The cloud-radiative forcing of the U.S. landfalling atmospheric rivers

Qianwen Luo
Purdue University

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THE CLOUD-RADIATIVE FORCING OF THE U.S. LANDFALLING
ATMOSPHERIC RIVERS

A Dissertation
Submitted to the Faculty
of
Purdue University
by
Qianwen Luo

In Partial Fulfillment of the
Requirements for the Degree
of
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For the degree of Doctor of Philosophy

Is approved by the final examining committee:
Qianlai Zhuang
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Jui-lin F. Li

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Approved by Major Professor(s): Wen-wen Tung

Approved by: Indrajit Chaubey
Head of the Departmental Graduate Program

09/28/2016
Date
For my parents, for their compassion and love.
ACKNOWLEDGMENTS

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Finally, I want to thank Google for making resources available.
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<td>AR</td>
<td>Atmospheric River</td>
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<td>CRF</td>
<td>Cloud-Radiative Forcing</td>
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<td>NE Pacific</td>
<td>Northeastern Pacific</td>
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<td>U.S.</td>
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<td>WRF</td>
<td>Weather Research and Forecasting</td>
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ABSTRACT

Luo, Qianwen Ph.D., Purdue University, December 2016. The Cloud-Radiative Forcing of the U.S. landfalling Atmospheric Rivers. Major Professor: Wen-wen Tung.

Atmospheric rivers (ARs) are narrow channels in the atmosphere that transport an enormous amount of moisture from the tropics to the higher latitudes. Streaks of highly reflective clouds are observed along with the ARs in satellite imagery. These clouds both influence the moisture transport of ARs, as well as modify the Earth-Atmospheric energy budget through pathways such as cloud-radiative forcing (CRF). This dissertation studies the CRF of the U.S. Landfalling ARs in weather and climate scales.

Three crucial questions are addressed. First, how do clouds produced by the ARs modulate the moisture and heat balance of the Earth-Atmospheric system? Even though studies of ARs date back to the 90s, past research has been primarily focused on their hydrological impacts. We addressed this research gap by comparing the dominant types of precipitating clouds and convection of two ARs. Through quantifying their effects on the energy balance in the midlatitudes, we found that when deep convection was the dominant cloud types of an AR, impressive CRF cooling was produced. Second, what are the sufficient climate conditions for the extensive CRF in the continental U.S.? We studied 60 ARs that reached the California coast (the Southwest ARs) and 60 ARs that reached Pacific Northwest during Nov – Mar, 2000 – 2008. It was found that when these West-Coast ARs were followed by the moisture surge from the Gulf of Mexico (the Gulf-Coast AR), it resulted in apparent statewide CRF. Such condition happened more frequently in the Southwest-AR scenario. Third, how does the subgrid-scale-convection-induced CRF influence the moisture transport of ARs? We ran two WRF ARW simulations for a Southwest-AR that was followed by
a Gulf-Coast AR. The only difference between the two simulations was one considered the CRF of subgrid-scale clouds while the other did not. By comparing the two simulations, we found that the subgrid-scale-convection-induced CRF helped prolong the lifespan of clouds in an AR, thus enabling moisture to be transported further downstream.

In short, this work helps improve our understanding of CRF of the U.S. landfalling ARs from both weather and climate perspectives. Our results are useful for validating the representation of clouds and radiation processes in weather and climate models, thereby help to improve AR predictions.
1 INTRODUCTION

Atmospheric rivers (ARs) are elongated channels of enhanced moisture flux in the atmosphere (Newell et al., 1992; Zhu and Newell, 1994). They perform ocean-to-ocean and ocean-to-land moisture transports (Newman et al., 2012), accounting for >90% of the total meridional moisture flux in the midlatitudes (Zhu and Newell, 1998). They are typically parts of the warm conveyor belts (WCBs), which are strongly ascending airstreams near winter extratropical cyclones. The WCBs transport and redistribute heat globally, and are featured by intense latent heat release and precipitation formation (e.g., Carlson, 1980; Browning, 1990; Eckhardt et al., 2004; Pfahl et al., 2014). When making landfall, they often lead to enhanced convective precipitation, disastrous flooding and induce extensive cloud coverage in the landfalling locations (Lavers et al., 2011; Gorodetskaya et al., 2014; Guan and Waliser, 2015; Eiras-Barca et al., 2016). Hence, realistic representation in model physics and accurate prediction of landfalling ARs are required to mitigate AR-related damages, particularly in the context of climate change (Dettinger et al., 2011a; Gao et al., 2015; Hagos et al., 2015; Radić et al., 2015; Warner et al., 2015).

Previous North American AR research has been focused on the hydrological impacts of ARs that make landfall on the West Coast (Ralph et al., 2011; Neiman et al., 2008a; Leung and Qian, 2009; Warner et al., 2012; Kim et al., 2013; Rutz et al., 2014; Payne and Magnusdottir, 2014) and recently in the Central-Eastern US (AR\textsubscript{GULF}) (Dirmeyer and Kinter, 2010; Moore et al., 2012; Lackmann, 2013; Lavers and Villarini, 2013, 2015; Mahoney et al., 2016). Numerous observational (e.g., Neiman et al., 2008b; Ralph et al., 2011; Newman et al., 2012) and model trajectory studies (e.g., Bao et al., 2006; Knippertz and Wernli, 2010; Ryoo et al., 2011; Sodemann and Stohl, 2013) have been performed to investigate the moisture sources as well as the transport mechanisms of ARs. Even though streaks of highly reflective clouds are observed
along with the ARs in satellite imagery (e.g., Luo, 2013), and persistent ice clouds are found in the WCB simulations (Madonna et al., 2014), it remains unclear how they influence the moisture transport of ARs, the ambient atmosphere, and the surface energy budget through pathways such as cloud-radiative forcing (CRF). Considering that CRF is an important forcing in the Earth system in weather and climate scales (Wetherald and Manabe, 1988; Randall et al., 1989; Stephens, 2005; Waliser et al., 2009; Fovell et al., 2010; Li et al., 2012; Bu et al., 2014; Li et al., 2016), we are motivated to fill the research gaps by studying the CRF of the U.S. landfalling ARs in weather and climate scales. In particular, we ask: 1) how do clouds produced by the ARs modulate the moisture and heat balance of the Earth-Atmospheric system? 2) what are the sufficient climate conditions for the extensive CRF in the continental U.S.? 3) how does the subgrid-scale-convection-induced CRF influence the moisture transport of ARs? These three questions are addressed in Chapters 2, 3 and 4, respectively. Discussions and implications for further research are in Chapter 5.
2 CASE STUDY OF MOISTURE AND HEAT BUDGETS WITHIN ATMOSPHERIC RIVERS

This chapter is adapted from Luo and Tung (2015), which for the first time quantified the CRF of the ARs. Specifically, we ask how the spatial characteristics, the amounts, and the dominant types of the precipitating clouds and convection in the ARs 1) influence the heat and moisture transport within ARs, and 2) modify the ambient atmospheric heat and moisture budgets. In probing these questions, we conducted a detailed case diagnosis on the precipitating systems of two ARs in January 2009, during the Year of Tropical Convection (YOTC; Waliser and Moncrieff, 2008; Moncrieff, 2010) over the Northeastern (NE) Pacific region (20° – 60° N, 180° – 120° W). The apparent heat source ($Q_1$) and apparent moisture sink ($Q_2$) were computed and analyzed for evidence of the dominant cloud types and to elucidate the collective impacts of subgrid-scale eddies on the surrounding environment. Then, comparisons among the column-integrated $Q_1$ and $Q_2$, the radiation budget, and surface heat fluxes were made.

This chapter is organized as follows. In section 2.1, we describe the data and methods employed for AR-structure construction as well as for heat and moisture budget residuals computations. In section 2.2, we offer a background overview of January 2009 and the synoptic-scale merging process of enhanced Integrated Water Vapor (IWV) bands. In section 2.3, we present the average vertical profiles of $Q_1$ and $Q_2$ around the local maxima of IWV or the AR ridge, and examine the vertically integrated heat and moisture budgets. Conclusions are in section 2.4.
2.1 Data & Methods

2.1.1 ECMWF YOTC data

This study used the European Centre for Medium-Range Weather Forecasts (ECMWF) high-resolution YOTC (e.g., Waliser and Moncrieff, 2008; Moncrieff, 2010) dataset as the primary data source. The YOTC dataset is available for a two-year period, from 1 May 2008 to 30 April 2010. It is produced by the ECMWF Integrated Forecast System, which comprises a 4D-Var data assimilation system, with high horizontal resolution of TL799 (∼25 km, May 2008 – December 2009) and TL1279 (∼16 km, January 2010 – April 2010), L91 model levels (91 vertical levels with the model top at 0.01 hPa).

The YOTC database includes 6-hourly global analysis (hereafter YOTC analysis) and up to 10-day forecasts (hereafter YOTC forecast, starting at 1200 UTC every day since 1 May 2008). The multiday forecasts are output at 3-h intervals. In this study, we retrieved the analysis and forecast data at 1°×1° horizontal grids with 25 pressure levels and surface level from http://apps.ecmwf.int/datasets/data/yotc_od/.

2.1.2 NOAA CMORPH precipitation

NOAA Climate Prediction Center morphing technique (CMORPH; Joyce et al., 2004) is one of the most commonly used and frequently validated dataset for global precipitation (e.g., Ebert et al., 2007; Tian et al., 2007; Sapiano and Arkin, 2009). This product is derived from four types of passive microwave measurements: Tropical Rainfall Measuring Mission Microwave Imager, Advanced Microwave Sounding Unit, the Special Sensor Microwave Imager, and the Advanced Microwave Scanning Radiometer. It uses the “morphing” techniques to solve the coverage gap problem and to provide microwave-derived precipitation fields with 0.25°×0.25° spatial resolution from 60° N to 60° S and 3-hourly temporal resolution from 1 December 2002 to the present. Because of its spatial and temporal resolutions, midlatitude data coverage,
as well as smooth spatial patterns, CMORPH is suitable for short-term precipitation detection (Tian et al., 2007). To compare CMORPH data with YOTC data, we downsample CMORPH to match the resolutions of YOTC.

2.1.3 GPCP daily precipitation

The Global Precipitation Climatology Project (GPCP) One-Degree Daily Precipitation version 1.2 dataset (Huffman et al., 2001) with $1^\circ \times 1^\circ$ spatial resolution is available daily from October 1996 to the delayed present. Using the threshold-matched precipitation index algorithm, precipitation between $40^\circ$ N and $40^\circ$ S are estimated from geosynchronous-orbit IR and low-orbit IR. The rescaled daily Television and Infrared Observation Satellite Operational Vertical Sounder and Atmospheric Infrared Sounder are used as the primary data sources for $40^\circ - 90^\circ$ in each hemisphere. In-between $40^\circ$ and $50^\circ$ in each hemisphere, smoothing is performed to solve the data inconsistency problem. We use the GPCP data mainly because of its high-latitude data coverage from $60^\circ$ to $90^\circ$ N.

2.1.4 AR definitions and the AR ridge construction

There exist several criteria to identify ARs. Zhu and Newell (1998) defined an AR as a filament-like structure of substantial moisture flux. Ralph et al. (2004) identified an AR as an elongated and narrow region ($\text{length} \geq 2000 \text{ km} \times \text{width} < 1000 \text{ km}$) with $\text{IWV} \geq 20 \text{ mm}$. We use the latter definition as the first step to identify AR events and then augment the identified ARs in space and time by loosing the geometrical constraint. At each instance, the broader definition captures a continuum of $\text{IWV} \geq 20 \text{ mm}$ over the NE Pacific region with an AR embedded, which we call the enhanced IWV surrounding ARs (AR-IWV hereafter). This enables us to consider the moisture reservoir of an AR and the time steps inclusively when an AR curved or deformed. Within each AR-IWV, we further define a narrow segment of maximum IWV by interpolating $2^\circ \times 2^\circ$ tiles, each of which is centered on a local IWV maximum.
This concentrated area is denoted as the AR ridge, which is marked with the thick contours in Figs. 2.1 and 2.2. The maximum precipitation (shaded area) tends to fall inside an AR ridge, which enables us to examine the propagation and evolution of precipitating systems embedded in ARs.

2.1.5 $Q_1$ and $Q_2$ calculations

In the absence of direct measurements of clouds and convective systems, their thermodynamic impacts on the ambient atmosphere are evaluated statistically using heat and moisture budget residuals ($Q_1$ and $Q_2$) computed with the $1^\circ \times 1^\circ$ YOTC analysis according to the thermodynamic and the moisture mass conservation laws. As suggested by the model trajectory study in Joos and Wernli (2012), net condensation, as well as deposition, contribute significantly to the total latent heating of WCBs, thus the $Q_1$ and $Q_2$ equations in Yanai et al. (1973) are modified to include the ice phase as follows (e.g., Johnson et al., 2016a):

$$Q_1 \equiv c_p \left( \frac{p}{p_0} \right)^\kappa \left( \frac{\partial \tilde{\theta}}{\partial t} + \mathbf{v} \cdot \nabla \tilde{\theta} + \omega \frac{\partial \tilde{\theta}}{\partial p} \right)$$

$$= Q_R + L_v (\bar{c} - \bar{e}) + (L_v + L_f) (\bar{d} - \bar{s}_s) + L_f (\bar{f} - \bar{m}) - \nabla \cdot \mathbf{s} \nabla \mathbf{v} - \frac{\partial s' \omega'}{\partial p}, \quad (2.1)$$

$$Q_2 \equiv -L_v \left( \frac{\partial \bar{q}}{\partial t} + \mathbf{v} \cdot \nabla \bar{q} + \omega \frac{\partial \bar{q}}{\partial p} \right)$$

$$= L_v (\bar{c} - \bar{e}) + L_v (\bar{d} - \bar{s}_s) + L_v (\nabla \cdot \bar{q} \nabla + \frac{\partial q' \omega'}{\partial p}) , \quad \text{and} \quad (2.2)$$

$$Q_1 - Q_2 \approx Q_R + L_f (\bar{d} - \bar{s}_s + \bar{f} - \bar{m}) - \left( \frac{\partial s' \omega'}{\partial p} + L_v \frac{\partial q' \omega'}{\partial p} \right), \quad (2.3)$$

where $\theta$ is the potential temperature, $q$ is the water vapor mixing ratio, $\mathbf{v}$ is the horizontal velocity, $\omega$ is the vertical $p$-velocity, $p$ is the pressure, $p_0 = 1000$ hPa, $\kappa = R/c_p$ with $R$ being the gas constant of dry air, $c_p$ is the specific heat capacity at constant pressure, $\nabla$ is the isobaric gradient operator, $Q_R$ is the radiative heating rate, $L_v$ and $L_f$ are the latent heat of vaporization and fusion, $c, e, d, s_s, f$ and $m$ are the
rates of condensation, evaporation, deposition, sublimation, freezing, and melting per unit mass of air, \( s \) is the dry static energy per unit mass of air. The overbar denotes the mean over a horizontal area equivalent to the mesh size of the gridded analysis, and the prime denotes the deviation from this mean, hence referring to subgrid-scale processes such as cloud convection, boundary layer fluxes, and turbulence. In deriving (3.1) and (3.2), the Reynolds conditions and their consequences are assumed to be accurate.

