Influence of Global Atmospheric Circulation Variations on Weather and Climate Extremes

Yen-Heng Lin
Utah State University

Follow this and additional works at: https://digitalcommons.usu.edu/etd

Part of the Climate Commons

Recommended Citation
https://digitalcommons.usu.edu/etd/7238

This Dissertation is brought to you for free and open access by the Graduate Studies at DigitalCommons@USU. It has been accepted for inclusion in All Graduate Theses and Dissertations by an authorized administrator of DigitalCommons@USU. For more information, please contact digitalcommons@usu.edu.
INFLUENCE OF GLOBAL ATMOSPHERIC CIRCULATION VARIATIONS ON WEATHER AND CLIMATE EXTREMES

by

Yen-Heng Lin

A dissertation submitted in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in

Climate Science

Approved:

Shih-Yu Wang, Ph.D.
Major Professor

Lawrence E. Hipps, Ph.D
Major Professor

Scott B. Jones, Ph.D.
Committee Member

Robert Davies, Ph.D.
Committee Member

Jin-Ho Yoon, Ph.D.
Committee Member

Mark R. McLellan, Ph.D.
Vice President for Research and Dean of the School of Graduate Studies

UTAH STATE UNIVERSITY
Logan, Utah

2018
ABSTRACT

Influence of Global Atmospheric Circulation Variations on Weather and Climate Extremes

by

Yen-Heng Lin, Doctor of Philosophy
Utah State University, 2018

Major Professor: Dr. Simon S.-Y. Wang
Department: Plants, Soils, and Climate

This dissertation aims to better understand the influences of global large-scale climate variations on extreme weather and climate events in order to improve subseasonal weather forecasts and future climate projections. Extreme weather and climate events can be influenced by local and teleconnection responses, such as El Niño and the Arctic Oscillation. The natural and anthropogenic variability of Arctic and tropical atmospheric features is an example of the importance of global circulations and weather extremes in climate research. This dissertation seeks to investigate the possible mechanisms of extreme events from the perspective of global circulation. This research first examines an extreme water shortage case study in Southeast Asia, where the Eastern Asian Summer Monsoon system and tropical climate variation strongly impact on precipitation. Next, extreme drought events in California over the winters of 2011–2012 to 2014–2015 are investigated to understand variations in circulation in drought winters. Last, the possible features of an Arctic subseasonal warming event are examined and found to be strongly associated with mid-latitude wintertime weather extremes.
Diagnostic results show that in early summer, the East Asian Summer Monsoon rainband has shifted from late May to early June since 1979. Variations in an upper-level cyclonic anomaly over eastern China and a lower-level anticyclonic anomaly in the subtropical Western Pacific are driving this timing shift, and anthropogenic global warming has been suggested as a possible cause. Furthermore, during the tropical cyclone season, peak low-level convergence over the Western North Pacific has a multidecadal-scale timing change, oscillating between July and August and influencing tropical cyclone genesis. However, this multidecadal variation with a frequency of ~20 years does not relate to any prominent atmospheric and oceanic variability.

In mid-latitude California, analysis results show that Californian drought winters are mainly modulated by the Pacific–North American teleconnection related to the El Niño–Southern Oscillation related and by the North Pacific Oscillation pattern, but these variabilities are not directly correlated with Californian droughts. Community Earth System Model Large Ensemble Porject projections show that drought-modulating circulations do not change under the global warming scenario. Lastly, an increase in the subseasonal Arctic tropospheric warming event, in contrast to the classic sudden stratospheric warming event, is observed over the past two decades. Arctic subseasonal temperature variation is associated with extreme winters in the mid-latitudes. The results of model simulation and historical data analysis indicate that sea-ice loss amplifies the subseasonal Arctic tropospheric warming event, and that the increase in deep convection over the tropical Pacific inducing the wave trend contributes to subseasonal Arctic warming.
Influence of Global Atmospheric Circulation Variations on Weather and Climate Extremes

Yen-Heng Lin

Global warming and climate change deeply influence weather and climate extremes, causing substantial property damage and loss every year around the world. Given the importance of heating differences between low-latitude and Arctic regions, which produce heat sources and cold sources that each influence global circulations, we investigate three extreme weather events in different regions in order to better understand the possible connections between extreme events and global circulation changes.

This study begins with climate variations in the low-latitude western North Pacific. In early summer, the timing of the wet season has shifted from late May to early June since 1979. This change influences the water supply in Southeast Asia. Our analysis results indicate that the increase in global temperatures is suggested to have induced this change. During the hurricane season, deep convection in the western North Pacific has a 20-year frequency of timing variations, oscillating between July and August and influencing hurricane activity. These variations have not been previously identified and do not have any driven forcings, but a precursor deep-convection signal is found in the spring.

Mid-latitude weather and climate can be influenced by tropical deep convection through the Pacific North American teleconnection. Our analysis results suggest that the wintertime Californian drought is mainly modulated by a teleconnection pattern from the tropics and natural variations in North Pacific circulation. Another key factor that influences mid-latitude circulation is Arctic temperature variations. We find an increase in the subseasonal Arctic warming event, suggesting more weather extremes in the mid-latitudes. Evidence suggests that sea-ice loss and the increase in tropical deep convection results in the increased likelihood of a subseasonal Arctic warming event.
ACKNOWLEDGMENTS

I would like to thank Dr. Simon Wang for making available to me the Climate Science Program. I would especially like to thank my committee members, Dr. Lawrence Hipps, Dr. Scott Jones, Dr. Robert Davies, and Dr. Jin-Ho Yoon for their support and assistance throughout the entire process.

I give special thanks to my wife, Amanda Chang, for her encouragement, moral support, and patience as I worked my way from the initial proposal writing to this final document. I could not have done it without all of you.

Yen-Heng Lin
CONTENTS

Page

ABSTRACT ........................................................................................................................................ iii
PUBLIC ABSTRACT ......................................................................................................................... v
ACKNOWLEDGMENTS ...................................................................................................................... vi
LIST OF TABLES ............................................................................................................................... ix
LIST OF FIGURES ............................................................................................................................. x

CHAPTER

I. INTRODUCTION ............................................................................................................................. 1

1.1 Introduction and Background ................................................................................................. 1
1.2 Tropical ENSO and MJO variability ...................................................................................... 3
1.3 Arctic Climate Variability ....................................................................................................... 4
1.4 Midlatitude Atmospheric Variations ....................................................................................... 6
1.5 Subseasonal and Intraseasonal Variability ........................................................................... 7
1.6 Objectives ................................................................................................................................. 7
References ........................................................................................................................................ 9
Figures ........................................................................................................................................... 14

II. INTERDECADAL CHANGES OF EARLY SUMMER PRECIPITATION
AND MULTIDECADAL TIMING VARIATIONS OF SUMMER LOW-
LEVEL CONVERGENCE ON THE EAST ASIAN SUMMER MONSOON ............................................. 17

Abstract ........................................................................................................................................ 17
2.1 Introduction ............................................................................................................................. 18
2.2 Data ......................................................................................................................................... 21
2.3 Interdecadal Variation of Active-phase Summer Monsoon ...................................................... 22
2.4 Multidecadal Variation of TC Activity .................................................................................... 27
2.5 Discussions and Summary ...................................................................................................... 33
References ....................................................................................................................................... 35
Figures and table ............................................................................................................................. 41

III. EMPIRICAL AND MODELING ANALYSES OF THE CIRCULATION
INFLUENCES ON CALIFORNIA PRECIPITATION DEFICITS .................................................... 57

Abstract ........................................................................................................................................ 57
3.1 Introduction ............................................................................................................................. 58
3.2 Data ......................................................................................................................................... 59
3.3 Results ...................................................................................................................................... 61
### Table of Contents

3.4 Discussions .......................................................................................... 66
3.5 Conclusions ......................................................................................... 69
References .................................................................................................... 70
Figures and Tables ..................................................................................... 75

**VI. ACCELERATED INCREASE IN THE ARCTIC TROPOSPHERIC WARMING EVENTS SURPASSING STRATOSPHERIC WARMING EVENTS DURING WINTER** ......................................................................................... 88

Abstract ....................................................................................................... 88
4.1 Introduction ............................................................................................. 89
4.2 Data and Model Experiments ................................................................. 91
4.3 Analysis and Results ............................................................................. 92
  4.3.1 Case Identification and Composite .................................................. 92
  4.3.2 Disparity in the Trends ..................................................................... 94
  4.3.3 Impact of External Forcing ............................................................... 95
  4.3.4 Impact of Tropical Forcing ............................................................... 98
4.4 Conclusions ....................................................................................... 100
References .................................................................................................. 101
Figures and Tables ..................................................................................... 106

V. Conclusions and Future Work .................................................................. 122

**APPENDICES** .......................................................................................... 126

**APPENDIX A. PERMISSIONS AND RELEASE LETTERS** .......................... 127

  License Agreement I: Materials for Chapter 2 ........................................ 127
  License Agreement II: Materials for Chapter 2 ....................................... 128
  License Agreement III: Materials for Chapter 3 ...................................... 129
  License Agreement IV: Materials for Chapter 4 ...................................... 130
  Permission to Reprint from Co-author I: Chapter 2 ............................. 131
  Permission to Reprint from Co-author II: Chapter 4 ............................ 132
  Permission to Reprint from Co-author III: Chapter 4 ........................... 133
  Permission to Reprint from Co-author IV: Chapter 4 ........................... 134

**CURRICULUM VITAE** ........................................................................... 135
<table>
<thead>
<tr>
<th>Table</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1 The correlation results between PCs and winter mean (November–March) of climate index over 18 California dry Winters. The highest corollary/anti-corollary index with PC1 or PC2 is shaded.</td>
<td>87</td>
</tr>
<tr>
<td>4.1 The Dates of Phase 1 Through Phase 7 of the Identified Cases.</td>
<td>121</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Schematic of the connection between global circulation and weather as well as climate extremes.</td>
</tr>
<tr>
<td>1.2</td>
<td>Schematic of ENSO-relative Pacific-North America (PNA) upper tropospheric geopotential height teleconnection anomalous pattern (solid lines) during a Northern Hemisphere winter which falls within an episode of warm sea surface temperature in the equatorial Pacific (shaded). The arrows in darker type reflect the strengthening of the subtropical jets in both hemisphere. The arrows in lighter type depict a min-tropospheric streamline as distorted by the anomaly pattern. Taken from Horel and Wallace [1981].</td>
</tr>
<tr>
<td>1.3</td>
<td>The structure and evolution of the Madden-Julian Oscillation (MJO) of DJF composite OLR and 850-hPa wind vector anomalies. Shading levels denote OLR anomalies less than -7.5, -15, -22.5, and -30 Wm⁻², respectively, and hatching levels denote OLR anomalies greater than 7.5, 15, and 22.5 Wm⁻², respectively. The magnitude of the largest vector is shown on the bottom right, and the number of days (points) from 1979 to 2001 falling within each phase category is given. The nominal time for transition between each of the numbered phases is 6 days but can vary from event to event. Taken from Wheeler and Hendon (2014).</td>
</tr>
<tr>
<td>2.1</td>
<td>(a) Long-term 15-day evolution of ΔOLR (235Wm⁻²-OLR) averaged in Taiwan (119° –122°E, 21° –25°N) from 1980 to 2010, following Wang and Chen (2008). The inset map depicts the geographical location of Taiwan (red). (b) Yearly distribution of daily ΔOLR applied with a 5-day moving average (shadings) overlaid with the linear trend contours from 1979 to 2014. A white and two yellow dashed lines indicate the period difference of the Meiyu as referred in the text.</td>
</tr>
<tr>
<td>2.2</td>
<td>Differences of ΔOLR (shadings) and (a) 250- and (b) 850-mb winds vectors between 1991–2002 and 1979–1990 for the 7 June–20 June period. (c–d) Same as (a–b) except for the differences between 2003–2014 and 1999–2002. (e) ΔOLR latitude (y-axis) and year (x-axis) distribution across the vicinity of Taiwan (white box in (b) and (d)) during the 7 June–20 June periods overlaid with the linear trend contours from 1979 to 2014. The latitudinal extent of Taiwan is shown by the green dashed lines.</td>
</tr>
<tr>
<td>2.3</td>
<td>Differences of ΔOLR (shadings) and (a) 250- and (b) 850-mb winds vectors between 1991–2002 and 1979–1990 for the 7 June–20 June period. (c–d) Same as (a–b) except for the differences between 2003–2014 and 1999–2002. (e) ΔOLR latitude (y-axis) and year (x-axis) distribution across the vicinity of Taiwan (white box in (b) and (d)) during the 7 June–</td>
</tr>
</tbody>
</table>
20 June periods overlaid with the linear trend contours from 1979 to 2014. The latitudinal extent of Taiwan is shown by the green dashed lines.

2.4 The EOF analysis of daily $\Delta$OLR from 1979 to 2014 for (a) EOFs 1, (b) PCs 1, (c) EOFs 2, (d) PCs 2, (e) reconstructed EOFs 1+2 and (f) reconstructed PCs 1+2. A 5-day moving average is applied prior to the EOF analysis.

2.5 Regression coefficients of the eddy streamfunction ($\psi_E$) June monthly mean regressed with PCs 1+2 for (a) 250 and (b) 850 hPa. Shadings indicate significance at p<0.05; arrows were added to illustrate the anomalous flow direction. H’s and L’s indicate anticyclonic and cyclonic anomalies, respectively. (c) Scatter diagram of the June eddy streamfunction ($\psi_E$) values at 250 and 850 hPa averaged from the respective domains outlined in (a) and (b) red boxes, with the last two digits of each year color-coded as indicated in the lower right.

2.6 Same as Figures 2.5a and 2.5b but regressed with PCs 1 for (a) 250mb and (b) 850mb; PCs 2 for (c) 250mb and (d) 850mb.

2.7 (a) The regression pattern of June $\psi_E$ 250 hPa with PCs 1+2 (contours) overlaid with the linear trend slopes of the June $\psi_E$ (shadings). Green-dotted areas indicate significance at p<0.05 for the regression. (b) Same as (a) but for the short-wave regime (i.e. zonal wavenumbers 5 and beyond) applied in each field, following the trending pattern as depicted by Wang et al. (2013b).

2.8 (a) Climatological 850-hPa velocity potential (VP) and divergent velocity (Vd) during Western North Pacific main tropical cyclone season (July-September), superimposed the locations of tropical cyclogenesis. (b) Yearly and seasonal distribution of the daily VP over convergent center (blue box in (a)) with the years in the y-axis and days in the x-axis. (c) Yearly distribution of the number of tropical cyclones with pentad scale with a 3-pentad moving average in Western North Pacific. (d) The time series of the monthly tropical cyclone number anomaly in July (blue) and August (red) relative to the1981-2010 with 5 years moving average. The shaded shows the deficit of two months. (e) same as (b) but with 20-100 days band pass filter and 5 years moving average. (f) the same as (c) but with 20-100 days band pass filter and 5 years moving average. In (e) and (f), positive/negative value presents convergent/divergent VP and high/low TC genesis anomalies.

2.9 (Yearly distributions of daily 850-hPa velocity potential (VP) with 30-90 days bandpass filter and 5 years moving averaged over convergent center in Western North Pacific for (a) NCEP1 reanalysis data with domain average over 125°E-165°E and 5°N-25°N, (b) JRA 55 reanalysis data with domain average over 120°E-150°E and 5°N-25°N , (c) NOAA 20th
2.10 (a)-(d) the EOF analysis of yearly distribution of daily 850-hPa velocity potential (VP) with 20-100 days band-pass filter, 5 years moving average, and 30 years moving average removed over the convergent center (Figure 1a, blue box) from 1 July to 10 September for NCEP1 (black line), JRA 55 (purple line), NOAA 20C (red line), and ERA 20C (blue line).

2.11 Power spectrum analysis of (a) NCEP1-PC1, (b) NCEP1-PC2, (c) NOAA 20C-PC1, (d) NOAA 20C-PC2, (e) ERA 20C-PC1, and (f) ERA 20C-PC2. Green line shows the 99% significance level.

2.12 The composites of 850-hPa velocity potential (shaded) and stream function (green line) at 2*10^6 m^2 s^-2 during 3 different periods when positive VP anomalies are in (a) July (1990-1996 and 2008-2014) and (b) August (1980-1986 and 1990-2005).

2.13 NCEP1 10-days average 850-hPa velocity potential (VP) regresses onto the (a) PC1 and (b) PC2, and 850-hPa stream function (ST) regressed on (c) PC1 and (d) PC2 from 1948 to 2016.

2.14 NCEP1-PC1 regress with every 2-month average of (a) 850-hPa velocity potential (VP), (b) Sea surface temperature (SST), and (c) 850-hPa streamfunction (ST) from 1948 to 2017.

2.15 (a) the sliding correlation with 61-day running window (from day-30 to day+30) between 850-hPa velocity potential (VP) over Western North Pacific (WNP) with MJO-RMM1 from 1979 to 2017. (b) the same as (a) but for MJO-RMM2 f. (c) The Yearly distributions of daily RMM1 (Top), RMM2 (bottom), and 850-hPa velocity potential (VP) from 1977-1991 with 20-100 days bandpass filter averaged over convergent center in Western North Pacific, applied with 5 years and 5 days moving average. (d) the same as (c) but for 2001-2015.

2.17 Same as Figure 2.1, but for CCSM4 historical simulation precipitation in recent 36 years with different forcing: (a,b) anthropogenic greenhouse gases (GHG), (c,d) natural including solar and volcanic forcing (Nat), and (e,f) anthropogenic aerosol (Aero). The yearly distribution of daily precipitation is the departure from seasonal means. Notice the rather weak Meiyu phase of rainfall than the observation, as well as the peak rainfall shift in (b) that is coincident with Figure 2.1b.

