, Huang et al. 2015). The linear trend in SST for 1900-2015 in panel (a) reveals unequivocal and widespread warming of the ocean surface in the post-industrial era. The trend is between 0–0.1°C per decade over almost all of the ocean surface over the full period. Note that more rapid warming has occurred in the second half of the observed period (not shown). Warming across much of the Pacific is less than in the Atlantic, Indian and Southern Oceans, as noted by Power et al. (2016). The spectral characteristics of global SST are compared in panels (b-d). The average spectral bandpower in each grid cell for the 2-7 year and 7-30 year frequency bands is computed from the power spectral density curve. The upper limit of 30 years is selected to be safely less than the period corresponding to the Nyquist frequency, which is the minimum rate at which a signal can be resolved without introducing aliasing. The global dominance of ENSO at interannual timescales is strongly apparent in the high 2-7 year variance in the eastern and central equatorial Pacific (panel b), with some areas of high 2-7 year variance apparent in the midlatitudes, particularly in the northwest Pacific. Variance in the 7-30 year frequency band is high across much of the central Pacific and in the North Pacific, with the variance maximum in the North Pacific further to the east than in the 2-7 year band. The ratio of the 7-30 year and 2-7 year frequency bands reveals a dominant pattern in the western Pacific and eastern Indian Ocean around the maritime continent, as well as a strong extension along the South Pacific Convergence Zone (SPCZ). The SPCZ is an expansive convective cloud band extending in a southeasterly direction from the western Pacific (Vincent 1994). The high ratio of the bands here reflects both the relatively low
variance in the interannual band and the decadal variability in the location of the SPCZ. The
SPCZ location is understood to be modulated by the PDV mean state (Folland et al. 2002;
Salinger et al. 2014). It is noted that the AMO, which has dominant periodicities at 20 years
and 45-65 years (Chylek et al. 2011; Chylek et al. 2012) is only partially covered by the
frequencies represented here. Similar results are obtained when the analysis is repeated using
the HadISST1 data (Figure S1, Supplementary Material).

The pattern of decadal to multidecadal sea surface temperature (SST) variability in the North
Pacific has been termed the Pacific Decadal Oscillation (PDO) (Mantua et al. 1997; Mantua &
Hare 2002). The Interdecadal Pacific Oscillation (IPO) is characterised by a Pacific-wide
decadally varying SST pattern (S. Power et al. 1999; Folland et al. 1999; Folland et al. 2002;
Parker et al. 2007; Christensen et al. 2013). The South Pacific counterpart of the PDO has also
been documented, and termed the South Pacific Decadal Oscillation (SPDO, Chen & Wallace
2015; Shakun & Shaman 2009). I note however that direct comparison between hemispheric
patterns is influenced by the differing distribution of observations in time and space, with a
higher number of observations in the northern hemisphere than the southern hemisphere,
partially prior to the satellite era.

Previous studies have used a variety of datasets and methods to calculate time series and
spatial patterns of the IPO and PDO, as summarised in Table 1 (adapted from Henley et al.
2015). Most have used Principal Component Analysis (PCA) and related techniques to define
PDV patterns. The method introduced by Henley et al. (2015) uses a linear combination of
mean SST in three large boxes in the Pacific.

Figure 2 compares spatial patterns and timeseries of the three major PDV patterns and Niño
3.4 using ERSSTv4. Both the unsmoothed and smoothed (13-year Chebyshev low-pass
filtered) versions of the timeseries are shown. For the IPO, the Tripole Index (TPI) of Henley
et al. (2015) is shown (calculated as described in Table 1). The PDO is the first principal
component (PC) of North Pacific SST poleward of 20°N (Mantua et al. 1997), the SPDO is the
South Pacific counterpart, the first PC of South Pacific SST poleward of 20°S (Chen & Wallace
2015). Note that poleward boundaries of 60° and 70° in both hemispheres have been used for
PDO and SPDO indices (Chen & Wallace 2015; Henley et al. 2015), however this has little
influence on the resulting patterns and timeseries. Here I use 70°N and 70°S. The Niño 3.4
index is the mean SST in the central Pacific (5°S to 5°N, 170°W-120°W). The patterns in panels
a, c, e and g are obtained by regression of global SST onto the filtered timeseries.

The IPO pattern has hemispheric quasi-symmetry and high variance in large regions in both
the western North and South Pacific. The PDO and SPDO have dominant variance in their
respective hemispheres, particularly the PDO. There are strong similarities between all four
patterns, especially between the IPO and Nino3.4, as noted by Newman et al. (2016), and also
the SPDO and Niño3.4. However, the dominance of the tropical variance over the
extratropical variance is only strongly apparent for Niño 3.4. The other patterns all have
relatively much stronger extratropical variability.
Turning to the temporal evolution of the patterns in panels b, d, f and h, the timing of the decadal variations is strikingly similar for all three PDV timeseries, with the well-established twentieth century shifts apparent around 1945, 1977 and 1999. The fact that the major PDV phase shifts are shared by the PDO and SPDO reveals strong hemispheric symmetry and is suggestive of a common element for these major shifts despite their data regions having no overlap. The decadal Niño 3.4 timeseries has similar timings for these shifts, but has additional zero crossings and substantially shorter durations between decadal oscillations. The PDV timeseries have periods of 30 or more years of persistent PDV phase, but this is not evident in the Niño 3.4 series. Although there is decadal variability in the Niño 3.4 region, the PDV timeseries exhibit significantly higher multidecadal persistence due to their incorporation of extratropical variability. Together, these observations suggest that the PDO and SPDO could be viewed as being part of a larger hemispherically symmetric PDV pattern. This pattern is strongly related to, but has higher memory on decadal timescales than, tropical variability alone. Therefore, diverging from the suggestion of Westra et al. (2015) that low-frequency ENSO and the IPO are equivalent, yet still acknowledging the similarities and known dynamical links between PDV and ENSO (as discussed in section 4), I find sufficient distinction between the spatial patterns and timeseries at decadal timescales to warrant the independence and use of the PDV series. The smoothed Niño 3.4 index is not a sufficient substitute for indices of Pacific decadal variability.