Equations (3.1) and (3.2) are calculated using the rhs of their first lines using the YOTC analysis, and are interpreted using the rhs terms in their second lines. Respectively, \( Q_1 \) represents the total effects of radiative heating, latent heat released due to microphysical phase changes, and the convergence of fluxes of sensible heat due to subgrid-scale eddies such as convection and turbulence, while \( Q_2 \) shows the total effects of net condensation and divergence of eddy moisture flux due to clouds and turbulence. In the presence of organized convection, the subgrid terms \(-\frac{\partial \overline{\omega}}{\partial p}\) and \(\frac{\partial \overline{\omega}}{\partial p}\) dominate the total eddy transports. The horizontal eddy transport terms \(-\nabla \cdot \overline{s \mathbf{v}}\) and \(\nabla \cdot \overline{q \mathbf{v}}\) may be ignored due to their typically small contributions (e.g., Arakawa and Schubert, 1974; Wu, 1994). Accordingly, the difference between \( Q_1 \) and \( Q_2 \) in (2.3) is indicative of net radiative heating, net latent heating associated with ice processes, and vertical eddy fluxes of moist static energy \( h = s + L_v q \).

Furthermore, (3.1)–(2.3) include processes transitioning in and out of the ice phase: \( d, \overline{s}, \overline{f}, \) and \( \overline{m} \). In convective updrafts, water vapor condenses rapidly and releases latent heat. Once the environmental temperature drops below 0\(^\circ\) C, the Bergeron process takes place; deposition outcompetes freezing and releases more latent heat. Among solid precipitation, high-density ice can survive above freezing temperatures long enough so that melting, rather than sublimation, likely contributes notably to diabatic cooling along with evaporation within downdrafts. In consequence, (3.1)–(2.3) can be approximated for convective systems as:
\[ Q_1 \approx Q_R + L_v(\bar{v} - \bar{a}) + L_f(\bar{d} - \bar{m}) - \frac{\partial \bar{s} \omega'}{\partial p}, \quad (2.4) \]

\[ Q_2 \approx L_v(\bar{v} - \bar{a}) + \frac{\partial q' \omega'}{\partial p}, \quad \text{and} \quad (2.5) \]

\[ Q_1 - Q_2 \approx Q_R + L_f(\bar{d} - \bar{m}) - \frac{\partial h' \omega'}{\partial p}. \quad (2.6) \]

The vertical structures of \( Q_1 \) and \( Q_2 \) that resulted from a subgrid cloud population that is predominantly deep convection are quite distinct from those that resulted from shallow convection or from convective organizations with prominent trailing stratiform clouds. Thus, the \( Q_1 \) and \( Q_2 \) profiles are commonly used to diagnose the dominant types of cloud systems (e.g., Nitta and Esbensen, 1974; Houze, 1989; Yanai and Johnson, 1993; Tung et al., 1999; Schumacher et al., 2008). For instance, as discussed in Johnson (1984), the deep convection type exhibits net heating (positive \( Q_1 \)) and drying (positive \( Q_2 \)) throughout the troposphere, with a primary \( Q_1 \) peak in the upper troposphere and a primary \( Q_2 \) peak in the lower troposphere; in contrast, the trailing stratiform type is associated with upper-tropospheric heating and drying due to condensation and deposition, and lower-tropospheric cooling and moistening contributed to melting and evaporation of precipitation.

### 2.1.6 Vertically integrated \( Q_1 \) and \( Q_2 \)

The vertical integrations of (2.4) and (2.5), have been widely used to determine the primary heat sources and moisture sinks (e.g., Luo and Yanai, 1984; Yanai and Tomita, 1998). In this study, the integrations are performed over the vertical range from \( p_T = 1 \) hPa to \( p_0 = 1000 \) hPa, denoted by

\[ < > = \frac{1}{g} \int_{p_0}^{p_T} \left( \right) dp. \quad (2.7) \]
The difference between \(< Q_1 >\) and \(< Q_2 >\) might be interpreted as follows (modified after, e.g., Luo and Yanai, 1984):

\[
< Q_1 > - < Q_2 > \approx < Q_R > + < Q_f > + F_S + F_{LH},
\]

(2.8)

where \(< Q_R >\), \(< Q_f >\), \(F_S\), and \(F_{LH}\) are the column-integrated net radiative heating rate, the net heating associated with the fusion \((L_f)\) term in (2.6) in the air column, the surface sensible heat flux, and the surface latent heat flux, respectively, per unit area. We further perform the spatio-temporal average (denoted by \([\ ]\)) for individual terms in (2.8) over various spatial domains, respectively, for the month of January, the AR1 event, and the AR2 event (see section 4c). The month of January has 124 time steps; AR1 has 19 time steps from 0000 UTC 04 January to 1200 UTC 08 January; and AR2 has 20 time steps from 0000 UTC 16 January to 1800 UTC 20 January. The \([< Q_1 >]\) and \([< Q_2 >]\) are computed from YOTC analysis. The \([< Q_R >]\), \([F_S]\), and \([F_{LH}]\) in section 5 are estimated from the YOTC forecast tendency data. This approach does not guarantee the budgets to be closed. However, as shown in section 5, the discrepancies remain one order of magnitude smaller than the leading terms.

### 2.2 Case overview


Figures 2.1 and 2.2 display the evolution of AR1 and AR2, respectively. As time elapsed, AR1 turned clockwise and had a west-east or southwest-northeast oriented AR1 ridge from 20° to 50° N. It made landfall from 1800 UTC 04 January to 1200 UTC 08 January (hereafter the AR1 period). Substantial oceanic precipitation was con-
centrated on the southwest side of AR1 (Fig. 2.1). In contrast, AR2 was south-north oriented roughly from 20° to 60° N. It first impacted land at 0000 UTC 16 January and left at 1800 UTC 20 January (hereafter the AR2 period). AR2 was narrower and was largely covered by enhanced precipitation (Fig. 2.2). In the late stages of both cases, a secondary IWV band was positioned to the west of the landfalling AR (Fig. 2.1 from 1200 UTC 07 January to 1200 UTC 08 January, and Fig. 2.2 at 1800 UTC 20 January). The secondary IWV band was typically co-occurred with another cyclone (Sodemann and Stohl, 2013), and would later catch up and merge with the remnants of the currently landfalling AR.

The two ARs were the most prominent cases in January 2009, which was an active month for landfalling ARs, featured by a weak La Niña condition. Figs. 2.3 and 2.4 show the 200-, 500-, and 850-hPa analyses for the mean state of January 2009 and the two ARs, respectively. At 200 hPa and 500 hPa, a ridge was located at around 135° W in the January mean state (Figs. 2.3a, b). At 850 hPa, a high-pressure center was at 35° N, 135° W while a low-pressure center was at 50° N, 170° W, with the strong southwest wind in between (Fig. 2.3c). The 850-hPa temperature was distributed in a similar pattern as the height fields. High 850-hPa specific humidity extended northeastward from the tropics to the midlatitudes (Fig. 2.3d).

During the AR1 period, the 200- and 500-hPa ridges were roughly at 140° W (Figs. 2.4a, b), to the west of those in the January mean state (Figs. 2.3a, b). At 850 hPa, enhanced 850-hPa west-southwest wind was between a high-pressure center at 32° N, 140° W and an elongated low-pressure system, which spanned from 45° N, 158° E to 63° N, 115° W (Fig. 2.4c). The 850-hPa specific humidity was west-east orientated. AR1 impacted the West Coast of North America from 40° to 50° N (Fig. 2.4d).

During the AR2 period, the 200- and 500-hPa ridges in south-north orientation were at around 122° W (Figs. 2.4e, f). The upper-level westerly jet split in the mid-Pacific into a northern branch into the Gulf of Alaska and a southern branch passing through Hawaii. The storm track was displaced along with the northern branch,
resulting in the north-south orientation of AR2. Pronounced 850-hPa southerly winds existed in the vicinity of an 850-hPa high pressure that centered at 40° N, 110° W, and an 850-hPa low pressure that centered at 52° N, 175° W (Fig. 2.4g). Correspondingly, the 850-hPa specific humidity band was south-north oriented. AR2 impacted the West Coast of North America from 55° to 60° N (Fig. 2.4h). It was narrower and shorter than its counterpart in the AR1 period.

The accumulated GPCP precipitation is illustrated in Fig. 2.5. AR1 had a broad precipitation band in a nearly west-east orientation. It made landfall in the province of British Columbia in Canada, and the states of Washington and Oregon in the U.S. AR2 had a narrow precipitation band in a south-north orientation. It penetrated inland Alaska and British Columbia. A comparison between specific humidity (Figs. 2.4d, h) and precipitation (Fig. 2.5) indicates that the likelihood of precipitation can be predicted from the high specific humidity content, especially over the midlatitude ocean. Over the ocean, accumulated precipitation with values ≥ 40 mm was mainly scattered on the southwest and northeast sides of AR1, whereas it was distributed continuously along the main path of AR2. In both ARs, precipitation was stronger in the landfalling regions than that over the ocean.

The large-scale convergence induced by merging multiple high IWV bands can increase the moisture content of an AR (e.g., Ralph et al., 2004; Bao et al., 2006). This process is often associated with multiple eastward propagating cyclones (Cordeira et al., 2013; Sodemann and Stohl, 2013). Clearly, the main merging location for AR1 was around 20° – 35° N, 180° – 165° W (the first two rows in Fig. 2.6a). In contrast, the location for AR2 extended to higher latitudes centered at 28° – 50° N, 160° – 140° W (the first row in Fig. 2.7a). After merging, the saturated and actual vapor pressures in the lower troposphere were largely enhanced. Fig. 2.8 depicts the differences in saturated vapor pressure of 2-m air between the AR periods and the no-AR period. The latter period is computed by excluding time steps in AR1 and AR2 periods from January 2009. Qualitatively similar features are observed for both ARs. First, positively saturated vapor pressure was mostly present inside the
enhanced IWV bands. Second, the local maxima of saturated vapor pressure were situated near the merging locations (Figs. 2.6 - 2.8), mostly outside of the maximum precipitation centers (Figs. 2.8 and 2.5). The positive differences imply an increase of the maximum possible amount of moisture that can exist in the lower-tropospheric air, owning to the conveyance and confluence of warm air by the ARs.

The merging process amplifies precipitation, which consumes moisture as well as releases latent heat in an AR. Figs. 2.6b and 2.7b present the average vertical profiles between 30° and 35° N for the wind, specific humidity and positive $Q_1$ for AR1 and AR2, respectively. Black and gray arrows mark the locations of the primary specific humidity peaks and the secondary specific humidity peaks, respectively. The peaks are in correspondence with the IWV bands in Figs. 2.6a and 2.7a. For AR1 at 0600 UTC 02 January, the primary specific humidity peak was located from 160° E to 170° W. The secondary specific humidity peak was at around 165° W. The $Q_1$ maximum was at around 170° E, 500 – 700 hPa, coinciding with substantial upward motions (the first row in Fig. 2.6b). As the high IWV bands merged and proceeded eastward, the primary specific humidity peak became narrower, and the associated heating intensified (the second and the third rows in Fig. 2.6). At 0600 UTC 05 January, specific humidity peak was at 170° W, with a weakening heating center (the fourth row in Fig. 2.6b). At 0600 UTC 06 January, the specific humidity peak propagated to around 160° W, while the $Q_1$ maximum became hard to discern (the last row in Fig. 2.6b). For all time steps in Fig. 2.6b, strong westerlies were present to the west of the primary specific humidity peaks, in contrast to the weak wind to the east of the primary specific humidity peaks. A similar evolution is observed in AR2 (Fig. 2.7b). However, the primary specific humidity peak was narrower than that in AR1.
2.3 Precipitating convection

2.3.1 Along-AR propagation

Figures 2.9 and 2.10 depict $Q_1$, $Q_1 - Q_2$, and CMORPH precipitation along the AR ridges from the southwest end to the northeast end. They are taken at time steps identical with those in Fig. 2.1 for AR1 and overlapped with those in Fig. 2.2 for AR2. The x-axis is the distance from the southwest end. As a reference for the spatial progression of the AR1 ridge, three vertical lines are drawn to show where AR1 intersected longitudes at 165°, 153°, and 141° W, respectively. For the mainly north-south oriented AR2, the vertical lines are for latitudes at 25°, 35°, and 45° N, respectively. The tropopause was around 200 hPa in the AR1 ridge and was around 300 hPa in the AR2 ridge (indicated by $1.5 \times 10^{-6} \text{ m}^2 \text{s}^{-1} \text{K kg}^{-1}$ or 1.5 PVU, not shown). The different tropopause heights between the two ARs likely resulted from their different latitudinal ranges.

The precipitating systems in AR1 underwent two distinct phases of development characterized by heating ($Q_1$) profiles, which are described as strengthening and weakening phases as follows. Figs. 2.9a–e show the strengthening phase of the precipitating systems in AR1. The system appears to be composed of two convective organizations, each with a spatial scale around 2000 km horizontally, evolving from vertical to tilted structures in roughly two days. At 0000 UTC 04 January (Fig. 2.9a), a 400 – 900 hPa heating with $Q_1 \geq 15 \text{ K day}^{-1}$ emerged at 30° – 35° N, 180° – 175° W. Maximum heating was near 500 hPa. Cooling existed below 950 hPa. At 1800 UTC 04 January (Fig. 2.9b), the heating became elongated and vertically tilted from 25° – 40° N, 180° – 166° W. Its west part showed net heating from 300 – 900 hPa. Its east part exhibited a mature mesoscale convective system type of upper-tropospheric heating (maximized near 450 hPa) and lower-tropospheric cooling (peaked at 800 hPa). At 0600 UTC 05 January (Fig. 2.9c), heating expanded eastward to 30° – 40° N, 178° – 160° W. At 1800 UTC 05 January, another system started, apparently mixed with shallower and deep convection types (left column of Fig. 2.9d), producing localized
intensive precipitation up to 3 mm hr$^{-1}$ at 29° – 36° N (right column of Fig. 2.9d). At 1800 UTC 06 January (Fig. 2.9e), an obviously tilted heating was at 26° – 44° N, 162° – 147° W. The tilting tendency, again, implies a transition of dominant cloud types from shallower convection on the west side to deeper convection and trailing stratiform on the east side.

Figures 2.9f–g are considered the weakening phases of the AR1 precipitating systems. The precipitating system propagated eastward and subsided (Figs. 2.9f, g), and its remnants produced less than 30 K day$^{-1}$ of $Q_1$ at 1200 UTC 07 and 1800 UTC 07 January. The $Q_1$ maxima were largely confined below 600 hPa. The strong $Q_1$ signals at the east end of the AR1 ridge (e.g., Figs. 2.9f) were produced by the landfalling precipitating systems; hence, they are not discussed here.

