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The worst North American drought year of the lastmillennium: 1934

Benjamin I. Cook
NASA Goddard Institute for Space Studies, New York, New York, benjamin.i.cook@nasa.gov

Richard Seager
Ocean and Climate Physics, Lamont-Doherty Earth Observatory, Palisades, New York

Jason E. Smerdon
Ocean and Climate Physics, Lamont-Doherty Earth Observatory, Palisades, New York

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Key Points:
- The 1934 drought was the most widespread and severe event of the last millennium
- Despite a strong La Niña, SST forcing played only a small role
- Atmospheric variability caused winter drying; dust intensified drying in spring

Supporting Information:
- Readme
- Figure S1
- Figure S2
- Figure S3
- Figure S4
- Figure S5
- Figure S6

Correspondence to:
B. I. Cook,
benjamin.i.cook@nasa.gov

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The worst North American drought year of the last millennium: 1934

Benjamin I. Cook[1,2], Richard Seager[2], and Jason E. Smerdon[2]

[1]NASA Goddard Institute for Space Studies, New York, New York, USA, [2]Ocean and Climate Physics, Lamont-Doherty Earth Observatory, Palisades, New York, USA

Abstract During the summer of 1934, over 70% of western North America experienced extreme drought, placing this summer far outside the normal range of drought variability and making 1934 the single worst drought year of the last millennium. Strong atmospheric ridging along the West Coast suppressed cold season precipitation across the Northwest, Southwest, and California, a circulation pattern similar to the winters of 1976–1977 and 2013–2014. In the spring and summer, the drying spread to the Midwest and Central Plains, driven by severe precipitation deficits downwind from regions of major dust storm activity, consistent with previous work linking drying during the Dust Bowl to anthropogenic dust aerosol forcing. Despite a moderate La Niña, contributions from sea surface temperature forcing were small, suggesting that the anomalous 1934 drought was primarily a consequence of atmospheric variability, possibly amplified by dust forcing that intensified and spread the drought across nearly all of western North America.

1. Introduction

The Dust Bowl drought of the 1930s was one of the worst environmental disasters in United States history, causing widespread crop failures, land degradation, human migration, and a permanent transformation in the social, economic, and agricultural structure of the Great Plains [*Hansen and Libecap*, 2004; *Worster*, 1979; *Seager et al.*, 2008]. The inception of this event is linked to sea surface temperature (SST) variations in the tropical Pacific [*Schubert et al.*, 2004a, 2004b; *Seager et al.*, 2008], while the intensity and atypical spatial pattern of the drought is attributed to random atmospheric variability [*Hoerling et al.*, 2009] and/or dust aerosol and land surface feedbacks [*Cook et al.*, 2008, 2009; *Schubert et al.*, 2004a]. Even by the standards set by the multiple years of the Dust Bowl drought, however, the summer of 1934 stands out as exceptionally severe.

While the term "Dust Bowl" was not coined until the following year, 1934 directly set the stage for 1935 and the subsequent years that would be characterized by extensive drought, widespread dust storms (including the Black Sunday storm of 14 April 1935), and the formation of the Soil Conservation Service [*Worster*, 1979]. The drying that fed the 1934 drought began the previous fall [*Murphy*, 1935], with major precipitation deficits in California, the Northwest, the Southwest, and across the Southern Plains. By the spring of 1934, the drying shifted to the Central Plains and Midwest. Importantly, these precipitation deficits occurred during the primary seasons of moisture supply for the various regions (i.e., winter and early spring in California and the West and spring over the Central Plains and Midwest), ensuring little likelihood of relief with the onset of summer. Accompanying the precipitation anomalies were five major dust storms originating in the Central Plains from November 1933 to May 1934 [*Mattice*, 1935], caused in part by anthropogenic land use practices [*Hansen and Libecap*, 2004; *McLeman et al.*, 2014]. Three of these events occurred in April 1934 from 9–12, 18–20, and 21–24, which collectively spread dust as far east as North Carolina and Florida, while the largest dust storm on 9–12 May spread dust across most of the United States east of the Central Plains [*Mattice*, 1935]. By the end of 1934, it was estimated that 65% of the total area of the Great Plains had been damaged by wind erosion, with 15% severely affected [*Hansen and Libecap*, 2004].