### 2.2. Influence of PDV on global surface temperature

A critically-important influence of PDV identified in recent years is its association with global mean surface temperature (GMST) variations (England et al. 2014; Maher et al. 2014; Kosaka & Xie 2013; Kosaka & Xie 2016; Fyfe et al. 2016; Meehl, Hu, Santer, et al. 2016; Henley & King 2017). Here I update the GMST and PDV timeseries by England et al. (2014) to include recent data and reorient the polarity of highlighted areas to designate the presence of positive phases (Figure 3). Periods of accelerated warming are associated with PDV positive phases (1925-1944, 1977-1998) and periods of slowdown in warming are associated with PDV negative phases (1945-1976, 1999-recent). The expected cooling effect of the negative phase from 1907-1926 is not apparent, which could be due to the uncertainty in IPO phase in this early data, or other non-PDV factors influencing global temperatures at this time. The record GMST years in 2015 and 2016 appear to indicate early signs of a resumption of accelerated warming. Although initialised decadal predictions are suggestive of an IPO positive transition (Meehl, Hu & Teng 2016), the accurate prediction of phase changes with multi-year lead times remains a research challenge. DelSole (2017) provides a review of decadal prediction of temperature, and points toward the additional use of simpler empirical models in decadal prediction. Several large research programs and studies are also focussed on decadal prediction and predictability (Meehl et al. 2014; Boer et al. 2016; Marotzke et al. 2016). Decadal prediction was the subject of a chapter in the fifth assessment report by the IPCC (Kirtman et al. 2013). Given these existing reviews, in the next section I focus on the estimation of the state of PDV in the recent past.
2.3. A tool to quantify the PDV state in the recent past

One of the challenges of tracking low-frequency phenomena is the question of the current state. Low-pass filters require windowing, which induces end effects at the start and termination of a time series. Common methods for handling end points include reflection or reduction of the filter order near the end points, but these induce end effects and can have spurious influences. In the case of a 13 year window (W), using annual data, for example, the last (and first) 6 years of the resulting smoothed series are influenced by reduced data available at the end point. A smoothed timeseries dated at the centre of the filter can have its most recent unaltered data point precisely 6 years prior to the last unfiltered data point. Typically, the most recent ~W/2 years of data are removed, or an alternative end-point estimate is used such as reflection, for which there is no likely physical justification. There is a continuum at the end points, whereby the state of the smoothed series is gradually less certain as the timeseries steps closer to the last data point, at which point it is most uncertain. Here I quantify this uncertainty using the TPI as an example.

I use an autoregressive (AR) model to simulate non-dynamically informed naïve future trajectories of the IPO, conditioned on the previous values. I emphasise that the aim here is only to provide an estimate of the recent state of the IPO, rather than any future state. I first fit a fifth-order autoregressive (AR5) process to the monthly TPI data. I then generate an ensemble of stochastic replicates of future unfiltered TPI, after initialising the model with the last 5 months of TPI data. The model order is optimised using Schwarz’s Bayesian Criterion (Schwarz 1978), an established formal statistical method of finding an optimum model given varying numbers of parameters and goodness of fit. Figure 4a shows the timeseries of unfiltered TPI data from 1990-2017, and the ensemble of low-pass filtered TPI. The ensemble of smoothed TPI series is obtained by stitching stochastic future trajectories from the initialised autoregressive process onto the past observed values and applying the low pass filter to smooth across the past-future boundary. This allows the filtered TPI to be computed up to the current data point, and estimates of the uncertainty of the recent TPI to be made under the assumption of AR red noise. Figure 4b shows the Brier skill score (BSS) for the IPO estimation tool, at lags relative to the current year. Positive values in BSS represent skill in the probability estimate above random chance (noting the influence of serial dependence; Wilks, 2010). Skill is computed for the period 1900-2000, during which the IPO state is known. Figure 4c shows the probability of IPO states for the period 2006-2017 using the IPO state estimation tool, conditioned on the most recent TPI data available (Apr, 2017) at the time of writing.

Most assessments of the IPO state truncate estimates of the low frequency series at around 7-10 years prior to the most recent data. The tool described here provides more information (data is available at www.esrl.noaa.gov/psd/data/timeseries/IPOTPI). It shows that, under the assumption of red-noise, there is an 80-90% probability that the IPO remained in its negative state until 2014-2015, after which point the probability of a positive IPO state is increasing. The tool provides a methodology for estimating the state of the IPO
probabilistically, and in doing so raises the confidence of our knowledge of the recent 1-6 years of the IPO. Better knowledge of the state of climate modes is likely to have benefits for the estimation of risk to water resources (Henley et al. 2013) and robust planning in an uncertain climate (Mortazavi-Naeini et al. 2015).

3. Palaeoclimate reconstructions

3.1. Review of PDV reconstructions

Instrumental PDV indices exhibit significant decadal-to-multidecadal variability. However, the brevity of the instrumental record relative to this decadal-scale variability leads to a high degree of uncertainty in the long-term behaviour of PDV. A better understanding of the frequency preference of PDV is needed. Since the early 2000s, numerous studies have sought additional insight into the pre-instrumental temporal evolution and frequency domain nature of PDV (primarily the IPO, PDO and smoothed ENSO) using palaeoclimate data. This section outlines and compares these reconstructions, extending the overview by Henley (2012).