The precipitating systems in AR2 covered a significant fraction of the AR ridge, which was, however, shorter than that of AR1 (see Figs. 2.9 and 2.10). In the strengthening phase of AR2 at 1200 UTC 17 January (Fig. 2.10a), the primary heating with $Q_1 \geq 15$ K day$^{-1}$ was to the south of 45° N from 400 to 900 hPa. It was vertically tilted over a horizontal distance around 2000 km, similar in structure and scale to the system in AR1. The entire precipitating systems strengthened and propagated northward, and at 0000 UTC 18 January (Fig. 2.10b), the heating center expanded from 23° – 48° N, 145° – 141° W. The tilt of $Q_1$ is suggestive of dominant cloud types transitioning from shallower to deep convection with trailing stratiform; however, it was not as pronounced a feature here as in AR1. In the weakening phase of AR2, precipitating systems still covered a substantial portion of the AR2 ridge (Fig. 2.10c at 1200 UTC 18 January, and Fig. 2.10d at 0000 UTC 19 January).

The vertical profiles of $Q_1 - Q_2$ appear to be complicated juxtapositions of latent heating and eddy transports associated with various cloud types. Equation (2.6) provides simplified guidelines to understand the $Q_1 - Q_2$ maps. The variable $Q_R$ in the troposphere tends to be on the order of -1 K day$^{-1}$. A pronounced feature in these maps is a dipole of a negative center immediately below a positive center. There can be two sets of dipoles stacked vertically in the low to midtroposphere, such as around
1500 km in Fig. 2.9b, around 1000 km in Fig. 2.9c, and between 1000 and 3000 km in Figs. 2.10a–c. These dipoles indicate divergence of moist static energy \((h)\) at the lower level and convergence at the upper level, which support strong evidence of vertical transport of moist static energy via convection. The two stacked dipoles are suggestive of the presence of at least two dominant convection cloud types detraining at different heights. Shallow cumulus and stratiform systems lack such a transport mechanism. Hence, they tend to yield a vertically uniform \(Q_1 - Q_2\) field. However, the trailing stratiform may be associated with a center of depositional heating above freezing level and melting cooling around the freezing level, according to (2.6). The depositional heating–melting cooling might have enhanced the dipoles between 400 and 800 hPa at 1500 km in Fig. 2.9b, at 1000 km in Fig. 2.9c, and at 3000 km in Fig. 2.10b. Last, under the undisturbed condition, the typical mixed layer turbulent mixing results in the convergence of heat and moisture, hence, positive \(Q_1 - Q_2 - Q_R\) in the PBL. Examples can be seen in the narrow positive zones in the lowermost troposphere in Figs. 2.9f,g.

In the weakening phase, AR2 did not exhibit the shallower-convection-dominant heating profile as AR1 did in Figs. 2.9f and g. This and the large precipitating convection coverage of the short AR2 ridge differentiate the two AR events. Considering that the precipitating convective organizations in both ARs developed to a similar spatial scale, one may conjecture that these systems drew moisture from the moisture reservoir in the tropics and depleted most of it on the spot. For AR1 to deliver moisture and make an impact upon landfall at a farther downstream location, extratropical heat and moisture sources must have been tapped into; therefore, the significant shallower cloud types and PBL mixing took place at its later stage.

### 2.3.2 Average \(Q_1\) and \(Q_2\) profiles

In Figs. 2.11 and 2.12, we conduct multiscale comparisons of the \(Q_1\) and \(Q_2\) inside the precipitating areas within the AR ridges, inside the AR-IWV, and outside
the AR-IWV in the NE Pacific region. Fig. 2.11 illustrates the average $Q_1$ and $Q_2$ profiles for the precipitating systems only in the AR ridge (hereafter the *precipitating* AR ridge). During the strengthening phase of precipitating systems in the AR1 ridge (Fig. 2.11a), a transition from deep convection type of low- to mid-tropospheric heating and drying (e.g., 0600 UTC 04 January and 1200 UTC 04 January) to the stratiform dominant type of midtropospheric heating and drying is observed ($\sim$550 hPa at 1800 UTC 04 January, see Johnson, 1984; Houze, 2004; Schumacher et al., 2008). However, the stratiform type of midtropospheric heating and drying quickly weakened. During the weakening stage (Fig. 2.11b), the main heating and drying peak shifted to around 800 hPa (e.g., 1800 UTC 06 January, 1200 UTC 07 January – 1800 UTC 07 January). This indicates the existence of low-level clouds, and the weakening of convective transport. On the contrary, the precipitating AR2 ridge had a robust deep convection type of heating throughout the strengthening phase. For example, at 1800 UTC 16 January – 0600 UTC 17 January in Fig. 2.11c, the $Q_1$ and $Q_2$ extrema were clearly separated, suggesting the presence of eddy vertical transport of moist static energy. Strong heating and drying remained prominent in the third landfall day of AR2 (e.g., 1200 UTC 18 January in Fig. 2.11d). The primary $Q_1$ peak was between 550 and 650 hPa, and the $Q_2$ peak was around 650 hPa during the weakening phase (Fig. 2.11d), implying the abundance of midtropospheric clouds. Notice that $Q_1$ profiles above 300 hPa in AR1 (Figs. 2.11a, b) and above 400 hPa in AR2 (Figs. 2.11c, d) are likely around and above the tropopause. The budget residuals are subjected to erroneous values arising from small perturbations in vertical velocity multiplied with a vertical temperature gradient under large static stability. The variations seen above these levels are therefore not interpreted.

Figure 2.12 compares the average $Q_{1f}$ (dark lines) and $(Q_1 - Q_2)_f$ (dark dashed lines) profiles for the AR-IWV average over the AR1 and AR2 periods. Recall that even though both AR ridges contained precipitating convective organizations, the one in the AR1 ridge was overtaken by shallower convection at its later stage prior to landfall. So, we anticipate that on average heating and drying associated with deep
convective systems affected AR1-IWV less than AR2-IWV. Indeed, such a difference is observed. As seen in Fig. 2.12a, negative \((Q_1 - Q_2)_I\) was present throughout most of the troposphere above 900 hPa in AR1-IWV, which could be largely explained by radiative cooling. In contrast, a double-peak structure of \((Q_1 - Q_2)_I\) was present in AR2-IWV, along with an overall stronger low- to mid-tropospheric positive \(Q_{1I}\) (Fig. 2.12b). As discussed in the previous subsection, the double-peak profile of \((Q_1 - Q_2)_I\) is likely the manifestation of vertical transports of moist static energy associated with convection detraining at different heights, overlaid with signals of depositional heating between 400 and 600 hPa and melting cooling around 600 hPa associated with a stratiform anvil.

Outside AR-IWV, large-scale downward motion largely dominated (not shown), along with heating \((Q_{1O}\) in Figs. 2.12a, b) and moistening due to PBL turbulence mixing, and minor midtropospheric drying in both ARs. However, the depth of the low-level heating and moistening was different between the two cases. The \(Q_{1O}\) profile for the AR1 period (Fig. 2.12a) exhibits heating below 900 hPa and cooling between 400 and 900 hPa, with a heating maximum of 3 K day\(^{-1}\) near the surface. Positive \((Q_1 - Q_2)_O\) existed below 750 hPa due to the prominent moistening in these levels. In contrast, a deeper layer of near-surface heating and moistening is observed during AR2, peaking at around 900 hPa (Fig. 2.12b). Moreover, radiative cooling dominated the heating profile above 800 hPa in the AR1 period (Fig. 2.12a), yet its impact seems to be offset by other sources of heating during the AR2 period (Fig. 2.12b).

### 2.3.3 Vertically integrated \(Q_1\) and \(Q_2\)

To yield insights into the net impacts of subgrid-scale processes on the surrounding environment, we compare \(<Q_1>\), \(<Q_2>\), \(<Q_R>\), \(F_{LH}\), and \(F_S\) averaged over the NE Pacific region, AR-IWV, outside AR-IWV, and the precipitating AR ridges during the January, AR1, and AR2 periods (Tables 2.1 and 2.2). To highlight results from the lhs of (2.8) in Tables 2.1, Fig. 2.13 shows the spatio-temporal average \(<Q_1>\) and
from YOTC analysis. The atmosphere over the NE Pacific was subjugated to a heat sink (negative spatio-temporal average $<Q_1>$, hereafter, $[<Q_1>]$) and moisture sink (positive $[<Q_2>]$) during the January and AR1 periods (Table 2.1), suggesting the importance of radiative cooling. However, it was a heat source and moisture sink in the AR2 period, implying the importance of latent heating and surface evaporation.

Meanwhile, weaker upwelling $[F_{LH}]$ and $[F_S]$ are found in the AR1 period than in the AR2 period for all regions (Table 2.2 and Fig. 2.14), even though AR1 and AR2 had comparable near-surface wind speed (not shown). Such a contrast amplified in IWV, with downward $[F_S]$ being observed in AR1-IWV. These suggest that the near-surface air in AR1-IWV was excessively warm and moist. In fact, calculations from YOTC analysis show that the mean temperature of the 2-m air was 0.61 °C higher than that of the sea surface in the precipitating AR1 ridge. Moreover, the differences in actual vapor pressure between the 2-m air and the sea surface were smaller in the precipitating AR1 ridge than those in the outside domains (not shown). Presumably, the downward $[F_S]$ and weak upwelling $[F_{LH}]$ in AR1-IWV are related to the lack of strong and persistent convection. Therefore, the warm air mass in AR1 was able to retain a large amount of moisture in propagation. In addition, the re-evaporation of precipitation under stratiform cloud decks likely served as an extra moisture source for the near-surface air (sections 2.3.2.3.1 and 2.3.2). In brief, precipitating systems in ARs appeared to play a crucial role in modifying the heat and moisture budget as well as air-sea interactions in ARs.

### 2.3.4 CRF Implications

Table 2.3 shows the shortwave cloud forcing ($C_s = S_{clr} - S_{cld}$, where $S_{clr}$ is the clear-sky reflected shortwave radiation, and $S_{cld}$ the cloudy-sky reflected shortwave radiation), longwave cloud forcing ($C_l = OLR_{clr} - OLR_{cld}$, where $OLR$ is the outgoing longwave radiation), and cloud radiative forcing ($CRF = C_s + C_l$) averaged over
AR1 and AR2 periods (see Ramanathan et al., 1989). Notice that 1800 UTC 20 January is omitted, so that both ARs have 10 time steps for the local morning and 9 time steps for the local evening.

The spatio-temporal average CRF or [CRF], is negative over all domains, an indication of the net cloud radiative cooling effect on the surface-atmosphere system. Moreover, AR1 had less [CRF] (-18.3 W m$^{-1}$) in AR-IWV than AR2 (-32.9 W m$^{-1}$). Such a contrast was amplified in the precipitating AR ridge. This is mainly because clouds in AR2-IWV had a stronger shortwave reflection at the top of atmosphere (more negative [C$_s$]) than those in AR1-IWV. Also, clouds in the AR2 ridge exhibited slightly weaker longwave forcing ([S$_{clr}$]) than those in the AR1 ridge. As discussed in section 2.3.1, the AR2 ridge had relatively more strong convection and a lower tropopause, with a strong heating center being confined below 400 hPa (Figs. 2.10a, b). Hence, clouds in AR2 were highly reflective. They effectively blocked the shortwave radiation from arriving the surface (not shown), and emitted more [OLR] (215.6 W m$^{-2}$) than in AR1. On the contrary, the AR1 ridge had less coverage of strong convection with a higher tropopause. Thus, convection in AR1 was able to develop to higher altitudes (e.g., the west end heating center with $Q_1 \geq 30$ K day$^{-1}$ reached upward to approximately 350 hPa in Fig. 2.9b, c), and likely retained more longwave radiation to the atmosphere (with [OLR]=193.3 W m$^{-2}$).

2.4 Conclusions

The oceanic precipitating systems of two strong landfalling ARs were studied. AR1 impacted British Columbia, Washington, and Oregon states ($40^\circ - 55^\circ$ N) during 04 January – 08 January 2009. AR2 influenced Alaska, U.S. and British Columbia in Canada ($55^\circ - 70^\circ$ N) during 16 January – 20 January 2009. The ECMWF YOTC data, CMORPH precipitation, and GPCP One-Degree Daily precipitation were used to construct the three-dimensional kinematic and thermodynamic fields. AR1 was between a high-pressure center at $32^\circ$ N, $140^\circ$ W and a low pressure system extending
from 45° N, 158° E to 63° N, 115° W, with an almost west-east orientation. AR2 was between an 850-hPa high pressure region centered at 45° N, 120° W and an 850-hPa low pressure region centered at 52° N, 175° W, with a south-north orientation downstream of a 200-hPa split jet.

Before landfall, merging sequential IWV bands over the ocean formed both ARs. This process increased the temperature and moisture content, as well as the actual and saturated vapor pressure of the near-surface air in the ARs. Meanwhile, precipitating convection was amplified. It consumed moisture as well as released latent heat in the path of AR propagation. In both ARs, precipitating convection often revealed vertically tilted $Q_1$ and $Q_2$ structures along a horizontal distance around 2000 km. This indicated a spatial transition from a predominant cloud population of shallower convection, to deep convection, to the trailing stratiform deck along the ARs.

However, clear differences existed between the two ARs. First, AR1 traversed a longer distance than AR2 from its tropical moisture reservoir to its midlatitude landfalling location. The precipitating systems in AR1 were mainly distributed on the southwest and northeast sides of the AR. They often revealed significant stratiform types of low-tropospheric level cooling and moistening characteristics of mature convective organizations. Whereas precipitating systems in AR2 continuously covered along the main path of the AR, most of the time exhibiting predominantly a deep convection type of heating throughout the troposphere. Second, AR1 had a higher tropopause (around 200 hPa in the AR1 ridge) than AR2 (around 300 hPa in the AR2 ridge). The primary $Q_1$ extrema in AR1 peaked at higher altitudes with taller heating centers than those in AR2.

In association with these distinctions, the NE Pacific region experienced a heat source and a moisture sink over the AR2 period, but underwent a heat sink and moisture sink despite AR1. It remained, on average, a heat sink and moisture sink in January 2009. This implies that although strong radiative cooling typically dominates the heat budget of the NE Pacific region in January 2009, it could be largely offset by latent heating through strong oceanic precipitation, as seen in AR2. In addition,
AR1-IWV had a weaker upward surface latent heat flux and a downward sensible heat flux, as compared to AR2-IWV. The contrast amplified in the AR ridges. This might be related to a relatively lack of strong and persistent convection in the excessive warm and moist AR1. Furthermore, dissimilarities in precipitation coverage and the dominant convection types between the two cases also translated to differences in CRF. In AR1-IWV, shortwave cloud forcing $C_s$ and longwave cloud forcing $C_l$ were comparable, resulting in small net CRF. In AR2-IWV, excessive shortwave reflection resulted in a more negative net CRF than AR1.

This study emphasized the roles of oceanic convection embedded in ARs. The cases studied showed that the convection not only impacted the moisture transport of ARs, but also modified the heat balance in the midlatitudes through latent heat release, convective heat transport, CRF, and air-sea interactions. Effects of these processes can be used to develop and validate physical parameterizations and simulations of tropical-extratropical interactions in weather and climate models. Therefore, as the next step, the statistical significance of the current results must be established and expanded in a climatological study.
Table 2.1: \([< Q_1 >]\) and \([< Q_2 >]\) in W m\(^{-2}\) averaged over NE Pacific region, AR-IWV, outside AR-IWV, and the precipitating AR ridge for January, AR1 and AR2 periods.

