A Millennial Proxy Record of ENSO and Eastern Australian Rainfall from the Law Dome Ice Core, East Antarctica

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(Manuscript received 22 December 2011, in final form 13 July 2012)

ABSTRACT

ENSO causes climate extremes across and beyond the Pacific basin; however, evidence of ENSO at high southern latitudes is generally restricted to the South Pacific and West Antarctica. Here, the authors report a statistically significant link between ENSO and sea salt deposition during summer from the Law Dome (LD) ice core in East Antarctica. ENSO-related atmospheric anomalies from the central-western equatorial Pacific (CWEP) propagate to the South Pacific and the circumpolar high latitudes. These anomalies modulate high-latitude zonal winds, with El Niño (La Niña) conditions causing reduced (enhanced) zonal wind speeds and subsequent reduced (enhanced) summer sea salt deposition at LD. Over the last 1010 yr, the LD summer sea salt (LDSSS) record has exhibited two below-average (El Niño-like) epochs, 1000–1260 AD and 1920–2009 AD, and a longer above-average (La Niña-like) epoch from 1260 to 1860 AD. Spectral analysis shows the below-average epochs are associated with enhanced ENSO-like variability around 2–5 yr, while the above-average epoch is associated more with variability around 6–7 yr. The LDSSS record is also significantly correlated with annual rainfall in eastern mainland Australia. While the correlation displays decadal-scale variability similar to changes in the interdecadal Pacific oscillation (IPO), the LDSSS record suggests rainfall in the modern instrumental era (1910–2009 AD) is below the long-term average. In addition, recent rainfall declines in some regions of eastern and southeastern Australia appear to be mirrored by a downward trend in the LDSSS record, suggesting current rainfall regimes are unusual though not unknown over the last millennium.

1. Introduction

The El Niño–Southern Oscillation (ENSO) is the dominant source of interannual climate variability. ENSO is characterized by a zonal seesaw in SST and atmospheric pressure in the tropical and subtropical Pacific and causes interannual climate and weather anomalies far beyond the Pacific margin. Observed changes to ENSO behavior in recent decades (e.g., Federov and Philander 2000; Ashok et al. 2007; Yeh et al. 2009) are difficult to assess in the context of either anthropogenic climate change or natural variability, as instrumental records for ENSO cover at most 150 yr. In the Australian region, declines in rainfall have been observed in recent decades in southwest Western Australia, southeastern Australia, and parts of

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DOI: 10.1175/JCLI-D-12-00003.1

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the eastern seaboard (e.g., Cai and Cowan 2006; Nicholls 2009; Cai et al. 2010; Gergis et al. 2011). It is also difficult to assess from short (100 yr) Australian rainfall records whether these declines, particularly in the ENSO-sensitive regions of eastern Australia, are due to ENSO-related natural variability or anthropogenic climate change.

To assess whether these phenomena are due to natural or anthropogenically induced variability, centennial-scale, annually resolved proxy records of ENSO are needed. The spatial signatures of individual ENSO events differ, so individual proxies can indicate varied responses to the same event. In the Pacific region, proxy records of regional climate and ENSO frequently disagree on past ENSO variability or whether different eras were characterized by predominantly warm (El Niño) or cold (La Niña) events. It is also unclear how ENSO expression varied through periods of well-known climate variability such as the Medieval Climate Anomaly (MCA; 800–1300 AD) and the Little Ice Age (LIA; 1500–1850 AD) [see, e.g., Khider et al. (2011) and Mann et al. (2009) versus Yan et al. (2011)]. Other studies suggest different factors, such as shifts in the position of the ITCZ, have affected climate variability in the tropical Pacific during the last millennium (e.g., Newton et al. 2006; Sachs et al. 2009). Using multiple diverse and spatially distributed records from tree rings, ice cores, and corals increases the reliability of ENSO reconstructions (Cobb et al. 2003; Braganza et al. 2009; Mann et al. 2009; McGregor et al. 2010). As understanding of low- to high-latitude Pacific Ocean teleconnections evolves, investigators are beginning to look to the data-scarce but potentially proxy-rich Antarctic high-latitude regions for ENSO signatures.

Most studies of Pacific extratropical teleconnections to high southern latitudes have focused on West Antarctica, where the extratropical signal is relatively strong and has been linked to variability in surface temperatures, precipitation, sea level pressures, and sea ice extent and thickness (e.g., Cullather et al. 1996; Bromwich et al. 2000; Kwok and Comiso 2002; Meyerson et al. 2002; Bertler et al. 2006; Yuan and Li 2008; Ding et al. 2011; for a review, see Turner 2004). In contrast, only a few studies have noted East Antarctic climate anomalies linked to tropical Pacific variability, including ENSO-related variability. Smith and Stearns (1993) observed an abrupt switch in sign of surface temperature and pressure anomalies across SOI minima along the Wilkes Coast region of East Antarctica. In conjunction with previous studies (Trenberth 1980; Mo et al. 1987), these authors suggested that these abrupt changes resulted in cold air outflows that affected blocking in the New Zealand region and ultimately the Southern Oscillation. L’Heureux and Thompson (2006) demonstrated a significant interaction between ENSO and the southern annular mode (SAM) during summer, showing the effect of ENSO is circumpolar and not confined to the South Pacific and West Antarctica.

Here, we describe a teleconnection linking ENSO with a seasonally resolved sea salt record from the summit of Law Dome (LD), East Antarctica. To date, studies of sea salts at LD have focused on annual or early winter sea salt [May–July (MJJ)] links to regional climate variability, in particular midlatitude wind speeds (Souney et al. 2002; Goodwin et al. 2004; Mayewski et al. 2004). In summer, midlatitude wind speeds (and sea salt deposition at LD) are greatly reduced; however, a clear correlation between LD summer sea salts, ENSO, and eastern Australian rainfall exists over the instrumental period. This study investigates a mechanism linking the equatorial western Pacific and Law Dome and explores changes in ENSO and rainfall variability in eastern Australia using a 1010-yr record of LD summer sea salts.