Beyond the dust storms and wind erosion, the drought had severe impacts on agriculture and water resources, as documented in a report issued by the federal government the following year [*Murphy*, 1935]. Through the summer of 1934, conditions for pasture, corn, and tame hay (hay cut from cultivated grasses) were considered extremely poor over most of western North America (WNA), from Texas and New Mexico and up through Montana and the Dakotas. Reservoir levels were low across WNA, averaging ~55% of capacity in Nevada, ~40% in California, ~22% in New Mexico, ~18% in Idaho, and ~13% in Colorado and

Wyoming. By the end of the summer of 1934, over 1100 counties had received an emergency drought designation. Ultimately, the effects of the drought were so severe that in June 1934, President Roosevelt requested $525 million in drought relief from Congress, over half ($275 million) of which was intended to relieve the livestock industry by means of providing emergency feed, buying starving animals from farmers, and slaughtering excess herds for food relief [*Worster*, 1979].

Previous studies have analyzed SST and land surface forcing during the persistent, multiyear Dust Bowl drought [*Cook et al.*, 2008, 2009; *Hoerling et al.*, 2009; *Schubert et al.*, 2004a, 2004b; *Seager et al.*, 2008], but no study to date has conducted a comprehensive analysis specifically for the exceptional drought year of 1934. Moreover, while regional drought variability in North America is understood reasonably well [e.g., *Schubert et al.*, 2008; 2009; *Seager et al.*, 2005a; *Seager and Hoerling*, 2014], we know much less about the dynamics underlying multiregion, pancontinental drought events [*Cook et al.*, 2014] like 1934; events that, because of their extent, often have much larger social impacts [e.g., *Hoerling et al.*, 2014]. Here we analyze the summer drought of 1934 by investigating the seasonal evolution of the underlying climate anomalies, placing this drought within the context of the paleoclimate record, and drawing lessons regarding what might be learned to better understand the severity and occurrence of current and future pancontinental droughts.

2. Materials and Methods

Our analysis is focused on the climate anomalies (temperature and precipitation) and dynamics during the "water year" (October 1933 to September 1934) leading up to and during the summer drought of 1934. The main seasons of moisture supply are different across the various regions of WNA: for example, most precipitation falls in winter in the Southwest and along the West Coast, while spring and summer dominate the Central Plains. It is therefore likely that the drought in each region of WNA will have a different meteorological and dynamical explanation, and thus, a consideration of the climate anomalies across the entire water year is required.

An updated version of the North American Drought Atlas (NADA) is used to characterize the 1934 drought and place this event within the context of North American drought variability over the last millennium [*Cook et al.*, 2004, 2010]. The NADA is a tree ring-based gridded reconstruction (0.5° lat/lon) of summer season (June-July-August; JJA) Thornthwaite-Based Palmer Drought Severity Index (PDSI). PDSI is a normalized index of drought, integrating changes in moisture supply (precipitation) and demand (evapotranspiration) with a memory timescale of about 12–18 months (meaning that summer season PDSI will depend, at least partially, on climate anomalies in prior months and seasons). Negative PDSI values indicate drier than average conditions; positive values indicate wetter than average conditions. In the NADA, all anomalies are relative to the average baseline for 1931–1990, and we use data from the NADA spanning 1000–2005 Common Era (C.E.).