<table>
<thead>
<tr>
<th></th>
<th>January</th>
<th>AR1</th>
<th>AR2</th>
</tr>
</thead>
<tbody>
<tr>
<td>NE Pacific</td>
<td>-11.9</td>
<td>-29.1</td>
<td>18.0</td>
</tr>
<tr>
<td>AR-IWV</td>
<td>148.7</td>
<td>41.9</td>
<td>201.7</td>
</tr>
<tr>
<td>([&lt; Q_1 &gt;]) Outside AR-IWV</td>
<td>-79.8</td>
<td>-86.9</td>
<td>-47.9</td>
</tr>
<tr>
<td>Prec AR ridge</td>
<td>624.7</td>
<td>709.9</td>
<td></td>
</tr>
<tr>
<td>NE Pacific</td>
<td>12.3</td>
<td>27.3</td>
<td>14.0</td>
</tr>
<tr>
<td>AR-IWV</td>
<td>177.2</td>
<td>121.3</td>
<td>236.9</td>
</tr>
<tr>
<td>([&lt; Q_2 &gt;]) Outside AR-IWV</td>
<td>-59.2</td>
<td>-52.5</td>
<td>-65.4</td>
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<tr>
<td>Prec AR ridge</td>
<td>724.7</td>
<td>770.9</td>
<td></td>
</tr>
</tbody>
</table>
Table 2.2.: Similar to Table 2.1, but for $[< Q_R >]$, $[F_{LH}]$, $[F_S]$, and 

$([< Q_1 >] - [< Q_2 >]) - ([< Q_R >] + [F_{LH}] + [F_S])$ in W m$^{-2}$.

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<th></th>
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<th>AR2</th>
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</thead>
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<td>AR-IWV</td>
<td>−132.8</td>
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<td>$[&lt; Q_R &gt;]$</td>
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<td>Prec AR ridge</td>
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<td>NE Pacific</td>
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<td>AR-IWV</td>
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<td>$[F_{LH}]$</td>
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<td>109.8</td>
</tr>
<tr>
<td>Prec AR ridge</td>
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<td></td>
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<td>NE Pacific</td>
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<td>AR-IWV</td>
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<tr>
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<td>37.6</td>
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<tr>
<td>Prec AR ridge</td>
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<td>NE Pacific</td>
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<td>−18.0</td>
<td>−23.6</td>
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<tr>
<td>AR-IWV</td>
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<td>−8.4</td>
<td>−16.6</td>
</tr>
<tr>
<td>$([&lt; Q_1 &gt;] - [&lt; Q_2 &gt;]) - ([&lt; Q_R &gt;] + [F_{LH}] + [F_S])$</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Outside AR-IWV</td>
<td>−19.5</td>
<td>−25.4</td>
<td>−0.8</td>
</tr>
<tr>
<td>Prec AR ridge</td>
<td>44.8</td>
<td>21.7</td>
<td></td>
</tr>
</tbody>
</table>
Table 2.3: Similar to Table 2.1, but for $[C_s]$, $[C_l]$, [OLR], and [CRF] in W m$^{-2}$.  

<table>
<thead>
<tr>
<th></th>
<th>AR1</th>
<th></th>
<th>AR2</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>NE Pacific</td>
<td>AR-IWV</td>
<td>Outside AR-IWV</td>
<td>Prec AR ridge</td>
</tr>
<tr>
<td>$C_s$</td>
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<td>-51.4</td>
<td>-33.5</td>
<td>-67.3</td>
</tr>
<tr>
<td>$C_l$</td>
<td>25.7</td>
<td>33.0</td>
<td>19.5</td>
<td>63.8</td>
</tr>
<tr>
<td>OLR</td>
<td>238.2</td>
<td>239.9</td>
<td>237.4</td>
<td>192.7</td>
</tr>
<tr>
<td>CRF</td>
<td>-16.0</td>
<td>-18.3</td>
<td>-14.0</td>
<td>-3.5</td>
</tr>
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</table>
Figure 2.1.: Time evolution of AR1 from 0000 UTC 04 January to 1200 UTC 08 January 2009: CMORPH precipitation in mm hr$^{-1}$ is shaded. The AR1 ridge in thick contour; IWV = 20 mm in thin contour. The figure is taken from Luo and Tung (2015).

Figure 2.2.: As in Fig. 2.1, but for AR2 from 0600 UTC 16 January to 1800 UTC 20 January 2009. The figure is taken from Luo and Tung (2015).
Figure 2.3.: Mean states for January 2009: (a) – (c) are for 200-, 500-, and 850-hPa wind vectors (m s$^{-1}$) and height contours with intervals of 20, 10, and 5 decametres, respectively. (d) shows 850-hPa temperature (contours intervals of 5° C) and specific humidity (shading intervals of 2 g kg$^{-1}$). The figure is taken from Luo and Tung (2015).
Figure 2.4.: Same as Fig. 2.3, but for AR1 (left) and AR2 (right). The figure is taken from Luo and Tung (2015).
Figure 2.5.: Four-day accumulated GPCP daily precipitation (shading intervals of 20 mm, white contour denotes 40 mm): (a) AR1 from 04 – 08 January 2009; (b) AR2 from 16 – 20 January 2009. The figure is taken from Luo and Tung (2015).
Figure 2.6.: Merging process for AR1 from 0600 UTC 02 January to 0600 UTC 06 January 2009: (a) in a plan view: CMORPH precipitation in mm hr$^{-1}$ is shaded. IWV = 20 mm in thick contour. 30° N and 35° N are marked by horizontal lines; (b) in a vertical section averaged over 30° – 35° N for $u$, $\omega$ vectors ($u$ in m s$^{-1}$ and $\omega$ in 10$^{-2}$ Pa s$^{-1}$), specific humidity with contour intervals of 2.5 g kg$^{-1}$, and positive $Q_1$ (shades, K day$^{-1}$). Black and gray arrows denote the primary and secondary specific humidity peaks, respectively. The figure is taken from Luo and Tung (2015).
Figure 2.7.: As in Fig. 2.6, but for AR2 from 0000 UTC 15 January to 0000 UTC 19 January 2009. The figure is taken from Luo and Tung (2015).
Figure 2.8.: Differences in saturated vapor pressure of 2-m air between AR period and no-AR periods (January 2009 excluding time steps during AR1 and AR2 periods) are shaded, with 2 hPa intervals: (a) AR1, and (b) AR2. White contours are four-day accumulated GPCP precipitation = 30 mm. Black contours are temporal averaged 850-hPa specific humidity = 4 g kg$^{-1}$. The figure is taken from Luo and Tung (2015).
Figure 2.9.: Vertical sections along the AR1 ridge for $Q_1$ (K day$^{-1}$, left), $Q_1 - Q_2$ (K day$^{-1}$, middle), and CMORPH precipitation (mm hr$^{-1}$, right). $Q_1$ and $Q_1 - Q_2$ are in contours with 15 K day$^{-1}$ intervals, with negative values being shaded. Horizontal dash lines mark temperature at 0°C. Top x-axis marks the coordinates where the AR1 ridge intersected 165° W, 153° W, and 141° W. Bottom x-axis is distance in km. The y-axes for $Q_1$ and $Q_1 - Q_2$ are pressure in hPa. Plotted from 0000 UTC 04 January to 1800 UTC 07 January 2009. The figure is taken from Luo and Tung (2015).
Figure 2.10.: Same as Fig. 2.9, but for the AR2 ridge. Three vertical lines mark where the AR2 ridge intersected 25° N, 35° N, and 45° N. Plotted from 1200 UTC 17 January to 0000 UTC 19 January 2009. The figure is taken from Luo and Tung (2015).
Figure 2.11.: Vertical profiles averaged over the precipitating AR ridges: lines are $Q_1$; dash lines are $-Q_2$. Line 1 – 3 in (a, c) are the strengthening time steps in AR1/AR2, line 4 – 6 in(b, d) the weakening phase of AR1/AR2. X-axis is the amplitude of $Q_1$ and $-Q_2$ in K day$^{-1}$, y-axis the pressure in hPa. The figure is taken from Luo and Tung (2015).

Figure 2.12.: Average vertical profiles during (a) AR1 and (b) AR2 periods. Dark lines (light lines with circles) are $Q_1$ inside (outside) AR-IWV, denoted as $Q_{1i}$ ($Q_{1o}$). Dark (light) dash lines are $Q_1 - Q_2$ inside (outside) AR-IWV, denoted as ($Q_1 - Q_2)_i$ (($Q_1 - Q_2)_o$). Vertical lines mark the 0 values. X-axis in K day$^{-1}$, y-axis the pressure in hPa. The figure is taken from Luo and Tung (2015).
Figure 2.13.: Spatio-temporal average \( <Q_1> \) and \(-<Q_2>\) for NE Pacific region in W m\(^{-2}\). The figure is taken from Luo and Tung (2015).
Figure 2.14.: Spatio-temporal average upward $F_S$, upward $F_{LH}$, and $<Q_R>$ for NE Pacific region (NE), AR-IWV (IWV), and the precipitating AR ridge (ARR) in W m$^{-2}$ for AR1 and AR2. The figure is taken from Luo and Tung (2015).
Chapter 2 and Luo and Tung (2015) pointed out that the ARs could modify the regional energy budget through pathways such as CRF in a weather-case study. However, the climatological impact of ARs through CRF remains unclear. This chapter is adapted from Luo and Tung (tted). We discuss the sufficient conditions for temporally persistent and spatially expansive CRF induced by ARs over the continental U.S. To explore this question, we compared and contrasted 60 ARs that made landfall in the Pacific Northwest (hereafter the AR$_{NW}$) with 60 ARs that influenced the California (hereafter the AR$_{SW}$) during Nov – Mar, 2000 – 2008 from various satellite and gridded-analysis observations. To discuss the relationship between the West-Coast ARs and Gulf-Coast ARs, we constructed the index for AR$_{GULF}$ that followed the arrivals of the West-Coast ARs. Several statistical methods such as singular value decomposition (SVD) were deployed to extract the spatio-temporal characteristics of different AR scenarios. Additionally, the apparent heat source ($Q_1$) and apparent moisture sink ($Q_2$) were computed and examined to explain the physical processes.

The chapter is organized as follows: data and methods are described in section 3.1. A background overview for the two West-Coast AR scenarios is provided in section 3.2. Section 3.3 compares the convection in the landfalling regions between AR$_{NW}$ and AR$_{SW}$. Section 3.4 quantifies the AR impacts on the Western U.S. and the Eastern U.S. Section 3.5 shows the dominant spatio-temporal characteristics associated with different AR scenarios. Section 3.6 discusses the background conditions for AR$_{GULF}$ occurrences. Conclusions are in section 3.7.
3.1 Data & Methods

3.1.1 ECMWF Interim

The European Centre for Medium-Range Weather Forecasts (ECMWF) Interim Reanalysis (Dee et al., 2011) is a global reanalysis product available from Jan 1, 1979 to present. It is produced by a T255 60-level atmospheric model and four-dimensional variational assimilation system. We retrieved the version of the ERA-I dataset from the National Center for Atmospheric Research Data Support Section, which is defined on a regular 512×256 N128 Gaussian grid at 37 standard pressure levels from 1000 to 1 hPa and at the surface. The data are available every 6 hours. To facilitate the calculation and analysis, we downsampled the data to daily, 1.5°×1.5° horizontal grids.

3.1.2 CERES SYN1deg

Clouds and the Earth’s Radiant Energy System (CERES) SYN1deg provides global estimates of the radiative fluxes and cloud properties measured by NASA satellites. It is available from March 1, 2000 to the delayed present, with 1°×1° spatial resolutions (Wielicki et al., 1996). We used the observed top-of-atmosphere fluxes under all-sky and clear-sky conditions in daily resolution, which are extensively validated and are highly stable, to compute the shortwave CRF (SWCRF) and the longwave CRF (LWCRF). Following Ramanathan et al. (1989), \( \text{SWCRF} = \text{SW}_{clr} - \text{SW}_{cld} \) and \( \text{LWCRF} = \text{OLR}_{clr} - \text{OLR}_{cld} \), where \( \text{SW}_{clr} \) is the clear-sky reflected shortwave radiation, and \( \text{SW}_{cld} \) is the cloudy-sky reflected shortwave radiation. \( \text{OLR} \) is the outgoing longwave radiation. We also used the daily Cloud Ice Water Path (IWP), Cloud Liquid Water Path (LWP), and Integrated Water Vapor (IWV).
3.1.3 GPCP daily precipitation

The Global Precipitation Climatology Project (GPCP) One-Degree Daily Precipitation Version 1.2 dataset (Huffman et al., 2001) provides daily global precipitation estimates at 1°×1° horizontal grids from October 1996 to the delayed present.

3.1.4 AR indices

ARs can be identified using either IWV or Integrated Water Vapor Transport (IVT, see mini review by Gimeno et al. (2014)). We used indices for ARs affecting the West Coast and the Gulf Coast. For the former, we used the one that was documented by Dettinger et al. (2011b). There were 60 AR\textsubscript{NW} reaching between 32.5° and 41°N, and 60 AR\textsubscript{SW} reaching between 41° and 52.5°N, respectively, during Nov – Mar, 2000 – 2008. For the latter, we only considered those that made landfall within 3 days upon the arrival of the West-Coast ARs. To do so, we followed the method in Lavers and Villarini (2013), but with modifications to match our data resolutions. First, we calculated the $IVT = \frac{1}{g} \int^{300}_P \rho v dp$, where $g$ is the acceleration due to gravity, $P_{sfc}$ is the surface pressure, $q$ is the water vapor mixing ratio; $v$ the horizontal velocity (Neiman et al., 2008a). Then, we extracted the maximum daily IVT at 40.5°N, 100.5° – 85.5°W for the entire datasets. We used the 85th percentile of the extracted values to be the threshold value of ARs, which is 304.92 kg m\textsuperscript{-1} s\textsuperscript{-1}. We retained the plumes with daily IVT ≥ 304.92 kg m\textsuperscript{-1} s\textsuperscript{-1} that spanned from 34.5° – 40.5°N between 100.5° and 85.5°W and were longer than 1000 km in a north-south direction.

3.1.5 $Q_1$ and $Q_2$ calculations

The apparent heat source ($Q_1$) and the apparent moisture sink ($Q_2$) (Yanai et al., 1973) were computed and interpreted as follows (Luo and Tung, 2015):
$$Q_1 \equiv c_p \left( \frac{p}{p_0} \right)^\kappa \left( \frac{\partial \bar{\theta}}{\partial t} + \nabla \cdot \nabla \bar{\theta} + \bar{\omega} \frac{\partial \bar{\theta}}{\partial p} \right)$$

$$\approx Q_R + L_v (\bar{e} - \bar{e} + \bar{d}) + L_f (\bar{d} - \bar{m}) - \frac{\partial s' \omega'}{\partial p}, \text{ and}$$

$$Q_2 \equiv -L_v \left( \frac{\partial \bar{q}}{\partial t} + \nabla \cdot \nabla \bar{q} + \bar{\omega} \frac{\partial \bar{q}}{\partial p} \right)$$

$$\approx L_v (\bar{e} - \bar{v} + \bar{d} + \frac{\partial q' \omega'}{\partial p}),$$

here, $\theta$ denotes the potential temperature, $\omega$ the vertical $p$-velocity, $p_0 = 1000$ hPa, $\kappa = R/c_p$ with $R$ the gas constant of dry air, $c_p$ the specific heat capacity at constant pressure, $\nabla$ the isobaric gradient operator, $Q_R$ the radiative heating rate, $L_v$ and $L_f$ the latent heat of vaporization and fusion, $c, e, d$ and $m$ the rates of condensation, evaporation, deposition, and melting per unit mass of air, $s$ the dry static energy per unit mass of air. The overbar is the mean over the horizontal grids. Prime is the deviation from the mean, thus referring to subgrid-scale processes such as cloud convection, boundary layer fluxes, and turbulence. The derivations of (3.1) and (3.2) assume the Reynolds conditions and their consequences are accurate.