Biondi et al. (2001) used 189 cores from 99 trees from 6 sites in Southern California (USA) and northern Baja California (Mexico) (31.0°-34.6°N, 115.5°-119.4°W) to develop a PDO reconstruction back to 1661. They had previously found that tree ring records from this region were better correlated with the PDO than ENSO. The species used were Jeffrey pine (Pinus jeffreyi) and big-cone Douglas fir (Pseudotsuga macrocarpa). They then performed a multivariate regression using as predictors the lag-one, concurrent, and lead-one values of the first EOF of the six unsmoothed precipitation-sensitive chronologies. Two 30 year periods of validation data were found to have correlations of 0.6-0.7 between the instrumental and reconstructed PDO. They also found good spectral coherence between their instrumental and proxy record, and a bidecadal frequency (period of around 17-28 years) for the PDO. They found a weakening of the bidecadal signal in the period 1760-1840, with the lower frequencies being restricted to the 20th century. Biondi et al. (2001) mention that although a 22-yr periodicity is also shown by solar activity, their PDO reconstruction showed no correlation with the Sun’s radiative forcing at either annual or decadal timescales. This led them to suggest that internal dynamics of the coupled ocean-atmosphere system have greater dendroclimatic significance than solar cycles at their study sites. Biondi et al. (2001) suggested that an expansion of their analysis using different networks and multiple proxies should be conducted to verify and extend their findings.

Gedalof & Smith (2001) compiled six tree ring-width chronologies since 1600 from stands of mountain hemlock (Tsuga mertensiana) from southern Oregon to the Kenai Peninsula, Alaska (42.6°-60.1°N, 122.1°-149.0°W), further north than the study by Biondi et al. (2001). Gedalof & Smith (2001) noted that while summer temperature is the dominant factor influencing the annual radial growth of mountain hemlock, at lower latitudes the growth was also sensitive to precipitation. They used Factor Analysis to combine the six chronologies. The leading eigenvector from the Factor Analysis explained 44% of the variance in ring-width and a linear...
regression against the PDO provided their PDO reconstruction dated back to 1600. Gedalof & Smith (2001) used wavelet analysis to analyse the evolution of the PDO, and in contrast to Biondi et al. (2001), they found that all of the interdecadal energy (30-70 years) is confined to the pre-1840 portion of the series, with nearly all of the interannual energy (< 8 years) occurring between 1840 and 1930. They found variability at 10-20 years present throughout the reconstruction. This suggests that the PDO is more likely to fluctuate on a decadal timescale, but can also exhibit a wide range of shorter and longer period durations.

D’Arrigo et al. (2001) used tree ring density and width data, also from sites further north than Biondi et al. (2001), including temperature sensitive sites in coastal Alaska (60.4°N, 145.4°W) and the Pacific Northwest (57.0°N, 135.1°W) as well as a precipitation sensitive site on the subtropical North American coast (45.0°N, 121.0°W) and two drought index sites in Mexico. After first screening the datasets for sites with a strong correlation with the PDO index of Mantua et al. (1997), D’Arrigo et al. (2001) produced a tree ring(PDO model using Principal Component Regression (PCR), after Cook et al. (1999). D’Arrigo et al. (2001) produced two PDO reconstructions extending back to 1700 and 1790, using different sites in the makeup of the indices and accounting for 53% and 44% of the variance in the instrumental PDO in their overlapping period. Again, unlike Biondi et al. (2001), the D’Arrigo et al. (2001) reconstructions suggested less pronounced PDO interdecadal variability since the mid-1800s. D’Arrigo et al. (2001) found significant (5% level) spectral power at periods of 25-50 years and around 2-3 years.

MacDonald & Case (2005) produced a millennium length PDO reconstruction covering the period 993-1996 CE using tree rings of both living and dead Limber pine (Pinus flexilis) specimens from the southern and northern areas of previous studies (southern California (34.1°N, 116.5°W) and Alberta in western Canada (52.0°N, 116.5°W)). The sites from these two areas are at opposite ends of the North American precipitation-PDO impact dipole (a positive PDO leads to higher precipitation to the south and decreased precipitation to the north). They used the post-1940 PDO index to calibrate a multiple linear regression model between the ring widths and the PDO. Their index agreed reasonably well with that of Biondi et al. (2001) (r=0.57), but fairly poorly with the reconstruction of Gedalof & Smith (2001) (r=0.19). The first 300 years (the medieval period) of the MacDonald & Case (2005) PDO index exhibits a pronounced negative PDO (993-1300 CE), with other significant century-scale shifts apparent in their index. This indicates that the variability of the PDO has possibly been unstable over previous centuries. Using wavelet analysis on a detrended form of the index they found evidence of 50-70 year periodicity of the PDO, varying significantly in strength over the reconstruction period. The 50-70 year frequency band was earlier reported by Minobe (1997). Hidalgo et al. (2001) and MacDonald & Case (2005) suggested that the relative sensitivity of different tree species to persistent drought could be a reason for the discrepancy between the prominence of the 50-70 year band in their study and the 12-28 year frequency band in earlier studies.
370 tree ring data from Asia, following evidence of interaction between the atmosphere over Asia
371 and decadal Pacific SST variability (Minobe 1997; Frauenfeld & Davis 2002; Schneider &
372 Cornuelle 2005). They used data from 17 sites in east Asia in their PDO reconstruction. This
373 subset was chosen from an initial set of 70 sites, based on their correlation with the PDO over
374 1900-1988 during March-May (just prior to the growing season) over a wide geographical
375 area. Their domain was from as far north as approximately 66.5°N in northern Russia and as
376 far south as 27.5°N in Bhutan, with several sites in Mongolia and Japan. D’Arrigo & Wilson
377 (2006) used PCR, and checked signal fidelity using a nested approach, removing the shortest
378 signal and recomputing statistics, then repeating. They concluded that their reconstruction
379 provides useful information about the Asian expression of the PDO at both interannual and
380 decadal time scales. D’Arrigo & Wilson (2006) reported spectral analysis results that revealed
381 peaks at 2-3 and 23 years, with some lower frequency modes, suggesting some commonality
382 with the work of Biondi et al. (2001) (17-28 years) and D’Arrigo et al. (2001) (25-50 years).
383 D’Arrigo & Wilson (2006) discussed broader linkages between the PDO and atmospheric
384 variables, the North Pacific Index (NPI), volcanic events, a coral record from the Indian Ocean
385 and the Asian monsoon. They suggested that to minimise disagreement between regional
386 PDV signals, future attempts at reconstructing the PDO may be optimised by including both
387 Asian and North American data.