3.1 Winter season (November–March) precipitation time series over California; the inset map shows the domain and winter mean precipitation in California. The red dots indicate the 18 dry winters in which precipitation was less then 0.7 standard deviation below average as
shaded. (b) The composite anomaly of the 250-hPa geopotential height of the 18 drought winters. Hatches indicate significant level at p<0.05 for the anomaly. ................................................................. 75

3.2 (a) The first two leading EOFs (shaded) of winter season (November–March) Z_{250} within the 18 dry winters, superimposed with their composite Z_{250} anomalies (contour). (b) The corresponding PCs in relation to each of the 18 dry winters................................................................................ 76

3.3 The same as Figure 3.2, the first two leading EOFs (shaded) and PCs of winter (November – March) geopotential height (Z) within the 18 dry winters, superimposed with dry winters’ Z anomalies (contour), but for (a) Z_{200} hPa and (b) Z_{500} hPa. ........................................................................................................... 77

3.4 The Z_{250} and SST patterns from the 18 California dry winters regressed upon PC1 (a,b), PNA index (c,d), PC2 (e,f), and negative NPO index (g,h). Hatches indicate significant level at p<0.05 for the regression............ 78

3.5 The influence ratio from the regression of 18 drought years precipitation onto normalized (range from -1 to 1) (a) PC1, (b) PC2, (c), PNA, and (d) negative NPO decided by the mean precipitation anomaly during drought years. (e) and (f) are the same as (c) and (d) but for 1948-49 ~ 2014-15 wintertime precipitation................................................................. 79

3.6 Winter season (November – March) precipitation in California (blue line) overlaid with (a) the PNA index (orange line) and (b) the inverted NPO index (pink line) from 1948-49 to 2014-15, superimposed with the 18 dry winters as vertically shaded................................................................. 80

3.7 (a)-(d) Same as Figure 3.5 but for the regression patterns of the PDSI with the PNA and NPO indices. (e) and (f) Long-term regressions of the PDSI with the PNA and the inverted NPO over the 1948-2015 period. Hatches indicate significant level at p<0.05 for the regression................................. 81

3.8 The 30 ensembles mean of the EOF analysis of 250-hPa HGT from each ensemble’s drought years. (a) The PNA-like EOFs of HIS run, (b) the negative NPO-like EOFs of HIS run, (c) the PNA-like EOFs of RCP run, and (d) the negative NPO-like EOFs of RCP run. The contour shows the drought years anomaly in HIS run (a,b) and RCP run (c,d). ........................................... 82

3.9 averaged Z_{250} hPa regression patterns from 30 ensembles in drought years by regressing with the PCs of (a) PNA-like EOFs from HIS run, (b) negative NPO-like EOFs from HIS run, (c) PNA-like EOFs from RCP run, and (d) negative NPO-like EOFs from RCP run. ................................. 83

3.10 averaged SST regression patterns from 30 ensembles in drought years by regressing with the PCs of (a) PNA-like EOFs from HIS run, (b) negative NPO-like EOFs from HIS run, (c) PNA-like EOFs from RCP run, and (d)
3.11 (a) The first two leading EOFs of normalized winter season (November–March) precipitation within the 18 dry winters over California, (b) The corresponding PCs in relation to each of the 18 dry winters, and (c) The Z250 patterns from the 18 California dry winters correlated upon PC1 and PC2 of California precipitation. ................................................................. 84

3.12 The precipitation changes in California simulated by CESM1 with 30 ensembles. (a) The ensemble mean of frequency distribution of November-March precipitation in California ($P_{CA}$). The historical scenario (HIS run) includes 74 years before 2005, and the RCP 8.5 scenario (RCP run) includes 74 years after 2005. (b) The correlation coefficient between $P_{CA}$ and PNA (left) and between $P_{CA}$ and negative NPO (right) within drought years on ensembles’ HIS/RCP runs and observational data (OBS). The blue circles show the correlation coefficient of ensemble mean or observation, and the gray bar indicates 50% of ensemble spread. (c) The 30 years window sliding correlation between PNA and $P_{CA}$ over simulation period. The black solid line is the ensemble mean, the shaded areas indicate 50% of ensemble spread, and the red line is the observational data. (d) The same as (c) but for $P_{CA}$ and negative NPO........ 85

4.1 (a) Daily Arctic Oscillation index from 1 November to 30 April with 5 day moving average; the shaded period indicates the rapid tropospheric warming (RTW) case. (b) Time-height cross section of daily polar cap height (PCH) over Arctic region (north of 65°N) with 5 day moving average using R2 data. (c) Same as Figure 1b but for air temperature anomaly. ..................................................................................... 106

4.2 300-hPa HGT plotted at 8800 m (thick black contour) overlaid with SLP at 1045 and 1055 hPa (blue contours) on 21 December 2015 (left) and 21 January 2016 (right). The associated extreme events are indicated in the right figure. ................................................................................................. 107

4.3 (a) Predicted T2m anomaly at 0-month lead by the National Multi-Model Ensemble (NMME) for January 2016 and (b) observed T2m anomaly of January 2016 derived from NASA GISS data. Notice the Eurasia-Siberia region in which temperature anomalies between the predicted and observed are opposite. ..................................................................................... 108

4.4 Polar Cap Height (PCH) plotted from 1953 to 2015, following Figure 1b. Data prior to (after) 1979 were derived from R1 (R2). Refer to Table 1 for cases identified......................................................................................................................... 109

4.5 (top) Composite PCH and (middle) air temperature spanning the seven phases of the index cycle (see text) averaged north of 65°N and (bottom) corresponding AO index for (a) the January 2016 RTW case, (b)
composite RTW with 30 cases, and (c) composite SSW with 28 cases since 1979, using R2 data. The averaged duration and 1 standard deviation (purple) of total cases are labeled, and the phases representing 0 and 8 were derived from 4 day averages before 1 and after 7, respectively. The green dashed lines in Figures 4.2b and 4.2c outline 95% confidence interval for the t distribution.

4.6 Composite seven phases of 250 hPa velocity potential (VP) and the eddy momentum component of the E-P flux (EPF) computed from (a) the January 2016 RTW case, (b) historical RTW cases, and (c) historical SSW cases. Hatched areas of VP and shaded areas of EPF in Figures 3b and 3c signify 95% confidence interval for the t distribution.

4.7 Composite 7 phases of 250-hPa streamfunction computed from (a) the January 2016 RTW case, (b) the RTW cases, and (c) the SSW cases, following Figure 4.6. Hatched areas in (b) and (c) are significant at the 95% confidence level.

4.8 (a) The case frequencies of RTW (blue) and SSW (light green) cases plotted at every 9 years during the November–March period and occurrence of Arctic region (north of 65°N) temperature anomalies with the daily long-term mean removed at (b) 50 hPa and (c) 1000–500 hPa average exceeding the color-coded thresholds.

4.9 (a) Ensemble mean of the January sea level pressure (SLP) anomalies simulated by (a) SIC run and (b) SST run subtracted from control run. (c) January 2016 SLP anomaly derived from R2. (d) Linear trends in SLP from 1979 to 2016 multiplied by 37 years by using R2; the green contours indicate 95% confidence interval for the t distribution. The sliding spatial correlation over 35°N–90°N with an 180° longitude range centered at the x axis value (i.e., 90° to the west and 90° to the east) between (e) the 2016 anomaly and the two ECHAM5 runs and (f) the post-1979 linear trend and the two runs.

4.10 Same as Figure 4.9 (a)-(d) but for T2m anomalies and trends.

4.11 The frequencies of (a) 1000-hPa HGT exceeding 160 m over the Arctic region and (b) T2m below -21.5°C over Siberia, as well as (c) power spectrum of Arctic 1000-hPa HGT, from the ensemble means of three ECHAM5 experiments.

4.12 (a) The 12 cases composite of the averaged 250-hPa velocity potential (VP250) during the RTW transit phases from 1979 to 1997. (b) the same as (a), but for the 19 cases from 1998 to 2016. (c) the difference between recent two decades (b) and previous two decades (a) of the composite of RTW VP850. (d) The linear trend of VP250 in winter after 1979.

4.13 The anomaly of the 9-year average of (a) wintertime (November-March)
250-hPa velocity potential (VP250), (b) wintertime VP250 variation, and (c) the frequency of VP250 that is less than 1.5 standard deviation in winter.........................................................118

4.14 The 300-hPa wave activity flux (WAF) and streamfunction anomaly (ST) of the RTW phases 4 and 7 in (a) 12 RTW cases averaged from 1979 to 1997 and (b) 19 RTW cases averaged from 1998 to 2016. .................................119

4.15 The composite analysis of 132 stronger convection (250-hPa velocity potential < 1.5 standard deviation) days in the eastern tropical Pacific (240°E-285°E, 10°S-20°N) from 1979 to 2015 winter in (a) 250-hPa velocity potential and (b) 300-hPa wave activity flux and streamfunction....120
CHAPTER 1

INTRODUCTION

1.1 Introduction and background

Climate variability and climate change profoundly influence human activities. During the past two decades, the acceleration of anthropogenic global warming has perturbed global atmospheric and oceanic circulations and climatological patterns [Cox et al., 2000; IPCC, 2014]. Both natural climate variability and anthropogenic global warming have been investigated extensively in numerous studies, but our understanding of the causes and mechanisms of climate variability are still fragmentary [Cox et al., 2000; Cohen et al., 2014; IPCC, 2014]. Due to successive weather and climate extremes that have caused life and property loss globally during the early twenty-first century [IPCC, 2014], this dissertation primarily focuses on understanding the possible causes of weather and climate extremes.

Weather and climate extremes can result from natural climate variability and anthropogenic climate change. Anthropogenic warming has dramatically changed global temperature distributions and zonal circulations since the twenty century [Cox et al., 2000; Lorenz and DeWeaver, 2007]. Over the Arctic, warming near-surface temperatures and sea-ice loss have been linked to an increase in extreme weather events [Cohen et al., 2014]. In addition, global warming has been shown to change tropical circulation [Vecchi et al., 2006; Knutson et al., 2010] and is associated with extreme weather over mid-latitude regions [Yoon et al., 2015]. Figure 1.1 illustrates the connections of climate
variability and extreme events in different regions. Because of the importance of the connection of global climate variability, in this dissertation we investigate climate variability and weather extremes from a global perspective in order to understand the causes and influences of different extreme events.

In terms of natural climate variability, because of the uneven heating of Earth’s surface by the sun, the tropics are the main heat source of global circulations, which play an important role in influencing both mid-latitude and Arctic subseasonal variability [Fukutomi and Yasunari, 2002; Ding et al., 2011]. For example, the El Niño–Southern Oscillation (ENSO), a phenomenon of sea surface temperature (SST) variability over the tropical Pacific, has been known to be a prominent source of atmospheric and oceanic variability that influences global climate [McPhaden et al, 2006]. Other sources of natural climate variability can also combine with ENSO variability to influence global atmospheric circulations, resulting in a complex climatic system. For instance, Pacific Decadal Oscillation and North Pacific SST decadal variability have similar atmospheric and oceanic patterns [Mantua, 1997], making forecasts and predictions difficult. On the other hand, the tropical Western Pacific Warm Pool and associated deep convection have multiple interactions with ENSO to influence global circulation, indicating that the impacts of different ENSO events are inconsistent and highly uncertain [Emerton et al., 2017]. Meanwhile the Arctic is a cold source of global atmospheric circulation, which is strongly associated with boreal wintertime weather extremes due to the variations of polar vortex [Thompson and Wallace, 1998]. The causes of Arctic circulation variation are diverse [Cohen et al., 2014]. Thus, complete and deliberate examination of weather and climate events can help us to understand the interaction of different climatic features.
The acceleration of anthropogenic global warming and rising temperatures has resulted in an increasing frequency of extreme weather events [IPCC, 2014; Yoon et al., 2016]. To understand the impact of warming temperatures on climatological circulations, we focus on diagnosing extreme events from the historical reanalysis data. Moreover, analysis of the effects of different global warming scenarios on the climate model and the Coupled Model Intercomparison Project (CMIP) will further help to predict future circulation changes under warming temperatures. In this dissertation, three extreme weather events, including 2014–2015 water shortage in Taiwan, 2011-2015 drought in California, and an extreme cold surge event in 2015-2016 winter, that caused severe property damage and loss are investigated to understand their causes and discuss the links between extreme weather and different sources of climate variability. The results of this study could help to improve future weather forecasting and climate projections of extreme events.

To understand the background of the research topics, literature reviews of tropical, Arctic, and mid-latitude climate variability, and their connection to atmospheric circulation variations will be described in the following four sections. The last section of this chapter outlines the objectives of this dissertation.

1.2 Tropical ENSO and MJO variability

Tropical convection anomalies induced by variations in SST over the tropical Pacific (also known as ENSO) dramatically influence tropical, mid-latitude, and Arctic circulations through different teleconnection mechanisms [Diaz et al., 2001; Ashok et al., 2007]. Another important ENSO-related circulation pattern is the Pacific–North
American (PNA) teleconnection pattern [Hotel and Wallace, 1981], which is shown in Figure 1.2. The PNA profoundly influences the weather and climate over North America [Ropelewski and Halpert, 1986]. The PNA circulation can also influence Arctic surface temperature warming [Riddle et al., 2013; Johnson and Kosaka, 2016]. However, the response of climate to the PNA pattern is highly variable between different ENSO years because of the significant influences of many other climatic factors [Kumar and Hoerling 1998]. Thus, research concerning ENSO-related circulation should consider other possible features in global circulation in order to make better predictions.

Another important tropical climate variation is the 30–60-day intraseasonal Madden–Julian Oscillation (MJO) [Madden and Julian, 1971; Wheeler and Hendon, 2014] with a large-scale eastward-moving oscillation of deep convection, shown in Figure 1.3. The MJO has been observed to impact global circulation [Zhang, 2013]. Specifically focusing on the Western Pacific, the MJO-related convection variability impacts variations in East Asian Summer Monsoon (EASM) rainfall and tropical cyclone (TC) activity [Ding and Chan, 2005; Chen et al., 2009]. Moreover, many studies have shown that the tropical deep convection enhanced by the MJO convection anomaly over the Indian Ocean and Western Pacific can influence Rossby wave behavior, and affect mid-latitude and Arctic circulations [Cassou, 2008; Seo et al., 2016; Goss et al., 2016].

1.3 Arctic climate variability

In contrast to the tropics, the Arctic is a source of cold air mass connecting to global circulation. Variations in Arctic circulation are associated with changes in Arctic tropospheric temperature and the polar vortex. The polar vortex in the troposphere (also
termed extratropical westerly jet streams) profoundly affects the weather over North America and other mid-latitude continents. However, the development process of a warming or cooling event in the Arctic is complex and varies in its degree of impact on weather.

On the other hand, Arctic Oscillation (AO) is a measure of surface temperature variation in the Arctic. AO can be interpreted as exchanges of air masses between mid-latitudes and high latitudes, and changes in AO are associated with variations in the tropospheric polar vortex [Thompson and Wallace, 2000]. A strong positive AO indicates that Arctic temperature is cooler than normal with a strong polar vortex, suggesting a less wavy jet stream and fewer weather events in mid-latitudes. On the contrary, a strong negative AO event is associated with a weaker tropospheric polar vortex, leading to warmer temperatures in the Arctic, colder air outbreaks into mid-latitude regions from the Arctic, and more extreme weather events over mid-latitude continents in the Northern Hemisphere [Cohen et al., 2010; L'Heureux et al., 2010; Kug et al., 2015].

Several types of atmospheric variability have been known to correlate with variation in Arctic tropospheric temperatures and mid-latitude weather extremes. The tropical intraseasonal variations associated with energized convective activity over the Western Pacific, such as the earlier-mentioned MJO and ENSO variations, can correlate with subsequent changes in the polar vortex and AO via teleconnection and moisture transport from low-latitude regions to the Arctic regions [Lee et al., 2011; Riddle et al., 2013]. Furthermore, the rate of temperature increase in the Arctic is twice as fast as the global average, owing to the positive feedback mechanism of albedo and sea-ice extent,
which is called Arctic amplification (AA) [Pithan and Mauritsen, 2014; Screen and Francis, 2016]. The warming Arctic and sea-ice losses are also connected to changes in Arctic circulations (Cohen et al., 2014). One goal of this study is to document how Arctic circulations connect to tropical heat forcings and further impact weather and extreme events in the mid-latitudes.

1.4 Midlatitude Atmospheric Variations

Variations in tropical Pacific and Arctic circulations strongly influence the mid-latitude weather and climate over the North American continent [Vimont et al., 2001; Riddle et al., 2013]. From the winters of 2011–2012 to 2014–2015, California suffered severe drought. A year later, one of the strongest El Niño events since the 1900s failed to bring California out of the drought condition. Previous studies and experiments suggested that an El Niño winter was associated with a wet southwestern U.S.A. [Ashok et al., 2007; Hoell et al., 2016], but did not bring any relief to the California drought in this case. Subsequently, California flipped to a very wet winter in 2016–2017 without any well-known tropical and oceanic forcings. This suggests that any connections between tropical relative forcings (ENSO) and MJO-induced deep convection are not linear and research must consider this interaction with other sources of climate variability in order to provide better predictions. In addition, the influences of the aforementioned Arctic variability similarly play an important role in mid-latitude climate. More diagnostic analyses of extreme weather events and their possible mechanisms are needed and are important for advancing our understanding of climate change.
1.5 Subseasonal Variability

Subseasonal variations embedded in the aforementioned long-term atmospheric and oceanic variability influence short-term weather and extreme weather events over different continents. The mid-latitude weather pattern variations are an example of typical subseasonal variability [Teng et al., 2013]. Over the mid-latitude regions, the subseasonal variations of the tropical convection, Arctic circulation, and mid-latitude internal variability appear to all combine to influence the mid-latitude weather and climate [Bjerknes and Solberg, 1922; Trenberth et al., 1998; Cohen et al., 2014]. In addition, the EASM exhibits pronounced intraseasonal variability and is the main climate system that influences the water supply over Eastern Asia. Extreme weather and TCs impact these regions from April through October, albeit with extensive temporal variation [Chang et al., 2011]. Furthermore, intraseasonal MJO-induced deep convection over the different tropical regions and subseasonal variations of the PNA teleconnection pattern have been linked to mid-latitude and Arctic circulation variation [Zhang, 2013]. Therefore, the study of the subseasonal variation of Arctic circulation should connect with tropical variability to better understand the interactions of critical atmospheric and oceanic variations.