Equations (3.1) and (3.2) were calculated using the rhs of their first lines from ERA-interim. Under convective conditions, they can be approximately interpreted as the linear combination of the rhs terms in the second lines. $Q_1$ shows the total effects of radiative heating, approximate the latent heat released during microphysical phase changes (Luo and Tung, 2015; Johnson et al., 2016b), and the convergence of fluxes of sensible heat caused by subgrid-scale eddies such as convection and turbulence. $Q_2$ represents the total effects of net condensation and divergence of eddy moisture flux owing to clouds and turbulence.

### 3.1.6 Percentile plots

To quantify the impacts of ARs on the U.S. continents and examine the duration of such impacts, we plotted the time evolutions of various parameters that were spatially
averaged over land for the Western U.S. (27.5° – 55.5°N, 130.5° – 106.5°W, region A in Fig. 3.4) and for the Eastern U.S. (27.5° – 55.5°N, 98.5° – 74.5°W, region B in Fig. 3.4). Specifically, we obtained 60 samples each for time steps of one-day before (Day–1), on the day of (Day+0), one-day after (Day+1), two-day after (Day+2), and three-day after (Day+3) the arrivals of the West-Coast AR, respectively. In each spatial domain, we computed the median and interquartile range for the 60 samples at each time step. Because of the skewness of the data, we used the one-tailed Wilcoxon Rank-Sum test to determine if each median of the AR\textsubscript{SW} is larger or smaller than that of the AR\textsubscript{NW} at the 5% significance level.

3.1.7 SVD

To capture the essential spatial distribution of CRF variability, we performed the singular value decomposition (SVD) on the observed SWCRF and LWCRF anomalies over the continental U.S. (27.5° – 55.5° N, 130.5° – 74.5° W) on Day+0 and Day+3:

$$X = U \Lambda V^T,$$

(3.3)

where $X$ is an $M \times N$ data matrix, with $M$ being the space dimension and $N$ being the time dimension. $X$ was calculated by subtracting the mean of Nov – Mar, 2000 – 2008 (Fig. 3.10) from the original data, and was then weighted by the cosine of the corresponding latitudes. We obtained $U$ as the Empirical Orthogonal Functions (EOFs), which is the spatial modes. The term $\Lambda V^T$ is the time-dependent principle components (PCs) for 60 cases, where $\Lambda$ is the matrix with the square root of the eigenvalues ($\lambda$) of both $XX^T$ and $X^TX$ along its diagonal. The values of $\lambda$ are therefore the variance each EOF is associated to. We reconstructed variables using the first through $k^{th}$ EOFs, where $k$ is the smallest integer satisfying

$$\frac{\sum_{i=1}^{k} \lambda_i}{\sum_{i=1}^{M} \lambda_i} \geq 50\%$$

(3.4)
We also included the \((k+1)^{th}\) EOF if it is degenerate with the \(k^{th}\) EOF (North et al., 1982). In order to identify the time steps with significant variability, we viewed \(PC_t\) as a \(k\)-dimensional vector at each time step \(t\), where \(t = 1, 2, \ldots, N\). We computed the norm \(\|PC_t\|\), and averaged the significant time steps when \(t\) satisfies
\[
\|PC_t\| \geq \frac{1}{N} \sum_{i=1}^{N} \|PC_i\|
\]  
(3.5)

Table 3.1 shows the cut-off \(k\) (first column), the fraction of total variance explained by the reconstruction (second column), and the number of significant time steps (third column).

3.2 Background overviews for AR\(_{SW}\) and AR\(_{NW}\)

Figure 3.1 illustrates the time mean of moisture transport (Fig. 3.1a) and cloud coverage (Fig. 3.1b) from Nov – Mar, 2000 – 2008. The Pacific Northwest between 39° and 50°N was moist (Fig. 3.1a) and cloudy (Fig. 3.1b) due to the apparent ocean-to-land moisture transport from the Northeastern Pacific Ocean. In contrast, it was dry (Fig. 3.1a) and mostly clear to the south of this region (Fig. 3.1b) due to the presence of subtropical high (not shown). The Central-Eastern half of the U.S. was more moist than the Western half, with IVT \(\geq 140\) kg m\(^{-1}\) s\(^{-1}\) extending northeastward from the Gulf of Mexico to the Great Lakes (Fig. 3.1a). Correspondingly, there was a narrow band of abundant mixed-phase clouds stretching northeastward from Texas at 25°N, 100°W to the north of the Great Lakes at 45°N, 70°W (Fig. 3.1b). In the tropics, a large amount of moisture was transported westward by the Caribbean Low-Level Jet (LLJ) from the Caribbean Sea to the Tropical Eastern Pacific (Fig. 3.1a).

The AR\(_{NW}\) and AR\(_{SW}\) were associated with different synoptic setup, moisture transport and cloud coverage (Figs. 3.2, 3.3). On Day -1 (Figs. 3.2 and 3.3a), the AR\(_{NW}\) was sandwiched between a strong subtropical high centered at 40°N, 110°W and an extratropical cyclone (hereafter the AR\(_{NW}\)-storm) centered at 50°N, 145°W. On the landfalling date of the AR\(_{NW}\) (Fig. 3.3b), the central pressure anomalies of the subtropical high amplified at 40°N, 110°W, whereas the AR\(_{NW}\)-storm drifted
northeastward to 56°N, 140°W and weakened. This drew a huge amount of moisture from the Northeastern Pacific Ocean to the Pacific Northwest between 40° and 55°N (Figs. 3.2b). An anomalously dry northeasterly flow was formed over 25° – 40°N, 110° – 90°W, on the downstream side of the ARNW. After the landfalling date, part of the ARNW-storm propagated inland (Fig. 3.3c) and quickly dissipated (Fig. 3.3d). The other part drifted away from the West Coast (Figs. 3.3c,d). Meanwhile, the subtropical high weakened and retreated southward (Figs. 3.3c,d). Accordingly, the ARNW was weakened from Day+0 and onward (Figs. 3.3c,d). The anomalously dry northeasterly flow was strengthened on Day+1 (Fig. 3.2c) and was weakened afterward (Fig. 3.2d).

In contrast, ARSW was associated with a deep extratropical cyclone centered at 47°N, 135°W (Figs. 3.3e). ARSW curved cyclonically, transporting less moisture to the West Coast from Day−1 to Day+1 (Figs. 3.2e−g). Meanwhile, there was a dry northeasterly transport from 40°N to 20°N in the downstream regions. From Day+1 onward (Figs. 3.3g,h), part of the ARSW-storm penetrated inland to the Central-Eastern U.S. and strengthened. Together with an apparent Bermuda High centered near 40°N, 70°W (Fig. 3.2h), it led to an anomalously moist southwesterly flow that spanned northeastward from 35°N, 100°W to 55°N, 70°W (Fig. 3.2h). At the same time, the other half of the ARSW-storm remained offshore the Pacific Northwest coast (Figs. 3.3g,h). It drew moisture from the Northeastern Pacific Ocean to Sierra Nevada (Figs. 3.2g,h).

The ARNW induced abundant mixed-phase clouds in Northeastern Pacific and in the landfalling regions on Day−1 and Day+0 (Figs. 3.3a,b), yet these clouds quickly disappeared after the landfall (Figs. 3.3c,d). In addition, the Chesapeake Bay and Quebec in Canada experienced increased cloudiness on Day−1 and Day+0 (Figs. 3.3a,b). On the contrary, even though ARSW produced less mixed-phase clouds over the ocean on Day−1 and Day+0 (not shown), it tended to lead to temporally prolonged and spatially extensive ice-cloud coverage in the Western U.S. (Figs. 3.3g,h). Moreover, enhanced mixed-phase clouds existed in the Central-Eastern U.S. three
days later, as indicated here with the ice water path in Fig. 3.3h. These suggest that the AR\textsubscript{SW} were associated with more extensive coverage of cloudiness in the U.S. We therefore hypothesized that there was additional moisture transported from the Gulf Coast to the Central-Eastern U.S., namely, the AR\textsubscript{GULF}, from the landfalling date of West-Coast AR onward.

3.3 The oceanic and continental clouds in AR\textsubscript{SW} and AR\textsubscript{NW}

Section 3.2 showed that AR\textsubscript{SW} was associated with longer-lasting IWP anomalies along the Pacific Coast than AR\textsubscript{NW}. This implies that the two scenarios might exhibit different convective heating or drying features along the West Coast. To quantify how such difference changed over time, here we compare and contrast the time evolvement of IWP along 124.5°W from 30° – 60°N (see line C–D in Fig. 3.4), as well as $Q_1$ and $-Q_2$ averaged over the landfalling regions (see region E in Fig. 3.4) between the two scenarios. Note that from Fig. 3.4, line C–D intersects the ocean between 30° and 47.5°N, and intersects land between 47.5° and 60°N. Fig. 3.5 is the IWP along line C–D, with “SW” and “NW” denoting the landfalling latitudes of the AR\textsubscript{SW} and the AR\textsubscript{NW}, respectively. For AR\textsubscript{NW} (Fig. 3.5a) on Day −1, the primary IWP peak was 345 g m$^{-2}$ over lands at 48°N. As the subtropical high shifted southward (Fig. 3.3b) and AR\textsubscript{NW} made landfall (Fig. 3.2b) on Day +0, the primary IWP peak increased to 360 g m$^{-2}$, with a secondary peak of 275 g m$^{-2}$ over the ocean at 42.5°N. However, the ice clouds in the West Coast quickly dissipated and the major IWP peak migrated northward to 50.5°N from Day +1 onward. In contrast, AR\textsubscript{SW} induced a broad cloud coverage along the West Coast. Figure 3.5b shows the IWP difference between AR\textsubscript{SW} and AR\textsubscript{NW}. The LWP difference between the two was qualitatively similar to that of the IWP. On Day −1, AR\textsubscript{SW} had more mixed-phase clouds offshore the California Coast near 41°N and fewer mixed-phase clouds to the north of 46°N than the AR\textsubscript{NW}. As the AR\textsubscript{SW} made landfall on Day +0 (Fig. 3.2), the north-south contrast in IWP difference between AR\textsubscript{SW} and AR\textsubscript{NW} was reduced (Fig. 3.5b). Nevertheless, a part
of the AR_SW-storm tended to remain offshore the Pacific Northwest (Figs. 3.3g,h) while the remnant of AR_NW-storm tended to retreat westward after landfall, there were more clouds associated with AR_SW on the West Coast on Day+2 and Day+3 (Fig. 3.5b).

The $Q_1$ and $-Q_2$ averaged over region E in Fig. 3.4 are shown in Fig. 3.6. Clearly, the convection associated with the AR_NW were strongest on Day−1. Yet they quickly decayed as time progressed (Fig. 3.6a). In contrast, AR_SW was associated with more persistent heating and drying along the West Coast, particularly from Day+0 onward (Fig. 3.6b). The relationships between latitudes and variables such as IWP, LWP, $Q_1$ and $Q_2$ were further analyzed (not shown), confirming the finding here that the convection associated with AR_NW were concentrated on the Northwest Coast (e.g., as suggested by Fig. 3.5), whereas those associated with AR_SW were spatially more expansive from Day−1 to Day+3.

### 3.4 The regional impacts of ARs on the U.S.

Section 3.3 examined the cloud type and convective heating or drying structures along the West Coast under two different AR scenarios. AR_SW was found to be associated with more ice clouds and more persistent heating and drying than AR_NW. Here, we further quantify the regional hydrological and cloud-radiative forcing impacts of the ARs. Figs. 3.7–3.8 present the time progression of the median and interquartile range for IWV, IWP, SWCRF and LWCRF averaged over the Western U.S. (land region A in Fig. 3.4) and the Eastern U.S. (land region B in Fig. 3.4), respectively, following the methods in see section 3.1.6.

From Day−1 onward, the AR_SW transported less moisture to the Western U.S. when compared with the AR_NW (Fig. 3.7a). However, the former produced significantly more ice clouds (Fig. 3.7b), stronger SWCRF (Fig. 3.7c) and LWCRF (Fig. 3.7d) than AR_NW. In the Eastern U.S. (Fig. 3.8), the AR_SW scenario was associated with less IWV (Fig. 3.8a), precipitation (not shown), IWP (Fig. 3.8b)
and LWP (not shown), and weaker SWCRF and LWCRF (Fig. 3.8d) on Day−1 and Day+0 than the ARNW. Presumably, this contrast was induced by the presence of the subtropical high in the Central-Eastern U.S. in the ARSW scenario (Figs. 3.3e,f). On Day+2 and Day+3, the opposite was true (Fig. 3.8), likely due to the strong low in the Central-Eastern U.S. in the ARSW scenario (Fig. 3.3). Interestingly, in both AR scenarios from Day−1 onward, the Western U.S. experienced NetCRF warming (Fig. 3.9a), whereas the Eastern U.S. experienced NetCRF cooling (Fig. 3.9b). One possible reason is clouds tended to form lower in the troposphere in the Eastern U.S. than in the Western U.S. This is implied by the primarily $Q_1$ and $-Q_2$ peaked in a lower altitude in the Eastern U.S. (not shown) than in the Western U.S. (e.g., Fig. 3.6). Correspondingly, a greater percentage of clouds formed in the Eastern U.S. was in the liquid phase (Fig. 3.9d), which exerted an obvious shortwave CRF. Nevertheless, the cloud hydrometeor phase alone did not explain the entire story. Specifically in the Western U.S., the ARSW scenario was associated with more LWP (not shown), more IWP (Figs. 3.7b), yet less LWP percentage (Fig. 3.9c) than the ARNW scenario. However, more NetCRF cooling was observed in the former scenario (Fig. 3.9a). These highlight the importance of considering the location and spatial coverage of the clouds as well as the cloud hydrometeor phase when examining the regional CRF impacts.

3.5 The dominant spatio-temporal variability of CRF

To reveal the mechanisms for the CRF differences between the two scenarios, we analyze the spatial distribution of the LWCRF and SWCRF climatology (Fig. 3.10), as well as the dominant CRF variability (Figs. 3.11–3.14). In the Western U.S., LWCRF warming maximized in the Pacific Northwest (larger than 30 W m$^{-2}$, Fig. 3.10a). The local maximum of SWCRF cooling was situated in the lower latitudes over the Cascades (greater than $-30$ W m$^{-2}$, Fig. 3.10b). For the eastern half of the country, obvious SWCRF was distributed in an elongated channel that spanned northeastward
from the Gulf Plain to the Great Lakes (−40 W m$^{-2}$ and more, Fig. 3.10b). It outweighed LWCRF by −15 W m$^{-2}$ and more, resulting in NetCRF cooling.