388 Shen et al. (2006) published a 530-year PDO reconstruction from 1470 using categorical
389 documentary drought/flood index records (a five-grade category index), from 28 of 58
390 available regions in eastern China (22-40°N, 110-122°E). The indices were derived from
391 Chinese historical documents (such as precipitation records and descriptions) by the China
392 Meteorological Administration (Zhang et al. 2003). Shen et al. (2006) used the 1925-1998 PDO
393 index of (Mantua et al. 1997) for calibration, and produced their PDO reconstruction using a
394 5-component Partial Least Squares (PLS) regression model. Their reconstruction explained
395 78% of the instrumental PDO variance, and spectral analysis showed a broad range of
396 frequency bands: 11-12, 23-28, 35-45, 50-70 and 75-115 year being statistically significant at the
397 5% or 10% level. The prominence of frequency bands fluctuated in spectral power throughout
398 the reconstruction period, with the 50-70 year band apparently appearing only since 1850.
399 Shen et al. (2006) suggested that lunar-solar tidal forcing contributed towards the pre-1850
400 oscillation of the PDO, and that since 1850 the dominant mode of oscillations of the PDO may
401 have been modulated by global warming caused by greenhouse gas forcing.

402 Linsley et al. (2008) used annual average oxygen isotope (d18O) time series from five Porites
403 lutea coral cores from Fiji and Tonga (16.8°-21.0°S, 179.2°-174.7°W) to produce their Fiji-Tonga
404 Interdecadal-Decadal Pacific Oscillation (F-T IDPO) index of low frequency Pacific climate
405 variability back to 1650. They exploited the expression of the IPO phenomenon on the low
406 frequency displacement of the SPCZ in coral isotopes in the South Pacific Convergence Zone
407 (SPCZ) salinity front region. The salinity front occurs due to oceanic circulation patterns and
408 a precipitation-evaporation imbalance, as identified by Gouriou & Delcroix (2002). Fresher
409 and warmer waters of the western Pacific lie to the west, and cooler, saltier waters lie to the
east of the salinity front. Coral d18O isotopes record the variation in these changes, with saltier water producing higher coral d18O values. The location of the SPCZ and the associated salinity front has been linked to ENSO (Gouriou & Delcroix 2002; Juillet-Leclerc et al. 2006; Linsley et al. 2006) and IPO variability (Folland et al. 2002). Linsley et al. (2008) used Singular Spectrum Analysis to resolve PDV in the 9-55 year band and computed their F-T IDPO index as the arithmetic mean of bandpass filtered d18O records. Their F-T IDPO index (from data south of the equator) was found to be anticorrelated (correlations of -0.41 and -0.51) and synchronous with two coral d18O interdecadal-decadal components from equatorial Pacific coral records, at Maiana and Palmyra, revealing a broader trend and providing some insight into oceanic mechanisms. Although extracted from data independent of the instrumental SST-based IPO index, the indices were highly correlated and in strong agreement regarding the timing of phase shifts. Linsley et al. (2008) noted distinct reductions in the amplitude of the interdecadal-decadal signal around 1685-1705 and 1740-1755 CE. Despite this apparent indication of the presence of 15-20 year neutral phases of the IPO, the F-T IDPO index also showed that a relatively regular progression of positive and negative IPO phases have occurred since 1650 in the SPCZ region, with a mean frequency of around 20 years and variance peaks near 11 and 35 years. Linsley et al. (2008) explained that this implied some degree of predictability in the IPO phenomenon and its proposed modulation of ENSO. A composite of South Pacific corals from Fiji, Tonga and Rarotonga back to 1791 by Linsley et al. (2015) revealed similar results, with relationships with the PDO and upper ocean heat content in the instrumental period. They found that decadal-scale changes in the South Pacific are a semiregular phenomenon with a mean period of around 25 years.

Verdon & Franks (2006) identified step changes in both IPO and PDO palaeoclimate and instrumental data sources. Their approach used the non-parametric Mann-Whitney U-test to identify differences between two halves of data in a 30-year moving window (Mauget 2003) in the reconstructions of (Biondi et al. 2001; D’Arrigo et al. 2001; Gedalof et al. 2002; MacDonald & Case 2005) and the coral record of (Linsley et al. 2000) at Raratonga. Verdon & Franks (2006) used the records to subjectively estimate the timing of PDO positive and negative phases, forming a composite categorical phase change PDO index for the period 1662-1992.