2. Methods and datasets

a. LD and sea salts

LD is a small coastal ice cap in Wilkes Land, East Antarctica. The local climate at the “Dome Summit South” (DSS) ice core site (66°46’11”S, 112°48’25”E, elevation 1370 m) is predominantly maritime with minimal katabatic influence and a generally southeasterly wind direction. DSS features high annual snowfall and relatively low mean wind speeds and surface temperatures (0.7 m ice-equivalent per year, 8.3 m s$^{-1}$, and $-21.6^\circ$C respectively) (Morgan et al. 1997; Palmer et al. 2001).

Sea salts preserved in ice core records are largely the result of wind speed–dependent aerosol injection from the ocean (bubble bursting), with some long-range transport effects (Gong et al. 1997). At LD, sea salt concentrations are related to large-scale, oceanic source strength features rather than local, meridional, or depositional factors. Annual LD sea salt concentrations are related to the strength of the large-scale circulation system over the Antarctic continent in winter (the Antarctic high) and its effect on the latitudinal belt occupied by the circumpolar trough (Souney et al. 2002). On a seasonal scale, early winter (MJJ) sea salts at LD are related to the midlatitude mean sea level pressure (MSLP) field in the southern Indian and southwest Pacific Oceans (Goodwin et al. 2004).

Brine crystals growing on sea ice (frost flowers) are another source of sea salt aerosols. Frost flower growth requires calm conditions, cold temperatures, and the formation of new sea ice; thus, sea-spray inputs from wind ablation of these structures are primarily significant during early winter (Rankin et al. 2000; Roscoe et al. 2011). New sea ice may form in the LD region in
March during cold outbreaks, although generally the sea ice growth season in this region is from April onward (R. Massom 2012, personal communication). Frost flower growth may conceivably contribute to sea salt aerosols at LD during the extended winter (December–March (DJFM)] considered here, with potential rare early inputs in March; however, we assume the bulk of the contribution is from open water. In any case, a sea salt contribution from frost flowers in the summer LD record is a result of wind speed–dependent ablation and is thus integrated into the observed wind proxy.

In coastal regions such as LD (within 200 km of the coast and 2000 m MSL) sea-spray atmospheric scavenging is dominated by wet deposition and not overly affected by the local orography (Benassai et al. 2005). The seasonal salt cycle from DSS has a flattened maximum during winter and a period of low concentrations during DJFM as a result of the seasonal frequency, position, and intensity of synoptic-scale cyclonic events and regional variations in wind speeds (Curran et al. 1998; Curran and Palmer 2001). Sea salt concentrations at LD have a long-tailed distribution. In this study, we use log-transformed values which give approximately normally distributed data and improved sensitivity to the typically low-level summer variability we are concerned with here.

b. Analysis and dating of the LD sea salt record

The high accumulation rate at LD allows sampling (and therefore dating) at subannual resolution. The record used in this study was compiled from a six-core composite chemistry record updated with short cores drilled over the last two decades to maximize overlap with the instrumental and satellite era. The LD \(^{18}\)O record is continuous and is the primary means of year margin dating via annual layer counting; trace chemistry species were used for confirmation of dating if necessary (see van Ommen and Morgan 1997; van Ommen et al. 2004; Plummer et al. 2012). Cleaned, high-resolution (2.5–5 cm) samples were analyzed for trace chemical species including sea salts via trace ion chromatography (see Curran et al. 1998; Curran and Palmer 2001; Plummer et al. 2012).

c. Instrumental climate data and LD sea salt analyses

We initially produced an instrumental period sea salt record with monthly resolution from the dated LD chloride ion (Cl\(^{-}\)) data spanning 1889–2009. The raw data record has an average temporal resolution of 19 Cl\(^{-}\) samples per year, but this varies depending on the sample size (sampling frequency) and annual layer thickness. To facilitate analysis, we resampled the raw record to 12 monthly values per year. The monthly record was log transformed and then seasonal correlations with ENSO and other climate indices were investigated by first averaging to summer (DJF) and extended winter (April–November) single value per year records. The two records were uncorrelated with each other, and only the summer record showed a significant correlation with annual (May–April) Southern Oscillation index [SOI; Australian Bureau of Meteorology (BOM)]. We then produced variations of the summer record over 2–4-month averaged bands from September to March for correlation analysis with similarly averaged records of the SOI; the four Niño SST regions of the equatorial Pacific, Niño-4 (5°N–5°S, 150°W–160°E), Niño-3.4 (5°N–5°S, 120°–170°W), Niño-3 (5°N–5°S, 90°–150°W), and Niño-1 + Niño-2 (0°–10°S, 80°–90°W) [the second Hadley Centre Sea Surface Temperature dataset (HADSST2); Rayner et al. 2006]; and the SAM index (Marshall 2003). Lead/lag correlations showed the DJFM sea salt series [the LD summer sea salt (LDSSS) record] to have the highest correlation with the ENSO indices (Table 1a), although correlations generally declined gradually as leads/lags were varied either side of those stated in Table 1a. We also investigated how the LDSSS record was related to local winds and wind directions in the Casey Station/Law Dome region. We found little to suggest that local meridional winds or depositional factors affected summer salt deposition, confirming the findings of Souney et al. (2002) that the sea salt record at LD is a broadscale, integrated zonal signal.

We also compared the LDSSS record to a SH blocking index (BI) spanning 1948–2009 (Pook and Gibson 1999; M. Pook 2011, personal communication). This index is calculated seasonally for every 2.5° longitude as follows:

\[
BI = 0.5(U_{25} + U_{30} + U_{55} + U_{60} - U_{40} - U_{50} - 2U_{45}),
\]

where, at a given meridian, \(U_X\) represents the zonal component of the mean 500-hPa geostrophic wind at latitude \(X\) (Pook and Gibson 1999). Analyses for each season (and extended summer season, DJFM) were used to identify longitudinal bands of high correlation with the LDSSS record. Where high correlations occurred, the BI was averaged across these longitudinal bands to produce the single correlation coefficients presented in Table 1.