Figures 3.11a–d present the first two EOFs and PCs for LWCRF on Day+0 of the AR$_{NW}$. EOFs 1 and 2 (Figs. 3.11a,c) explain the comparable fraction of the total variance. EOF1 and PC1 (Figs. 3.11a,b) show a strong positive LWCRF pattern, which extended inland from the Pacific Northwest to the Great Plains around 98°W, along with a weaker negative LWCRF over the southwestern U.S. EOF2 (Fig. 3.11c) captures a significant negative pattern to the east of the lower Mississippi River near 95°W and to the south of the Great Lakes near 44°N. Together with PC2 (Fig. 3.11d), it indicates that the Eastern U.S. often experienced weaker LWCRF than the climatology (Fig. 3.10a) when AR$_{NW}$ made landfall. Three days later (Figs. 3.11e–h), EOF1 (Fig. 3.11e) captures a local minimum stemming inland from the Panhandle of Florida to the Great Lakes. EOF2 (Fig. 3.11k) depicts a wide-spread structure of variability to the south of 45°N, with a local minimum in the Gulf of California.

Figures 3.12a–d are the first two EOFs and PCs for LWCRF on Day+0 of the AR$_{SW}$. EOF1 (Fig. 3.12a) explains around 28% of the variance. It shows a considerable variability in the landfalling regions to the west of the Rocky Mountain near 103°W. PC1 (Fig. 3.12b) is predominately positive, meaning that the LWCRF was typically stronger than the climatology in the Western U.S. (Fig. 3.10a). EOF2 (Fig. 3.12c) captures a local maximum emanating inland from the Gulf of California to around 42°N, 95°W. On Day+3 (Figs. 3.12e–h), EOF1 and PC1 (Figs. 3.12e,f) imply that enhanced LWCRF warming often manifested in an elongated pattern stretching northeastward from the lower Mississippi River basin to the Great Lakes. EOF2 (Fig. 3.12g) shows a large spatial variability spreading northeastward from the Gulf of California to the Great Plains around 52°N, 95°W. PC2 (Fig. 3.12h) is predominately positive. Isentropic analysis for individual AR$_{SW}$ cases confirms that moisture originated in the Tropical Eastern Pacific Ocean frequently transversed the Gulf of California and fueled the lower-level trough in the Western and Central U.S. (refer to chapter 4).
Figure 3.13 is the reconstructed LWCRF variability, based on methods in section 3.1.7. On Day+0 of AR$_{NW}$ (Fig. 3.13a), a zone of enhanced LWCRF warming extended eastward from the landfalling location around 45°N, 125°W to the Great Plains around 49°N, 102°W. In addition, the Atlantic Coast underwent LWCRF warming, with a local maximum in Georgia. On Day+3 (Fig. 3.13b), however, weak positive LWCRF variability was in the Great Plains and the Central Plains. The Gulf and Atlantic Plain underwent negatively anomalous LWCRF. When the AR$_{SW}$ made landfall (Fig. 3.13c), significantly enhanced LWCRF variability was to the west of the Rocky Mountains around 107°W. It maximized in the Sierra Nevada and the Intermountain regions. This pattern continued to exist on Day+3 (Fig. 3.13d), owing to the persistent low in the offshore of Pacific Northwest (Fig. 3.3h). Meanwhile, the Eastern U.S. underwent LWCRF warming. It was distributed in an elongated channel extending from the Lower Mississippi River near 31°N, 95°W to the Great Lakes. This pattern resembles the anomalous IVT pattern in the same region (Fig. 3.2h).

Figure 3.14 is the reconstructed SWCRF variability. Figure 3.14a displays a north-south contrasting SWCRF pattern on Day+0 of the AR$_{NW}$. While the landfalling sites experienced weak SWCRF cooling, the regions that were dominated by the subtropical high (to the south of 48°N, 122°W, Fig. 3.2b) experienced positively anomalous shortwave CRF with respect to climatology (Fig. 3.13b). On Day+3 (Fig. 3.14b), the Gulf Plain underwent a strong positively anomalous SWCRF. On Day+0 of AR$_{SW}$ (Figs. 3.14a), strong SWCRF cooling was located to the west of the Rocky Mountains around 107°W. It amplified in the Sierra Nevada. On Day+3 (Fig. 3.14d), negative SWCRF variability covered a large part of the U.S. It was most apparent in the Eastern U.S., spanning from Texas and Louisiana to the Great Lakes.

3.6 The AR$_{GULF}$ after the West-Coast ARs

The elongated structure of the strong SWCRF and LWCRF variability in the Eastern U.S. on Day+3 of AR$_{SW}$ was found to be related to the AR$_{GULF}$. Thus, here
we compare the frequency, magnitudes and background conditions of AR\textsubscript{GULF} in different West-Coast AR scenarios. As seen in Fig. 3.15, there were 35 AR\textsubscript{GULF} incidents in the AR\textsubscript{SW} scenario, much more than those in the AR\textsubscript{NW} scenario. Specifically, 24 out of 35 AR\textsubscript{GULF} incidents took place two and three days after AR\textsubscript{SW} made landfall.

Figure 3.16 shows the composite mean fields of 250-hPa geopotential height, 250-hPa wind speed, GPCP rain rate, 1000 – 500-hPa thickness, 850-hPa wind vectors, mean sea-level pressure, IVT, IWP, and LWP when the West-Coast ARs and the AR\textsubscript{GULF} occurred in sequence. On Day+0 in AR\textsubscript{NW} scenario, there was a strong 250-hPa ridge extending from 60°N, 110°W to 35°N, 125°W. A jet streak was situated on the upstream side of this ridge, with its core spanning from 45°N, 135°N to 50°N, 115°W (Fig. 3.16a). The AR\textsubscript{NW} was positioned between a deep zonally-stretched Aleutian Low centered near 60°N, 145°W and a subtropical high centered at 40°N, 110°W in the Western U.S. (Fig. 3.16b). The AR\textsubscript{NW} produced enhanced precipitation (not shown) and abundant mixed-phase clouds (Fig. 3.16c) in Pacific Northwest. When AR\textsubscript{GULF} arrived, there was a 250-hPa trough in 30° – 50°N, 105°W, and a ridge in 25° – 50°N, 80°W (Fig. 3.16d). A 250-hPa jet streak was exiting the ridge axis (Figs. 3.16d), with the AR\textsubscript{GULF} positioned in the right entrance region. The latter spanned northeastward from 35°N, 90°W to 45°N, 80°W (Figs. 3.16d,e). Precipitation was organized along the AR\textsubscript{GULF}. Extensive cloudiness indicated by LWP and IWP spanned northeastward from Texas, the Great Lakes, to Quebec (Fig. 3.16f). The Caribbean LLJ was active in the AR\textsubscript{NW} scenario. It transported a significant amount of moisture from the Caribbean Sea to the Tropical Eastern Pacific (Figs. 3.16b,c).

Unlike the AR\textsubscript{NW}, the AR\textsubscript{SW} was associated with a weaker-amplitude 250-hPa ridge when making landfall (Fig. 3.16g). The Aleutian Low was shallower. It was centered at 60°N, 145°W, stretching in a northwest-southeast direction over the ocean (Fig. 3.16h). The AR\textsubscript{SW} was weaker due to a weaker pressure gradient between the Aleutian Low and the Pacific High (Fig. 3.16h). A subtropical high was in the Eastern U.S., centered at 35°N, 80°W (Fig. 3.16h). When AR\textsubscript{GULF} struck the Central-Eastern U.S., a prominent 250-hPa trough was situated in the western half of the
country, with its axis at 110°W. An amplified 250-hPa ridge was in the Eastern U.S., with its axis at 80°W (Fig. 3.16j). An apparent 250-hPa jet was entering the ridge (Fig. 3.16j), producing upper-level divergence to its right (not shown). Compared with the pattern in Fig. 3.16e, the surface low centered at 45°N, 95°W was deeper, and the Bermuda High was stronger and stretched along the southwest-northeast direction, magnifying the pressure gradient and winds in between (Fig. 3.16k). The upper-level divergence and the strong pressure gradient amplified the ARGULF (Fig. 3.16k) and the associated precipitation (Fig. 3.16j). Ample mix-phase clouds were distributed along the pathway of the ARGULF and to the north of the Great Lakes (Fig. 3.16l).

3.7 Conclusions

We compared 60 ARs that made landfall in Southwestern U.S. (ARSW) with 60 ARs that made landfall in the Northwestern U.S. (ARNW) during Nov – Mar, 2000 – 2008. Their differences in moisture transport and CRF were addressed. The ARNW was stronger and associated with an amplified Pacific High in the Western U.S. It delivered plentiful moisture to and produced strong precipitation along the Pacific Northwest from Day−1 to Day+1 (Fig. 3.2). However, after part of the ARNW-storm propagated inland on Day+1, the western half drifted offshore (Fig. 3.3). The moist advection ceased (Fig. 3.2). In contrast, the ARSW transported less moisture to the landfalling regions on Day−1 and Day+0 (Fig. 3.2). Yet, it was associated with a persistent low offshore the Pacific Northwest (Fig. 3.3). The latter led to persistent moisture transport to the West Coast (Fig. 3.2), as well as extensive cloudiness and CRF in the Western U.S. (Figs. 3.5–3.7, 3.9, 3.13, 3.14).

Zero to three days after the arrivals of West-Coast ARs, a secondary AR ascending from the Gulf of Mexico (ARGULF) may occasionally penetrate deeply into the Central-Eastern U.S. There was more and stronger ARGULF in the ARSW scenario than in the ARNW scenario (Fig. 3.15). Such concurrence induced strong CRF variability in the eastern half of the country (Figs. 3.8, 3.9, 3.13, and 3.14).
Further analyses suggest that the synoptic setup in the AR\textsubscript{SW} scenario tended to trigger enhanced AR\textsubscript{GULF} (Fig. 3.16). The key findings were summarized in Fig. 3.17: On the landfalling dates of the AR\textsubscript{NW}, there was a high-amplitude 250-hPa jet streak and a subtropical high in the Western U.S., a Bermuda High in the East Coast offshore, and a strong Caribbean LLJ (CLLJ in Fig. 3.17). On the landfalling dates of the AR\textsubscript{GULF} that followed the AR\textsubscript{NW}, the subtropical high weakened, while the Bermuda High strengthened. This AR\textsubscript{GULF} was developed in the vicinity of an extratropical cyclone in the Central Plains. In contrast, on the arrival dates of the AR\textsubscript{SW}, the Pacific High was located offshore the West Coast with a more zonal jet stream. A subtropical high maximized in the Central-Eastern U.S. On the arrival dates of AR\textsubscript{GULF}, the extratropical cyclone in the Central Plains was deeper. The Bermuda High offshore the East Coast became stronger. These led to a stronger pressure gradient and hence the stronger AR\textsubscript{GULF}. In addition, the 250-hPa jet streak was entering the upper-level ridge, producing divergence over the Central-Eastern U.S. These acted to enhance the development of the precipitation and clouds, leading to abundant mix-phase clouds along the pathway of the AR\textsubscript{GULF} and to the north of the Great Lakes.

In summary, we looked for the often happened scenario that was sufficient for extensive CRF of ARs over the U.S. We found that the synergy between the West-Coast ARs and Gulf-Coast ARs led to enhanced CRF over the continental U.S. Such synergy was observed more frequently associated with the ARs that reached the southwestern U.S.
Table 3.1.: SVD reconstruction

<table>
<thead>
<tr>
<th>Variable</th>
<th>$k$</th>
<th>Explained Variances</th>
<th>Time steps</th>
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<td>LWCRF</td>
<td>5</td>
<td>60.5%</td>
<td>26</td>
</tr>
<tr>
<td></td>
<td>6</td>
<td>59.0%</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>5</td>
<td>55.9%</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td>5</td>
<td>52.9%</td>
<td>25</td>
</tr>
<tr>
<td>SWCRF</td>
<td>5</td>
<td>62.0%</td>
<td>21</td>
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<td>55.2%</td>
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<td>50.9%</td>
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<td>4</td>
<td>51.2%</td>
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</tr>
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Figure 3.1.: Average over Nov – Mar, 2000 – 2008 (a) IVT (kg m$^{-1}$ s$^{-1}$) and (b) IWP in shades (g m$^{-2}$), LWP = 60, and 75 g m$^{-2}$ in black contours.
Figure 3.2.: The composite mean of anomalous IVT vectors in kg m$^{-1}$ s$^{-1}$ for (a)–(d) AR$\text{NW}$ and (e)–(h) AR$\text{SW}$. Anomalies that are significantly different from climatology (>95% confidence) are shaded. White contours indicate IVT anomalies = 200kg m$^{-1}$ s$^{-1}$.
Figure 3.3.: Similar to Fig. 3.2, but for IWP anomalies in shades (g m$^{-2}$), and pressure anomalies at the mean sea level in black contours with 2 hPa intervals.
Figure 3.4.: Regions A and B for the Western U.S. and the Eastern U.S. Line C-D is along 124.5°W from 30° – 60°N. E is the landfalling regions for the West-Coast ARs.
Figure 3.5.: Time evolution of IWP along line C-D in Fig. 3.4. (a) is for AR\textsubscript{NW} and (b) for AR\textsubscript{SW}-AR\textsubscript{NW}. Lines with the lightest color are Day−1, and darkest color are Day+3. Day+0 is the landfalling time. “SW” and “NW” marks the landfalling latitudes of the AR\textsubscript{SW} and the AR\textsubscript{NW}.
Figure 3.6.: Similar to Fig. 3.5, but for $Q_1$ (solid line) and -$Q_2$ (dashed line) averaged over region E in Fig. 3.4 in K day$^{-1}$.
Figure 3.7.: Time evolution of (a) IWV in cm, (b) IWP in g m$^{-2}$, (c) SWCRF and (d) LWCRF in W m$^{-2}$ average over the Western U.S. (land region A in Fig. 3.4). Black lines are AR$_{NW}$, dashed blue lines the AR$_{SW}$. Thick lines show the median, thin lines represent the 25th and 75th percentiles, respectively. The "*" marks the days when the median values for the two scenarios were different at 5% significance level.
Figure 3.8.: Similar to Fig. 3.7, but for the Eastern U.S. (land region B in Fig. 3.4).
Figure 3.9.: Similar to Fig. 3.7, but contrasting (a and b) the NetCRF in W m$^{-2}$ and (c and d) the percentage of LWP in the total cloud water path between the Western U.S. (first column) and the Eastern U.S. (second column).
Figure 3.10.: As in Fig. 3.1, but for (a) LWCRF and (b) SWCRF in W m$^{-2}$. 
Figure 3.11.: EOFs and PCs for LWCRF produced by 60 AR\textsubscript{NW} on (a)–(d) Day+0 and (e)–(h) Day+3. (a and e): EOF1; (c and g): EOF2; (b and f): PC1; (d and h): PC2. The x-axis of PCs shows individual AR cases.
Figure 3.12.: Similar to Fig. 3.11, but for LWCRF induced by 60 AR$_{SW}$. 
Figure 3.13.: Reconstructed LWCRF for AR\textsubscript{NW} on (a) Day+0 and (b) on Day+3, and for AR\textsubscript{SW} on (c) Day+0 and (d) on Day+3.
Figure 3.14.: As in Fig. 3.13, but for SWCRF.