Mann et al. (2009) reconstructed global surface temperature fields over the period 500-2000 CE using a multiproxy palaeoclimate network of 1138 records. Their dataset was made up of 1036 tree-ring records, 3 marine sediment series, 14 speleothem series, 19 lacustrine series, 32 ice core series, 15 marine coral series and 19 historical documentary series. They used the regularised expectation maximisation (RegEM) climate field reconstruction (CFR) procedure of (Schneider 2001; Mann et al. 2007) where high-frequency (period < 20 years) and low-frequency (period > 20 years) components of the reconstruction are calibrated separately. The regional average temperature in region associated with the PDO (22.5N-57.5N, 152.5E-132.5W) was extracted from the surface temperature field reconstruction for the period 500-2000 to obtain their PDO index. There is a strong resemblance between their global mean temperature series and their PDO index. The authors note that the low value of their
truncation parameter $K$ may lead to greater apparent levels of similarity between regions than exists in the true spatial temperature pattern. They acknowledge that this likely leads to their calculated indices, including the PDO, exhibiting an artificially high level of similarity than their true underlying counterparts, particularly prior to 1600 CE (Mann et al. 2009).

The study by McGregor et al. (2010) merged 10 ENSO reconstructions into a unified ENSO proxy (UEP) using PCA to extract and reconstruct the joint features of the component reconstructions. McGregor et al. (2010) smoothed their UEP reconstruction to obtain an estimate of low frequency ENSO variability, finding similarity with the IPO and smoothed PDO indices in the instrumental period. They found an increase in ENSO variance since 1900, which could have been related to reduced signal-to-noise back in time, possibly related to dating uncertainties.

Henley et al. (2011) consolidated the 7 PDO and IPO reconstructions of Biondi et al. (2001), D’Arrigo et al. (2001), Gedalof & Smith (2001), MacDonald & Case (2005), D’Arrigo & Wilson (2006), Shen et al. (2006) and Linsley et al. (2008) to form a combined palaeo IPO index (CPIPO) over the period 1570-2000. They used an objective smoothing procedure to fit low pass filters to individual reconstructions, and combined these filtered series using an averaging scheme that weighted each series to the goodness of fit to instrumental data. They found that over the 440 years of the reconstruction the IPO had a broad frequency range, with a positively skewed distribution of phase durations varying between 3 and 33 years and a mean of 15 years. Similarly to McGregor et al. (2010) an increase in variance was observed since 1900.

Vance et al. (2015) developed a millennium-length reconstruction of the IPO over the period 1000-2003 CE using an annually resolved ice core record from Law Dome in Antarctica. They used two nonlinear multivariate regression methods (decision tree, DT and piecewise linear fit, PLF) to regress ice accumulation and log-transformed sea salt concentration onto the instrumental IPO index. Vance et al. (2015) observed a bias towards more positive IPO (23 positive compared to 13 negative) phases in their reconstruction, particularly prior to around 1200 CE, in contrast to the negative PDO tendency found by MacDonald & Case (2005) in that period. Vance et al. (2015) also reported an average positive phase duration of 14 years compared to 9 years for the negative phase.

3.2. Comparison of PDV reconstructions

Here I compare the temporal and spectral features of the 12 PDV reconstructions reviewed in section 3.1. Figure 5 shows the time series of low-pass reconstructions, for the period 1500-2000 CE, shown as shaded anomalies above and below zero, to aid visual comparison of the timing of PDV phase shifts. The unsmoothed time series are also shown where available.

The three major PDV phases in the twentieth century are captured fairly reliably by most of the reconstructions, with the exception of the Mann et al. (2009) PDO index, which is out of phase with all 11 other reconstructions during the negative and positive PDV phases in the
mid and late twentieth century. It is worth noting that this record was not directly calibrated to an instrumental PDV index, unlike all other reconstructions. There are large discrepancies between the palaeoclimate PDV reconstructions prior to the instrumental period, such that the ensemble could be colloquially termed ‘palaeo-spaghetti’. Table 2 shows the Spearman correlations between each pair of reconstructions during their common period of overlap, highlighting statistically significant correlations after adjusting for serial autocorrelation. The strongest correlation is between two of the North American tree ring reconstructions (Biondi et al. 2001; MacDonald & Case 2005). A total of 112 of the 144 pairwise correlations between smoothed series are below 0.2. The composite reconstructions (Henley et al. 2011; McGregor et al. 2010; Verdon & Franks 2006) have generally high correlation to the other members, as would be expected. Of the non-composites, the reconstructions of Vance et al. (2015), D’Arrigo et al. (2001) and Biondi et al. (2001) have the highest correlation to the other members. The Mann et al. (2009) index is the only reconstruction with a negative mean correlation with the other members, suggesting that it is not a reliable PDO reconstruction. The reconstructions which extend significantly into the first half of the millennium (MacDonald & Case 2005; Mann et al. 2009; Vance et al. 2015) bear little or no temporal resemblance to each other.

Figure 6 presents the power density spectra of each of the 12 reconstructions for the period 1700 CE to the most recent data available, using the multi-taper method (MTM, Thomson (1982). First order autoregressive models are fitted to unfiltered data where available, and their spectra added to the figure. The spectra reveal a wide range of frequency characteristics. Most reconstructions reveal very little high frequency variance at periods of <10 years, with the possible exceptions of Shen et al. (2006) with peaks 2, 3, 5, and 7-10 years, and D’Arrigo et al. (2001) at < 3 years. For low frequency variability, (Linsley et al. 2008) and (Gedalof & Smith 2001) have high variance in the 10-15 year band. Multidecadal variability in the 20-50 year band is observed commonly, however consistency in the spectral characteristics of the reconstructions is limited.