All correlation coefficients in this study were calculated on detrended time series with bootstrap confidence intervals according to Mudelsee (2003). Levels of significance were determined from the number of effective degrees of freedom \(N^{eff}\) calculated from lag-1 autocorrelation of each series according to the method of Bretherton et al. (1999). Rainfall data are from the BOM high-quality station network (http://www.bom.gov.au/climate/change/datasets/datasets) and the Australian Water Availability Project (AWAP). An interdecadal Pacific oscillation
### Table 1. (a) Seasonal correlations between detrended LDSSS and ENSO indices from 1889–2009, longitudinal sectors of a SH blocking index (1948–2009; Pook and Gibson 1999), and the SAM index (Marshall 2003) are shown. (b) ENSO correlations using full-year averages (May–April) are shown. Correlations significant at 99% are in bold, at 95% are shown in normal type, and <95% are shown in italics. Errors are bootstrap confidence intervals (Mudelsee 2003) and effective degrees of freedom N_{eff} are calculated from lag-1 autocorrelation of both series.

**(a) Seasonal corr Index (detrended)**

<table>
<thead>
<tr>
<th>Index</th>
<th>Range</th>
<th>r, lead/lag (months)</th>
<th>95% CI</th>
<th>Significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>SOI (NDJF)*</td>
<td>1889–2009</td>
<td>0.336, 1</td>
<td>[0.201; 0.456]</td>
<td>&lt;0.0001, N_{eff} = 121</td>
</tr>
<tr>
<td>Niño-4 (SO)*)</td>
<td>1889–2009</td>
<td>−0.300, 3</td>
<td>[−0.447; −0.105]</td>
<td>0.001, N_{eff} = 118</td>
</tr>
<tr>
<td>Niño-3.4 (SO)*</td>
<td>1889–2009</td>
<td>−0.289, 3</td>
<td>[−0.428; −0.107]</td>
<td>0.001, N_{eff} = 122</td>
</tr>
<tr>
<td>Niño-3 (ND)*</td>
<td>1889–2009</td>
<td>−0.246, 1</td>
<td>[−0.375; −0.055]</td>
<td>0.007, N_{eff} = 121</td>
</tr>
<tr>
<td>Niño-1 + Niño-2 (DJFM)</td>
<td>1951–2009</td>
<td>−0.289, 0</td>
<td>[−0.508; −0.084]</td>
<td>0.03, N_{eff} = 59</td>
</tr>
<tr>
<td>BI DJFM 150°–170°W</td>
<td>1948–2009</td>
<td>−0.267, 0</td>
<td>[−0.481; −0.021]</td>
<td>0.04, N_{eff} = 62</td>
</tr>
<tr>
<td>BI SON 90°–170°W*</td>
<td>1949–2009</td>
<td>−0.402, 3</td>
<td>[−0.574; −0.197]</td>
<td>0.002, N_{eff} = 60</td>
</tr>
<tr>
<td>SAM (DJFM)</td>
<td>1958–2009</td>
<td>0.264, 0</td>
<td>[0.007; 0.438]</td>
<td>0.07, N_{eff} = 51</td>
</tr>
<tr>
<td>SOI (NDJF)*</td>
<td>1958–2009</td>
<td>0.444, 1</td>
<td>[0.251; 0.606]</td>
<td>0.001, N_{eff} = 52</td>
</tr>
</tbody>
</table>

*(Index leads LDSSS.)*

**(b) ENSO corr Index (detrended)**

<table>
<thead>
<tr>
<th>Index</th>
<th>Range</th>
<th>r</th>
<th>95% CI</th>
<th>Significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>SOI (May–April)</td>
<td>1889–2009</td>
<td>0.305</td>
<td>[0.156; 0.426]</td>
<td>0.0008, N_{eff} = 120</td>
</tr>
<tr>
<td>Niño-4 (May–April)</td>
<td>1889–2009</td>
<td>−0.289</td>
<td>[−0.431; −0.086]</td>
<td>0.002, N_{eff} = 117</td>
</tr>
<tr>
<td>Niño-3.4 (May–April)</td>
<td>1889–2009</td>
<td>−0.287</td>
<td>[−0.421; −0.091]</td>
<td>0.002, N_{eff} = 120</td>
</tr>
<tr>
<td>Niño-3 (May–April)</td>
<td>1889–2009</td>
<td>−0.240</td>
<td>[−0.378; −0.036]</td>
<td>0.009, N_{eff} = 120</td>
</tr>
<tr>
<td>Niño-1 + Niño-2 (May–April)</td>
<td>1951–2009</td>
<td>−0.161</td>
<td>[−0.389; 0.156]</td>
<td>0.24, N_{eff} = 58</td>
</tr>
</tbody>
</table>

(IPO) index was used to investigate decadal to multidecadal variability in the observed correlations between ENSO, LD, and Australian rainfall using an austral summer centered, single annual value time series from HADSS2 observations (Parker et al. 2007). The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kistler et al. 2001) was used to investigate SST, wind, and geopotential height (GPH) anomalies associated with the relationship between LD and ENSO. All reanalysis datasets have high-latitude data problems related to data scarcity and trend effects in the SH, particularly prior to 1979 (Hines et al. 2000; Kistler et al. 2001). Data coverage improves greatly after 1979; thus, we used composites from 1979 to 2009 only and assess anomalies in broad terms only.

**d. Producing the 1010-yr LDSSS record**

Given the DJFM series had the highest correlation with ENSO indices over the instrumental period, this method was extended to produce a full 1010-yr LDSSS record. As resolution (number of samples per year) decreases slightly with depth/age, we incorporated the following checks for resolution fidelity over the full 1010-yr record. To check if sea salt concentrations from shoulder seasons contributed to overestimation of summer Cl− by leaking into the DJFM period in low resolution/accumulation years (i.e., <12 samples per year), the full raw dataset (which averaged 12 samples per year) was resampled down to six samples per year. Comparison of this artificial low accumulation record with the original confirmed no significant effect of low accumulation years at the sample densities used.

Small physical gaps in the Cl− record during summer, even in high accumulation years, could also have an overcontribution from spring/autumn concentrations. Where more than one of the four specified summer months (DJFM) was not represented in the raw Cl− record we opted to not use the averaged summer value, which resulted in 37 missing summers in the final 1010-yr LDSSS record. No missing summers occurred within the defined instrumental period or in more than three consecutive years. Missing summers were filled using linear interpolation for spectral analysis.