1.6 Objectives

The aim of this dissertation is to investigate extreme weather events and their possible connections to climate processes. To accomplish this, the diagnostic analysis begins by analyzing the observed reanalysis and climate model data of extreme weather events. We examine the natural climate variability and variability under global warming temperatures in order to better predict future changes in climate. Because of the
importance of the tropical heat forcing, Arctic cold source, and mid-latitude climate on global circulation, which induce extreme weather events and influence the water supply in different regions, a diagnostic analysis of possible mechanisms and signal precursors is imperative. The following research case studies are employed to accomplish the primary aim: (1) variations in the EASM on the timing changes of early summer rainfall and peak low-level convergence during July and August, (2) possible impact features on California drought variations, and (3) Arctic subseasonal temperature variation and its possible connections.

The first case study was motivated by the 2014–2015 water shortage in Taiwan due to inactive TC activity in August 2014 and late monsoon active-phase rainfall in the early summer of 2015. The period from May to October is rainy season in Taiwan. The variations of summer monsoon rainfall and tropical cyclone rainfall in the season influence annual water resources in Taiwan. The objective of this study is to investigate the possible climate factors that are associated with water shortages. The findings will assist decision-makers to manage the risk of water resources. We hypothesized that, besides the interannual variability, long-term climate variability such as anthropogenic warming and natural variability could contribute to extreme weather events. By analyzing the observed reanalysis data, climate variation during these two seasons was examined. The details of this study are found in Chapter 2.

The second study was inspired by the five-year drought in California from 2011 to 2015, when a strong ridge of high pressures covered the western United States, thus inducing the droughts. However, each year, the ridge of high pressure presented various large-scale circulation patterns, indicating that different climate processes could
contribute to the drought in California. From the analysis of historical data and climate model projections, we seek to investigate the diversity of drought circulation with a different approach to previous studies about drought variation in California, which is shown in Chapter 3.

The last study is motivated by extreme cold surge events in January 2016 over the East American and East Asian continents. The events even influenced weather over the tropical Indo-China Peninsula. To investigate the causes of this extreme event and the possible climate processes, we hypothesized that similar extreme weather and circulation variations have happened before, and some important factors trigger subseasonal weather extremes under the warming Arctic climate. We aim to determine the impact processes in order to improve weather forecasting and to better understand the future tendency of extreme weather events. These findings are described in Chapter 4. The final discussions and conclusion are presented in Chapter 5.

References


Madden, R. A., and P. R. Julian (1971), Detection of a 40–50 day oscillation in the zonal


Figure 1.1 Schematic of the connection between global circulation and weather as well as climate extremes.
Figure 1.2 Schematic of ENSO-relative Pacific-North America (PNA) upper tropospheric geopotential height teleconnection anomalous pattern (solid lines) during a Northern Hemisphere winter which falls within an episode of warm sea surface temperature in the equatorial Pacific (shaded). The arrows in darker type reflect the strengthening of the subtropical jets in both hemisphere. The arrows in lighter type depict a min-tropospheric streamline as distorted by the anomaly pattern. Taken from Horel and Wallace [1981].
Figure 1.3 The structure and evolution of the Madden-Julian Oscillation (MJO) of DJF composite OLR and 850-hPa wind vector anomalies. Shading levels denote OLR anomalies less than -7.5, -15, -22.5, and -30 Wm$^{-2}$, respectively, and hatching levels denote OLR anomalies greater than 7.5, 15, and 22.5 Wm$^{-2}$, respectively. The magnitude of the largest vector is shown on the bottom right, and the number of days (points) from 1979 to 2001 falling within each phase category is given. The nominal time for transition between each of the numbered phases is 6 days but can vary from event to event. Taken from Wheeler and Hendon (2014).
CHAPTER 2

INTERDECADAL CHANGES OF EARLY SUMMER PRECIPITATION AND MULTIDECADAL TIMING VARIATIONS OF SUMMER LOW-LEVEL CONVERGENCE ON THE EAST ASIAN SUMMER MONSOON

Abstract

In the Western North Pacific (WNP), the East Asian Summer Monsoon (EASM) is found to have both interdecadal and multidecadal timing variations. EASM has shown a marked timing shift of active convection since 1979 in early summer, and a multidecadal timing variation of atmospheric peak low-level convergence during July-August. Diagnostic analysis indicates that active convection over Taiwan has occurred later in the season, from late May to early June, with a tendency of increasingly intense rainfall. This timing shift of convection results from a southward migration of Meiyu rainband, driven by an upper-level cyclonic anomaly over eastern China and a lower-level anticyclonic anomaly in the subtropical Western Pacific. These two circulation patterns enhance both the moisture transport and baroclinic forcing. The role of Western Pacific warming and anthropogenic greenhouse gases in these changes is suggested.

During the tropical cyclones season, the atmospheric low-level convergence is the main factor influencing the tropical cyclone (TC) genesis. It is observed that the peak seasons

---

This chapter includes two studies one was published, and another one is in preparation:


of TC genesis and low-level convergence have shifted since 1998, from mid-August to late-July, and this shift affects the low-frequency change in the TC number. A multidecadal frequency peak at 20 years is revealed in the timing variation the tropical intra-seasonal oscillation (ISO). This multidecadal variation is correlated to a large-scale low-level convergence anomaly in the Western Pacific in spring. Diagnostic analysis did not identify any prominent atmospheric and oceanic variations associated with this multidecadal variation in the seasonality of low-level convergence. The possible mechanism might result from the low-frequency natural variability in the timing of the ISO.

2.1 Introduction

The East Asian summer monsoon (EASM) undergoes an active–break–revival sequence and the associated migration of the rainbands makes the timing of each phase geographically unique [Chen et al., 2004]. This distinct lifecycle of EASM regulates rainfall and water supply in several Asian countries, including Taiwan. Located in the central region of EASM, Taiwan covers 36,000 km² of complex terrain with a population approaching 24 millions (location shown in Figure 2.1a, inset). The active phase of EASM (Meiyu) produces the first influx of substantial water for agricultural, industrial and residential uses. Wang and Chen [2008] indicated that the active-phase EASM (interchangeable with Meiyu hereafter) contributes to ~60% of Taiwan’s early-summer rainfall. On the other hand, the revival phase of EASM is dominated by tropical cyclones. The Western North Pacific (WNP) features the highest population of tropical cyclones (TCs) compared to any other ocean basin. Climatologically, the TC frequency
peaks in the month of August with about 6 cyclones. This seasonal increase in the August TC activity accompanies the revival phase of the East Asia-WNP summer monsoon lifecycle [Chen et al., 2004]. The phases of EASM relative to Taiwan are displayed in Figure 2.1a by the outgoing longwave radiation (OLR) averaged within $119^\circ -122^\circ E, 21^\circ -25^\circ N$; here, OLR is shown as departure from $235Wm^{-2}$ to approximate convective rainfall regime, denoted as $\Delta$OLR (=235 – OLR). This feature is critical because, as of April 2015, Taiwan underwent the most severe drought in its 67 years of recorded history and yet, the arrival of Meiyu mitigated the drought situation. Moreover, in August 2014, WNP underwent a remarkably quiet TC season without a single TC. The dramatic reduction in the TC activity created a “typhoon drought” in some WNP islands, such as Taiwan that relies on TC rainfall. Subsequently, the water deficit in 2014 TC season and the late of active-phase summer monsoon resulted in severe water shortage (http://www.latimes.com/world/asia/la-fg-taiwan-drought-20150510-story.html).

In terms of the active-phase summer monsoon, predicting the timing and strength of active-phase EASM at longer range (>2week) remains a challenge, making drought adaptation and planning difficult [M.-M. Lu, Central Weather Bureau, 2015, personal communication]. The Meiyu rainband is driven by the mid-tropospheric warm advection and transient eddies that are steered by the westerly jet, and these circulations induce instability and adiabatic ascent while the tropical warm pool supplies the moisture [Chen et al., 2004; Sampe and Xie, 2010]. Previous studies have indicated that interannual variability of the EASM circulations is linked to the Tibetan Plateau thermal conditions and Indian Ocean sea surface temperature (SST) anomalies [Li and Yanai, 1996; Zhao et
al., 2010; Liu and Wang, 2011; Hu and Duan, 2015]. These processes are complicated by the varying mid-tropospheric temperature advection within the Meiyu rainband [Kosaka et al., 2011; Okada and Yamazaki, 2012]. However, few studies have focused on the interdecadal variability of Meiyu. Among these, Li et al. [2010] found that EASM has shifted southward since 1958 probably due to the meridional asymmetric warming between the South China Sea (SCS) and East Asian continent. Luo and Zhang [2015] reported that peak Meiyu rainfall in southern China has tended to arrive later since 1993 due to weakened low-level southwesterly winds. Focusing on Taiwan, Huang and Chen [2014] observed a transition of Meiyu rainfall from the predominately frontal regime to an increase in the diurnal convection regime. Regardless, a mechanistic explanation of the Meiyu’s interdecadal variation is lacking; this is analyzed herein.

With regard to the 2014 August TC reduction, Hong et al. [2016] suggested that the positive sea surface temperature anomaly (SSTA) in the eastern North Pacific reduced the Walker circulation and the resultant subsidence anomaly over the WNP is enhanced by an eastward propagating intraseasonal oscillation (ISO) episode, forming a marked divergent anomaly that suppressed TC formation. The relative inactivity in TCs during the recent decade is noted by Liu and Chan [2013] and Yeh et al. [2008], who suggested that the strengthened vertical wind shear and intensified subtropical high may be to blame.

Intrigued by the remarkable inactivity of TCs in August 2014 and how this event is linked to the recent decline in the WNP TCs, this study is focused on the timing/seasonality change of TC genesis. TC genesis in the WNP is affected by multiple factors including low-level convergence, the monsoon trough, warm sea surface, and
vertical wind shear [Harr and Elsberry, 1995; Chen et al., 2004; McBride, 1995]. The seasonal variations of low-level convergence and TCs are modulated by the tropical ISO [Sobel and Maloney, 2000; Huang et al., 2011; Maloney and Dickinson, 2003; Chen et al. 2009; Nakano et al. 2015]; meanwhile, the ISO and TC activities are affected by interannual variations like the El Niño–Southern Oscillation (ENSO) [Chen et al., 2006; Pohl and Matthews, 2007; Li et al. 2012; Kim et al., 2008]. Consequently, both the intraseasonal and interannual variations, as well as lower-frequency or decadal variability, collectively affect the strength and position of the monsoon trough and subsequent TC geneses [Sobel and Maloney, 2000; Wang and Chan, 2002; Camargo et al., 2010].

Amid these previous studies of the Meiyu rainband and WNP TC activity, it appears that none has investigated the systematic timing change. This study seeks to investigate possible climate variability that affects water resources over the EASM, and this research reports apparent interdecadal timing shift in the active-phase monsoon rainband and multidecadal timing fluctuations in the TC genesis. The study is arranged as follows: Section 2 presents data sources; Section 3 shows the analysis results of interdecadal variation of Meiyu; Section 4 shows the analysis results of multidecadal variation of WNP TC activity; Section 5 is related discussions and summary.

2.2 Data

The following data sets are utilized: (1) the 1.0°-resolution daily OLR Version 1.2 produced by the National Oceanic and Atmospheric Administration (NOAA) [Lee and NOAA CDR program, 2011] from 1979 to 2014 with the missing values during May–
June 1985 filled by the NOAA interpolated OLR [Liebmann and Smith, 1996], (2) the monthly NOAA Extended Reconstructed SST Version 4 [smith et al., 2008], (3) Joint Typhoon Warning Center (JTWC) best tracks (4) the ECMWF post-1979 reanalysis data at a 1.0° resolution (ERA-Interim) [Dee et al., 2011], (5) the ECMWF 40-year Reanalysis (ERA-40) from 1958 to 1979 at a 2.5° resolution [Uppala et al., 2005] to merge with ERA-Interim for a longer-term analysis, (6) the National Centers for Atmospheric Research/National Center for Atmospheric Research reanalysis (NCEP1) daily variables from 1948 to 2016 [Kalnay et al., 1996], and (7) For the comparison of long-term variability among the different reanalysis products, we also analyzed the Japan Meteorological Agency Japanese 55-year Reanalysis (JRA-55) [Kobayashi et al., 2015], National Oceanic and Atmospheric Administration 20th Century Reanalysis Version 2 (NOAA 20C) [Compo et al., 2006, 2011; Whitaker et al., 2004], and European Centre for Medium-Range Weather Forecasts Re-Analysis 20th Century (ERA 20C) reanalysis data [Poli et al., 2016].

2.3 Interdecadal variation of active-phase summer monsoon

The long-term change in the active-phase EASM is examined by analyzing the daily $\Delta$OLR in Taiwan from mid-May to mid-July (x-axis) for each year from 1979 to 2014 (y-axis); this is plotted in Figure 2.1(b). The use of $\Delta$OLR compensates for the lack of long, stable record of daily precipitation. Here, $\Delta$OLR is subject to a 5-day and 5-year running mean to focus on the predominant intraseasonal variability that drives the EASM lifecycle [Chen et al., 2004]. The peak of $\Delta$OLR has undergone a timing shift from mostly late May before the 1990s to predominantly early June. There is also a tendency
for ΔOLR to become stronger and more concentrated in mid-June (10th–15th) after 2003. To illustrate this change, we compute the linear trend of ΔOLR for each day from 1979 to 2014 and superimpose it on Figure 2.1(b) as contours. Apparently, ΔOLR has decreased by 20 Wm\(^{-2}\) in late May accompanied by an increase of 30 Wm\(^{-2}\) in mid-June, estimated from the linear trend. Noteworthy is the change in the convective time span that has reduced from 3 weeks before 2003 to less than 2 weeks afterwards, suggesting more intense rainfall occurring within a shorter period of time. This feature echoes the finding of Huang and Chen [2014] that frontal rainfall regime in Taiwan has gradually been replaced by diurnal convection regime in May and June.

The large-scale circulation and precipitation anomalies associated with the timing shift of Meiyu are examined by two epoch differences of the 250- and 850-hPa winds and ΔOLR: (1) between 1991–2002 and 1979–1990 to depict the timing shift and (2) between 2003–2014 and 1991–2002 to depict the precipitation intensification (these periods are indicated by arrows in Figure 2.1b), in June. The circulation and ΔOLR anomalies during 7th–20th June are plotted in Figure 2.2. In the earlier period, a robust upper-level cyclonic anomaly forms over eastern China (Figure 2.2a, ‘L’) while a marked low-level anticyclonic anomaly extends from the SCS across the Philippine Sea (Figure 2.2b, ‘H’). Combined, these circulations induce strong southwesterly flows coupled with upper-level westerlies, promoting baroclinic instability in and around Taiwan. Correspondingly, a substantial increase in ΔOLR is observed in the northern SCS stretched across Taiwan, signifying an intensification of frontal rainfall regime. These circulation anomalies possibly reflect a stationary wave pattern superimposed on the westerly jet stream that was found to be influenced by the mechanical effect of the
Tibetan Plateau [Wu and Chou, 2013].

For the latter period (after 2003), Figure 2.2c shows that the upper-level westerly winds enhance slightly, while a low-level cyclonic circulation appears in the vicinity of Taiwan (Figure 2.2d, ‘L’). Combined, these circulation changes delineate a meridional migration of Meiyu in the context of interdecadal variation. The change in ΔOLR is also substantial as it is shifted further south adjacent of the Philippines covering only the southern part of Taiwan. To clarify this implication, we plot in Figure 2.2e the latitude-time section of ΔOLR across Taiwan during 7th–20th June. Apparently, positive ΔOLR north of Taiwan has migrated southward from 26° to 20°N. Consequently, what used to be a relatively dry spell in Taiwan (i.e. between 18° and 24°N) has become increasingly convective in recent years. As is shown in Figure 2.3, Supporting Information, the earlier period of 24 May–6 June undergoes a decrease in convective activity as a result of this ΔOLR migration. These results provide a geographical reference for the timing change of Meiyu.

In order to connect the reported timing shift with the large-scale circulation change, we adopt a method designed to delineate the yearly evolution of a daily variable, following Wang et al. [2014]. This method uses the empirical orthogonal function (EOF) of the covariance matrix of ΔOLR over Taiwan, by treating ΔOLR’s daily interval as the vector and its yearly interval as the time series. After applying a 5-day moving average (to capture the predominant intraseasonal variability of EASM), we obtain a set of EOFs representing the daily variation of ΔOLR and a set of principal components (PCs) representing the yearly variation. The first two EOFs are shown in Figures 2.4a–d representing the amplification of the temporally displaced ΔOLR, constituting
collectively 32.7% of the total variance. The EOF 1 (Figure 2.4a) and EOF 2 (Figure 2.4c) show positive values in mid-June with an increasing trend of the PCs (Figures 2.4b and 2.4d), suggesting a tendency for enhanced convective activity and its timing shift in Taiwan. Next, we combine these two leading modes to reconstruct the ΔOLR changes in Taiwan while filtering out less relevant signals [Van den Dool, 2007]. The combinations of EOFs/PCs 1+2 are shown in Figure 2.4e and 2.4f. The distribution of EOFs 1+2 indicates maximum ΔOLR in mid-June and minimum ΔOLR in late May, and this feature has intensified as shown by the increasing trend in PCs 1+2 (significant at p<0.05). Consequently, PCs 1+2 form an index enabling us to compare the change in subseasonal variability against interannual variations of any given variable.