We calculate temperature and precipitation anomalies from the latest version (3.21) of the climate grids provided by the Climate Research Unit (CRU) at the University of East Anglia [*Harris et al.*, 2014]. The CRU climate grids are monthly average fields of climate variables that are interpolated from station observations to a uniform 0.5° lat/lon grid, with temporal coverage spanning the entire twentieth century (1901–2012). SST anomalies are taken from the HadISST data set [*Rayner et al.*, 2003] (1870 to present). We use 500 hPa geopotential heights from version 2 of the Twentieth Century Reanalysis (1871–2012) to investigate the dynamics [*Compo et al.*, 2011]. For a comparison with the winter of 2013–2014, we also use dynamical fields from the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis [*Kalnay et al.*, 1996] (1948 to present). To make things consistent with the base period in the NADA, all anomalies are calculated relative to the mean climatology for 1931–1990, except for the NCEP-NCAR reanalysis, which uses a baseline of 1948–1990.

3. Results

3.1. The Drought

Drought during the summer of 1934, as reflected in the PDSI anomalies, was intense and widespread across much of North America (Figure 1), extending from the West Coast to the Great Lakes and from Texas to Montana. The tree ring-based PDSI estimates in the NADA (Figure 1, left) are largely corroborated by PDSI values independently calculated from the CRU climate grids (Figure 1, right). Area-averaged NADA PDSI for western North America (WNA: 125°W–85°W, 30°N–49°N) in 1934 was −4.11, the single most intense

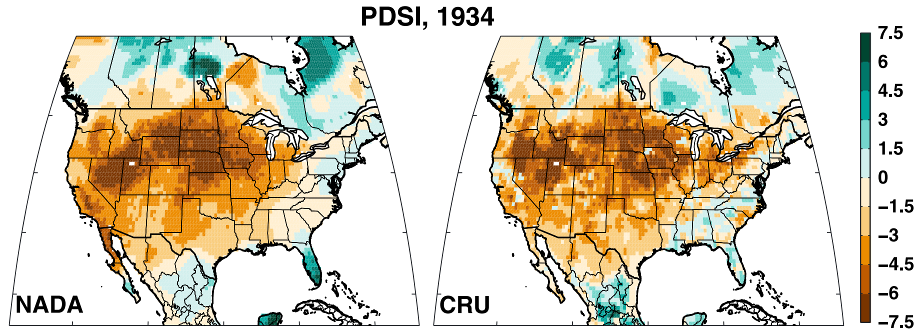

Figure 1. Summer season (JJA) Palmer Drought Severity Index (PDSI) for 1934 from (left) the North American Drought Atlas and calculated from (right) the CRU 3.21 climate grids.

drought year of the last millennium (Figure 2, top). This magnitude of drought in WNA was a departure of over 3.5 standard deviations from the long-term mean (1000–2005 C.E.) and exceeded the magnitude of the second most intense drought year in WNA (1580, PDSI = −3.19) by almost 30%.

In addition to being exceptionally intense, 1934 was also an outlier in terms of drought extent, with extreme drought (PDSI ≤ -3) covering 71.6% of WNA, the most extensive summer of extreme drought by a broad margin (Figure 2, bottom). By comparison, the second most extensive extreme drought year was 1956, when 49.3% of WNA was affected. The NADA ends in 2005 and therefore cannot be used to compare 1934 to more recent extreme drought years, such as 2011 and 2012. The National Climatic Data Center (NCDC) nevertheless tracks the percent area of the United States in "moderately to extremely dry" conditions over the instrumental period (1895 to present; see Acknowledgements for data URL). According to this metric, drought extent was above 70% for May through September 1934, averaging 77.2% in JJA. By comparison, the average for JJA of 2012 was only 59.7%, with a maximum extent in July of 61.8%. The average extent in JJA of 2011 was even lower at only 28.9%, indicating that 2011, while extreme, was highly localized (predominantly in the Southwest and Southern Plains) [*Seager et al.*, 2014]. These results confirm the NADA analysis, further supporting our observation that 1934 was the worst single year of drought in WNA of the last 1000 years.