Average summer (DJFM) Cl− concentrations (in μM) over the 1010-yr LD record are significantly lower than April–November concentrations (cf. 1.73 ± 0.84 μM and 4.38 ± 1.35 μM, respectively), and there is no correlation between the summer and winter records. The summer concentrations are also well within analytical detection limits (0.006 μM) (Curran and Palmer 2001).

**3. Instrumental period observations**

**a. ENSO: 1889–2009**

The LDSSS record is significantly correlated (anti-correlated for SST) with the SOI as well as SST in the Niño-4, Niño-3.4, Niño-3, and Niño-1 + Niño-2 regions. At maximum correlation LDSSS lags SOI (r = 0.336; P < 0.0001; N_{eff} = 121) by 1 month and the four Niño SST indices by up to 3 months (Table 1a). The strength of the
anticorrelation between LD<sub>SSS</sub> and the Niño SST regions dissipates gradually from west to east across the Pacific and the lag shortens, suggesting the progenitor of the mechanism linking ENSO and LD<sub>SSS</sub> is the area of the central-western equatorial Pacific (hereafter referred to as CWEP) containing the Niño-4 SST region (5°N–5°S, 170°W–160°E; depicted in Fig. 2). Correlation values using annualized (May–April) ENSO indices show generally similar but lower r values and significance (Table 1b) except for the Niño-1 + Niño-2 region, which on annual values is not significantly correlated with LD<sub>SSS</sub> (r = −0.161; P = 0.24; N<sub>eff</sub> = 58), indicating the seasonal nature of ENSO anomalies and event maturation. The seasonal lags identified assisted with understanding the nature of the mechanism connecting the CWEP and LD outlined in section 4.

b. SH blocking: 1948–2009

Atmospheric blocking is a significant feature of SH circulation, particularly in the South Pacific southwest of New Zealand (Renwick 1998; Pook and Gibson 1999). Broad regions of the South Pacific show increased frequency of blocking during ENSO warm events, particularly in austral spring and summer. In the southwest Pacific, ENSO warm events coincide with suppressed blocking during June–August (JJA) and then enhanced blocking during September–November (SON) (Renwick 1998). Blocking has a major influence on where cyclones cross the Antarctic coast, with subsequent precipitation effects as relatively warm, moist air is advected toward the interior of the Antarctic continent (Pook and Gibson 1999; Massom et al. 2004). It is unknown at this point what effect blocking-related incursions have on precipitation at LD or whether there is any related sea salt signal. Massom et al. (2004) suggest there is a major contribution to precipitation at low accumulation inland sites in East Antarctica related to blocking in the South Tasman Sea (140°–150°E) and suggests there may be some precipitation effect at coastal sites including LD.

However, we do not see a blocking signature for the South Tasman region (140°–150°E) in the LD<sub>SSS</sub> record. This does not rule out a South Tasman blocking signature in LD sea salts in other seasons. There is, however, a small correlation between summertime blocking in the southwest Pacific southeast of New Zealand (150°–170°W) and the LD<sub>SSS</sub> record, which is discussed further in section 4 (Table 1). Also, a much stronger correlation is observed with springtime blocking in a broader region (90°–170°W) across the South Pacific, consistent with the observations of Renwick (1998) of ENSO-forced variability in zonal winds and atmospheric heights in this region (Table 1a). The 3-month lag between this springtime blocking and LD<sub>SSS</sub> suggests this observation is related to the transmission of the ENSO signal to LD via the South Pacific.

c. ENSO event years

The LD<sub>SSS</sub> record, like any paleoclimate proxy, is noisy (Fig. 1). High and low summer sea salt years can occur because of factors unrelated to ENSO. The two highest summer sea salt concentrations in the instrumental LD<sub>SSS</sub> record occur during ENSO neutral years (1891 and 1901), while average LD<sub>SSS</sub> values can occur in very high SOI years. Numerous events stand out as low summer sea salt years (e.g., the 1940/41, 1941/42, 1982/83, and 1987/88 El Niño events; Fig. 1); however, the LD<sub>SSS</sub> record does not unfailingly capture all known ENSO warm or cool events.

Analysis of equatorial Pacific SST anomalies clearly indicate the development of an ENSO event, with El Niño (La Niña) events typically showing patterns of warming (cooling) in the central and eastern Pacific and cooling (warming) in the CWEP. These anomalies and their related lower atmosphere phenomena cause the far-reaching weather patterns well known across the Pacific Basin, as well as teleconnections beyond the tropical–subtropical margin. We observe the essentially opposite patterns in SST that occur in El Niño versus...
La Niña years in composite maps of summer (DJFM) SST anomalies constructed using lower- and upper-quartile LDSSS years from 1979 to 2009 (Fig. 2). Mechanisms driving the correlations and SST patterns we observed using the LDSSS record are described in the next section.

4. Mechanisms during the instrumental/satellite periods

a. The ENSO extratropical teleconnection

Numerous studies have identified a teleconnection between the tropical Pacific Ocean and southern high latitudes. The second and third EOFs of 500-mb GPH variability in the SH identify an atmospheric bridge, the Pacific and South American pattern (PSA) (Mo and Paegle 2001). The PSA varies on an interannual basis but is on average the SH response to ENSO warm or cold events. During the austral winter prior to a mature El Niño (La Niña) event, a Rossby wave pattern of high–low–high (low–high–low) tropospheric pressure anomalies propagates from the CWEP through the South Pacific toward South America and the Antarctic Peninsula in a great-circle pattern. This alternating pressure pattern affects the Amundsen Sea low (ASL; a quasi-stationary low pressure (LP) system in the South Pacific), which can be linked to the tropical ENSO anomalies with strong correlations observed between SST in the Niño-3.4 region and central pressures of the ASL (Karoly 1989; Mo and Paegle 2001; Jin and Kirtman 2009). The PSA is variable on an interannual basis, due primarily to its propagation being dependent on the exact pattern and zonal distribution of tropical Pacific SST anomalies and the behavior of the zonal wind in the South Pacific (which is affected by other climate modes, such as SAM) (Harangozo 2004; Turner 2004; Lachlan-Cope and Connelly 2006). Other authors have argued that the SH response leads ENSO, since the Rossby wave pattern precedes the ENSO peak season; however, the physical processes forcing this remain unclear (Trenberth and Shea 1987; Jin and Kirtman 2009).