By regressing PCs 1+2 upon the eddy streamfunction field (i.e. removing the zonal mean) for the month of June, the resultant regression coefficients depict the anomalous circulations accompanying the increased ΔOLR in Taiwan during mid-June. Figures 2.5a and 2.5b show such circulation patterns at 250 and 850 hPa, depicting westerly (southwesterly) anomalies that prevail over Taiwan at the upper (lower) level. By comparison with Figure 2.2, these circulation features correspond well with the upper trough in eastern China and the Western Pacific anticyclone. The similarity of circulation patterns between Figures 2.5a and 2.5b and Figure 2.2 also suggests that the anomalous circulations leading ΔOLR to become more active in mid-June resulted from two sources: deepening of the upper-level trough northwest of Taiwan and strengthening of the anticyclone in the subtropical western Pacific. In Figure 2.6, we show the individual regressions of PCs 1 and 2 and their resultant circulation pattern, which reveal similar synoptic processes.
We next compare the interannual variations of the June stream function between
the upper-level cyclone and the lower-level anticyclone, using values averaged from their
center areas (domain outlined in Figure 2.5a and 2.5b). The variations of these two
circulation features are not correlated (r <0.16), as illustrated by the scatter plot of Figure
2.5c. In other words, the circulation patterns in response to PCs 1+2 (i.e. increased
ΔOLR) could only appear in the second quadrant of the scatter diagram, i.e. when
negative 250 hPa values (trough) and positive 850 hPa values (ridge) coexisted.
However, by adding the years onto the scatters, there is a discernible change in that the
concurrence of the strengthened Western Pacific anticyclone with the deepened eastern
China cyclone has increased after 1997 (indicated as red). This result is intriguing in that,
although these two levels of circulation do not correlate, in the long run they have
become increasingly cohesive in producing precipitation along the SCS-Taiwan corridor
in the month of June.

A tendency has been observed in June for the low-level anticyclonic anomaly in
subtropical Western Pacific and upper-level cyclonic anomaly in eastern China to occur
together more frequently. This feature promotes frontal instability and subsequent
convection in early June over Taiwan, delaying its Meiyu season. For the upper level,
previous studies analyzing the change in mid-latitude stationary waves have noted an
amplified short-wave regime and associated increases in weather extremes [Screen and
Simmonds, 2013; Teng et al., 2013; Wang et al., 2013b; Screen and Simmonds, 2014].
Other research [Wang et al., 2013b; Cho et al., 2015] has indicated an intensification of
the Eurasia-South Asia short-wave train in the month of June [Yasunari et al., 1991;
Ding and Wang, 2005]. This reported wave pattern consists of (from west to east) a
deepened trough in western Nepal, a strengthened ridge over Bhutan and an enhanced trough over eastern China – these are shown in Figure 2.7. The cause of this changing wave-train pattern is under debate, and our testing of SST regression with PCs 1 + 2 (not shown) does not reveal any robust linkage with any known climate mode. However, the SST in subtropical Western Pacific has tended to warm by 40% associated with the 30-W m$^{-2}$ increase of ΔOLR in June, based on linear regression. This increase in local SST coincides with the ongoing warming trend in the Western Pacific.

2.4 Multidecadal variation of TC activity

Climatologically, the low-level convergent circulation, illustrated in Figure 2.8a by the mean 850-hPa velocity potential (VP) during June-October and the divergent winds, exhibits a spatial correspondence of the convergent center with the TC genuses. To examine the variation in the timing of the WNP’s low-level convergence, we plot in Figure 2.8b the yearly and seasonal distributions of the daily 850-hPa velocity potential (VP) averaged over the convergent center (10°N-30°N, 120°E-160°E; Figure 2.8a blue box) during June-October over the period of 1948-2017 (with the years in the y-axis and days in the x-axis). The corresponding 5-day-mean (pentad) TC counts are plotted in Figures 2.8c; here, a TC genesis on January 2nd is counted into the January 1-5 period, and so forth. Figure 2.8c also reveals the post-1999 decrease in the August-September TC numbers as noted by earlier studies [Liu and Chan, 2013; Yeh et al., 2008]. Visual examination of Figures 2.8b and 2.8c reveals a discernable timing shift within the timescale manifest of a multidecadal variation. Within the season (x-axis), the timing shifts in TC genesis and VP resemble episodes of intraseasonal variation such as
one indicated by the green dashed lines. Combined with the yearly evolution (y-axis), these intraseasonal episodes revealed in TC genesis and VP appear to have advanced earlier since the late 1990s. Such timing variation implies a change in the peak TC season from August to July. By plotting the July and August anomalies of TC genesis in Figure 2.8d (applied with a 5-year moving average), there appears to be a multidecadal variability not only in the TC number but also the seasonality in which July TCs tend to oscillation oppositely with August TCs every decade or two. The timing of VP and TC genesis tends to persistently advance or delay every few years, with a tendency towards an advancement in the timing of TC genesis in the recent decade.

Next, to highlight the predominant intraseasonal variability, both the VP and TC numbers are bandpass-filtered with 20-100 days and applied with a 5-year moving average to focus on the lower frequency; this is shown in Figures 2.8e and 2.8f in the context of anomalies. The timing variation in TC genesis is coherent with the intraseasonal episodes of VP, supporting the inference made from Figures 2.8b and 2.8c. The result thus far suggests that the seasonality of the large-scale, low-level convergence in the WNP may slowly become weaker in August yet stronger in July during some 20 years, and vice versa during other decades. This tendency is supported by the multiple reanalysis data sets as shown in the supplemental Figure 2.9 with JRA-55, NOAA 20C, and ERA 20C, showing consistent seasonality variation in the VP anomaly.

To more quantitatively depict this seasonality variation, the empirical orthogonal function (EOF) analysis is conducted on the filtered VP (i.e. using data from Figure 2.8e) to examine how the time-mean circulation anomalies affect the seasonality of weather patterns, following the method of Wang et al. [2014, 2016]. Here, the EOF analysis is
performed by treating VP’s daily interval as a spatial dimension and its yearly interval as time. The result leads to a set of EOFs representing the daily variation and another set of principal components (PCs) representing the yearly variation. By focusing on July 1 through September 10, the first two EOFs of filtered VP represent the amplification of geographically displaced climatology of VP.

As shown in Figure 2.10a, EOF 1 presents a predominant intraseasonal variation with an opposite phase between July and August, which is temporarily in-quadrature (out-of-phase) with the climatological VP evolution (shaded curve) and thereby suggests either an advancement or delay. EOF 2 (Figure 2.10c) is in-phase with the climatological VP and so, depicts either an amplification or weakening of the climatological variation. Judging from the yearly evolution in the PC time series (Figures 2.10b and 2.10d), the timing of peak convergent VP has first shifted to early August during the early 2000s (PC2) and further advanced into late July after year 2010 (PC 1). These results are consistent among the four reanalysis datasets analyzed, as shown in Figure 2.10. By subjecting these PC time series to the power spectral analysis, the result (Figure 2.11) shows that the low-frequency shifts of the VP seasonality occur every 20 years or so.

To investigate the influence of low-level convergence in different periods on the TC related circulation, Figure 2.12 illustrates the multidecadal change in the divergent circulation and the corresponding monsoon trough position, represented by the 850-hPa VP and ST during two episodes: (A) 1990-1996 and 2008-2014 (with positive VP anomalies in July; Figures 2.10a and 2.10b) and (B) 1980-1986 and 1990-2005 (with negative VP anomalies in July). The evolution of the WNP monsoon lifecycle is
illustrated by three periods from mid-July to late July and then to mid-August (dates indicated in the figure). During episode A (early WNP monsoon), the low-level convergence is most enhanced in late July associated with a robust monsoon trough in the WNP. During episode B (late WNP monsoon), the late-July convergence center is weakened but it lasts longer, through mid-August alongside a strengthened monsoon trough. During either episode, the eastern extension of monsoon trough fluctuates with the variation of the VP center; these features apparently module the seasonality of TC genesis.

In the subsequent analysis, we examine the physical meaning of the multidecadal variation reflected in the PC time series by examining their associated change in the large-scale mean state. By regressing the bandpass-filtered VP with the PCs, the resultant pattern reveals an eastward propagation of low-level convergence (Figures 2.13a and 2.13b) similar to the basic features of the Madden-Julian oscillation (MJO) [Madden and Julian, 1971;1972]. The regressed streamfunction pattern (Figures 2.13c and 2.13d) also shows the MJO-like teleconnections, consistent with Knutson and Weickmann [1987]. In addition, studies have shown that the MJO activity varies with the mean-state variations in the WNP VP/TC [Maloney and Hartmann, 2001; Hsu and Wang, 2001].

To investigate the possible forcing of the multidecadal variability in the timing of VP and TCs, Figure 2.14 shows the regressions of the 2-month mean VP, SSTA, and the 850-hPa streamfunction (ST) with the same PC1 time series, performed during March through July. (The regression variables are applied the same filters in the multidecadal PC1.) Depending on the polarity, the regressed VP delineates the predominant pattern leading up to the shifted timing of peak convergence either in July or in August. As
shown in Figure 2.14a, a zonal wavenumber-1 structure emerges in April and persists through June with a center in the western Pacific warm pool, which dissipates by June. This result implies that there is a springtime “precursor” in the large-scale divergent circulation that affects summertime intraseasonal variation. However, the SST regressions in Figure 2.14b do not reveal any outstanding pattern in any season, suggesting that the more striking VP pattern and its associated variation of Pacific subtropical height (Figure 2.14c) during March-May is mostly of atmospheric origin. Furthermore, this multidecadal variation modulates the westward extension of subtropical high and monsoon trough during July-August (Figure 2.14c), echoing the circulation variations in different timing of peak low-level convergence in Figure 2.12.

In order to evaluate the relationship between the WNP VP and MJO, we compute the sliding correlation between the Real-time Multivariate MJO index 1 and 2 (RMM1 and RMM2) [Wheeler and Hendon, 2014] and the area averaged WNP VP (Figure 2.8a blue box), with a 61-day running window throughout the year as the x-axis for each year from 1979 to 2017 in the y-axis (Figures 2.15a and 2.15b). Since the RMM1-related VP variation (represented by outgoing longwave radiation (OLR) in Wheeler and Hendon [2004]) is located at the Maritime Continent near our studying region, the sliding correlations are significantly ($r =+/- 0.32, p < 0.05$) positive in all seasons. On the other hand, the RMM2 sliding correlation reveals an annual cycle that has a positive (negative) correlation in boreal summer (winter), suggesting that the phase of RMM2 is partially related to the variation of WNP VP comparing to RMM1 that has high correlation annually. The possible explanation is that the RMM2-related VP variations (represented by OLR in Wheeler and Hendon [2004]) cover part of the Western Pacific with weak
variation comparing to the main variation over the Indian Ocean (Wheeler and Hendon [2004]). Moreover, Zhang and Dong [2014] indicates the MJO variation is associated with the Intertropical Convergence Zone (ITCZ), suggesting the weak correlation between RMM2 and WNP VP presents an annual cycle and only shows significant when the ITCZ reaches farthest north and south.

Next, to investigate the possible connection between the timing variation of WNP VP in summer and the precursor low-level convergence in spring, we plot the yearly distribution of daily RMM1, RMM2, and WNP VP from January to December during two periods when the low-level convergence gradually shift from July to August (Figure 2.15c) and from August to July (Figure 2.15d). As the results in Figures 2.15a and 2.15b, the RMM1 has similar distributions to the WNP VP, and RMM2 only shows similar patterns in summer. We further close examine the intraseasonl variations of them; from the indicating solid line (convection anomaly) and dashed line (divergence anomaly) in Figures 2.15c and 2.15d, we can find that the distinct intraseasonal variation of WNP VP evolves in spring and continues into summer. Moreover, the phases of RMM1 and WNP VP keep consistent from spring to summer, implying the phases of intraseasonal WNP VP and RMM1 in spring can provide an estimate to the phases in summer. On the other hand, previous studies indicate the tropical large-scale convection and tropical low-level heating have been widely regarded as to impact the MJO [Zhang, 2005; Miura et al., 2007; Zhang and Song, 2009; Ling and Zhang 2011]. The multidecadal large-scale low-convergence single in spring (Figure 2.14) is apparently relevant to the consistent phase of WNP VP and RMM1 since spring (Figures 2.15c and 2.15d).
2.5 Discussions and Summary

The variation of lower-level circulations in the Western Pacific has been widely documented. Yet, most studies only focused on the typical summer season of June–August, rather than the seasonal transition of May or June as well as July or August. Nevertheless, those studies have uniformly found an interdecadal variation of active-phase monsoon rainband and a multidecadal timing variation of peak low-level convergence in WNP. In the early summer, there is a link between the strengthened North Pacific subtropical anticyclone the increased SST under the anthropogenic global warming, which leads to late peak precipitation. The result also shows in the model simulation with the historical single-forcing experiment of the Community Climate System Model Version 4 (CCSM4) derived from the CMIP5 archive [Taylor et al., 2009]. By reproducing Figure 2.1 using daily precipitation output of CCSM4, which is shown in Figure 2.16, it is observed that only the anthropogenic greenhouse gases (GHG) forcing simulates the timing shift of the active-phase EASM in a way similar to the observation. Neither the natural forcing nor the aerosol forcing generated any persistent change in the occurrence of peak rainfall. The preliminary result of Figure 2.16 suggests a possibility that anthropogenic GHG can influence the timing change of Meiyu rainfall in Taiwan, which is consistent with early studies that the strengthened subtropical anticyclone adds thermal contrast between land and ocean [Li et al., 2012] and further warms the northern Indian Ocean [He and Zhou, 2015] while enhancing thermal contrast in the subtropical Western Pacific [Wang et al., 2013a]. These reported changes in oceanic thermal property and land-sea contrast have a detectable anthropogenic footprint and could be linked to the finding of this study.
On the other hand, during July-August, the analysis result shows that the timing of peak low-level convergence has a multidecadal variation shifting between July and August with a frequency of ~20 years, affecting the seasonality of TC genesis. The low-level convergence has shifted from August to July since 1998, inducing the remarkable inactivity of TCs in August 2014 and water shortage in WNP islands. The timing variation of low-level convergence is associated with a corresponding timing shift in the tropical ISO (MJO). Moreover, the resultant regression maps show this multidecadal variation is correlated to a large-scale low-level convergence anomaly in the Western Pacific in spring and early summer, but the sea surface temperature does not reveal any eminent patterns. The multidecadal variation presented here did not identify any linkage of the known atmospheric and oceanic variations in the seasonality of low-level convergence. We did find that the MJO phase in late spring can be a precursor of the peak low-level convergence to occur either in July and in August.

The diagnostic results of the multidecadal and interdecadal variations profoundly influence the timing of peak precipitation and the water management in the EASM regions. These variations result from anthropogenic and natural climate variability changing the timing of upper trough of high pressure system and tropical low-level convergence respectively. Because of the high density of population and the complicated circulation system in EASM regions, more research is needed to improve the forecast. Subsequent analysis using the full archive of CMIP5 outputs will be the focus of future studies to estimate the future changes and understand the possible mechanism of the natural low-frequency variation. Moreover, EASM system has been shown to associated with mid-latitude and Arctic circulation. The teleconnection impacts form EASM should
be investigated to better understand the global circulation system.

Reference


Yasunari T., A. Kitoh, and T. Tokioka (1991), Local and remote responses to excessive
snow mass over Eurasia appearing in the northern spring and summer climate – a study with the MRI GCM, *Journal of the Meteorological Society of Japan*, 69(4), 473–487.