Figure 3.15.: Distribution of AR\textsubscript{GULF} 0–3 days after the West-Coast ARs.
Figure 3.16.: (a)–(f) AR_{NW} followed by AR_{GULF}. (g)–(l) AR_{SW} followed by AR_{GULF}. (a)–(c), (g)–(i) the landfalling days of the West-Coast ARs. (d)–(f) and (j)–(l) the landfalling days of the AR_{GULF}. First column: 250-hPa geopotential height (dam) in black contours, 250-hPa wind speed (m s$^{-1}$) is shaded, GPCP rain rate $\geq$ 8 mm day$^{-1}$ in blue hatches; Second column: 1000–500-hPa thickness (dam) is shaded, 850-hPa wind vectors in m s$^{-1}$, mean sea-level pressure (hPa) in black contours, IVT $\geq$ 304.92 kg m$^{-1}$ s$^{-1}$ in green hatches; Third column: IWP (g m$^{-2}$) is shaded and LWP = 70 g m$^{-2}$ in blue contours.
Figure 3.17.: Summarized schematic for (a and b) AR\textsubscript{NW} and (c and d) AR\textsubscript{SW}. (a and c) represent the arrival time of West-Coast ARs, (b and d) represent the arrival time of AR\textsubscript{GULF}. The gray contours are the 250 hPa geopotential height. The black contours outline strong mean sea surface high or low pressure.
4 THE IMPACTS OF SUBGRID-SCALE-CONVECTION-INDUCED CRF ON THE MOISTURE TRANSPORT OF ATMOSPHERE RIVERS

Chapter 3 showed that a strong AR\textsubscript{GULF} tended to occur upon the arrival of the AR\textsubscript{SW} (Fig. 3.15). In examining the IVT for individual AR\textsubscript{GULF} cases, it was found that AR\textsubscript{GULF} could draw moisture from both the Gulf of California and the Gulf of Mexico to the Central-Eastern U.S. Such moisture transport process led to multistate cloudiness (e.g., Fig. 3.3), which modulated the energy budget of the Earth-Atmosphere system through CRF (Figs. 3.7 – 3.14) and surface heat fluxes (Luo and Tung, 2015).

The research questions in this chapter are 1) How may the landfalling process of the AR\textsubscript{SW} precondition the occurrence of a strong AR\textsubscript{GULF}? 2) How do the interactions between clouds and radiation affect the moisture transport of the AR and the surface energy budget?

To address these questions, a relatively recent case was selected from the AR\textsubscript{GULF} index in section 3.1.4 for a detailed investigation. The AR\textsubscript{SW} made landfall on Mar 6, 2006, and the AR\textsubscript{GULF} took place 2 to 4 days later on Mar 8 – 10, 2006. The latter produced statewide severe storms and was documented as one of the billion-dollar disaster events by \url{http://www.ncdc.noaa.gov/billions/events}. We used observations from ERA-interim, NEXRAD (Next-Generation Radar), and NLDAS (North American Land Data Assimilation Systems) in concert with two numerical simulations facilitated by Weather Research and Forecasting (WRF) Advanced Research WRF (ARW) model. In particular, we first delineated from observations how the synoptic backgrounds of the AR\textsubscript{SW} influence the moisture sources of the AR\textsubscript{GULF}. Then, we compared the two WRF-ARW simulations and validated them against the NLDAS to infer the radiation feedbacks of subgrid-scale clouds on ARs’ moisture transport as well as on surface energy balance. The chapter is organized as follows: data and methods are described in section 4.1. A background overview for the AR
case is provided in section 4.2. Sections 4.3 and 4.4 discuss the influences of synop-
tic setup on the moisture transport of AR_GULF. Section 4.5 address the impacts of
subgrid-scale-convection-induced CRF on the moisture transport in two WRF-ARW
simulations. Conclusions are in section 4.6.

4.1 Data & Methods

This chapter uses ERA-interim 6-hourly data and CERES SYN1deg 3-hourly data
mentioned in section 3.1. In addition, we also used observations from NLDAS and
NEXRAD.

4.1.1 NLDAS

The NLDAS provides quality-controlled and spatially and temporally consistent
land-surface model datasets from best available observations and model output. The
NLDAS has a domain covering the conterminous U.S., offering vegetation, green-
ness fraction, soil properties, and elevation information. It is specifically made to
reduce the errors in the storage of soil moisture and energy which are often present
in Numerical Weather Prediction. We used NLDAS-2 Forcing products (Xia et al.,
2012), which offer hourly data in 0.125° grid spacing over conterminous U.S. from
Jan 1979 to the present. We used the precipitation, Convective Available Poten-
tial Energy (CAPE), surface downwelling shortwave and longwave radiation, surface
temperature, and surface wind to understand the bias of the WRF-ARW simulations.

4.1.2 NEXRAD Composite

In addition to the NLDAS, the NEXRAD Level III precipitation estimates were
used. NEXRAD is a network of 160 high-resolution Doppler weather radars (Fulton
et al., 1998), which is a 10 cm wavelength radar that operates at a frequency between
2,700 and 3,000 MHz. Here, we downloaded the radar mosaic images from Iowa Envi-
The images are available from Jan 01, 1995 to present every five minutes with pixel resolution about 1 km.

### 4.1.3 WRF ARW setup

To understand the effects of radiatively active subgrid clouds on ARGULF, we conducted two WRF ARW simulations from 1200 UTC 05 Mar to 0000 UTC 10 Mar, 2016. Both simulations used WRF-ARW model version 3.7 (Skamarock et al., 2008) without data assimilation. They were run on a single domain with 27-km horizontal grid spacing and 57 vertical layers. This domain covered the entire conterminous U.S. and some surrounding oceanic areas. Domain top was 10 hPa. The integral time step was 120 seconds. Both simulations used WRF single-moment 5-class microphysics scheme (Hong et al., 2004), Rapid Radiative Transfer Model for global (RRTMG, Iacono et al., 2008) shortwave and longwave schemes, Noah Land Surface model (Niu et al., 2011), MM5 similarity (Monin and Obukhov, 1954), Yonsei University planetary boundary layer scheme (Hong et al., 2006), and the Kain-Fritsch convective parameterization scheme (Kain, 2004). RRTMG was called every 20 minutes. The initial conditions and boundary conditions were derived from the National Centers for Environmental Prediction Global Forecast System Final Operational Global Analyses. The simulated fields were output hourly. The only difference between the “BASE” case and the “MOD” case was the latter allowed the subgrid cloud fraction to interact with radiation (Alapaty et al., 2012).

### 4.2 Case overview

On Mar 4 and 5, an extratropical cyclone was propagating eastward towards the West Coast of the North America. It attained its lowest surface pressure over the ocean at 984 hPa between 0600 UTC and 1500 UTC on Mar 5, as indicated by the Weather Prediction Center’s surface chart (not shown). Starting from 0000 UTC Mar 6,
this extratropical cyclone drifted northeastward to 47°N, 129°W at 1200 UTC 06 Mar (Fig. 4.1a) as it approached land. It arrived at the coast of British Columbia on 0600 UTC Mar 7 and at the western border of Alberta, Canada on 1200 UTC Mar 7 before it completely dissipated.

The AR_SW was formed on the southeast side of this extratropical cyclone (hereafter the AR_SW-storm). It reached its peak intensity over the NE Pacific Ocean between 0000 UTC and 1200 UTC on Mar 5 (e.g., 0600 UTC 05 Mar in Fig. 4.3a). After that, the AR_SW started to make landfall near 35°N, 120°W and produced enhanced cloudiness in the Western U.S. (1200 UTC 06 Mar in Fig. 4.3b). It induced precipitation over California from 1800 UTC Mar 5 to 1800 UTC Mar 6 (e.g., 1200 UTC 06 Mar in Fig. 4.1b). With the dissipation of the AR_SW-storm on Mar 7, the AR_SW drifted southward to the Gulf of California and weakened (e.g., 1800 UTC 07 Mar in Fig. 4.3c).

On Mar 8, a strong AR_GULF was formed between the surface low centered near 38°N, 102°W (hereafter the AR_GULF-storm) and a high centered near 32°N, 84°W (not shown). As the surface low deepened and moved eastward to 37°N, 96°W, the AR_GULF propagated eastward (e.g., 1800 UTC 09 Mar in Figs. 4.2a and 4.3c). This AR_GULF induced severe thunderstorms and heavy multistate precipitation from Louisiana to Ohio (Figs. 4.4 and 4.2b). The moisture content of this AR remained high even after it propagated offshore to the Western Atlantic Ocean on 1800 UTC 10 Mar (not shown). There existed a dryline over Texas between Mar 05 and Mar 09. This dryline peaked between 1200 UTC Mar 07 to 0000 UTC Mar 09 (e.g., 1800 UTC 07 Mar in Fig. 4.5). It was synoptically-active (Hane, 2004), and propagated eastward with an intense low, which later became the AR_GULF-storm.
4.3 Upper-level and Low-level Jet

The climatology study in chapter 3 revealed that the location of the upper-level jet streaks influenced the moisture transport of and the amount of cloudiness in the AR_{GULF}. To understand this impact further, here the locations of 250-hPa and 850-hPa jet as well as the 850-hPa geopotential height are examined. On 1800 UTC Mar 04 over the NE Pacific Ocean, a 250-hPa jet streak was entering the strong cyclonically curved flow near 45° N, 135° W. This produced a strong along-flow ageostrophic wind that pointed towards the upstream direction at the bottom of the trough (not shown). Accordingly, the convergence on the upstream side of the 250-hPa trough and the divergence on the downstream side of the trough were enhanced (e.g., Bjerknes and Holmboe, 1944; Beebe and Bates, 1955; Uccellini and Kocin, 1987; Moore and Van-knowe, 1992). At the same time, the AR_{SW}-storm at 850 hPa was located beneath the 250-hPa trough. It induced a strong LLJ to its right and a weak LLJ to its left (not shown). As the 250-hPa trough amplified, the ageostrophic wind at the bottom of the trough was enhanced. Correspondingly on 0000 UTC Mar 05 (not shown), the 250-hPa jet streak was “broken into” two parts, with one on the upstream side of the trough and one on the downstream side of the trough. The later intensified, leading to the impressive divergence at 250 hPa (not shown). This thus enhanced the strength of an 850-hPa LLJ off the West Coast on 0600 UTC Mar 05 (Fig. 4.6a, Shapiro, 1982). Meanwhile, an 850-hPa trough extended southwestward from 50° N, 104° W to 30° N, 107° W, inducing a weak LLJ on its downstream side. As time evolved, the 250-hPa trough and the AR_{SW}-storm over the NE Pacific Ocean weakened and moved northward. On 1200 UTC Mar 06 (Fig. 4.6b), the latter produced a weak 850-hPa LLJ along the coast of Cascades. On 1800 UTC Mar 07, the 250-hPa trough was positively tilted, extending southwestward from 55° N, 115° W to 20° N, 145° W. On its downstream side (Fig. 4.6c), two parallel 250-hPa jet streaks stretched northeastward from California, Baja California to about 40° N, 105° W. In the exit region of these parallel jet streaks, an 850-hPa trough was formed. It spanned southward from
Manitoba, Canada to New Mexico and led to an LLJ to its east. From Mar 8 to 9, the two parallel 250-hPa jet streaks were merged, strengthened and became more elongated (not shown). Prominent 250-hPa divergence was observed on the equatorward side of the merged jet streak. This divergence favored the development of a strong and lengthened 850-hPa LLJ and a deep 850-hPa trough over the central U.S. (not shown). On 1800 UTC Mar 09, the 250-hPa trough became neutrally tilted, with its half wavelength decreased (not shown). This intensifying trough helped amplify the downstream 250-hPa jet streak that spanned from 42° N, 95° W to 20° N, 110° W (Fig. 4.6d). A broad band of strong divergence was situated along and to the right of the jet axis and a narrow band of convergence was located to the left of the jet (not shown). This convergence-divergence dipole structure enhanced the magnitude of the 850-hPa LLJ, which was rooted in the Gulf of Mexico and extended along the Appalachian Mountains. The 850-hPa trough beneath the 250-hPa jet deepened as well. It stretched southwestward from the Great Lakes to Texas and to the Gulf of California (Fig. 4.6d). On 0000 UTC Mar 10 (not shown), the above mentioned 250-hPa trough tilted negatively and quickly propagated eastward.

4.4 Moisture transport

The location and amplitude of the LLJ and lower-tropospheric trough influenced the origin, the pathway, and the intensity of moisture transport of the ARGULF. To gain a 3D visualization of the moisture transport, the 300 and 315 K isentropic surfaces are presented in Figs.4.7 and 4.8, respectively. A caveat here is the mixing ratio on the isentropic surfaces can be underestimated/overestimated in regions where strong diabatic heating/cooling occur. However, the pattern revealed by the analysis provides a good qualitative understanding of the 3D moisture transport.

On 0600 UTC Mar 05 on the 300 K isentropic surface (Fig. 4.7a), the ARSW-storm was at 44° N, 135° W. In association with its WCB, warm and moist air mass ascended from 800 hPa in the NE Pacific Ocean at 25° N, 150° W to 600 hPa in the
Western U.S. near 37° N, 125° W. An anticyclone was situated in the Gulf of Mexico centered at 25° N, 90° W. It induced a warm and moist channel to its west, which drew warm and moist air from 900 hPa in the Gulf of Mexico to 500 hPa near 50° N, 100° W. On 1200 UTC Mar 06 (Fig. 4.7b), moisture ascending from the NE Pacific Ocean traversed California and reached 500 hPa at Saskatchewan, Canada. It induced precipitation in California (Fig. 4.1b), along with increasing cloudiness (Fig. 4.3b). On the other hand, moisture from the Gulf of Mexico propagated eastward towards the Atlantic Plain (Fig. 4.7b). On 1800 UTC 07 Mar (Fig. 4.7c), the warm and moist air associated with the ARSW continued to propagate towards the Canadian Shield between 500 and 800 hPa, where it met the strong moisture surge from the Gulf of Mexico. Increasing cloudiness was observed along the moisture channel from the NE Pacific (Fig. 4.3c). On 1800 UTC Mar 09, abundant moisture was brought primarily from the Gulf of Mexico to the Great Lakes and the Appalachian Mountains (Fig. 4.7d). Heavy precipitation (Fig. 4.2b) and enhanced cloudiness (Fig. 4.3d) covered a large portion of the central-eastern U.S.