In summary, despite significant efforts over the past 20 or more years, palaeoclimate reconstructions don’t yet provide a consistent picture of the history of Pacific Decadal Variability in the past millennium, or even in the past 300-400 years when data availability is greatly increased. Discrepancies between reconstructions remain unresolved. These are likely due to: regional or continental limitations of most reconstructions in tandem with the non-stationarity of regional teleconnections (Gallant et al. 2013; Battehup et al. 2015), spectral biases of palaeo archives (Franke et al. 2013) and/or the particularities of reconstruction methods and data treatments. With these discrepancies, despite consistent evidence of pre-instrumental teleconnections with regional rainfall (Gergis & Henley 2016), there is insufficient coherent evidence across reconstructions and regions to be confident in the temporal history or dominant frequencies of PDV in the pre-instrumental period.
4. Tropical-extratropical interactions

Mechanisms of PDV have been widely studied, with several mechanisms proposed to explain PDV and particularly the high decadal variance in the extra-tropical Pacific Ocean. These mechanisms can mostly be grouped into two broad schools of thought with regards to the key driving force: tropical processes and extratropical processes. I note that several mechanisms remain plausible, including the phenomenon of slow westward propagating oceanic Rossby waves in the Pacific which have a time scale near a decade, as highlighted by Meehl & Hu (2006). It is possible that these Rossby waves interact with the meridional overturning cells. It also appears possible that anthropogenic aerosols have an influence on PDV, with increased aerosol levels potentially causing a negative PDV response and a future decrease in anthropogenic aerosols therefore potentially leading to a positive PDV phase (Smith et al. 2016). I refer the reader to the review of the dynamics of interdecadal climate variability by Liu (2012) for a historical examination of proposed mechanisms.

Here I draw attention to a possible PDV process that couples the tropics and extratropics via both atmospheric and oceanic responses. The process is a bi-hemispheric extension of the mechanism introduced by Farneti et al. (2014). Figure 7 schematically illustrates the mechanism, combining diagrammatic elements adapted from Farneti et al. (2014), Lu et al. (1998) and England et al. (2014). Following the numbered items in the figure, the mechanism of is as follows:

1. Tropical SST anomalies commence in their anomalously cool phase, for example, as observed in the recent negative PDV phase.
2. The atmospheric response to these cooler tropical SSTs is a weakening of the thermally-driven rising branch of the Hadley circulation, resulting in a slowdown and equatorward meridional migration of the Hadley cells (the atmospheric ‘bridge’).
3. This leads to a reduction in the strength of the extratropical trade winds, and the generation of a wind stress curl of anomalous sign to climatology. Note that an equatorward meridional migration of the Hadley cells leads to an increase in the zonal winds within 10° of the equator, consistent with the findings of England et al. (2014).
4. The Subtropical Gyres (STG) are weakened by these anomalous winds, through reduced Ekman transport and subsequently reduced downwelling, resulting in reduced equatorward mass flux in the subtropical meridional overturning cells (STC) (the oceanic ‘tunnel’).
5. This reduced STC mass flux also transmits a reduced meridional heat transport and is associated with a reduction in the eastward flowing equatorial undercurrent (EUC, not shown).
6. Equatorial upwelling is therefore reduced, generating a warm SST anomaly at the equator, which is of opposite sign to the original cool SST anomaly. Thus, a negative feedback is induced in the surface layer in the tropics on decadal timescales.
Most studies to date have sought evidence for a single tropical or extratropical forcing process for PDV, and many studies highlight the role of stochastic atmospheric forcing. Some studies have used simple stochastic characterisations of proposed causal links between ENSO and PDV, however similarly simple hierarchical models that propose a PDV forcing of interannual variability might be equally well validated by observational data. Despite substantial advances in recent decades, we currently lack sufficient data and dynamical evidence to conclusively reject any of the major proposed mechanisms. Statistical verification of a particular model scheme, modelling evidence of isolated mechanisms and the reporting of lead-lag correlations do not guarantee the validity of a causal dynamical relationship. A number of causal mechanistic PDV arrangements therefore remain plausible.

The mechanism outlined here suggests that the cool and warm phases of PDV are not driven by tropical or extratropical variability alone, but are instead the result of coupled and continuous tropical-extratropical interactions on decadal timescales, with a continuum of responses from positive to negative PDV phases, influenced strongly by the tropical influences of ENSO. This interpretation is consistent with the findings of Newman et al. (2016) that there are a number of separate northern and southern hemisphere ocean and atmosphere processes that contribute to differences between the hemispheric responses, such as the Kuroshio current and the Aleutian Low for the northern hemisphere. But, it is plausible that the similarities between northern and southern hemisphere PDV patterns arise from basin-wide quasi-symmetric interactions between the tropics and extratropics, in which the tropics would act as a strong synchronising mechanism for both hemispheres. This mechanism contributes towards an explanation of Pacific-wide coherent patterns of SST variability on decadal timescales.

5. Conclusions and Future Research

Decadal variability plays a critical role in the earth’s climate system. Given the association between Pacific decadal variability (PDV) and global mean surface temperature (Meehl, Hu, Santer, et al. 2016), and the high contribution of internal variability to projections of near-term climate (Hawkins & Sutton 2009), PDV is a major potential avenue for constraining climate projections and quantifying non-stationary risks. It is also crucial that we improve our understanding and resilience against severe and persistent multidecadal phenomena such as megadroughts (Ault et al. 2016).