Zhao P., S. Yang, and R. Yu (2010), Long-term changes in rainfall over eastern China and large-scale atmospheric circulation associated with recent global warming, *J. Clim.*, 23(6), 1544–1562.
Figure 2.1. (a) Long-term 15-day evolution of ΔOLR (235Wm$^{-2}$-OLR) averaged in Taiwan (119° –122°E, 21°–25°N) from 1980 to 2010, following Wang and Chen (2008). The inset map depicts the geographical location of Taiwan (red). (b) Yearly distribution of daily ΔOLR applied with a 5-day moving average (shadings) overlaid with the linear trend contours from 1979 to 2014. A white and two yellow dashed lines indicate the period difference of the Meiyu as referred in the text.
Figure 2.2. Differences of ΔOLR (shadings) and (a) 250- and (b) 850-mb winds vectors between 1991–2002 and 1979–1990 for the 7 June–20 June period. (c–d) Same as (a–b) except for the differences between 2003–2014 and 1999–2002. (e) ΔOLR latitude (y-axis) and year (x-axis) distribution across the vicinity of Taiwan (white box in (b) and (d)) during the 7 June–20 June periods overlaid with the linear trend contours from 1979 to 2014. The latitudinal extent of Taiwan is shown by the green dashed lines.
Figure 2.3. Differences of ΔOLR (shadings) and (a) 250- and (b) 850-mb winds vectors between 1991–2002 and 1979–1990 for the 7 June–20 June period. (c–d) Same as (a–b) except for the differences between 2003–2014 and 1999–2002. (e) ΔOLR latitude (y-axis) and year (x-axis) distribution across the vicinity of Taiwan (white box in (b) and (d)) during the 7 June–20 June periods overlaid with the linear trend contours from 1979 to 2014. The latitudinal extent of Taiwan is shown by the green dashed lines.
Figure 2.4. The EOF analysis of daily ΔOLR from 1979 to 2014 for (a) EOFs 1, (b) PCs 1, (c) EOFs 2, (d) PCs 2, (e) reconstructed EOFs 1+2 and (f) reconstructed PCs 1+2. A 5-day moving average is applied prior to the EOF analysis.
Figure 2.5. Regression coefficients of the eddy streamfunction ($\psi_E$) June monthly mean regressed with PCs 1+2 for (a) 250 and (b) 850 hPa. Shadings indicate significance at $p<0.05$; arrows were added to illustrate the anomalous flow direction. H’s and L’s indicate anticyclonic and cyclonic anomalies, respectively. (c) Scatter diagram of the June eddy streamfunction ($\psi_E$) values at 250 and 850 hPa averaged from the respective domains outlined in (a) and (b) red boxes, with the last two digits of each year color-coded as indicated in the lower right.
Figure 2.6. Same as Figures 2.5a and 2.5b but regressed with PCs 1 for (a) 250mb and (b) 850mb; PCs 2 for (c) 250mb and (d) 850mb.
Figure 2.7. (a) The regression pattern of June $\psi_E$ 250 hPa with PCs 1+2 (contours) overlaid with the linear trend slopes of the June $\psi_E$ (shadings). Green-dotted areas indicate significance at p<0.05 for the regression. (b) Same as (a) but for the short-wave regime (i.e. zonal wavenumbers 5 and beyond) applied in each field, following the trending pattern as depicted by Wang et al. (2013b).
Figure 2.8. (a) Climatological 850-hPa velocity potential (VP) and divergent velocity (Vd) during Western North Pacific main tropical cyclone season (July-September), superimposed the locations of tropical cyclogenesis. (b) Yearly and seasonal distribution of the daily VP over convergent center (blue box in (a)) with the years in the y-axis and days in the x-axis. (c) Yearly distribution of the number of tropical cyclones with pentad scale with a 3-pentad moving average in Western North Pacific. (d) The time series of the monthly tropical cyclone number anomaly in July (blue) and August (red) relative to the 1981-2010 with 5 years moving average. The shaded shows the deficit of two months. (e) same as (b) but with 20-100 days band pass filter and 5 years moving average. (f) the same as (c) but with 20-100 days band pass filter and 5 years moving average. In (e) and (f), positive/negative value presents convergent/divergent VP and high/low TC genesis anomalies.
Figure 2.9. Yearly distributions of daily 850-hPa velocity potential (VP) with 30-90 days bandpass filter and 5 years moving averaged over convergent center in Western North Pacific for (a) NCEP1 reanalysis data with domain average over 125°E-165°E and 5°N-25°N, (b) JRA 55 reanalysis data with domain average over 120°E-150°E and 5°N-25°N, (c) NOAA 20th Century Reanalysis with domain average over 122°E-152°E and 8°N-24°N, and (d) ERA 20th Century Reanalysis with domain average over 115°E-155°E and 5°N-25°N.
Figure 2.10. (a)-(d) the EOF analysis of yearly distribution of daily 850-hPa velocity potential (VP) with 20-100 days band-pass filter, 5 years moving average, and 30 years moving average removed over the convergent center (Figure 1a, blue box) from 1 July to 10 September for NCEP1 (black line), JRA 55 (purple line), NOAA 20C (red line), and ERA 20C (blue line).
Figure 2.11. Power spectrum analysis of (a) NCEP1-PC1, (b) NCEP1-PC2, (c) NOAA 20C-PC1, (d) NOAA 20C-PC2, (e) ERA 20C-PC1, and (f) ERA 20C-PC2. Green line shows the 99% significance level.
Figure 2.12. The composites of 850-hPa velocity potential (shaded) and stream function (green line) at 2*10^6 m^2 s^{-2} during 3 different periods when positive VP anomalies are in (a) July (1990-1996 and 2008-2014) and (b) August (1980-1986 and 1990-2005)
Figure 2.13. NCEP1 10-days average 850-hPa velocity potential (VP) regresses onto the (a) PC1 and (b) PC2, and 850-hPa stream function (ST) regressed on (c) PC1 and (d) PC2 from 1948 to 2016.
Figure 2.14. NCEP1-PC1 regress with every 2-month average of (a) 850-hPa velocity potential (VP), (b) Sea surface temperature (SST), and (c) 850-hPa streamfunction (ST) from 1948 to 2017.
Figure 2.15. (a) the sliding correlation with 61-day running window (from day-30 to day+30) between 850-hPa velocity potential (VP) over Western North Pacific (WNP) with MJO-RMM1 from 1979 to 2017. (b) the same as (a) but for MJO-RMM2. (c) the Yearly distributions of daily RMM1 (Top), RMM2 (bottom), and 850-hPa velocity potential (VP) from 1977-1991 with 20-100 days bandpass filter averaged over convergent center in Western North Pacific, applied with 5 years and 5 days moving average. (d) the same as (c) but for 2001-2015.
Figure 2.16. Same as Figure 2.1, but for CCSM4 historical simulation precipitation in recent 36 years with different forcing: (a,b) anthropogenic greenhouse gases (GHG), (c,d) natural including solar and volcanic forcing (Nat), and (e,f) anthropogenic aerosol (Aero). The yearly distribution of daily precipitation is the departure from seasonal means. Notice the rather weak Meiyu phase of rainfall than the observation, as well as the peak rainfall shift in (b) that is coincident with Figure 2.1b.
CHAPTER 3

EMPIRICAL AND MODELING ANALYSES OF THE CIRCULATION INFLUENCES ON CALIFORNIA PRECIPITATION DEFICITS

Abstract

Amplified and persistent ridges in western North America are recurring features associated with drought conditions in California. The recent drought event (2012–2016) lasted through both La Niña and El Niño episodes, suggesting additional climate drivers are important in addition to the commonly perceived El Niño-Southern Oscillation. Diagnostic analyses presented here suggest that, while the Pacific North American (PNA) and North Pacific Oscillation (NPO) do not directly cause drought in California, the relationships between them and with the upper air circulation pattern do modulate the spatial drought pattern. The positive PNA relative circulation leads drier northern California, and (-NPO) relative circulation leads southern California to be drier. The types of drought in this region emerge mostly from the combination of two PNA and NPO relative oceanic and atmospheric oscillations. At present, climate model projections do not indicate any significant change in these particular drought-modulating processes.

The material for this chapter was recently published as:

3.1 Introduction

During the winters from 2011-2012 to 2014-2015, a persistent upper tropospheric ridge developed over the Northeastern Pacific, and this anomalous ridge prohibited much of the rain-producing weather disturbances from reaching California. The reduction in the rainy-season precipitation and warmer temperature led to a major drought with declined snowpack and subsequently less water during the dry seasons. While the occurrence of drought is not uncommon in California [Department of Water Resources, 1978, 1993, 2015], the fact that this recent drought episode has lasted four consecutive years was unprecedented in a 1200-year reconstructed history [Robeson, 2015].

Various climate modes impact the winter precipitation in California, such as the North Pacific Oscillation (NPO) or West Pacific [WP; Linkin and Nigam, 2008], the Pacific-North American [PNA; Renwick and Wallace, 1996], and the El Niño-Southern Oscillation (ENSO) patterns. Even though California droughts are closely associated with an amplified and stagnant ridge over the western U.S., the formation mechanism of the ridge itself is elusive [Wang et al. 2014], and the interpretation of the causes of historical droughts in California has been inconsistent. The notable 1976-77 California drought winter was reported to associate with the El Niño [Namias, 1978]. However, this was contradicted by the argument of Seager et al. [2015] that the 2012 California drought was initiated by the 2011 La Niña. Apart from the classic view that ENSO and the Pacific Decadal Oscillation (PDO) collectively contribute to California’s dry winters [McCabe and Dettinger, 1999; Kam et al., 2014], recent studies that focused on the post-2012 drought identified other unique atmospheric and oceanic features. For example, H. Wang and Schubert [2014] suggested that the 2013 sea surface temperature (SST)
anomalies in the North Pacific produced a predilection for the ensuing California drought. Swain et al. [2014] pointed out the record-setting ridge in the upper troposphere as the cause of drought, while Wang et al. [2014] reported the associated geopotential height dipole (with a trough counterpart over the Great Lakes) was linked to an ENSO precursor. Subsequent studies [Hartmann 2015; Lee et al. 2015] also linked the 2013-14 SST pattern to the North Pacific Mode [Deser and Blackmon, 1995] with an amplitude modulation from reduced sea ice content in the Arctic.

How do we reconcile these different observations and interpretations about which climatic features and variability influence drought conditions in California? In addressing this question, we shifted our attention on the atmospheric circulation and SST settings that affect the pattern of drought. The hypothesis addressed is that there are different dynamical processes that govern events leading to several distinct patterns of drought. This differs from previous studies searching for simply the causes of drought, implying that all droughts are the same. Additional understanding of the different types of atmospheric patterns associated with patterns of dry episodes could help society anticipate or mitigate the next drought.

3.2 Data

To depict the winter season from November to March precipitation deficit over California, we utilized the Parameter-elevation Regressions on Independent Slopes Model (PRISM) [Daly et al., 2008] precipitation at 4km horizontal resolution developed by Oregon State University (http://prism.oregonstate.edu). For the atmospheric circulation, we analyzed the monthly National Centers for Environmental Prediction
(NCEP)/National Center for Atmospheric Research (NCAR) Reanalysis data with a spatial resolution of $2.5^\circ \times 2.5^\circ$ from 1948 to 2015 [Kalnay et al., 1996]. To explore the sea surface temperature (SST) pattern, we analyzed the monthly NOAA Extended Reconstructed Sea Surface Temperature version 3 with a $2^\circ \times 2^\circ$ spatial resolution [Smith et al., 2008]. For the purpose of documenting the drought connection to climate patterns, we also examined existing climate indices archived by the NOAA Earth System Research Laboratory: http://www.esrl.noaa.gov/psd/data/climateindices/list/. Due to the maximum extent of NCEP datasets, the analysis period of observational data is from 1948 to 2015. To reveal the drought distribution, the PRISM-derived Palmer Drought Severity Index (PDSI) [Palmer, 1965] data with 4km horizontal resolution were also utilized. The PDSI is constructed by taking into account water supply, water demand, and soil moisture information influenced by surface temperature, and is used commonly to show the global and regional drought features [Dai et al., 2004; Heim, 2002].

To further examine the variations of climate variability under external climate forcing and understand long-term changes, the historical and future simulations of Community Earth System Model version 1 (CESM1) [Hurrell et al. 2013] with CAM5 for the atmospheric component were used. The model version and setting follow Yoon et al. [2015a]. Thirty ensemble members of CESM1 with $0.9^\circ \times 1.25^\circ$ horizontal resolution through the Large Ensemble Project [Kay et al. 2015] were utilized. The historical forcing scenario (HIS run), including greenhouse gases, aerosols, volcanic activity, ozone, land use change, and solar, covers 1920-2005 period, and the future RCP8.5 forcing scenario (RCP run) that represents a high-emission pathway [Taylor et al. 2012] covers 2006-2080 period. The climate indices, such as PNA and NPO, are computed by
correlating the 250-hPa geopotential height onto corresponding PNA (0°E – 360°E, 20°N – 90°N) and NPO (165°E – 90°W, 10°N – 70°N) loading patterns during the winter season in each ensemble. The loading patterns are generated by regressing NCEP’s 250-hPa geopotential height onto observational PNA and NPO indices. The ensemble spread of initial conditions is generated by the ‘round-off differences’ method [Key et al. 2014].

3.3 Results

To understand the circulation variations in drought years in California, we first identify precipitation deficit events. Based upon the winter (November–March) precipitation in California (P_{CA}), which is shown in Figure 3.1a from 1948-1949 to 2014-2015, we defined the occurrence of drought to be when P_{CA} exceeded 0.7 standard deviation below the 57 years winter mean, a threshold to balance water deficit intensity with a sufficient number of cases. This definition of drought led to the inclusion of recent severe droughts of 1976-1977, 2011-2012 and 2014-2015, as well as other major low-precipitation years, isolating 18 drought events as indicated by red dots. Figure 3.1b shows the composite 250-hPa geopotential height anomalies of these 18 drought winters (as a departure from the 1948-2015 mean), depicting a prominent high-pressure anomaly centered off NW US and western Canada, but covering much of the western US including California. Such a ridge will prohibit the occurrence of winter storms that produce rainfall in California [Swain et al. 2014]. A discernable yet weak wave train emerges in the upstream region over the North Pacific. Given its upstream source near the western North Pacific, this wave train appears to be different from the ENSO-induced teleconnection forced by the central/eastern tropical Pacific heating anomalies.
[Schonher and Nicholson, 1989]. Another notable feature in the downstream side over northeastern North America is a robust anomalous trough that, together with the western ridge, forms the so-called North American dipole [Wang et al. 2015].

To analyze the extent to which these 18 droughts differ case-to-case, we first applied the empirical orthogonal function (EOF) to depict the variation of the November–March 250-hPa geopotential height ($Z_{250}$) within these 18 drought events. The analysis here was focused on the region encompassing the composite ridge anomaly (165°E – 90°W, 10°N – 70°N). The loading patterns of the first two leading modes (EOF 1 and 2) are shown in Figure 3.2a as shadings, which are superimposed on composite $Z_{250}$ contours for comparison. These two EOFs constitute collectively about 70% of the total variance, meaning that the first two leading modes are the major circulation variations. It is noteworthy that they are also about the same fraction of variance, and thus importance. EOF1 depicts a northeastern extension of the anomalous ridge in the positive phase (according to the principal component or PC; Figure 3.2b) and a westward extension over the Gulf of Alaska in the negative phase. In EOF2, the anomalous ridge would expand mostly towards the northwest the Bering Sea and into the southwestern U.S. as well. This pattern has shown prominence during the recent (2013-2014) drought as noted by Wang et al. [2014]. Similar results are revealed in the upper troposphere at 200-hPa and 500-hPa, as shown in Figure 3.3. These results are empirical evidence of the existence of two distinct climate circulation schemes that affect the pattern of drought in this region.

The intensity of these two EOFs was proportional to the amplitude of the PC values. Therefore, we correlated the PC of 18 drought winters with different climate
indices of the same 18 drought winters; these correlation coefficients are summarized in
the supplementary material Table 3.1. The results illustrate that PC1 has a high positive
correlation ($r=0.90$) with the PNA index [Barnston and Livezey, 1987], while PC2 has a
high negative correlation ($r=-0.83$) with the North Pacific Oscillation (NPO). The NPO is
a leading atmospheric variation mode determined as the 2nd PC of the November–March
1000-hPa height anomalies over North Pacific [Rogers, 1981]. Likewise, the linear
regression patterns of $Z_{250}$ with PC1 (Figure 3.4a) and PNA (Figure 3.4c) for the 18 dry
winters appear to be similar, while the corresponding SST regressions (Figures 3.4b and
d) reveal an ENSO-like pattern. This is not surprising since ENSO is the prime forcing of
the PNA teleconnection [Yu and Zwiers, 2007], even though ENSO does not connect as
prominently to the PC series of $Z_{250}$ (Table 3.1). Meanwhile, the height and SST
regression patterns with PC2 (Figure 3.4e) and negative NPO (Figure 3.4g) both present
a distinct high-latitude seesaw from tropics to the Bering Sea and its associated “triband”
SST anomalies (Figures 3.4f and h) as noted by Linkin and Nigam [2008]. We note that
the NPO’s associated SST pattern (Figure 3.4h) is also similar to the “North Pacific
Mode” of SST identified by Deser and Blackmon [1995], which depicts the SST
counterpart of the sea level pressure-based NPO. Given that PNA and NPO represent
atmospheric modes, there may be intra-seasonal variability that is overlooked in this
seasonal mean analysis.

The results shown in Figure 3.4 suggest that different climate forcing sources may
influence the distribution and intensity of droughts by modulating the drought-inducing
ridge. To examine these teleconnection impacts on the drought pattern, we show in
Figures. 3.5a and c the association of PC1 and PNA with precipitation in the drought
winters. Here, the values represent influence ratio on precipitation, that is calculated from the regressions of precipitation onto each normalized index divided by the mean precipitation anomaly within the 18 dry winters, ranging between -1 and 1. Positive cases of PC1 and PNA are associated with drier conditions in the Pacific Northwest, leaking into northern California, and less dry conditions in southern California and western/southwest coast of California. This shift of drought pattern is caused by the anomalous ridge being extended further northeastward (EOF1; Figure 3.2a). For PC2 and the negative NPO (-NPO), impacts on precipitation (Figures 3.5b and d), southern California and some of the Southwest U.S. experience more severe drought conditions, and correspondingly northern California and some of Northwest U.S. exhibit less intense drought with the southward extension of the high-pressure ridge (EOF2; Figure 3.2a). Although the influence fractions of PNA and NPO on California drought winters’ precipitation do not show big differences (about 10-20%) in Sierra Nevada, the connections are larger in the coastal and agriculture intensive (valley) regions (Figures 3.5b and d). Moreover, it is instructive to consider that drought may not be defined only by precipitation. The regression results of PDSI in drought winter onto PC1/PNA and PC2/(-NPO) (Figures 3.6a-d) show the PNA and NPO are related to a measure of drought intensity, especially in southern California, with changes of over 1 point of PDSI value.

It appears that both the PNA and NPO, in addition to ENSO, modulate the drought pattern in California, as well as the western U.S., but they do not directly cause the drought (i.e., due to their weak direct relationship with precipitation in California). This latter point is demonstrated in Figures 3.5e and f showing the influence ratio on
precipitation by regressing wintertime precipitation onto normalized PNA and NPO indices during the entire 1948-2015 period divided by the winter mean precipitation. It is clear that the patterns for the entire period (Figures 3.5e and f) resemble in general terms those of the 18 drought years (Figures 3.5c and d). The record low snowpack in the Sierra Nevada in 2013-2014 that was identified as the lowest in the past 500 years [Belmecheri et al. 2015] coincided with a negative PNA and a strong negative NPO phases (Figure 3.6). This combined influence on the extreme low snowpack further intensified the drought making it harder to recover. However, even though the precipitation pattern bears resemblance with the variations within the 18 drought winters, neither the PNA nor NPO shows statistically significant correlations with precipitation within California. A similar lack of correlation was also found between the PNA/NPO and the winter PDSI (Figure 3.7). Altogether, these results suggest that the PNA and NPO play more of a modulating role of drought in California rather than causing it.

Further validation was carried out by analyzing the CESM1 Large-Ensemble simulations of HIS and RCP runs. By applying the same analysis as in the observational data, the simulated geopotential height anomalies in each of the 30 members reveal similar PNA and NPO features, both in the leading EOFs and regression patterns. In Figure 3.8 we show the ensemble mean of PNA-like and NPO-like EOFs of each scenario. The results show that the HIS and RCP runs have similar variances in the first two leading modes, suggesting that the CESM1 simulations agree that PNA and NPO are key circulation features that modulate the drought pattern in California. Figure 3.9 and Figure 3.10 show the averaged regression patterns of each ensemble’s PC onto CESM1’s $Z_{250}$ and SST, suggesting that the influence of modeled PNA (NPO) on California
droughts slightly increase (decrease) in RCP run. Recent studies [Zhou et al. 2014; Yoon et al. 2015a] have indicated the anthropogenic warming would change North Pacific circulation and, in turn, would influence climate conditions in North America. The result from CESM1’s RCP run suggests that this influence would be realized through the modulations of PNA and NPO.