As noted previously, most studies on the effect of Pacific extratropical teleconnections in the Antarctic region have focused on West Antarctica. While the strongest low- to high-latitude ENSO teleconnection occurs via the PSA in the austral winter (JJA) prior to an ENSO event, a clear circumpolar signal is seen in the SH in correlations of summer SOI [November–February (NDJF)] and 500-mb GPH (Fig. 3a). This pattern manifests not only as the expected relationship between SOI and trade winds in the tropical Pacific (strengthened southeastern flow during high SOI years) but as a strong positive relationship between the Southern Oscillation and circumpolar high-latitude westerly flow (strengthened westerly flow in high SOI years) (Fig. 3b). This is a distinctly SAM-like pattern and not surprising given ENSO is known to contribute 25% of the variance in SAM during austral summer (L’Heureux and Thompson 2006). We see a lower correlation between the SAM (DJFM) and LDSSS than between the SOI (NDJF) and LDSSS over the 1958–2009 period (Table 1a), suggesting the LDSSS record reflects the variance in SAM that is related to ENSO as shown by L’Heureux and Thompson (2006).

b. The ENSO–Law Dome mechanism

Composite maps of 500-mb GPH anomalies from the upper- and lower-quartile summers of both SOI and LDSSS (1979–2009) are broadly similar despite the noisiness of the ENSO signal in the LDSSS record (Fig. 4). During El Niño–like summers (lower-quartile LDSSS/ SOI composites), there is a weaker South Atlantic LP
anomaly and an approximately 30\" westward shift of the South Pacific HP anomaly in the LDSSS composite (cf. Fig. 4d with Fig. 4b). The La Niña–like (upper-quartile SOI/ LDSSS) composites (Figs. 4a,c) have similar spatial patterns; however, the height anomalies are weaker in the LDSSS composite (Fig. 4c). In addition, both upper-quartile (La Niña–like) composites (Figs. 4a,c) show anomalies considerably weaker than their El Niño–like counterparts (Figs. 4b,d), indicating a lesser response to La Niña compared to El Niño in SH middle to high latitudes that is reflected in the LDSSS record. These dissimilarities between the upper and lower composites of each record point to the influence of regional drivers on the variability of sea salt concentrations at LD.

The patterns observed in the high- and low-quartile SOI/LDSSS composites reflect changes in zonal surface winds over broad parts of the SH (Fig. 5). In high SOI/La Niña–like years there is a slight strengthening of the zonal westerly flow at high latitudes in the Southern Ocean (Fig. 5a), and this strengthening can also be seen in the upper-quartile LDSSS composite (Fig. 5c). In low-quartile SOI/LDSSS years (Figs. 5b,d), the wind anomalies are stronger but in the opposite direction. Mean zonal wind speeds are reduced by up to 2 m s\(^{-1}\) in a circumpolar pattern, reducing sea salt aerosol production and resulting in reduced sea salt concentrations in LD precipitation. This finding is analogous to previous studies on sea salt variability at LD, which showed sea salt concentrations reflected broadscale source strength signals of mid- to high-latitude zonal winds (Souney et al. 2002; Goodwin et al. 2004).

In addition to these large-scale drivers of sea salt concentration, regional influences in the LD sector likely play a role. One regional influence on sea salt concentrations at LD during summer may be synoptic-scale phenomena in the southwest Pacific that have been attributed to short-term, high-frequency ENSO variability. Using a simple thickness model, Smith and Stearns (1993) observed a distinct switch in sign of surface temperature/pressure anomalies across SOI minima along the Wilkes Coast (120\°W–30\°E), the region of East Antarctica between the Ross and Amery ice shelves that contains LD. They attributed this switch to an outflow of cold, relatively dense air from a ridgelike pattern over Wilkes Land. Using findings from previous studies, they proposed the cold outflow could enhance midlatitude blocking southeast of New Zealand (Mo et al. 1987; Parish and Bromwich 1987). Blocking in this region is also associated with a cutoff low, which has been suggested may influence the western branch of the Southern Oscillation (Trenberth 1980). If so, this blocking presumably links high-latitude ENSO-related anomalies back to the low-latitude regions that initially forced them. As mentioned in section 3, there is a small but significant and coincident correlation between LDSSS and midlatitude blocking southeast of New Zealand (150\°–170\°W; Table 1). The work in this study does not allow us to say whether this signal results from the direct influence of cold outflows from Wilkes Land or is merely another reflection of the ENSO signal at LD. However, the 30\° westward shift of the main center of anomalous high pressure in the South Pacific as well as the regional concentration of reduced midlatitude zonal winds southeast of New Zealand in the El Niño–like LDSSS composite compared to the El Niño–like SOI composite (cf. Figs. 4d, 5d with Figs. 4b, 5b) is curious. It suggests the LDSSS
record is sensitive not only to ENSO-forced springtime blocking across a broad region of the South Pacific but also to a coincident blocking feature southeast of New Zealand, which in turn may affect the western arm of the Southern Oscillation (Trenberth 1980).

5. The 1010-yr Law Dome summer sea salt record

In this study, variability in the LDSSS record over the last 1010 yr is discussed in terms of three distinct epochs (Fig. 6). Epochs 1 (1000–1260 AD) and 3 (1920–2009 AD) are periods of below-average LDSSS values. Lying between these is epoch 2 (1260–1860 AD), a long period of average to above-average values with a short departure below average around 1720–50 AD and relatively high values around 1840–60 AD.

Spectral analysis of the 1010-yr LDSSS record (Fig. 7) reveals a number of significant periodicities. The analysis employed the multitaper technique (Park 1992; Mann and Lees 1996) with a resolution parameter of two and three tapers and significance levels computed from a red-noise model fit to the input series. For further details on the software used, see the University of California, Los Angeles (UCLA) Singular Spectrum Analysis-Multitaper Method (SSA-MTM) Toolkit (http://www.atmos.ucla.edu/tcd/ssa/).

As expected from the correlation analyses, significant spectral features are seen in the 2–7-yr ENSO band,
particularly at 2.8, 4.4, and 6 yr (significant at 99% level). Other periodicities significant at the 95% level are seen at 7.4, 12.5, and 17 yr, and there are other periodicities that just reach 95% significance.