In this article I reviewed the patterns and timeseries of PDV and found strong similarities between the IPO, PDO and SPDO, which all express high variance in the extratropical Pacific Ocean. The major PDV phase shifts in the observed period are near identical for the PDO and SPDO, which also exhibit strong hemispheric symmetry in their spatial patterns, despite their data regions having no geographical overlap. This is suggestive of a common element for these major shifts. Acknowledging the known dynamical links between PDV and ENSO, I find ample distinction between PDV and Niño 3.4 (both patterns and timeseries) to conclude that a smoothed Niño 3.4 index is not a sufficient substitute for the indices of PDV.
I develop a tool to provide uncertainty quantification of the recent state of PDV, using the TPI (Henley et al. 2015) as an example. I use the tool to provide reliable estimates of IPO state up to 2 years prior to the present, finding that there is an 80-90% probability that the IPO remained in its negative state until 2014-2015.

I then turn to the pre-instrumental period and review annual resolution palaeoclimate reconstructions of PDV over the last 1000 years, documenting their source data, reconstruction methods and key findings. I then compare the reconstructions in the time and frequency domains, finding that the reconstructions depict a large range of temporal and spectral features, and have generally poor agreement. This reduces the evidence for either: the quality of regionalised palaeoclimate reconstructions as recorders of large-scale modes, their consistency across archive types and methods, stationarity in regional teleconnections to PDV and/or the existence of spatially and temporally coherent PDV in the past millennium.

Additional future efforts in data collection, including the development of new proxy data methods, and reconstruction, are critical to resolving these issues. This is particularly true in the Southern Hemisphere, where palaeoclimate data is much sparser than in the Northern Hemisphere.

I then draw the reader’s attention to a Pacific-wide tropical-extratropical mechanism that suggests that the cool and warm phases of PDV are not driven by tropical or extratropical variability alone, but are instead the result of continuous tropical-extratropical interactions on decadal timescales. It is conceivable that the similarities between northern and southern hemisphere PDV patterns arise from basin-wide quasi-symmetric interactions between the tropics and extratropics, in which case the tropics act as a strong synchronising mechanism for both hemispheres. I therefore reason that a range of mechanistic PDV arrangements remain plausible.

There continue to be substantial challenges to research in this field. A significant challenge with the study of PDV is that a high level of expertise is required in multiple fields such as palaeoclimatology, oceanography and atmospheric science. Most palaeoclimatologists are experts in one proxy and its meticulous collection, but have less expertise on large scale climate dynamics. Most dynamical experts and climate modellers are focussed on shorter timescales and don’t have a comprehensive understanding of the utility, purpose, uncertainty and signal in palaeoclimate proxy records. Such are the challenges of many open questions in science and society – multidisciplinary coordination and expertise is a critical feature for future success.

Another significant challenge to the comprehensive picture of PDV is the relatively short and hemispherically asymmetric coverage of observations. In particular, our observations of the intermediate depth ocean are very sparse and short prior to the availability of sub-700m depth ARGO ocean floats in 2005 (Henley et al. 2017). Progress on the synthesis of depth profile data into gridded subsurface datasets (Good et al. 2013; Schmitdko et al. 2013) will
continue to inform our understanding of PDV. The consistency, quality, continuation and expansion of the observational networks that underpin these analyses is of high importance.

There remain extensive opportunities for improvements in, and renewed efforts in, model development targeting PDV and decadal variability more generally. I would recommend moving beyond EOF patterns, semantics and isolated mechanisms and towards a coherent, focussed and more systematic assessment of the mechanisms of PDV using multiple approaches. Significant opportunities exist in the identification of trigger mechanisms for PDV shifts (Meehl & Teng 2014; Meehl et al. 2015), and more comprehensive modelling studies that consider interactions between the tropics and extratropics, rather than searching for a single driving mechanism acting in isolation.

Although the slowdown in global warming was widely reported, too little attention has been given to the alternating attenuation and amplification effects of decadal modes on global surface temperatures. Natural variability has cushioned the rise in global temperature since around 2000, and an acceleration in the rate of global warming is likely in the coming decades. Although the slowdown in GMST did not surprise most scientists, it had a detrimental impact on the public perception of our ability to project global temperature. It gave fodder to science contrarians and conspiracy theorists, and temporarily dented the wider public credibility of science. Recall also that the slowdown was at least partly due to PDV. Hence, one might infer that a better understanding of PDV has an impact on the public credibility of science as a whole.

Given the global implications of decadal to multidecadal climate variability on human and natural systems, for scientists and managers in climate, water resources, earth science, palaeoclimatology, geography, biology, ecology, oceanography and atmospheric science, it is critical that we bolster efforts to understand decadal climate variability. Pacific decadal variability should therefore be a critical focus area for several disciplines in the climate sciences. In the coming years and decades, we need a bolstering of the efforts to understand, model and measure the past and future trajectory of Pacific decadal climate variability and its impacts on the future climate.
### Table 1. Summary of definitions, datasets and analysis methods of IPO and PDO indices, following Henley et al. (2015)

