Synoptic-to-Regional-Scale Analysis of Rainfall in the Atacama Desert (18°–26°S) Using a Long-Term Simulation with WRF

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ABSTRACT

In this study, reanalysis data and a long-term simulation with the regional climate model WRF (1982–2017; 10 km resolution) is used to analyze synoptic and regional processes associated with rainfall events in the Atacama Desert. Five composites, each with 10 WRF-simulated rainfall events, are studied. They are selected based on a clustering and comprise the top winter events in South Atacama (23°–26°S), Southeast Atacama, and North Atacama (18°–23°S), and the top summer events in North Atacama and Northeast Atacama. Winter rainfall events in South Atacama are mostly associated with strong low pressure systems over the southeast Pacific and atmospheric rivers at their foreshore, while cutoff lows occurring anomalously far north facilitate strong rainfall in North Atacama. Accordingly, tropical continental areas and the remote tropical and subtropical Pacific are identified as primary moisture sources, and moisture transport toward the Atacama Desert mainly takes place in the free troposphere (above 800 hPa). Strong summer rainfall events in North Atacama and Northeast Atacama are associated with a southward displaced Bolivian high. During rainfall events in North Atacama the high is shifted westward when compared to the Northeast Atacama events. Consequently, northern Chile is located at the northern periphery of the Bolivian high and the resulting strong easterlies may push strong convective systems from the Altiplano, toward the Atacama coast. Detailed analyses of individual rainfall events reveal that the most important synoptic patterns associated with rainfall not only control the synoptic-scale moisture transport into the Atacama Desert, but also decisively influence the regional atmospheric circulation.

KEYWORDS: Atmosphere; South America

1. Introduction

The Atacama Desert (approximately 18°–26°S, 70.5°–68.5°W) in northern Chile is considered to be the driest desert on Earth. It is bordered by the southeast Pacific and a steep coastal cliff in the west and the Andes in the east, and is thus characterized by strong orographic gradients (Fig. 1a). The climate in the Central Valley below 2300 m above mean sea level (MSL) is hyper-arid, with annual precipitation of less than 1 mm in some parts of the Atacama Desert, while mean rainfall increases up to 300 mm yr^{−1}

toward the western Cordillera (eastern part in Fig. 1a) and Altiplano (northeastern part in Fig. 1a) (Houston 2006).

The hyper-arid climate in the Central Valley is the result of a unique and complex interplay of several processes in this region. The Atacama Desert is located in the subtropics at the descending branch of the Hadley cell (e.g., Rondanelli et al. 2015). The resulting large-scale subsidence over the southeast Pacific, which is intensified in austral summer by the response to tropical convection over the South American continent (Rodwell and Hoskins 2001), leads to the formation of a subsidence inversion and a high pressure system over the subtropical southeast Pacific and drives southerlies along the coast of the Atacama Desert. These southerlies induce advection of cold waters from higher latitudes and upwelling of cold deep water. The low sea surface temperatures enhance the subsidence inversion, which, due to the orographic characteristics (see Fig. 1a), decouples the moist marine boundary layer from the boundary layer inland. Furthermore, insolation effects over the western slopes of the Andes result in strong

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FIG. 1. (a) Model domain and topographical height (in m) of the WRF simulation with a horizontal resolution of 10 km. The labels “North Atacama,” “NE Atacama,” “South Atacama,” and “SE Atacama” refer to the naming convention for the four cluster domains (see Fig. 2a) used throughout the manuscript. The red dot represents the location of Antofagasta. (b) WRF-simulated (shaded surfaces) and observed (shaded points) mean annual rainfall (in mm yr^{-1}) for the period 1982–2017.

daytime upslope flows that induce zonal divergence with a subsidence return flow over the near-coastal Central Valley and thus contribute to the hyper-aridity in this region (Rutllant et al. 2003, 2013). Interestingly, there is no consensus about the role of the Andes in blocking moisture advection from the Amazon region. In a climate model study, Garreaud et al. (2010) found that a reduction of the mean height of the central Andes leads to a drying of the continental plains, but does not affect the precipitation in the Atacama.