3.4 Discussions

The PNA and NPO relative circulations are associated with the spatial distribution of precipitation during drought in California. Meanwhile, when applying the EOF analysis on the 18 drought winters’ normalized precipitation over California region (Figure 3.11a and b), the first two leading modes also have about 70% of total variance, which is similar with the EOF analysis of Z250-hPa. But, the 1st mode of EOF for precipitation shows no distinct spatial pattern, while the 2nd mode does show a strong north-south pattern. The correlation patterns of Z250-hPa onto two leading PCs for precipitation do not show any significant correlation coefficients (at p<0.05 level) (Figure 3.11c). It means these two orthonormal eigenvectors are not associated with a specific circulation pattern.

If one looks more closely, the 1st mode of EOF for the 18-drought-winters precipitation is associated with the 1st and 2nd EOFs of Z250-hPa being opposite in sign (Figure 3.2). Recall the value of the 1st mode of Z250-hPa is associated with the PNA, and the value of the 2nd is associated with (−NPO). Since the regressions of PNA and (−NPO) with precipitation have opposite dipole patterns near California (Figures 3.5c and d), the result is that the two regressions effectively counter each other and no dipole
pattern is observed. So the first mode of precipitation is associated with PNA and (-NPO) having the same sign, it displays no spatial pattern. In contrast, the 2nd leading mode of EOF for California 18-drought-winters precipitation appears to result from a constructive effect of PNA and (-NPO) (Figures 3.5c and d) to enhance the dipole pattern. This is because, in this case the PC1 and PC2 of the Z250-hPa, associated with PNA and (-NPO), have opposite signs, so the respective regressions with precipitation have similar spatial patterns, that re-enforce each other. Therefore, it is the combination of the signs of the circulation EOFs, related to those of the PNA and (-NPO) that relate to spatial patterns of drought year precipitation, and several types of droughts in California.

The forcing sources of the PNA are manifold, and previous studies have indicated that the intensity of the PNA is associated with the eastern tropical Pacific SST [Straus and Shukla, 2002; Yu and Zwiers, 2007] and the East Asian Jet Stream [Wallace and Gutzler, 1981; Leathers and Palecki, 1992]. This explains the 2nd-highest correlation coefficient of the Niño indices with PC1 as shown in Table 3.1. The NPO’s role in the modulation of the California drought does connect to ENSO, since the NPO acts as an ENSO precursor (i.e. no direct correlation) through interactions with tropical SST and wind anomalies across the equatorial Pacific. The NPO’s role in triggering ENSO occurs under the so-called “Seasonal Footprinting Mechanism” (SFM), from which the NPO imparts a surface wind stress to change the surface heat fluxes and underlying SST [Vimont et al. 2003; Alexander et al. 2010]. These features supplement the common perception that ENSO and its different phases are responsible for California drought. Recent studies [Yoon et al. 2015a,b] projected that both intense drought and excessive flooding in California may increase by 50% towards the end of the 21st century, and this
projection is based upon a strengthened relation to the ENSO cycle—not only through its warm and cold phases, but also its precursor (transition) patterns.

The long-term precipitation regression results with PNA and NPO indices from 1948-1949 to 2014-1915 show much less significance than the relationship of variations within California (Figures 3.5e and f). For instance, the winters of 1975-1976 and 1976-1977 were associated with a distinct opposite phase of the PNA pattern and El Niño SST anomalies, while in 2011-2012 a La Nina SST pattern prevailed. The recent record droughts in 2013-14 and 2014-15 were associated with an amplified NPO (Figure 3.2b) without the presence of a mature-phase El Niño. The lack of association between the California drought and the PNA and the NPO is evident in Figure 3.6 indicating a mixture of phases in either index during the 18 dry winters. These reported features and the association with the NPO make simulation and/or prediction for California’s winter climate difficult, since the majority of models do not simulate the ENSO precursors (i.e. the NPO) so well [Wang et al. 2015; Yoon et al. 2015a]. Further research is needed in identifying the source of variability and predictability in the drought-producing ridge off the Northwest U.S. region as revealed in Figure 3.1b.

Analysis of the CESM1 Large-Ensemble simulations with historical and RCP forcing scenarios is supportive of the respective roles of PNA and NPO in modulating the drought pattern. This finding led us to question the extent to which the relations between PNA/NPO and California’s precipitation may change in the future climate. To answer this question, Figure 3.12a shows the frequency distribution of California winter precipitation superimposed with the CESM1’s simulations. While the HIS run presents a normal distribution, the RCP run shifts the wet tail substantially and the dry tail slightly
(far left). These results are consistent with the finding of the increased water cycle extremes in California projected by Yoon et al. (2015a), although the annual precipitation may not change with human-induced climate change [Pierce et al., 2013]. Further, the correlation coefficients of PNA/NPO and California precipitation within the drought winters (Figure 3.12b) and the sliding correlation coefficients of PNA/NPO and California precipitation with a 30-year window over all simulation period (Figure 3.12c and d) both show that, despite of the projected change in the frequency distribution of California precipitation, its association with either PNA or NPO remains very weakly and insignificantly correlated.

3.5 Conclusions

The upper-atmospheric high-pressure (ridge) anomaly that accompanies drought in California exhibits several patterns, which modulates the spatial distribution of precipitation during drought in this region. These findings suggest that the variations of the ridge are collectively modulated— but not directly caused— by a combination of geopotential heights, and the synchronization of the signs of the PNA and NPO. Neither the PNA nor the NPO appears to directly contribute to the formation of drought, at least for the winter season. Rather, they alter the pattern of drought. There are several different combinations of forcing factors (Z250-hPa, PNA, NPO) that are associated with drought in CA.

The analysis of CESM1 simulations indicates that these modulations of PNA and NPO will not change significantly in the future, although there is a projected increase in the extreme wet/dry anomalies in California.
In terms of future research, a couple of unsolved questions are worth pursuing: (a) investigating whether the appearance of drought-producing ridge anomaly is actually forced or caused by any prominent mode of climate variability, or is purely due to random changes in atmospheric states, and (b) examining the impacts of constructive and destructive superposition of the PNA and NPO and how well they can be simulated in seasonal predictions. Lastly, even though this study is focused on precipitation, the effect of temperature on exacerbating drought cannot be discounted; this temperature effect was recently demonstrated using paleoclimate record for the California drought [Griffin and Anchukaitis, 2014]. The effect of anthropogenic warming as suggested by recent studies on the increasing chances of low-precipitation years in California [Yoon et al., 2015a; AghaKouchak et al., 2014; Diffenbaugh et al., 2015] and the associated dynamic processes warrants further investigation.

Reference


Dai, A., Trenberth, K. E. and Qian, T. A. (2004), global dataset of Palmer Drought
Severity Index for 1870–2002: Relationship with soil moisture and effects of surface warming, J. Hydrometeorol. 5, 1117–1130.


**Figures and Tables**
Figure 3.1. Winter season (November–March) precipitation time series over California; the inset map shows the domain and winter mean precipitation in California. The red dots indicate the 18 dry winters in which precipitation was less than 0.7 standard deviation below average as shaded. (b) The composite anomaly of the 250-hPa geopotential height of the 18 drought winters. Hatches indicate significant level at p<0.05 for the anomaly.
Figure 3.2. (a) The first two leading EOFs (shaded) of winter season (November–March) $Z_{250}$ within the 18 dry winters, superimposed with their composite $Z_{250}$ anomalies (contour). (b) The corresponding PCs in relation to each of the 18 dry winters.
Figure 3.3. The same as Figure 3.2, the first two leading EOFs (shaded) and PCs of winter (November – March) geopotential height (Z) within the 18 dry winters, superimposed with dry winters’ Z anomalies (contour), but for (a) Z200 hPa and (b) Z500 hPa.
Figure 3.4. The Z250 and SST patterns from the 18 California dry winters regressed upon PC1 (a,b), PNA index (c,d), PC2 (e,f), and negative NPO index (g,h). Hatches indicate significant level at p<0.05 for the regression.
Figure 3.5. The influence ratio from the regression of 18 drought years precipitation onto normalized (range from -1 to 1) (a) PC1, (b) PC2, (c) PNA, and (d) negative NPO decided by the mean precipitation anomaly during drought years. (e) and (f) are the same as (c) and (d) but for 1948-49 – 2014-15 wintertime precipitation.
Figure 3.6. Winter season (November – March) precipitation in California (blue line) overlaid with (a) the PNA index (orange line) and (b) the inverted NPO index (pink line) from 1948-49 to 2014-15, superimposed with the 18 dry winters as vertically shaded.
Figure 3.7. (a)-(d) Same as Figure 3.5 but for the regression patterns of the PDSI with the PNA and NPO indices. (e) and (f) Long-term regressions of the PDSI with the PNA and the inverted NPO over the 1948-2015 period. Hatches indicate significant level at $p<0.05$ for the regression.
Figure 3.8. The 30 ensembles mean of the EOF analysis of 250-hPa HGT from each ensemble’s drought years. (a) The PNA-like EOFs of HIS run, (b) the negative NPO-like EOFs of HIS run, (c) the PNA-like EOFs of RCP run, and (d) the negative NPO-like EOFs of RCP run. The contour shows the drought years anomaly in HIS run (a,b) and RCP run (c,d).
Figure 3.9. Averaged Z250 hPa regression patterns from 30 ensembles in drought years by regressing with the PCs of (a) PNA-like EOFs from HIS run, (b) negative NPO-like EOFs from HIS run, (c) PNA-like EOFs from RCP run, and (d) negative NPO-like EOFs from RCP run.
Figure 3.10. averaged SST regression patterns from 30 ensembles in drought years by regressing with the PCs of (a) PNA-like EOFs from HIS run, (b) negative NPO-like EOFs from HIS run, (c) PNA-like EOFs from RCP run, and (d) negative NPO-like EOFs from RCP run.
Figure 3.11. (a) The first two leading EOFs of normalized winter season (November–March) precipitation within the 18 dry winters over California, (b) The corresponding PCs in relation to each of the 18 dry winters, and (c) The Z250 patterns from the 18 California dry winters correlated upon PC1 and PC2 of California precipitation.
Figure 3.12. The precipitation changes in California simulated by CESM1 with 30 ensembles. (a) The ensemble mean of frequency distribution of November-March precipitation in California ($P_{\text{CA}}$). The historical scenario (HIS run) includes 74 years before 2005, and the RCP 8.5 scenario (RCP run) includes 74 years after 2005. (b) The correlation coefficient between $P_{\text{CA}}$ and PNA (left) and between $P_{\text{CA}}$ and negative NPO (right) within drought years on ensembles’ HIS/RCP runs and observational data (OBS). The blue circles show the correlation coefficient of ensemble mean or observation, and the gray bar indicates 50% of ensemble spread. (c) The 30 years window sliding correlation between PNA and $P_{\text{CA}}$ over simulation period. The black solid line is the ensemble mean, the shaded areas indicate 50% of ensemble spread, and the red line is the observational data. (d) The same as (c) but for $P_{\text{CA}}$ and negative NPO.
Table 3.1. The correlation results between PCs and winter mean (November–March) of climate index over 18 California dry Winters. The highest corollary/anti-corollary index with PC1 or PC2 is shaded.
CHAPTER 4

ACCELERATED INCREASE IN THE ARCTIC TROPOSPHERIC WARMING EVENTS SURPASSING STRATOSPHERIC WARMING EVENTS DURING WINTER

Abstract

In January 2016, a robust reversal of the Arctic Oscillation took place associated with a rapid tropospheric warming in the Arctic region. This was followed by the occurrence of a classic sudden stratospheric warming in March. The succession of these two distinct Arctic warming events provides a stimulating opportunity to examine their characteristics in terms of similarities and differences. Historical cases of these two types of Arctic warming were identified and validated based upon tropical linkages with the Madden-Julian Oscillation and El Niño as documented in previous studies. The analysis indicates a recent and seemingly accelerated increase in the tropospheric warming type versus a flat trend in stratospheric warming type. The shorter duration and more rapid transition of tropospheric warming events may connect to the documented increase in midlatitude weather extremes, more so than the route of stratospheric warming type. Model simulations with an atmospheric general circulation model suggest that the

This chapter includes two studies one was published, and another one is a conference paper and in preparation:

reduced Arctic sea ice contributes to the observed increase in the tropospheric warming events and associated remarkable strengthening of the cold Siberian high manifest in 2016. Observed evidence shows that the increase in the frequency and amplitude of deep-convection over the central and eastern tropical Pacific enhance the chances of teleconnection wave trend and influence the midlatitude and the Arctic circulations.

4.1 Introduction

In January 2016, the Arctic Oscillation (AO) index underwent a drastic phase reversal (Figure 4.1a) with excursions exceeding positive 2 standard deviations (σ) transitioning to negative 2σ within 20 days. Vertical profiles of the polar cap height (PCH) anomalies in Figure 4.1b, referencing the standardized geopotential height (HGT) averaged north of 65°N [e.g., Kim et al., 2014], and the Arctic air temperature anomalies from long-term mean (Figure 4.1c) both show that the troposphere warmed and expanded rapidly in the first half of January. The upper tropospheric flows transitioned correspondingly from a zonal pattern in December to high-amplitude semistationary waves (Figure 4.2 in the supporting information), accompanied by extreme weather events worldwide including severe flooding in the UK and Ireland (early January), Winter Storm Jonas that “rivals biggest East Coast snowstorms on record” [The Weather Channel, 2016], and a tremendous buildup of the Siberian high (late January; Figure 4.2) leading to record cold spells in Taiwan with unprecedented 84 hyperthermia fatalities and massive agricultural damages [TIME, 2016]. However, operational multimodel ensemble models did not predict the widespread cold temperature anomalies across northern Europe and Eurasia even at zero-month lead (Figure 4.3), presenting a challenge
This January 2016 event, referred herein as the rapid tropospheric warming (RTW), is distinct from the well-known sudden stratospheric warming (SSW) that propagates downward [Limpasuvan et al., 2004; Butler et al., 2015], of which a classic example was observed closely following the RTW during March and April (Figure 4.1). It is known that SSW acts as a precursor to AO phase change through downward propagation of the stratospheric polar vortex variation [Baldwin and Dunkerton, 1999; Tripathi et al., 2015], though this process was not observed in the January RTW. A weakened polar vortex (negative AO) induces more cold air outbreaks [Thompson and Wallace, 1998] and cold surges in Asia [Park et al., 2011], and strong negative AO events produce colder-than-normal winters throughout the Northern Hemisphere [Honda et al., 2009; Cohen et al., 2010; L’Heureux et al., 2010; Park et al., 2011; Kug et al., 2015]. AO can be amplified by sea ice fluctuations [Wang and Ikeda, 2000; Rigor et al., 2002; Overland and Wang, 2005], the 2016 RTW and SSW events did coincide alongside record warmth and low sea ice concentration (SIC) in the Arctic that lasted through spring [Overland and Wang, 2016]. Moreover, previous research links subseasonal Arctic warming events to tropical convective activities in the Pacific region [Jahnson and Feldstein, 2010; Lee et al., 2011]. The effect of Arctic warming on the change in midlatitude weather extremes has been an area of active research as was reviewed by Overland et al. [2016], Screen [2017], and Cohen et al. [2014]. To understand the differences and the long-term changes between SSW and the reported RTW, we decided to examine their characteristics and development processes in terms of similarities and differences by using reanalysis data and a global climate model.


4.2 Data and Model Experiments

Daily variables derived from the National Centers for Environmental Prediction (NCEP)/Department of Energy Reanalysis II (R2) data [Kanamitsu et al., 2002] and the NCEP/National Center for Atmospheric Research Reanalysis data [Kalnay et al., 1996] were used for the analysis of post-1979 and post-1950 atmospheric variations, respectively. The sea surface temperature (SST) data were obtained from the Extended Range SST version 3 and the SIC data from the NOAA Optimum Interpolation SST and Sea Ice version 2 [Reynolds et al., 2007]. The AO index was obtained from the NOAA Climate Prediction Center. We conducted atmospheric general circulation experiments by using the European Centre Hamburg Atmosphere Model version 5 (ECHAM5) [Roeckner et al., 2003], with T42 horizontal resolution and 31 vertical sigma levels. All simulations used initial and boundary conditions from the R2 reanalysis. Following the model setup described in Lee et al. [2015]. The control simulation was driven by the observed climatological monthly SST and SIC (over the 1998–2010 period) while two experiments were performed. One set of simulations denoted as the SST experiment combined monthly varying SST (differing for every month and year) with climatological SIC (constant for each month and year). A second set of simulations denoted as the SIC experiment inverts the forcing, with climatological SST combined with monthly varying SIC boundary conditions. Each experiment included 30 member simulations from October to March. The model simulation setup and discusses its performance with a reference to Saha et al. [2014].
4.3 Analysis and Results

4.3.1 Case Identification and Composite

We first begin with an analysis of the observations to identify historical RTW events. These events require four conditions be met with the following: (1) positive PCH/HGT anomaly north of 65 N only happens in the troposphere (beneath 200 hPa), (2) PCH/HGT anomaly has a magnitude greater than 1.5σ, (3) PCH anomalies in the stratosphere remain neutral to negative, and (4) a corresponding AO phase reversal must be present to reflect the surface climate anomaly; however, the magnitude of AO was not considered. To reduce high-frequency weather signals, a 5 day moving average was applied to the PCH and the AO. Stratospheric warming (positive PCH at 70 hPa and above) must not be present both during the tropospheric warming episode and in the 15 days leading up to it (as was the case in the January 2016 RTW). We used an established definition for the SSW [Limpasuvan et al., 2004; Charlton and Polvani, 2007; Butler et al., 2015]: (1) the 50 hPa PCH with its 5 day running mean reaches an anomaly greater than 1.5σ, (2) this increased PCH then propagates downward below 200 hPa, and (3) the duration of any given event is determined by the timespan between the first appearance of warming in the stratosphere and the last appearance of warming below 200 hPa. The case selection was conducted manually; each identified case is dated in Table 4.1 with 31 RTW and 28 SSW cases. For the reader’s reference, each year’s observed PCH profiles during the 1950–2016 period are shown in Figure 4.4.