3.2. La Niña Forcing

Cold SST anomalies in the eastern tropical Pacific developed in the fall of 1933 (Figure S1), a transition to La Niña conditions typically associated with suppressed precipitation across the Southwest and Southern Plains [*Ropelewski and Halpert*, 1987; *Seager et al.*, 2005b, 2014]. During October–December (OND) the NINO 3.4 index anomaly was −0.98 K, weakening slightly to −0.85 K during January–March (JFM) in 1934. Despite the moderately strong La Niña, circulation over North America during OND (Figure S2) and JFM (Figure S3) diverged substantially from the canonical La Niña response. During a typical La Niña, a broad, zonally oriented band of high pressure extends across the Southwest and Southern Plains [*Seager et al.*, 2010, 2014; *Trenberth et al.*, 1998]. Circulation during the fall and winter of 1933–1934, however, was characterized by an upper level ridge centered over the West Coast and an accompanying trough over the Northeast (Figure 3). This suggests that despite its strength, the 1933–1934 La Niña event was a relatively minor contributor to the precipitation deficits preceding the 1934 drought, even though persistent La Niña forcing was likely important to establishing the Dust Bowl as a long-term, multiyear event [e.g., *Schubert et al.*, 2004a; *Seager et al.*, 2008].

3.3. Winter and Early Spring

Cold season precipitation was reduced across much of WNA during the winter of 1933–1934 (Figure 3, left column), with major deficits in California, the Northwest, and the eastern half of the United States. Along with the drying, warm temperatures prevailed across most of WNA, while most of eastern North America was anomalously cold (Figure 3, middle column). Such a dipole configuration, characterized by a ridge over the Gulf of Alaska or along the West Coast and a trough over northeast North America, would have acted to suppress precipitation over much of WNA and favored the advection of cold air from high latitudes into the eastern part of the continent. These circulation and climate anomalies strongly resemble the most recent winter of 2013–2014 (Figure S4) when a similar, but more persistent, circulation pattern intensified the

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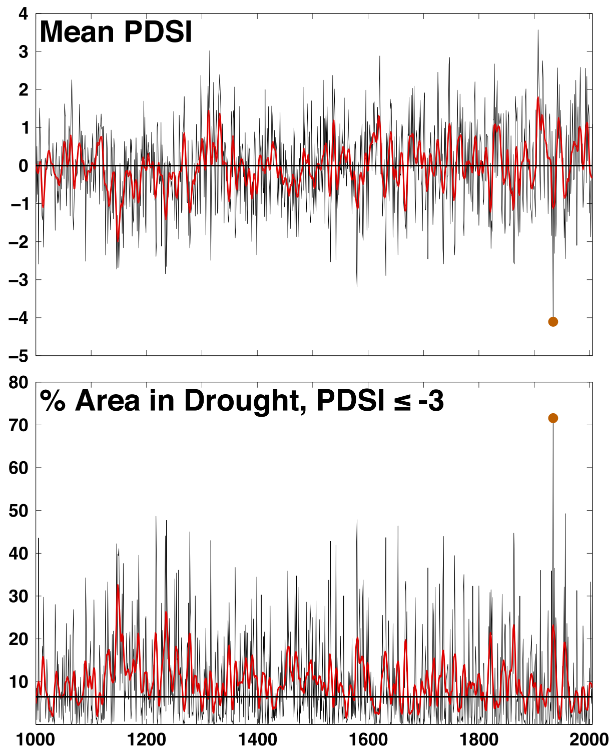

Figure 2. For western North America (125°W–85°W, 30°–49°), (top) area-averaged PDSI and (bottom) percent area with PDSI anomalies less than or equal to −3, indicating extreme drought.

ongoing drought in California and the Southwest and caused extreme cold conditions in eastern North America [*Wang et al.*, 2014b].