An evolutive spectrum reveals the persistence of these features (Fig. 7b), which may be considered in terms of the three epochs outlined previously: 1000–1260 AD, 1260–1860 AD, and 1920–2009 AD. The high LDSSS epoch 2 shows several spectral peaks between 6 and 10 yr, which are muted or absent in epochs 1 and 3. Epoch 2 also has strong spectral peaks at 12.5- and 25-yr periods, which are absent in epochs 1 and 3. These observations are also evident from separate spectra computed for the three epochs (not shown).

A further prominent variation in spectral content is seen at the 2–2.5-yr range. The early part of the record (to 1450 AD) has more power in this band than the later record. This is clearly not related to a high or low summer sea salt period. We rule out an artifact where high-frequency power is lost because of undersampling, as the record is well sampled throughout, and if anything better sampled through the period of lower variability after 1450 AD. Likewise, there is a feature around 17 yr that is intermittent but more prominent in the period up to 1800 AD.

**FIG. 5.** Composite anomaly maps of surface zonal winds from (a) upper- and (b) lower-quartile summer SOI (NDJF) and (c) upper- and (d) lower-quartile LDSSS (DJFM) using detrended data from 1979 to 2009. Contour intervals are 1 m s⁻¹. Composites are from the season specified: that is, NDJF (SOI) and DJFM (LDSSS).
Considering specifically the variability associated with ENSO, we undertook coherence analyses between the LD<sub>SSS</sub> and SOI/Niño-4 SST series over the instrumental period (Table 2). These results show bands of strong coherence. In-phase coherence between LD<sub>SSS</sub> and both series is seen at periods of 2.5, 3.6 (for SOI), and 3.9 yr (for Niño-4). High coherence is also seen at a 4.7-yr period with LD<sub>SSS</sub> lagging SOI and Niño-4 SST by 5 and 7 months, respectively. There is a broad band of coherence between LD<sub>SSS</sub> and Niño-4 SST at 10 yr, which is interesting, as this is not typically regarded as a band of ENSO variability. The phase difference of the two signals at 10 yr is large, with LD<sub>SSS</sub> leading by 20 months. Alternatively, it could be that LD<sub>SSS</sub> lags by 100 months or that there is a 60-month (5 yr) offset to these lags and an antiphase relationship. This 10-yr period may be related to the interaction between ENSO and SAM, as a similar period has been found in middle to high latitudes possibly related to the SAM (Yuan and Yonekura 2011).

Taken together, we see that the relationship between LD<sub>SSS</sub> and ENSO is confirmed in the spectral analysis and that there is some indication that the variability itself shifts with the ENSO state. The high LD<sub>SSS</sub> (ENSO cool) period (epoch 2) is marked by muted variability in the 4–5-yr band (with the exception of the fifteenth century) and a shift to longer-period variability in the 6–10-yr band and at 12 and 25 yr. The 10-yr periodicity that correlates with Niño-4 SST is predominantly a feature of the low LD<sub>SSS</sub> (ENSO warm) periods. The modern epoch (since 1920) bears some similarity to the period from 1000–1260, with a tendency toward ENSO warm conditions and stronger variability at periods around 4–5 and 10 yr.

6. Decadal-scale variability

A growing body of literature suggests that the tropical–extratropical connection between ENSO and the South Pacific and Antarctica is highly variable, depending on the exact pattern and zonal distribution of tropical SST anomalies and the zonal wind in the South Pacific (Harangozo 2004; Turner 2004; Lachlan-Cope and Connelly 2006). L’Heureux and Thompson (2006) found ENSO variability affects the SAM during DJF, yet Fogt and Bromwich (2006) found this connection varied on decadal scales due to interaction with the SAM during the prior spring, which either reinforced or opposed the high-latitude teleconnection. Recently, Fogt et al. (2011) found this connection to be phase dependent; when a La Niña (El Niño) event occurs during a SAM positive (SAM negative) period during summer (DJF), the resulting anomalous transient eddies in the South Pacific reinforce each other, and this also seems to be the case during periods of weak/neutral SAM. Other phases (e.g., La Niña–SAM negative) are characterized by opposing transient eddies, and the high-latitude teleconnection is weakened. This obviously has implications for the strength and stability of the ENSO–LD<sub>SSS</sub> proxy, although decadal analysis suggests a further reason for temporal variability in the ENSO–LD teleconnection, as described below.

IPO modulation of the ENSO signature at Law Dome

The source of decadal variability in ENSO is debated; however, one possible interaction is with the IPO, a coherent pattern of SST variability over the Pacific Ocean that is thought to modulate the influence of ENSO on Australian climate (Power et al. 1999; Salinger et al. 2001; Rissbey et al. 2009). Our observations (Table 1) suggest the source of the correlation between ENSO and LD is the CWEP region (section 3). As the IPO has a decadal to multidecadal period and to be consistent with previous studies on ENSO–IPO interactions (e.g., Power et al. 1999; Cai et al. 2010), we used 13-yr window sliding correlation analyses and 13-yr Gaussian smoothing of an annual IPO index to explore temporal variability in the teleconnection between LD<sub>SSS</sub> and Niño-4.
The Niño-4 SST–LDSSS correlation shows large decadal-scale changes in strength over the last ~140 yr (Fig. 8a). Two periods occur where the 13-yr running correlation coefficient between LDSSS and Niño-4 is reduced to around zero; from approximately 1894 to 1906, and from approximately 1948 to 1974 (there is also a period at the beginning of the instrumental record up to 1883). When a 13-yr smoothed index of the IPO is plotted with the sliding correlation time series of LDSSS–Niño-4 SST an interesting relationship emerges. Virtually without fail, when the IPO is positive, the LDSSS–Niño-4 correlation is significant; when the IPO flips to a negative state, the correlation breaks down. Our results suggest the IPO is responsible for the background decadal-scale variability in the teleconnection between LD and ENSO. This variability in teleconnection strength has implications for how the LDSSS record may be interpreted as a record of ENSO-related variability.