To depict the common characteristics within a given AO/RTW episode, we adopted the “index cycle” approach of the AO that aligns the same phase of each oscillation episode, following Tanaka and Tokinaga [2002] and Wang et al. [2014]. The
transition from positive to negative status was evenly divided into seven phases: phase 1 designates the maximum AO, and phase 7 represents the minimum AO, each phase comprising 5 days centered on the third day. We then constructed the composites of the vertical PCH profiles, air temperature anomalies, and corresponding AO indices; these are shown in Figure 4.5 (the 2016 case was removed from the composite for the “leave one out” verification). It was found that the average AO transition in SSW takes an average of 48 days ± 15 days (1σ), more than twice as long than RTW that averaged 19 days (±5 days); these are also shown in Figure 4.5. (In Figure 4.5 and ensuing composite analysis, significance level was computed based on Student’s t test.)

Next, the dynamical aspects of RTW and SSW were compared with the previous observations of the tropical intraseasonal variations. One of the documented variation modes that can modulate the Arctic temperature and AO is the Madden-Julian Oscillation (MJO) [L’Heureux and Higgins, 2008; Yoo et al., 2012; Goss et al., 2016]. Figure 4.6a shows the 250 hPa velocity potential (VP) during the 2016 RTW event, which reveals a zonal wave-1 structure of VP that apparently propagated eastward up to phase 6. Such a wave-1 pattern and its eastward propagating signal are indicative of an MJO, of which a strong event occurred during January 2016 (http://www.bom.gov.au/climate/mjo/). In the RTW composite (Figure 4.6b), the VP propagation is even more pronounced and is phase-consistent with the 2016 case. In the SSW composite (Figure 4.6c), the VP pattern is rather disorganized and weak, suggesting minimal MJO influence. The analysis of the 250 hPa streamfunction (Figure 4.7) outlines the corresponding stationary wave anomalies with the 2016 case, in which a Pacific-North America (PNA) type of teleconnection pattern can be seen.
We further computed the Eliassen-Palm Flux [Edmon et al., 1980] and plotted the zonal-mean eddy momentum component with latitude in Figure 4.6, denoted as EPF. A zonally averaged EPF diagnoses the impact of transient eddies on the time mean flow and, in turn, delineates the large-scale and fast responses in the planetary wave trains that connect between the tropics and the Arctic [Trenberth, 1986]. As shown in Figure 4.6 (right axis of each panel) by the positive EPF that persisted throughout phases 3–7, RTW is more pronouncedly affected by tropical teleconnections; in contrast, positive EPF in SSW only is present in phases 6 and 7. This result is in good agreement with the corresponding streamfunction anomalies in Figure 4.7 showing a larger amplitude of the midlatitude stationary waves in the RTW composite, as well as the previous studies that identified the MJO influence on heat and moisture fluxes into the Arctic [Yoo et al., 2012; Park et al., 2015].

4.3.2 Disparity in the Trends

Encouraged by the consistency between the composite RTW and the January 2016 case presented in Figure 4.5, we proceeded to examine the long-term frequencies of cases beginning in 1953, with a 9 year interval (that allows for equal number of years up to 2016. As shown in Figure 4.8a, the frequency of RTW underwent a marked increase since the 1990s, more than doubled in the last decade, while the SSW frequency reveals a flat trend. The number of the RTW cases appears to exceed that of the SSW cases in the recent decade (though the exact numbers are likely dependent on the event definition). This marked increase in RTW may signify the emergence of the disproportionate Arctic warming (relative to midlatitudes) from the noise of natural
variability since the late 1990s [Serreze et al., 2009]. However, the increase in frequency should not reflect the Arctic warming, since the daily long-term trends of PCH and AO have been removed in the case selection.

To illustrate the frequency changes of RTW and SSW, Figures 4.8b and 4.8c show the daily distribution of extreme Arctic temperature anomalies from the 50 hPa level and the 1000–500 hPa average, respectively, during November–March starting in 1979 (using R2, with the daily long-term trend removed). The daily temperature thresholds used for the depiction of extreme warming are indicated in the legend. It is apparent that only the occurrence of tropospheric warming has increased and the cluster of changes has taken place in the beginning of the 21st century.

4.3.3 Impact of External Forcing

The concurrence of the January 2016 RTW event with the strong El Niño led to a speculation concerning the extent to which tropical sea surface temperature anomalies affect the Arctic warming. Previous studies [Ineson and Scaife, 2009; Butler et al., 2014; Johnson and Kosaka, 2016] have identified a stratospheric pathway from which El Niño affects the Arctic circulation and temperature. However, the January 2016 RTW case was not preceded by any stratospheric warming (Figure 4.1) and, as was stated in Overland and Wang [2016], the 2016 El Niño did not contribute to the AO change. Hence, we show in Figures 4.9a and b the differences in the January sea level pressure (SLP) of the SIC experiment and the SST experiment from the control experiment to assess their respective role. The corresponding T anomalies are shown in Figure 4.10. The SLP patterns in the two experiments are apparently opposite over Eurasia and
Siberia, where the impact from SIC anomalies in the Barents-Kara (B-K) Seas is known to be pronounced [Inoue et al., 2012; Park et al., 2015]. Compared with the 2016 anomaly (Figure 4.9c), the SIC experiment produced a similar SLP pattern in Eurasia and Siberia, while the SST experiment captured the classic Pacific-North America (PNA) teleconnection. Given the short duration of RTW (~20 days), PCH anomalies in the troposphere may reflect synoptic or intraseasonal variability more than a modulation of annular mode by some external forcings [Löptien and Ruprecht, 2005]. In the case of January 2016, a series of intense North Atlantic storms did contribute to the Arctic warming [Kim et al., 2017]. In terms of air temperature, SIC experiment led to substantial cooling in Siberia, while SST experiment generated a pan-Arctic cooling instead (Figure 4.10).

While the examination of monthly or seasonal means provides useful information to predominant climate anomalies, the fact that RTW events take place within one month signifies a need for further investigation on the intraseasonal variability. Here, we calculated from each member (i) the frequency of the Arctic HGT (1000 hPa, north of 65°N) being greater than 160 m and (ii) the frequency of the Siberian 2-meter air temperature (T2m; 80°E-120°E and 40°N-65°N) reaching below -21.5°C; both of which were determined during the period from December 15 to January 31. These thresholds were based upon ±1σ of the 30-member average. Figures 4.11a and b display the HGT and T2m frequencies of the three runs, respectively (numbers reflect the accumulated days from each member). The SIC experiment produced considerably more extreme events of Arctic high pressure buildup and cold Siberia than SST and control experiments did. We also computed the power spectrum of Arctic 1000-hPa HGT from
November 1 to February 28, showing their 30-member ensemble mean in Figure 4.11c. While the SIC experiment exhibits a consistently higher power in the 30-60-day frequency band than control experiment, the SST experiment 3 does not show any noticeable difference. This result suggests that the reduction of SIC can amplify intraseasonal variability in the Arctic region. Possibly, the slowdown of the polar vortex creates highamplitude Rossby waves (e.g., Figure 4.2 right), which are known to exhibit an amplified frequency in the intraseasonal timescale.

To more quantitatively describe the pattern difference between SIC and SST experiments, we computed the sliding spatial correlation between the experiments and observed 2016 anomaly, with a circumglobal 180° longitude range (from 35°N to 90°N centered at the longitude of x axis). As shown in Figure 4.9e, the sliding correlations delineate an opposite response from the two experiments, where the SST experiment produced a PNA-like response in the western hemisphere primarily in North America, while the SIC experiment generates a response in the eastern hemisphere encompassing Siberia and Eurasia. The SIC experiment produced the documented connection of the B-K Seas’ low sea ice with Siberia’s abnormally cold winters [Honda et al., 2009; Petoukhov and Semenov, 2010; Kim et al., 2014]. Moreover, the SIC decline in 2016 apparently led to a high-pressure response over Siberia, and this corresponds to the post-1979 trend in SLP (Figure 4.9d), a change that was driven by sea ice loss [Screen and Simmonds, 2013; Vihma, 2014]. However, we did not observe any upward propagation of eddy heat flux into the stratosphere during the January 2016 RTW (not shown).

Butler et al. [2014] have indicated that El Niño can increase temperature in high-latitude North America through a tropospheric pathway; here the SST experiment
produced a marked Arctic cooling instead (Figure 4.10). As a further examination, the sliding spatial correlation between the two experiments and the observed SLP trends (Figure 4.9f) shows that SLP trends in the Arctic region and Siberia are highly correlated with SLP anomalies in both the SIC experiment and January 2016 anomaly, suggesting that El Niño’s effect on the Arctic circulations was largely offset by that of SIC reduction over the eastern hemisphere and Siberia in 2016.

4.3.4 Impact of Tropical Forcing

To examine the circulation changes associated with the increase in RTW cases in the past decade, we show the average 250-hPa velocity potential (VP250) in the transit phases of RTW cases in two eras. First, the earlier era is from 1979 to 1987 with 12 RTW cases, and the recent era is from 1998 to 2015 with 19 RTW cases (Figures 4.12a and b). The results reveal deep convections (upper-troposphere divergence) in the Pacific in both eras. However, the difference between the two reveals an eastward shift of the deep convection to the eastern Pacific as shown in Figure 4.12c. A similar difference is seen in the linear trend of VP250 in winter since 1979 (Figure 4.12d).

The changes in deep convection of RTW events lead to the next analysis of investigating the decadal variations of the wintertime VP250 which is shown in Figure 4.13. The variations reveal an increase in divergence anomaly (Figure 4.13a) in the eastern tropical Pacific in the last two decades. The central Pacific region also exhibits a divergent anomaly during the same period. This change is further elucidated by the variance of wintertime VP250 (Figure 4.13b), which shows higher VP250 variances in the central and eastern tropical Pacific in the recent era. Moreover, Figure 4.13c shows
the frequency anomalies of the strong VP250 divergence (VP250 is less than 1.5 standard deviation) in winter. The results echo the changes of mean value and variance that the tropical Pacific in the recent era has more active convections than the early era.

Lee et al. [2011] indicates that the intraseasonal tropical convective precipitation can affect the Arctic warming. To examine how the increase in the tropical deep convection is relevant to the increased frequency of RTW in the recent era, we further conduct the wave activity flux (WAF) [Takaya and Nakamura, 2001] analysis on the composite of RTW cases in the early era (12 cases) and recent era (19 cases) to depict the change of tropical deep-convection forcing and its impact pathway, which is shown in Figure 4.14. In the recent era, the 300-hPa WAF (Figure 4.14b) reveals a northeastward wave trend generated from the Eastern Pacific. This wave trend forms a high-pressure anomaly in the southwest U.S., a low-pressure anomaly in the eastern U.S., and a high-pressure anomaly in the northern Atlantic from phase 4. This feature is not discernable in the early era (Figure 4.14a). This wave trend combines with the Pacific Northern American (PNA) teleconnection pattern from the central Pacific to induce high-amplitude wavy jet streams and cause high-pressure (warming) anomalies in Alaska and the Greenland regions.

To further evaluate the impact of deep convection in the eastern tropical Pacific, we construct the composite of upper-troposphere VP and WAF when the VP250 index is less than 1.5 standard deviation (Figure 4.15). The result shows a similar PNA teleconnection pattern and a wave trend over the U.S., which is consistent with previous modeling studies [Hoskins and Ambrizzi, 1993; Branstator 2014].
4.4 Conclusion

This study was motivated by the observation of two distinct Arctic warming events occurring in succession during early 2016, one confined in the troposphere with a shorter duration (RTW) and the other being a classic SSW with a longer time span. Given the high-impact consequence of the January 2016 RTW and the challenge it presented to subseasonal prediction (Figure 4.3), their differences and frequencies were examined. Composite analysis based on PCH and AO criteria was compared with those documented in the literature, including tropical influence, ENSO impact, and SIC effects. Subsequent analysis uncovered distinct trends in the frequencies of RTW and SSW events: whereas the frequency of SSW has not changed in any significant way, the frequency of RTW has increased dramatically and appears to accelerate since the 1990s, surpassing SSW events in the recent decade.

Forced experiments using ECHAM5 indicated that the loss of sea ice (as was the case during 2016’s record low SIC) can amplify intraseasonal variations in the high-latitude circulations which, according to the literature [Harnack and Crane, 1984; Horel and Mechoso, 1988; Athanasiadis et al., 2014], can translate to increased atmospheric blocking during winter season and subsequent cold surges. Moreover, the recent increase in the frequency of subseasonal Arctic rapid tropospheric warming events is associated with the increase in variations of tropical deep convection in the central and eastern tropical Pacific. The active tropical convection induces higher chances of teleconnection patterns to influence the midlatitude and the Arctic circulations, resulting in more weather extremes and more RTW events in the boreal winter. Moreover, the impact pathway from the eastern tropical Pacific to the U.S. and to the Northern Atlantic
becomes more important to influence northern hemisphere circulations because of the increase in the frequency of the deep convections in the eastern tropical Pacific in recent decades. These results are substantial in that RTW and associated fast AO transition involve more pronounced influxes of Rossby wave energy and moisture from the tropics than of SSW; this further suggests an increased risk of cold-season extreme weather events accompanying RTW.

Recent studies have unequivocally indicated that the loss of sea ice profoundly influences temperatures and circulations over the Arctic region which, in turn, modulates the midlatitude extreme weather [Cohen et al., 2014; Jung et al., 2014; Kug et al., 2015]. This study moves one step further by highlighting the existence of, and the increase in, the RTW type of Arctic events that may possess a greater threat in the form of midlatitude extreme weather. Future work requires high-performance climate modeling to properly represent the tropical deep-convection perturbation and regional dynamic processes revealed in the 2016 RTW and SSW cases to predict similar cases beyond weather forecasting timescale.

Reference


Overland, J., and M. Wang (2016), Recent extreme Arctic temperatures are due to a split polar vortex, J. Clim., 29(15), 5609–5616.


Serreze, M., A. Barrett, J. Stroeve, D. Kindig, and M. Holland (2009), The emergence of surface-based Arctic amplification, Cryosphere, 3(1), 11–19.


TIME (2016), East Asia hit by record snowfalls and cold weather (Jan 24, 2016), TIME. [Available at http://time.com/4192220/weather-eastasia-taiwan-japan-cold-winter/].