To further quantify the similarity between circulation during the winters of 1933–1934 and 2013–2014, we computed centered anomaly correlations (ACc) over North America (170°W–50°W, 20°N–70°N) between monthly 500 hPa height fields for these two periods. Of all months during the water year of 1933–1934, November 1933 and March 1934 were the strongest circulation analogues to this most recent winter. Highest correlations with November 1933 are in November 2013 (ACc = +0.51), December 2013 (ACc = +0.58), January 2014 (ACc = +0.74), and March 2014 (ACc = +0.57). For March 1934, correlations are similarly strong with December 2013 (ACc = +0.56), January 2014 (ACc = +0.68), and March 2014 (ACc = +0.64). November 1933 also correlates strongly with circulation during October–December 1976 (ACc = +0.64 to +0.71), the beginning of one of the most severe Californian droughts in the historical record (1976–1977). While further analysis is beyond the scope of this paper, these correlations do suggest that wintertime ridging along the West Coast may play a recurring role in driving precipitation deficits during major West Coast droughts.

The 1933–1934 geopotential height and climate anomalies are not consistent with a La Niña SST forcing signal; *Wang et al.* [2014b], however, linked the 2013–2014 winter circulation pattern to SST variability preceding an El Niño event in the western tropical Pacific (110°E–160°E, 10°S–20°N). Compared to the winter of 2013–2014, SST anomalies in this region were relatively weak during 1933–1934. To more explicitly test the SST forcing hypothesis, however, we analyzed ensemble simulations of two atmospheric models forced by observed SSTs, simulations that will thus contain responses to precursor SST anomalies, including La Niña conditions and other global SST patterns. These models (the Community Climate Model Version 3 (CCM3) and the NASA Goddard Earth Observing System model version 5 (GEOS-5)) have been used previously to investigate SST forcing of North American drought variability [e.g., *Seager et al.*, 2005a; *Wang et al.*, 2014a]. For 1933–1934, the ensemble mean in each model (Figure S5) shows a circulation response much more consistent with the expected La Niña pattern than the observations (Figures S2 and S3). This confirms our previous analysis, indicating little role for SST forcing of the drought during the 1933–1934 winter.

3.4. Late Spring and Summer

The center of drying and warming of the 1934 event moved to the Central Plains and Midwest in April and May, before shifting to the Southern Plains in June and July (Figure 4). From May afterward, the warm temperature anomalies were centered in regions where precipitation was below normal, likely reflecting the typically strong inverse relationship between warm-season temperature and precipitation in the Central Plains [*Madden and Williams*, 1978]. Circulation during the spring and summer was characterized by a persistent ridge centered firmly over the Northwest and Northern Plains, a feature that would have inhibited convection and reduced moisture advection from the Gulf of Mexico [*Brönnimann et al.*, 2009; *Cook et al.*, 2011], suppressing precipitation over the Plains. This circulation feature is absent in the SST-forced CCM3 and GEOS5 ensembles (Figure S5), suggesting little role for SST forcing during the spring and summer.