<table>
<thead>
<tr>
<th>Authors and year</th>
<th>Primary Base Dataset(s)</th>
<th>Spatial resolution</th>
<th>Domain</th>
<th>Time resolution</th>
<th>Period of Analysis</th>
<th>Filtering</th>
<th>Method</th>
<th>Index Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mantua et al. (1997)</td>
<td>HSSTD / OISST</td>
<td>5° x 5° / 1° x 1°</td>
<td>Poleward of 20°N</td>
<td>Monthly</td>
<td>1900-1993</td>
<td>None</td>
<td>PCA</td>
<td>PC1 of SST over Pacific domain</td>
</tr>
<tr>
<td>Zhang et al. (1997)</td>
<td>COADS / HSSTD</td>
<td>2° x 2°, aggregated to 4° x 6°</td>
<td>Poleward of 20°N</td>
<td>Monthly</td>
<td>1900-1993</td>
<td>6-yr low pass</td>
<td>PCA</td>
<td>PC1 of low pass filtered SST over Pacific domain</td>
</tr>
<tr>
<td>Folland et al. (1999)</td>
<td>MOHSS16C/NMAT (Parker et al. 1995)</td>
<td>Equal areas, 10° x 12° at equator</td>
<td>Global</td>
<td>Seasonal</td>
<td>1861-1996</td>
<td>13-yr low pass</td>
<td>PCA</td>
<td>Projection of unfiltered MOHSS16C onto low-pass filtered EOF2 for 1861-1996, then filtered to obtain low frequency version</td>
</tr>
<tr>
<td>Power et al. (1999)</td>
<td>MOHSST6 / NMAT</td>
<td>Various</td>
<td>Global and North Pacific</td>
<td>Seasonal and Annual</td>
<td>1856-1998</td>
<td>13-yr low pass</td>
<td>PCA</td>
<td>PC1</td>
</tr>
<tr>
<td>Parker et al. (2007)</td>
<td>HadCRUT3 Rayner et al. (2006)</td>
<td>Equal areas, 10° x 12° at equator</td>
<td>Global</td>
<td>Seasonal</td>
<td>1850-2006</td>
<td>11-yr low pass Chebyshev</td>
<td>PCA</td>
<td>Projection of unfiltered HadSST3 onto low-pass filtered EOF2 for 1891-2001, then filtered to obtain low frequency version</td>
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</table>

*Principal Component Analysis*
Table 2. Correlations between IPO and PDO reconstructions. Spearman correlations computed for full overlapping period between paired IPO and PDO reconstructions; p-value shown in parentheses is adjusted for serial autocorrelation; bold entries indicate statistical significance at the 5% level.

<table>
<thead>
<tr>
<th>Recon</th>
<th>Bion01</th>
<th>Geda01</th>
<th>DArr01</th>
<th>MacD05</th>
<th>Shen06</th>
<th>DArr06</th>
<th>Verd06</th>
<th>Lins08</th>
<th>Mann09</th>
<th>McGr10</th>
<th>Henl11</th>
<th>Vanc15</th>
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<tbody>
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<td>0.16 (0.00)</td>
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<td>0.33 (0.00)</td>
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<td>0.00 (0.97)</td>
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</tbody>
</table>
Figure 1. Trend and persistence in global sea surface temperature for 1900-2016. a) Trend in global SST (linear regression coefficient), b) average bandpower in the 2-7 year and c) 7-30 year frequency bands and d) ratio of bandpower between the two bands.
Figure 2. Pacific SST regression patterns and timeseries 1900-2016. a) IPO (TPI) pattern and b) timeseries, c) PDO pattern and d) timeseries, e) SPDO pattern and f) timeseries, g) Niño 3.4 pattern and h) timeseries; Patterns are the regression of global SST onto the unfiltered indices; unfiltered timeseries are shown as shaded anomalies; smoothed series use a 13-year Chebyshev low-pass filter.
Figure 3. Global mean surface temperature anomaly and the IPO (1900-2016). Observed annual global mean surface temperature anomaly (red bars, HadCRUT4, 1850-1900 baseline); Major decadal IPO negative and positive periods are shown as alternating pink (IPO positive) and white (IPO negative) bands, adapted from England et al. (2014) and Henley & King (2017).
Figure 4. A scheme to quantify the IPO state in the recent past. a) Unfiltered TPI timeseries (blue, 1990-2017) and an ensemble distribution of the low-pass filtered TPI (red, 1990-2017) using observed data for the past values and stochastic simulations for the future. Uncertainty is represented using 1000 replicates of a monthly AR(5) model (order optimised using Schwarz’s Bayesian Criterion; Schwarz 1978) conditioned on the most recent data available; a 13-year low-pass Chebyshev filter is applied across the past-future boundary; b) Brier skill score for the IPO state estimation scheme in each year relative to the current year, skill is computed for the period 1900-2000; positive values represent skill in the probability estimate relative to random chance (note the influence of serial dependence; Wilks, 2010); c) Probability of IPO states 2006-2017 using the IPO state estimation scheme conditioned on the most recent TPI data available.
Figure 5. Palaeoclimate reconstructions of the PDO and IPO for 1500-2000. Low frequency reconstructions shown in blue and yellow shaded anomalies (around zero line); unsmoothed series shown in black where available; Reconstructions based on North American tree rings (Biondi et al. 2001; Gedalof & Smith 2001; D’Arrigo et al. 2001; MacDonald & Case 2005), Asian tree rings (D’Arrigo & Wilson 2006), Documentary records (Shen et al. 2006), multiproxy composites of site records or reconstructions (Verdon & Franks 2006; Mann et al. 2009; McGregor et al. 2010; Henley et al. 2011), South Pacific Coral ε18O (Linsley et al. 2008) and the Law Dome ice core from Antarctica (Vance et al. 2015)
Figure 6. Power spectra of IPO and PDO reconstructions (1700 CE - recent). Multi-taper method (MTM) power spectra and 95% confidence intervals of IPO and PDO reconstructions (red/pink); AR(1) spectra and 95% confidence intervals shown for reconstructions where unfiltered indices are available (orange solid and dashed);
Figure 7. Schematic of a PDV mechanism with tropical-extratropical interactions on decadal timescales. Idealised Hadley cell is shown in yellow, ocean surface layer gyres in blue: Subtropical Gyre (STG), Subpolar Gyre (SPG) and Antarctic Circumpolar Current (ACC); Vertical overturning cells in the ocean shown in green: Subtropical Cell (STC), Subpolar Cell (SPC); diagram is an adaptation from Farneti et al. (2014), England et al. (2014) and Lu et al. (1998);
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