Figure 4.1. (a) Daily Arctic Oscillation index from 1 November to 30 April with 5 day moving average; the shaded period indicates the rapid tropospheric warming (RTW) case. (b) Time-height cross section of daily polar cap height (PCH) over Arctic region (north of 65°N) with 5 day moving average using R2 data. (c) Same as Figure 1b but for air temperature anomaly.
Figure 4.2. 300-hPa HGT plotted at 8800 m (thick black contour) overlaid with SLP at 1045 and 1055 hPa (blue contours) on 21 December 2015 (left) and 21 January 2016 (right). The associated extreme events are indicated in the right figure.
Figure 4.3. (a) Predicted T2m anomaly at 0-month lead by the National Multi-Model Ensemble (NMME) for January 2016 and (b) observed T2m anomaly of January 2016 derived from NASA GISS data. Notice the Eurasia-Siberia region in which temperature anomalies between the predicted and observed are opposite.
Figure 4.4. Polar Cap Height (PCH) plotted from 1953 to 2015, following Figure 1b. Data prior to (after) 1979 were derived from R1 (R2). Refer to Table 4.1 for cases identified.
Figure 4.5. (top) Composite PCH and (middle) air temperature spanning the seven phases of the index cycle (see text) averaged north of 65°N and (bottom) corresponding AO index for (a) the January 2016 RTW case, (b) composite RTW with 30 cases, and (c) composite SSW with 28 cases since 1979, using R2 data. The averaged duration and 1 standard deviation (purple) of total cases are labeled, and the phases representing 0 and 8 were derived from 4 day averages before 1 and after 7, respectively. The green dashed lines in Figures 4.2b and 4.2c outline 95% confidence interval for the t distribution.
Figure 4.6. Composite seven phases of 250 hPa velocity potential (VP) and the eddy momentum component of the E-P flux (EPF) computed from (a) the January 2016 RTW case, (b) historical RTW cases, and (c) historical SSW cases. Hatched areas of VP and shaded areas of EPF in Figures 3b and 3c signify 95% confidence interval for the t distribution.
Figure 4.7. Composite 7 phases of 250-hPa streamfunction computed from (a) the January 2016 RTW case, (b) the RTW cases, and (c) the SSW cases, following Figure 4.6. Hatched areas in (b) and (c) are significant at the 95% confidence level.
Figure 4.8. (a) The case frequencies of RTW (blue) and SSW (light green) cases plotted at every 9 years during the November–March period and occurrence of Arctic region (north of 65°N) temperature anomalies with the daily long-term mean removed at (b) 50 hPa and (c) 1000–500 hPa average exceeding the color-coded thresholds.
Figure 4.9. (a) Ensemble mean of the January sea level pressure (SLP) anomalies simulated by (a) SIC run and (b) SST run subtracted from control run. (c) January 2016 SLP anomaly derived from R2. (d) Linear trends in SLP from 1979 to 2016 multiplied by 37 years by using R2; the green contours indicate 95% confidence interval for the t distribution. The sliding spatial correlation over 35°N–90°N with an 180° longitude range centered at the x axis value (i.e., 90° to the west and 90° to the east) between (e) the 2016 anomaly and the two ECHAM5 runs and (f) the post-1979 linear trend and the two runs.
Figure 4.10. Same as Figure 4.9 (a)-(d) but for T2m anomalies and trends.
Figure 4.11. The frequencies of (a) 1000-hPa HGT exceeding 160 m over the Arctic region and (b) T2m below -21.5°C over Siberia, as well as (c) power spectrum of Arctic 1000-hPa HGT, from the ensemble means of three ECHAM5 experiments.
Figure 4.12. (a) The 12 cases composite of the averaged 250-hPa velocity potential (VP250) during the RTW transit phases from 1979 to 1997. (b) the same as (a), but for the 19 cases from 1998 to 2016. (c) the difference between recent two decades (b) and previous two decades (a) of the composite of RTW VP850. (d) The linear trend of VP250 in winter after 1979.
Figure 4.13. The anomaly of the 9-year average of (a) wintertime (November-March) 250-hPa velocity potential (VP250), (b) wintertime VP250 variation, and (c) the frequency of VP250 that is less than 1.5 standard deviation in winter.
Figure 4.14. The 300-hPa wave activity flux (WAF) and streamfunction anomaly (ST) of the RTW phases 4 and 7 in (a) 12 RTW cases averaged from 1979 to 1997 and (b) 19 RTW cases averaged from 1998 to 2016.
Figure 4.15. The composite analysis of 132 stronger convection (250-hPa velocity potential < 1.5 standard deviation) days in the eastern tropical Pacific (240°E-285°E, 10°S-20°N) from 1979 to 2015 winter in (a) 250-hPa velocity potential and (b) 300-hPa wave activity flux and streamfunction.
<table>
<thead>
<tr>
<th>Case</th>
<th>RTW Cases: Phases 1–7</th>
<th>SSW Cases: Phases 1–7</th>
</tr>
</thead>
<tbody>
<tr>
<td>6</td>
<td>10 Feb 1988 to 27 Feb 1988</td>
<td>9 Dec 1984 to 19 Jan 1985</td>
</tr>
<tr>
<td>7</td>
<td>26 Dec 1990 to 12 Jan 1991</td>
<td>17 Jan 1987 to 21 Mar 1987</td>
</tr>
<tr>
<td>12</td>
<td>28 Feb 1997 to 20 Mar 1997</td>
<td>9 Nov 1993 to 10 Jan 1994</td>
</tr>
<tr>
<td>14</td>
<td>27 Feb 2000 to 16 Mar 2000</td>
<td>7 Nov 1996 to 30 Dec 1996</td>
</tr>
<tr>
<td>15</td>
<td>11 Nov 2001 to 7 Dec 2001</td>
<td>9 Dec 1997 to 10 Jan 1998</td>
</tr>
<tr>
<td>16</td>
<td>18 Nov 2002 to 7 Dec 2002</td>
<td>10 Dec 1998 to 11 Jan 1999</td>
</tr>
<tr>
<td>17</td>
<td>5 Jan 2005 to 26 Jan 2005</td>
<td>25 Jan 1999 to 10 Mar 1999</td>
</tr>
<tr>
<td>19</td>
<td>8 Nov 2006 to 24 Nov 2006</td>
<td>24 Jan 2001 to 24 Mar 2001</td>
</tr>
<tr>
<td>20</td>
<td>19 Jan 2007 to 7 Feb 2007</td>
<td>15 Feb 2002 to 23 Mar 2002</td>
</tr>
<tr>
<td>21</td>
<td>18 Nov 2008 to 30 Nov 2008</td>
<td>24 Nov 2003 to 16 Jan 2004</td>
</tr>
<tr>
<td>22</td>
<td>15 Nov 2010 to 26 Nov 2010</td>
<td>10 Jan 2006 to 21 Mar 2006</td>
</tr>
<tr>
<td>23</td>
<td>2 Dec 2010 to 17 Dec 2010</td>
<td>21 Feb 2008 to 29 Mar 2008</td>
</tr>
<tr>
<td>24</td>
<td>27 Dec 2010 to 10 Jan 2011</td>
<td>10 Jan 2009 to 13 Feb 2009</td>
</tr>
<tr>
<td>25</td>
<td>14 Nov 2012 to 29 Nov 2012</td>
<td>2 Nov 2009 to 4 Jan 2010</td>
</tr>
<tr>
<td>26</td>
<td>16 Dec 2013 to 5 Jan 2014</td>
<td>17 Jan 2010 to 23 Feb 2010</td>
</tr>
<tr>
<td>28</td>
<td>10 Feb 2014 to 28 Feb 2014</td>
<td>9 Jan 2013 to 20 Mar 2013</td>
</tr>
<tr>
<td>29</td>
<td>3 Nov 2014 to 13 Nov 2014</td>
<td></td>
</tr>
<tr>
<td>30</td>
<td>17 Feb 2016 to 4 Mar 2016</td>
<td></td>
</tr>
<tr>
<td>2015-16&lt;sup&gt;a&lt;/sup&gt;</td>
<td>22 Dec 2015 to 15 Jan 2016</td>
<td>13 May 2016 to 27 Apr 2016</td>
</tr>
</tbody>
</table>

<sup>a</sup>Not used in the composite. The period of 2015–2016 shows the date of the recent extreme cases that are included in the composite analysis.

**Table 4.1.** The Dates of Phase 1 Through Phase 7 of the Identified Cases.
CHAPTER 5

CONCLUSIONS AND FUTURE WORK

Over the past two decades, the increase in extreme weather events has caused substantial property damage and loss globally. The causes of weather and climate extremes are varied. This dissertation sought to examine the influences of variations in natural and anthropogenic climate variability on extreme weather and climate events, in order to better understand the connection between global circulation and extreme events and to improve the predictability of extreme weather and future climate. To investigate possible causes and processes of climate variability, various statistical methods were used to diagnose extreme weather events. This dissertation attempted to investigate extreme weather events in different regions from the perspective of global large-scale circulation to examine the possible links between extreme events and different sources of climate variability. Three specific topics were addressed: (1) Chapter 2 investigated a water shortage case due to the influences of both multidecadal and interdecadal variability of the East Asian Summer Monsoon (EASM) during early and late summer, (2) Chapter 3 addressed the Californian drought variations and its corresponding atmospheric and oceanic connections, and (3) Chapter 4 evaluated the subseasonal Arctic tropospheric warming events and their possible external forcings.

Chapter 2 examined a severe water shortage case in the Western North Pacific (WNP) island of Taiwan in 2014–2015. Different climate variability phenomena were
found to influence the EASM system and contribute to drought in the early and late summer. During the early summer season, monsoon precipitation has shifted to later in the season. This change is due to the intensive upper-level cyclonic anomaly over Eastern China and the low-level anticyclonic anomaly over the WNP, with the anthropogenic warming effect being a possible mechanism. On the other hand, during the late summer (July and August) the low-level deep convergence over the WNP shows a multidecadal (~20 years) timing variation, in which the low-level deep convergence anomaly oscillates between August and July. The change was shown in different reanalysis data and is associated with tropical cyclone activity. Evidence showed that the large-scale low-level convergence reveals similar multidecadal variations during spring, but sea surface temperature does not show relative forcing. Moreover, this 20-year oscillation does not match other atmospheric and oceanic variabilities, indicating that the variation stems from internal (natural) atmospheric variability. The connection between the timing variation of WNP low-level convergence and large-scale low-level convergence in spring may result from consistent variation in the tropical intraseasonal Madden–Julian Oscillation between spring and early summer. Thus, the results depict a new climate variation and provide a possible single precursor, which helps improve understandings of the EASM and facilitates better prediction of climate variation during summer over the WNP.

Anthropogenic warming has been linked to the increase in weather and climate extremes, but the mechanisms and influences of natural climate variability are not fully understood. Motivated by the fourth year of a record-breaking drought in California from the winters of 2011–2012 to 2014–2015, Chapter 3 focused on investigating circulation
variations during the Californian wintertime drought to investigate its possible causes. The diagnostic results showed that, during the California drought winters, ENSO-related Pacific–North American (PNA) teleconnection and the atmospheric internal North Pacific Oscillation (NPO) dominate circulation variations. Analyses suggest that Californian drought results from the influences of different kinds of climate variability. Positive PNA-related circulation leads to drier conditions in northern California and negative NPO-related circulation induces drier conditions in southern California. The combined influences of these two climate variations produce varied drought levels in California, suggesting that the mechanisms causing the Californian drought are different every year. Moreover, the CESM Large Ensemble projection results to 2100 indicate that the impacts of these two climate variations do not change significantly, although future projections of Californian climate show more extreme wet/dry anomalies. The results of this study indicate that more research is necessary to investigate the causes of Californian drought and provide better seasonal forecasts.

Arctic circulation significantly impacts mid-latitude weather in the boreal winter. Arctic surface air temperatures are warming faster than the rest of the planet on average. To investigate the mid-latitude weather and climate extremes from the change in Arctic circulation, Chapters 4 further studied the subseasonal Arctic tropospheric warming event and its connection to sea-ice loss and deep convection over the tropical Pacific. The subseasonal Arctic temperature warming event is associated with weather extremes in the mid-latitudes. Analysis results showed an increase in Arctic tropospheric warming events, implying that more extreme weather events in the mid-latitudes were likely to have occurred during the past two decades as a result. Model simulations indicated that
sea-ice loss could amplify Arctic tropospheric warming events. Further analysis suggested that the increasing frequency of tropical deep convection over the western and eastern Pacific is associated with Arctic warming events. A tropical deep convection could induce Rossby wave trains to influence high-latitude circulation and contribute to Arctic warming.

This dissertation provided a framework of how long-term variability of large-scaled circulations connect to weather and climate extremes for future research to understand how anthropogenic and internal climate variations affect weather and climate extremes. Nevertheless, the connections between extreme weather and circulation patterns in different regions are important and need further study to improve weather forecasting and climate prediction. Future work could follow the results of this dissertation to further focus on physical mechanisms of subseasonal variations in tropical deep, the causes of Arctic tropospheric warming, and the combined influences of tropical and extratropical circulation on mid-latitude extreme weather. More diagnostic analysis should be conducted to uncover additional climate variations. Additionally, model simulations to investigate the contribution percentage of tropical forcings on extratropical circulations will be a major development in future research for understanding climate dynamics.
APPENDICES
Appendix A

PERMISSIONS AND RELEASE LETTERS

License Agreement I: Materials for Chapter 2

Welcome to RightsLink

This article is available under the terms of the Creative Commons Attribution License (CC BY) (which may be updated from time to time) and permits use, distribution and reproduction in any medium, provided that the Contribution is properly cited.

For an understanding of what is meant by the terms of the Creative Commons License, please refer to Wiley's Open Access Terms and Conditions.

Permission is not required for this type of reuse.

Wiley offers a professional reprint service for high quality reproduction of articles from over 1400 scientific and medical journals. Wiley's reprint service offers:

• Peer reviewed research or reviews
• Tailored collections of articles
• A professional high quality finish
• Glossy journal style color covers
• Company or brand customisation
• Language translations
• Prompt turnaround times and delivery directly to your office, warehouse or congress.

Please contact our Reprints department for a quotation. Email corporate.sales.europe@wiley.com or corporate.sales.usa@wiley.com or corporate.sales.SC@wiley.com.
License Agreement II: Materials for Chapter 2

Copyrights

Copyright and Licensing

For all articles published in MDPI journals, copyright is retained by the authors. Articles are licensed under an open access Creative Commons CC BY 4.0 license, meaning that anyone may download and read the paper for free. In addition, the article may be reused and quoted provided that the original published version is cited. These conditions allow for maximum use and exposure of the work, while ensuring that the authors receive proper credit.

In exceptional circumstances articles may be licensed differently. If you have specific condition (such as one linked to funding) that does not allow this license, please mention this to the editorial office of the journal at submission. Exceptions will be granted at the discretion of the publisher.
License Agreement II: Materials for Chapter 3
License Agreement III: Materials for Chapter 4

Open Access Article

This article is available under the terms of the Creative Commons Attribution Non-Commercial No Derivatives License CC BY-NC-ND (which may be updated from time to time) and permits non-commercial use, distribution, and reproduction in any medium, without alteration, provided the original work is properly cited and it is reproduced verbatim.

For an understanding of what is meant by the terms of the Creative Commons License, please refer to Wiley's Open Access Terms and Conditions.

Permission is not required for non-commercial reuse. For commercial reuse, please hit the "back" button and select the most appropriate commercial requestor type before completing your order.

If you wish to adapt, alter, translate or create any other derivative work from this article, permission must be sought from the Publisher. Please email your requirements to RightsLink@wiley.com.
Permission to Reprint from Co-author I: Chapter 2

07-25-2018

Chi-Hua Wu
128 Academia Road, Section 2,
Nankang, Taipei, 115, Taiwan

Dear Dr. Chi-Hua Wu:

I am in the process of preparing my dissertation in the Plants, Soils, and Climate department at Utah State University, in fulfillment of my degree program for Ph.D., Climate Science. I am requesting your permission to include as a chapter in my dissertation the paper: Interdecadal change of the active-phase summer monsoon in East Asia (Meiuy) since 1979, published in Atmospheric Science Letters, Letters, which you are listed as co-author. A footnote with citation information will appear on the first page of this chapter, and a copy of this permission letter will be included as an appendix to the dissertation. Please indicate your approval of this request by filling in the form below granting permission to include this work in my dissertation. If you have any questions, please email or call me at the number below.

Thank you for your cooperation,

Yen-Heng Lin
(435) 213-7804

I hereby give permission to Yen-Heng Lin to reprint the following material in his dissertation. (Include full bibliographical information, including page numbers and specifications (e.g., table numbers, figure numbers, direct quotation of lines).


Signed: ________________________________
Permission to Reprint from Co-author II: Chapter 4

07-25-2018

Ming-Ying Lee
64, Gongyuam Road, Taipei 10048, Taiwan

Dear Dr. Ming-Ying Lee:

I am in the process of preparing my dissertation in the Plants, Soils, and Climate department at Utah State University, in fulfillment of my degree program for Ph.D., Climate Science.

I am requesting your permission to include as a chapter in my dissertation the paper: Accelerated increase in the Arctic tropospheric warming events surpassing stratospheric warming events during winter, published in Geophysical Research Letters, Letters, which you are listed as co-author. A footnote with citation information will appear on the first page of this chapter, and a copy of this permission letter will be included as an appendix to the dissertation. Please indicate your approval of this request by filling in the form below granting permission to include this work in my dissertation. If you have any questions, please email or call me at the number below.

Thank you for your cooperation,

Yen-Heng Lin
(435) 213-7804

---

No problem, I approve.

---

I hereby give permission to Yen-Heng Lin to reprint the following material in his dissertation. (Include full bibliographical information, including page numbers and specifications (e.g., table numbers, figure numbers, direct quotation of lines).


Signed: Ming-Ying Lee

---

Ming-Ying Lee 李明婷
Associate Technical Specialist 學士
Long-Range Forecast Section
Central Weather Bureau 中央氣象局/長期預報課
TEL: +886-2-23691214 FAX: +886-2-23691180
64 Gongyuam Road Taipei, Taiwan 臺北市公園路64號
E-mail: plato@cbwb.gov.tw
Permission to Reprint from Co-author III: Chapter 4

07-25-2018

Jonathan D. D. Meyer,
4825 Old Main Hill, Logan, UT 84322

Dear Dr. Jonathan D. D. Meyer:

I am in the process of preparing my dissertation in the Plants, Soils, and Climate department at Utah State University, in fulfillment of my degree program for Ph.D., Climate Science. I am requesting your permission to include as a chapter in my dissertation the paper: Accelerated increase in the Arctic tropospheric warming events surpassing stratospheric warming events during winter, published in Geophysical Research Letters, Letters, which you are listed as co-author. A footnote with citation information will appear on the first page of this chapter, and a copy of this permission letter will be included as an appendix to the dissertation. Please indicate your approval of this request by filling in the form below granting permission to include this work in my dissertation. If you have any questions, please email or call me at the number below.

Thank you for your cooperation,

Yen-Heng Lin
(435) 213-7804

I hereby give permission to Yen-Heng Lin to reprint the following material in his dissertation.
(Include full bibliographical information, including page numbers and specifications (e.g., table numbers, figure numbers, direct quotation of lines).


Signed: ________________________________
Dear Dr. Philip Rasch:

I am in the process of preparing my dissertation in the Plants, Soils, and Climate department at Utah State University, in fulfillment of my degree program for Ph.D., Climate Science.

I am requesting your permission to include as a chapter in my dissertation the paper: Accelerated increase in the Arctic tropospheric warming events surpassing stratospheric warming events during winter, published in Geophysical Research Letters, which you are listed as co-author. A footnote with citation information will appear on the first page of this chapter, and a copy of this permission letter will be included as an appendix to the dissertation. Please indicate your approval of this request by filling in the form below granting permission to include this work in my dissertation. If you have any questions, please email or call me at the number below.

Thank you for your cooperation,

Yen-Heng Lin
(435) 213-7804

________________________________________________________________________


Signed: Philip Rasch ______________________
Education

Ph.D. Student 8/2014 – present
*Climate Science, Department of Plants, Soils, and Climate. Utah State University*
Concentrations: Arctic Climate, Tropical Meteorology, Climate change, Monsoon

*Istitute of Atmospheric Physics, National Central University, Taiwan*
Thesis: The study of intra-seasonal variation of India summer monsoon in 1979

Bachelor of Science 9/1999 – 6/2003
*Department of Atmospheric Sciences, National Central University, Taiwan*

Publications


**Conference Presentations:**


Teaching Experience

PSC-6123 Climate Data Analysis: 2017 Fall Semester Graduate Teaching Assistant

Academic Working Experiences

Research Assistant 08/2014 – 08/2018
Dr. Shih-Yu Wang, Climate Science, Department of Plants, soils, and Climates.
Utah State University

Research Assistant 08/2012 - 08/2014
Dr. Min-Hui Lo, Department of Atmospheric Science, National Taiwan Univ., Taiwan

Honors

Presidential Doctoral Research Fellowships 2014- 2018
Utah State University