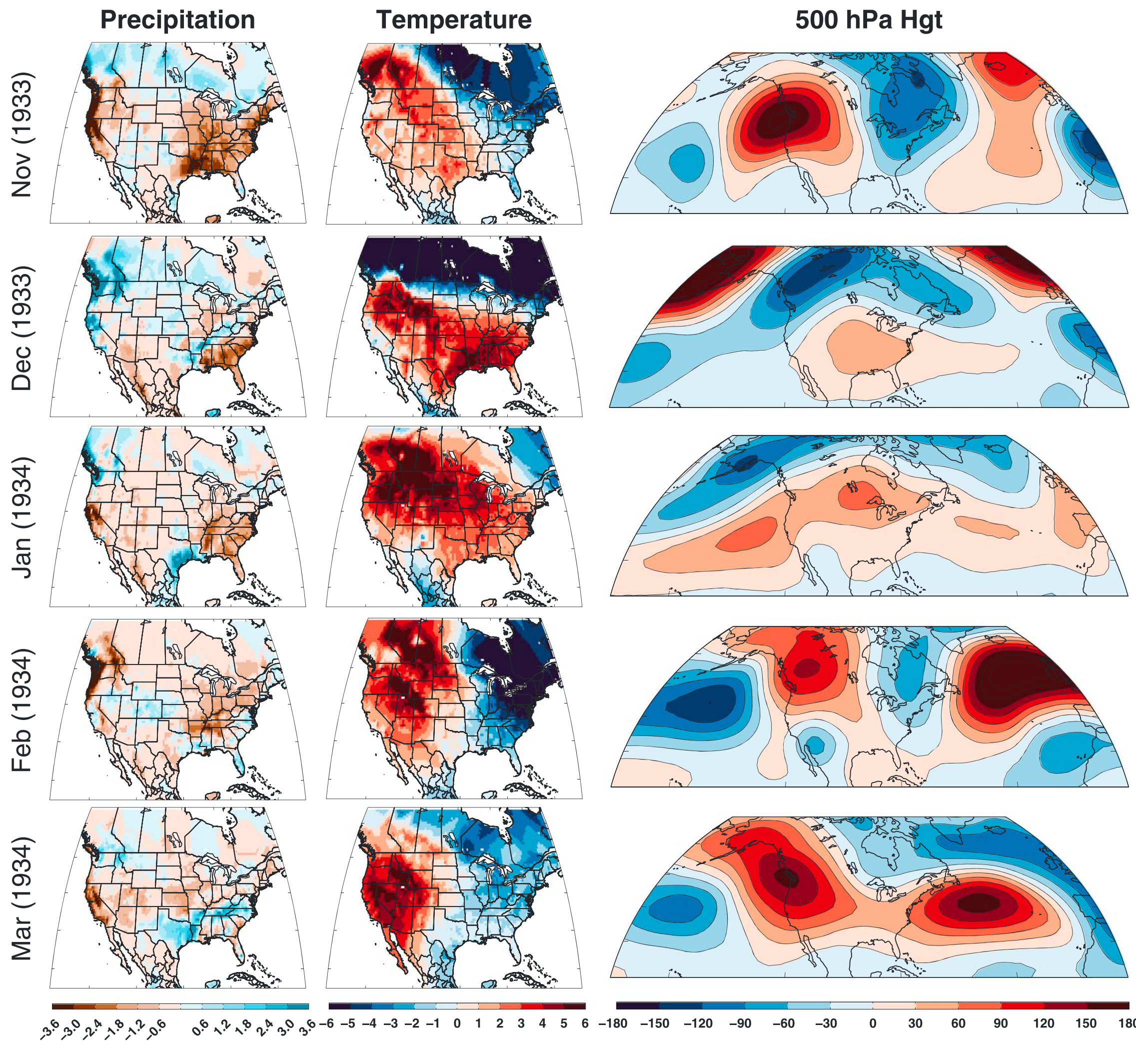

Figure 3. (left column) Precipitation (mm d^{-1}), (middle column) temperature (K), and (right column) 500 hPa height (m) anomalies during the fall and winter of 1933–1934.

The most intense precipitation deficits, however, occurred downwind from the Plains across the Midwest, Ozarks, and the Ohio River Valley. As discussed in section 1, large-scale dust storm events originated in the Central and Northern Plains regions (e.g., North and South Dakota, Kansas, and Nebraska) during April and May 1934 [*Mattice*, 1935], arising from a combination of reduced precipitation and/or poor land use practices [*Hansen and Libecap*, 2004]. These events were large enough to spread dust across much of the Midwest and eastern United States (e.g., Illinois, Iowa, Missouri, Ohio, and Pennsylvania) [*Mattice*, 1935], regions that saw some of the largest precipitation deficits during these same months (Figure 4). This coherence is consistent with the hypothesis that the historically unprecedented dust storms during the Dust Bowl amplified and spread drying [*Cook et al.*, 2008, 2009, 2011].