Interestingly on the 315 K isentropic surface (Fig. 4.8) before Mar 08 – the onset of ARGULF – moisture from the Gulf of Mexico was confined to below 500 hPa to the south of 35° N (Figs. 4.8a–c). Yet as the ARGULF-storm developed and propagated eastward (Fig. 4.6d), the moisture surge from the Gulf of Mexico penetrated inland to the Northeastern U.S. near 47° N, 75° W (Fig. 4.8d). Moreover, 0 – 4 days after the landfall of ARSW, moisture originated in the Tropical Eastern Pacific Ocean near 15° N, 120° W traveled through the Mexican Plateau and fueled the moisture surge from the Gulf of Mexico (e.g., Figs. 4.8b–d). The moisture from the Tropical Eastern Pacific Ocean was induced by the deep trough over the central-eastern U.S. as depicted in Figs. 4.6d and 4.8d. This indicates that the synoptic background of the ARSW has downstream impacts on the moisture transport of ARGULF through affecting the locations and intensity of the lower-tropospheric trough.
4.5 The impacts of subgrid-scale-convection-induced CRF on AR$_{GULF}$

In addition to the impacts of synoptic backgrounds on the moisture transport of AR$_{GULF}$, we also want to understand the feedbacks of subgrid-scale cloudiness on AR$_{GULF}$ and the regional energy budget through CRF. Here, we compare the BASE and MOD simulations. As shown in Fig. 4.9, the SWCRF cooling dominated the net CRF budget along the pathway of the AR$_{GULF}$ from Louisiana to the Great Lakes in both simulations. However, the BASE case simulated a weaker net CRF cooling (by about 20 – 60 W m$^{-2}$, Fig. 4.9), and more surface downwelling shortwave (by about 40 – 120 W m$^{-2}$, Fig. 4.11b) in the upstream (to the south of 38° N) pathway of AR$_{GULF}$. Accordingly, the BASE case was associated with an increase of surface temperature up to about 0.5°C and an increase of 850-hPa temperature up to 0.3°C along the AR$_{GULF}$ (not shown). This extra warming reduced the northeast-southwest temperature gradient, thereby weakening the 850-hPa wind speed from 33° N, 100° W to 41° N, 81° W up to about 1.2 m s$^{-1}$ (Fig. 4.10b). The additional warming in the BASE case in the lower troposphere increased the upstream 850-hPa equivalent potential temperature $\theta_e$ by more than 0.3°C (Fig. 4.11), while reducing the downstream $\theta_e$. Such a contrast was amplified in 700 hPa (not shown). This implies that the upstream AR$_{GULF}$ in the BASE run was more unstable and moister than that of the MOD run. Indeed, despite some spatial variability, the BASE case had stronger hourly CAPE (by 100 to 300 J kg$^{-1}$) and weaker convective inhibition than the MOD case in the upstream pathway (not shown). Accordingly, the BASE case had more mixed-phase clouds (Figs. 4.12a,b) and water vapor mixing ratio (Fig. 4.12c) than the MOD case to the south of 35° N, as illustrated by the vertical profiles that intersect the AR$_{GULF}$ in the local afternoon (along 92° W at 2000 UTC 09 Mar in Fig. 4.12). The BASE case also produced a deeper convection near 34.7° N, and less low- and mid-level clouds below 700 hPa between 35° and 36.5° N than the MOD case (Figs. 4.12a,b). Thus, stronger upstream precipitation was observed in the BASE case (greater than 0.3 mm hr$^{-1}$, Fig. 4.13a). This not only dried the
upstream atmosphere quickly but also caused the early demise of convection. Hence, less subgrid-scale clouds and moisture were transported downstream, resulting in less downstream subgrid-scale (Fig. 4.13a) and grid-scale (Fig. 4.13b) precipitation.

The daily-averaged heat fluxes illustrate the impacts of the subgrid-scale-convection-induced CRF on the surface energy budget. Compared with the MOD case, an additional upstream precipitation in the BASE case enhanced the surface latent heat flux by about 5 – 30 W m\(^{-2}\) (Fig. 4.14a). Meanwhile, the extra upstream surface downwelling shortwave radiation in the BASE case increased the surface sensible heat fluxes by about 5 to 10 W m\(^{-2}\) (Fig. 4.14b).

Validating the WRF simulated fields against those from the NLDAS (Fig. 4.15) shows that the WRF underestimated the total upstream precipitation, while producing too much downstream rain (Fig. 4.15c). A further comparison shows that both the WRF runs had smaller CAPE values than that from NLDAS. This suggests that both runs underestimated the upstream buoyant energy. Unlike the results in Alapaty et al. (2012); Herwehe et al. (2014), which show an improvement of summer precipitation when accounting for the subgrid-scale-convection-induced CRF, our MOD run did not simulate the rain produced by AR\text{GULF} better. Nevertheless, the comparison between the two runs shed light on how CRF may affect the AR\text{GULF} and the surface energy budget. First, mainly through reducing the surface insolation and lower-tropospheric temperature, the subgrid-scale-convection-induced CRF helps to reduce the upstream buoyant energy. This delays the transition from shallower convection to deep convection, as well as the onset of subgrid-scale precipitation. Hence, more clouds and moisture can be transported to the downstream regions. Strong precipitation is likely to be triggered in later times in the downstream sites, and the lifespan of an AR is prolonged. Second, by reducing the upstream surface solar insolation and precipitation, the subgrid-scale-convection-induced CRF reduces the surface temperature and moisture, thereby decreasing the surface latent heat flux and upward sensible heat flux. Third, as the longevity of the upstream subgrid-scale clouds increases, the spatial and temporal coverage of low- and mid-level clouds increase (Figs. 4.12a,b).
Presumably, the extensive coverage of low- and mid-level clouds would further enhance the SWCRF cooling of the ARs, thereby further increasing the lifetime of ARs. In short, it is possible that the strong net CRF cooling of ARs favors the long-range moisture transport of a landfalling AR.

4.6 Conclusions

In this chapter, we discussed how the synoptic background upon the arrival of AR_{SW} and the subgrid-scale-convection-induced CRF of AR_{GULF} influenced the moisture transport of AR_{GULF}. When a 250-hPa jet streak merged with another one, or when the half wavelength of the 250-hPa trough reduced (Newton and Trevisan, 1984), the upper-level divergence intensified. This enhanced divergence magnified the LLJ that was associated with the AR_{GULF}-storm. It thus favored strong moisture surge from the Gulf of Mexico. At the same time, there was a deep trough in the central-eastern US (e.g., on 850 hPa, it spanned southwestward from the Great Lakes to Texas and to the Gulf of California in Fig. 4.6d). It induced southwesterly on the downstream sides. Moisture originated in the Tropical Eastern Pacific Ocean near 15° N, 120° W ascent along this southwesterly and propagated towards the Great Lakes (Figs. 4.7d and 4.8d). In addition to these two synoptic-scale mechanisms, the subgrid-scale-convection-induced CRF appeared to increase the longevity of the AR_{GULF}. In our MOD run, the SWCRF cooling outweighed the LWCRF warming. Through reducing the surface insolation by about 40 – 120 W m\(^{-2}\) (Fig. 4.10b), the subgrid-scale clouds cooled the surface by up to 0.5°C in the upstream AR_{GULF}. This reduced the horizontal temperature gradient and thereby the wind speed of the LLJ by up to 1.2 m s\(^{-1}\) (Fig. 4.10b). More importantly, the surface shortwave cooling reduced the upstream buoyant energy. As a result, the onset of deep convective precipitation was delayed in the MOD simulation. As a consequence, less drying (more than 0.3 mm hr\(^{-1}\) in Fig. 4.13a) occurred in the upstream regions where the subgrid-scale clouds originated. The spatial and temporal coverage of shallower convection
increased (Figs. 4.12a,b). More moisture was transported to the downstream sites and was removed in later times (Fig. 4.13). Furthermore, through reducing the upstream precipitation and surface insolation, the existence of the subgrid-scale cloudiness decreased the surface latent heat fluxes by about 5 to 30 W m$^{-2}$ (Fig. 4.14a), and the surface sensible heat fluxes by about 5 to 10 W m$^{-2}$ in the upstream areas (Fig. 4.14b). It is noteworthy that the additional low- and mid-level clouds in the MOD run may further reinforce the net CRF cooling, and therefore helped to enhance the longevity of the AR$_{GULF}$. 
Figure 4.1.: On 1200 UTC 06 Mar 2006: (a) Weather Prediction Center’s surface chart; (b) NEXRAD map from Iowa Environmental Mesonet.

Figure 4.2.: As in Fig. 4.1, but on 1800 UTC 09 Mar 2006.
Figure 4.3.: CERES IWP (g m\(^{-2}\)) is shaded, IWV = 2, 3 cm is in magenta contours. Time evolution from (a) 0600 UTC 05 Mar to (d) 1800 UTC 09 Mar 2006.
Figure 4.4.: SPC storm report, from

http://www.srh.noaa.gov/bmx/?n=event_03092006_weatherdata.

Figure 4.5.: Surface chart valid at 1800 UTC 07 Mar 2006.
Figure 4.6.: As in Fig. 4.3, but for 850-hPa geopotential height (m) in black contours, 850-hPa isotach $\geq 20$ m s$^{-1}$ at 5 m s$^{-1}$ interval are in the red contours, and 250-hPa isotach (m s$^{-1}$) in shades. Brown dash indicates 850-hPa toughs.
Figure 4.7.: As in Fig. 4.3, but for ERA-Interim Montgomery streamfunction at $10^3$ J kg$^{-1}$ interval in black contours, pressure (hPa) in red contours, and mixing ratio (g kg$^{-1}$) in shades on the 300 K isentropic surface.
Figure 4.8.: As in Fig. 4.7, but on the 315 K isentropic surface.

Figure 4.9.: Daily mean net CRF on Mar 9, 2006 for BASE case in shades, and for BASE–MOD case at 20 W m\(^{-2}\) intervals are contoured. The zero contours are omitted.
Figure 4.10.: Daily mean variables on Mar 9, 2006 for (a) BASE case; (b) BASE–MOD case: surface downward shortwave is shaded in W m$^{-2}$. The 850-hPa wind speed at (a) 5 m s$^{-1}$ intervals and at (b) 0.6 m s$^{-1}$ intervals are contoured. The zero contours are omitted.

Figure 4.11.: As in Fig. 4.9, but the 850-hPa theta-e for BASE case is shaded, and for BASE–MOD case at 0.3 °C intervals is contoured. The zero contours are omitted.
Figure 4.12.: Vertical profile along 92° W at 2000 UTC Mar 9 for mixing ratio of (a) cloud drops, (b) cloud ice and (c) water vapor. BASE case is in dark contours. MOD case is in light contours. BASE–MOD case is shaded. Contour intervals are (a) (b) 0.1 g kg⁻¹ and (c) 1 g kg⁻¹. “L” marks the center of a surface low.
Figure 4.13.: Similar to Fig. 4.9, but (a) is the subgrid-scale precipitation and (b) the grid-scale precipitation. Contour interval is 0.3 mm hr\(^{-1}\). The zero contours are omitted.

Figure 4.14.: As in Fig. 4.13, but (a) is the surface upwelling latent heat flux and (b) the surface upwelling sensible heat flux. Contour interval is 10 W m\(^{-2}\).
Figure 4.15.: Daily mean total precipitation on Mar 9, 2006 for (a) NLDAS, (b) BASE case, (c) BASE–NLDAS in shades and BASE–MOD case at 0.3 mm hr$^{-1}$ intervals in contours, (d) the differences between Root-mean-square error of BASE with respect to NLDAS and Root-mean-square error of MOD with respect to NLDAS at 0.5 mm hr$^{-1}$ intervals in contours. The zero contours are omitted.
5 DISCUSSIONS & IMPLICATIONS

This dissertation was guided by three main questions. First, how do clouds induced by the ARs modulate the moisture and heat budget of the Earth-Atmospheric system? Second, what are the sufficient climate conditions for the extensive CRF in the continental U.S.? Third, how does the subgrid-scale-convection-induced CRF influence the moisture transport of ARs?

Three interesting observations had been made in this study. First, the precipitating systems in ARs often revealed a vertically tilted structure (chapter 2). This implied the existence of various kinds of convection from the shallower ones, to deep convection, and to the trailing stratiform deck. The variations in spatial coverage and cloud types were translate to the differences in radiative forcing and surface heat fluxes, thus affecting the regional heat and moisture balance differently. Second, when the West-Coast ARs and the AR$_{\text{GULF}}$ occurred in sequence, it favored extensive cloudiness therefore CRF in continental U.S. (chapter 3). Such scenario tended to occur upon the arrival of AR$_{\text{SW}}$, with the synoptic setup playing an important role. Third, the subgrid-scale clouds in the ARs could modulate moisture transport of the ARs directly and indirectly via CRF. Through precipitation, the clouds removed moisture from distant regions such as NE Pacific Ocean and Gulf of Mexico (chapters 2 and 4). Meanwhile, it introduced localized moisture sources to the ARs via precipitation re-evaporation and surface evaporation (chapter 2). These are the direct pathways. In addition, the presence of the subgrid-scale convection reduced the surface shortwave heating in the ARs. This appeared to alleviate the upstream deep convective precipitation in our MOD run (chapter 4). Our WRF simulations and observations suggested that such process increased the upstream low- and mid-level clouds, thus providing a possible feedback loop to further reduce the shortwave heating in the ARs (Fig. 5.1). Even more, when the upstream subgrid-scale precipita-
tion was reduced, more moisture could be transported to the downstream sites by the ARs, with a reduction in the upstream surface latent heat flux (Fig. 5.1). Considering that the ARs are defined by the moisture content, it is possible that the former would favor a long-lived AR, while the latter might counteract the positive effects of the former (Fig. 5.1). Further research is needed to understand the complex interactions among clouds, radiation, and the moisture transport of the ARs, as well as to reveal the relative importance of different feedback mechanisms (Fig. 5.1).

Five questions could be asked based on this work. First, what roles do local topography, water bodies and aerosols play in modulating the magnitudes of CRF (Neiman et al., 2013; Creamean et al., 2015; Ralph et al., 2016; White et al., 2015)? Second, how may the extensive CRF induced by the AR_{SW} favor the incidences of the long-lasting and strong AR_{GULF}? Third, how sensitive is the CRF coupled with ARs to the cloud types? Fourth, how does the CRF of the ARs fluctuate the regional climate? Five, how well do our contemporary weather and climate models represent the multiscale interactions among CRF, surface fluxes, and atmospheric moisture transport in various AR scenarios? For questions 2–4, we expect that the difference in net CRF between the Western US (i.e., net warming) and the Central-Eastern US (i.e., net cooling) may complicate the questions. In order to further examine these questions, particularly for questions 2 and 3, we plan to perform a sensitivity test on the initial conditions. Specifically, we are initiating the WRF simulation three days prior to the landfall of the West-Coast AR, and concluding the simulation after the Gulf-Coast AR demise. To explicitly resolve the convection and the associated CRF for the Gulf-Coast AR, we will use three nested domains. The inner-most one will cover the Central-Eastern U.S. at 30° –45° N, 100° –85° W with ≤3 km horizontal grid spacing.

In conclusion, landfalling ARs often lead to extreme precipitation, widespread flood, mudslides, and avalanches over a large area, which strongly impacts the human dimension. To facilitate the AR-related disasters prevention, we need to improve our understanding of the conditions that contributes to strong landfalling ARs. Further-
more, by better understanding the physical mechanisms through which ARs affect land surface energy budget through CRF, we may better predict the changing AR impacts as the climate change. Our work not only filled the current research gap but also have implications for improving AR predictions. Our results are readily usable for validating and improving weather and climate models. More broadly, when combined with human-dimension data, our findings can be applied to assess regional/global climate risks. The dissertation thereby serves as a basis for informing the public and private sector decision making in disaster and water management.

Figure 5.1.: The hypothesized feedback loops based on WRF simulations and observations. Line with arrow head indicates positive coupling. Line with an open circle indicates negative coupling.
REFERENCES


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EDUCATION

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