Regionality: The case of eastern Australian rainfall

ENSO is the key driver of rainfall in Australia in terms of broad influence and impact (Risbey et al. 2009). However, normal indices of ENSO strength are not good predictors of the effect of a given ENSO event on eastern Australian rainfall, as the rainfall response is only linear for La Niña events and not El Niño events (Wang and Hendon 2007). This asymmetry has been variously attributed to changes in the zonal distribution of SST anomalies (Wang and Hendon 2007), decadal variability of ENSO (e.g., the IPO) (Power et al. 1999, 2006; Risbey et al. 2009; Cai et al. 2010), and interactions with other modes such as the SAM or the Indian Ocean dipole (Risbey et al. 2009, and references therein).
Table 2. Period and phase offset (expressed as LDSSS lag) for significant coherence between LDSSS and SOI or Niño-4 over the period 1889–2009. Periods in bold are significant at 99%; others are significant at 95%. The third column, LDSSS–Niño-4 INV, references LDSSS to an inverted Niño-4 index for comparison with SOI.

<table>
<thead>
<tr>
<th>Period (yr)</th>
<th>LDSSS–SOI*</th>
<th>LDSSS–Niño-4*</th>
<th>LDSSS–Niño-4 INV*</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>—</td>
<td>120°/40</td>
<td>—</td>
</tr>
<tr>
<td>4.7</td>
<td>30°/15</td>
<td>225°/35</td>
<td>45°/7</td>
</tr>
<tr>
<td>3.9</td>
<td>—</td>
<td>180°/23</td>
<td>In phase</td>
</tr>
<tr>
<td>3.6</td>
<td>In phase</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>2.5</td>
<td>In phase</td>
<td>180°/15</td>
<td>In phase</td>
</tr>
</tbody>
</table>

* Phase lag of LDSSS.

In eastern Australia, ENSO and blocking collectively account for 50% of the variance in the rainfall record (Risbey et al. 2009). However, the asymmetric ENSO–rainfall relationship in this region means that the strength of a given El Niño event often bears little resemblance to how dramatic rainfall declines are in a given season. Rather, it appears eastern Australian rainfall, in particular during spring, is most sensitive to SST variability in the CWEP, rather than the absolute magnitude of SST variability in the eastern Pacific (Wang and Hendon 2007). This has implications for recent below-average rainfall in eastern Australia (e.g., southeast Queensland) where short instrumental records limit the ability to infer whether rainfall declines are the result of natural variability or related to anthropic climate change (Heinrich et al. 2009; Cai et al. 2010). Furthermore, eastern Australia, particularly the subtropical and southeastern region, is virtually bereft of long rainfall proxies, so it is difficult to determine whether the current declines are unusual beyond the context of the last century.

b. Law Dome and eastern Australian rainfall in the instrumental period

We investigated whether the CWEP–LDSSS relationship would result in a connection between LD and Australian rainfall. Monthly rainfall records from the Bureau of Meteorology (BOM) high-quality station dataset were compared to the instrumental period of the LDSSS record. Analysis of seasonal and annual leads and lags revealed a pattern of strong correlation between annual (January–December) rainfall and LDSSS (rainfall lead) in the eastern subtropical and temperate region. Table 3 shows correlations between LDSSS and the four BOM high-quality stations with the longest continuous monthly records from a grid spanning 25°–30°S and 148°–152°W.

Cai et al. (2010) found that the decline in southeast Queensland rainfall since the 1980s was due to a flip in state of the IPO in the late 1970s from a “negative” or La Niña–like state to a “positive” or El Niño–like state, a finding consistent with the work of Power et al. (1999) and Risbey et al. (2009). Cai et al. (2010) relate other flips in state earlier in the instrumental record to similar declines in rainfall in southeast Queensland. Their study is in agreement with earlier work suggesting that the breakdown of the ENSO–eastern Australian rainfall relationship is due to zonal shifts in the equatorial Pacific SST anomalies (Wang and Hendon 2007). It is clear from Fig. 8a that the LDSSS–Niño-4 relationship breaks down during periods of negative IPO. Sliding correlations (again, 13-yr band) of LDSSS and the four high-quality rainfall records shown in Table 3 from eastern Australia have periods of high and low correlation strength virtually identical to the periods observed in the LDSSS–Niño-4 relationship (Figs. 8a,b). Figure 8b is particularly interesting, because it shows that, unlike the LDSSS–Niño-4 relationship, the LDSSS–rainfall relationship (at least at these four stations) rarely breaks down completely.

Spring Niño-4 SST anomalies (September–October) are strongly anticorrelated with annual (January–December) rainfall in eastern Australia (Fig. 8c; note that the color bar of the Niño-4 map has been inverted for clarity). A very similar spatial pattern emerges when the Niño-4 record is replaced with the LDSSS record (Fig. 8d). The same variance (and in certain regions of eastern Australia more) is explained by the LDSSS–east Australian rainfall correlation as by the LDSSS–Niño-4 correlation (Tables 1, 3 and Fig. 8d), which is surprising given the convoluted path of the CWEP–LD teleconnection. We suggest this is a result of reinforcement by regional drivers that affect both the LD region and Australian rainfall. It is likely most of this reinforcement is the result of the SAM and the projection of the extratropical signature onto the SAM during DJF as noted by L’Heureux and Thompson (2006). The significant correlation with rainfall in southeast Australia and southwest Western Australia, regions known to be influenced by variability in the SAM (Nicholls 2009; Risbey et al. 2009), also suggests this to be the case (Fig. 8d). The correlation between blocking in the South Pacific in spring and LDSSS is unlikely to be a reinforcing mechanism (Table 1) as blocking in this region is too far to the east to directly affect Australian rainfall (Risbey et al. 2009) and, as noted previously, is more likely associated with the propagation of ENSO-forced anomalies through the South Pacific.

c. 1010 yr of rainfall variability in eastern Australia

Despite the IPO-related correlation variability identified above, the LDSSS record provides insight into rainfall variability in eastern Australian over the last
1010 yr at the decadal to centennial scale (Fig. 6). The LDSSS record has a maximum around 1840–60 but is below average from ~1920 to the present. The three epochs defined in section 6 suggest annual rainfall in eastern Australia is currently lower than the long-term average but that a similar period of below-average rainfall occurred during 1000–1260.