Mineral dust aerosols typically have high shortwave albedo and will reduce net radiation at the surface and the top of the atmosphere. To compensate for this energy deficit, circulation will adjust through anomalous subsidence, inhibiting convection, cloud development, and precipitation. This circulation response was previously observed in model simulations of the Dust Bowl that included dust storms and aerosols, experiments that produced a higher-fidelity simulation of the long-term, persistent (1932–1939) Dust Bowl drought than experiments using SST forcing alone [*Cook et al.*, 2008, 2009, 2011]. Similarly, the vertical velocity anomalies for May 1934 (Figure S6) show pronounced subsidence over the Midwest, corresponding to the same region of major precipitation deficits and dust aerosol loading during this month, a pattern similar to the dust-induced subsidence characterized in the Dust Bowl simulations [*Cook et al.*, 2011, Figure 14]. These

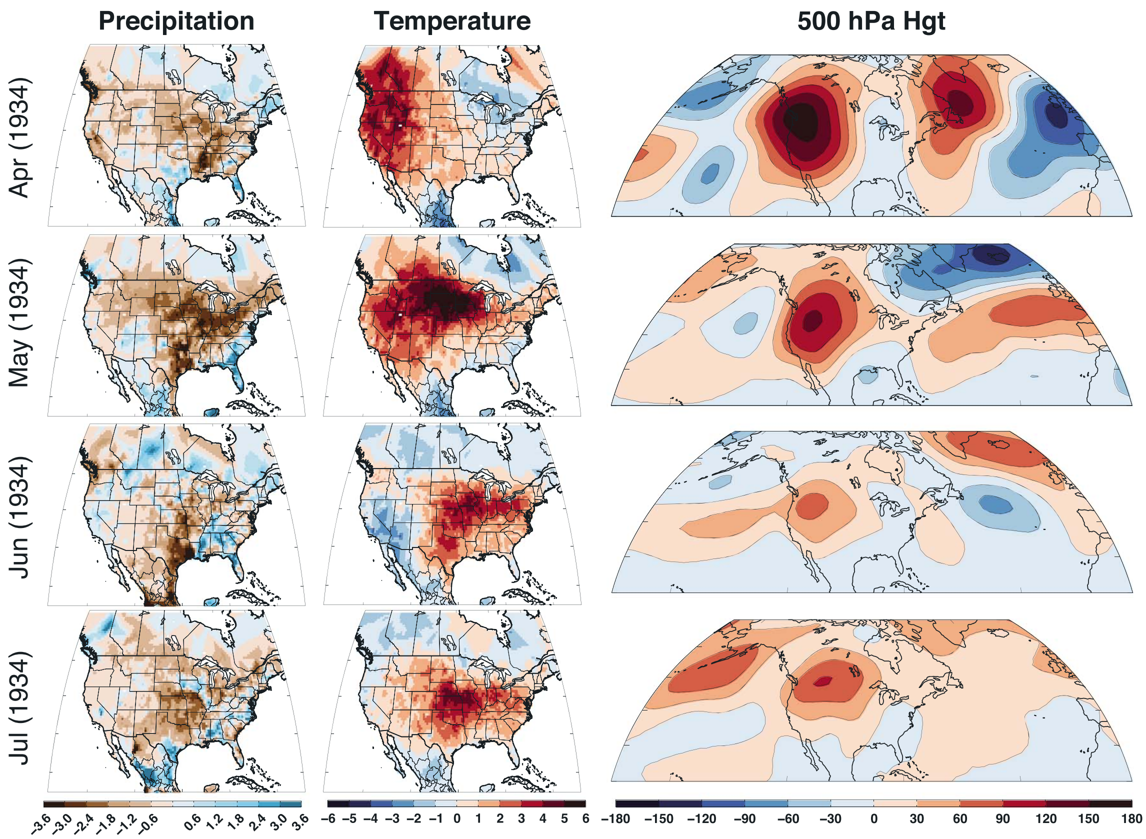

Figure 4. (left column) Precipitation (mm d^{-1}), (middle column) temperature (K), and (right column) 500 hPa height (m) anomalies during the spring and summer of 1934.

previous experiments constrained the dust forcing using soil loss and wind erosion estimates from the Soil Conservation Service (SCS). Unfortunately, there are no large-scale, quantitative estimates of soil losses and wind erosion available specifically for 1934 (the SCS was not fully active until 1935); it is therefore impossible to conduct new simulations targeting drought attribution specifically for this year. However, these new observations, combined with previous modeling work and the highly anomalous nature of the 1934 event compared to the last millennium, suggest the possibility of anthropogenic dust aerosol forcing playing a role in amplifying the drought.

4. Discussion and Conclusions

Over the last decade, WNA experienced several droughts that highlighted the vulnerability of this area to water resource availability [*Hoerling et al.*, 2014; *Seager et al.*, 2014; *Wang et al.*, 2014a]. Improving our understanding of drought evolution and dynamics is therefore a high research priority, especially for pancontinental events that span multiple regions and often have larger social and ecological impacts compared to regional droughts [*Cook et al.*, 2014; *Hoerling et al.*, 2014]. In 1934, precipitation deficits coinciding with the main seasons of moisture supply in each region turned this year into the single most extensive and intensive year of extreme drought in WNA during the last thousand years, helping to transform the persistent Dust Bowl drought into a national environmental, social, and economic crisis. We argue that this event likely developed through an unfortunate confluence of internal variability and anthropogenic forcing that allowed drought to spread across nearly the entirety of WNA.

During the winter, precipitation along the West Coast and in the Southwest was suppressed by a ridge centered on the coast, a pattern similar to the persistent circulation anomaly that dominated North American climate during the winter of 2013–2014 [*Wang et al.*, 2014b] and in the fall preceding the severe 1976–1977 California drought. The recurrence of this ridging pattern during some of the worst historical West Coast droughts suggests that it may be a robust feature driving precipitation and drought variability in this region.

Despite SSTs indicating a moderate La Niña, and SST-forced general circulation model (GCM) experiments simulating the expected dynamical response to La Niña over North America, the observed climate anomalies suggest that this climate mode played only a minor role in the 1934 drought. However, uncertainties in the underlying SST data set prior to 1950 may somewhat limit the ability to accurately capture the response to, for example, extratropical SST forcing in the GCM experiments [*Kennedy*, 2014].

In the spring, precipitation deficits occurred in the Central Plains and Midwest, allowing the drought to spread across nearly the entire expanse of WNA, establishing the event as one of the few pancontinental droughts of the twentieth century. Coincident with these precipitation deficits were dust storms, a common occurrence in WNA during droughts [e.g., *Goudie and Middleton*, 1992; *Neff et al.*, 2008]. The scale of wind erosion and aeolian activity during the Dust Bowl (including 1934) was nevertheless unprecedented since at least the Medieval Era [e.g., *Miao et al.*, 2007], most likely amplified by human land use practices at the time [*Hansen and Libecap*, 2004]. Moreover, much of the drying in April and May 1934 occurred not where the dust storms originated but rather downwind where dust aerosol loading in the atmosphere was observed [*Mattice*, 1935]. Because we could not conduct a formal attribution study, our analysis does not enable us to definitively attribute the warm-season drying to the dust storms at the time. However, when combined with previous modeling work demonstrating the role of dust aerosols in suppressing precipitation during the Dust Bowl [*Cook et al.*, 2008, 2009], our results are suggestive of the possibility that dust aerosols would have amplified the drying in 1934, pushing this drought beyond the normal range of variability observed in the paleoclimate record.

It is likely that given their rarity, pancontinental droughts do not all share the same underlying dynamical causes. For example, the severe decadal drought of the 1950s, which includes the widespread 1956 pancontinental event, can be reproduced in GCM simulations from SST forcing alone [e.g., *Hoerling et al.*, 2009; *Seager et al.*, 2005a], suggesting a weaker role for either local feedbacks or internal atmospheric variability. The 1934 event nevertheless serves as a useful exposition of how compounding events may give rise to widespread drought across North America. While not all of the 1934 conditions will be applicable in the future, the well-known impact of tropical Pacific SSTs and our emerging understanding of the West Coast ridging and associated circulation dipole are two components of variability in the climate system that should be further tracked and studied as important underlying dynamics contributing to pancontinental drought events.

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