Because of the scarcity of good-quality precipitation proxy records in the Australian region, eastern mainland Australia has very few precipitation records or reconstructions. Nonetheless, a few records do exist, though to our knowledge none covering the period 1000–present. Lough (2011) used coral records from the Great Barrier Reef to reconstruct 330 yr of summer rainfall for northeast Queensland (northeast tropical Australia), a region that receives predominantly tropical summer rain (DJF). These authors found a distinct dry period, with reduced and less variable summer rainfall from the mid-1700s to the mid- to late 1800s. In contrast, Gergis et al. (2011) used lake levels and documentary evidence to reconstruct rainfall since European settlement (1788) in southeastern Australia (1783–1988) and found predominantly wet conditions from the mid-1780s to the mid-1830s. This apparent disparity between the tropical north and subtropical/temperate south is possibly explained by a southward displacement of the ITCZ during this time (attributed to the end of the LIA period) (Newton et al. 2006). Such southward displacement of the ITCZ is thought to have resulted in tropical aridity and wetter conditions farther south of the equator. In the LDSSS record, ~1750–1870 is a period of generally above-average LDSSS values, implying above-average rainfall for eastern Australia, particularly around 1830–60 (Fig. 6). This is in general agreement with the findings of enhanced aridity in the tropics (Lough 2011) and wetter conditions farther south (Gergis et al. 2011), as the significant spatial representation of rainfall in Australia by the LDSSS record is generally restricted to the subtropical and temperate regions of eastern mainland Australia.

Numerous studies have attributed seasonal declines in rainfall in various regions of Australia in recent decades to natural variability and/or anthropogenic climate change. The most prominent is the 15%–20% decline in

![Decadal-scale variability in the LDSSS-east Australian rainfall relationship illustrated by plotting the IPO index (red; 13-yr Gaussian smooth) with 13-yr running correlation time series records of LDSSS and (a) September–October Niño-4 SST (blue) and (b) four rainfall stations: Wallangra Station (red), Taroom post office (PO) (magenta), Surat PO (yellow), and Miles PO (black). January–December AWAP rainfall data correlation from 1910–2009 is shown with (c) September–October Niño-4 SST (concurrent year) and (d) LDSSS (rain leads by 11 months). Note that colors have been inverted in (c) for clarity (blue is a negative correlation). Significance at 95% occurs at the second contour ($r = \pm 0.200$). Locations of the stations in (b) are shown in (d).]
winter rainfall since the 1970s in southwest Western Australia (SWWA) (e.g., Cai and Cowan 2006). More generally, southern Australia has seen seasonal rainfall declines in the last few decades (Nicholls 2009), and the variability and declines in southeast Queensland have already been discussed (Wang and Hendon 2007; Cai et al. 2010). The LD snow accumulation record has been linked to winter rainfall decline in SWWA (van Ommen and Morgan 2010). This work suggested the SWWA decline may be unusual in the last 700 yr and related to anthropogenic climate change. Our findings may appear comparable; however, our results should not be considered linked to the findings of van Ommen and Morgan (2010) as, despite being derived from the same ice core record, they use a different dataset (accumulation) related to different mechanistic drivers that are linked to climatically distinct regions of Australia. The LDSSS record indicates that modern rainfall amounts in eastern Australia may be lower than the millennial average of annual rainfall. While rainfall in eastern Australia may be lower than normal and unusual, this kind of variability is not unknown in the last 1010 yr.

8. Summary

This study identifies a highly significant signature of ENSO during the instrumental period (1889–2009) of a 1010-yr record of summer sea salts from Law Dome, East Antarctica. The signature is related to SST anomalies during the previous spring from the CWEP, a region known to produce ENSO-related extratropical teleconnections to high southern latitudes. We attribute this correlation to Rossby wave activity known to be the SH expression of ENSO. A further significant correlation \( r = 0.402 \) with blocking in the South Pacific “downstream” from the CWEP region confirms the CWEP to South Pacific pathway of the teleconnection. This results in circumpolar GPH and zonal wind anomalies during austral summer that are known to influence the SAM and ultimately delivery of sea salt aerosol to coastal Antarctica (Karoly 1989; Pook and Gibson 1999; Mo and Paege 2001; L’Heureux and Thompson 2006). The ENSO–LD teleconnection is decadal variable during the instrumental period, with oscillations that are broadly coincident with variability in SSTs in the equatorial Pacific attributed to phase changes of the IPO (Power et al. 1999; Salinger et al. 2001; Risbey et al. 2009).

Strong ENSO band features (2–7 yr) and intermittently persistent decadal-scale (10–25 yr) features are seen in spectral analyses of the 1010-yr LDSSS record. The main epoch of higher than average (La Niña–like) LDSSS values (1260–1860 AD) is associated with longer ENSO-band periodicities of 6–7 yr. Two periods exhibit below-average LDSSS values (1000–1260 AD and 1920–present) and are associated with shorter ENSO-band periodicities (2–5 yr) attributed to more El Niño–like regimes.

The LDSSS record is significantly correlated with annual rainfall in eastern Australia. This relationship appears to be reinforced by the interaction between SAM and ENSO during austral summer but is similarly constrained by decadal variability associated with phase changes of the IPO as observed in other studies (e.g., Wang and Hendon 2007; Cai et al. 2010). The LDSSS record suggests periods of below-average rainfall similar to the declines currently experienced in some parts of Australia are unusual but not unknown over the past 1010 yr and are associated with more El Niño–like regimes. The LDSSS record currently spans 1000–2009 AD. Future work may be able to extend the record beyond ~2 kya, allowing further insight into past changes in ENSO and eastern Australian rainfall.

Acknowledgments. This work was supported by the Australian Government’s Cooperative Research Centres Programme, through the Antarctic Climate and Ecosystems Cooperative Research Centre (ACE CRC). NCEP correlations and composites produced using NOAA/ESRL PDI data (http://www.esrl.noaa.gov/psd/) and Panoply version 3.0 (http://www.giss.nasa.gov/tools/panoply). SST data are from the Met Office (http://www.metoffice.gov.uk/hadobs). We thank M. Pook, J. Risbey,
G. Meyers, and three anonymous reviewers for helpful comments and suggestions.

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