In a hyper-arid environment like the Atacama Desert, rainfall events and episodes of increased water availability leave long-lasting geomorphological traces and have a strong impact on the biota. For example, high accumulated precipitation and extreme rainfall events may lead to activating germination of many species (Plischoff et al. 2017) and result in spectacular “blooming desert” events in the Atacama Desert (Chávez et al. 2019). Walk et al. (2020) found that coastal alluvial fan morphodynamics in northern Chile are primarily controlled by climate processes that are associated with precipitation. Moreover, heavy rainfall events of the recent past triggered debris flows along the coast of the Antofagasta region (Vargas et al. 2006). Hence, rainfall is one of the main controlling factors for the

geomorphological and biological evolution in the Atacama Desert. Consequently, large efforts have been made in establishing precipitation archives for this region in the recent past (Ritter et al. 2019; Diederich et al. 2020). For a reliable interpretation of these archives a good knowledge of the processes controlling the variability and heterogeneity of rainfall is required.

Previous studies showed that rainfall in the Atacama Desert is related to large-scale processes and tropical–subtropical teleconnections (e.g., Garreaud and Rutllant 1996). Houston (2006) analyzed the impact of ENSO on Atacama rainfall based on historical station measurements. For El Niño events he found positive precipitation anomalies along the coast during summer and throughout central and southern Atacama during winter. In contrast, drier than normal conditions occur during these events in the Altiplano. La Niña events are associated with positive precipitation anomalies in the western Cordillera and the Altiplano in summer, although not obviously in winter. For northern Chile, Vargas et al. (2006) found that debris flows associated with strong winter rainfall occurred during the development phases of El Niño events accompanied by a northward displacement of subtropical troughs. These troughs are either related to an equatorward-shifted blocking high

(corresponding to a positive phase of the Pacific–South America pattern; [Vargas et al. 2006](#)) that forces low pressure systems northward, or to a ridge embedded in a deep trough (negative phase of the Pacific–South America pattern) that may generate cutoff lows off northern Chile. It should be noted that there is obviously a sharp transition between the processes associated with rainfall over central Chile and those associated with rainfall in northern Chile and the Atacama Desert. For instance, using monthly gridded rainfall datasets [Barrett and Hameed \(2017\)](#) found that precipitation in central Chile and in northern Chile are reversely correlated with the position and the intensity of the South Pacific high.

A close relationship between the phases of the Madden–Julian oscillation (MJO) and precipitation anomalies over central Chile has been documented by [Barrett et al. \(2012\)](#). [Juliá et al. \(2012\)](#) conclude that the majority of strong precipitation events at the coast in La Serena (30°S) come along with an active MJO near the central equatorial Pacific. At the Chajnantor plateau located in the eastern Atacama Desert at above 4800 m MSL, composites of different MJO phases agree well with the precipitable water vapor (PWV) composites, thus indicating that the MJO may modulate the PWV content over the Atacama Desert ([Marín and Barrett 2017](#)). [Meseguer-Ruiz et al. \(2019\)](#) applied a *K*-means clustering to predictors from a reanalysis dataset to classify four weather regimes that describe the atmospheric circulation over northern Chile in austral summer, and these weather regimes are strongly linked to precipitation anomalies at stations in the study area. For example, positive precipitation anomalies in the central zone of northern Chile are related to a negative upper-tropospheric geopotential height anomaly centered south of 30°S, while a weather regime representing the Bolivian high shifted toward central Chile is associated with positive rainfall anomalies in high altitudes in the northeast of the study area. The latter corresponds to the findings of [Garreaud et al. \(2003\)](#), who concluded that the position and the intensity of the Bolivian high influence the upper-air circulation and thus the intra-seasonal rainfall variability over the Altiplano.

While the large-scale mechanisms facilitating rainfall in the Atacama Desert are meanwhile well understood, studies dealing with the synoptic-to-regional-scale processes involved in precipitation events and the rainfall heterogeneity are relatively few, particularly for the hyper-arid core of the desert. [Bozkurt et al. \(2016\)](#) analyzed in detail a single extreme precipitation event that occurred in March 2015 (the so-called Atacama Flood). This event was associated with a cutoff low off the coast of northern Chile accompanied by positive sea surface temperature anomalies over the eastern tropical Pacific.

Due to this constellation, anomalously high precipitable water content over the Peruvian Bight region was advected southeastward ahead of this cutoff low toward northern Chile ([Barrett et al. 2016](#)). Aside from reanalysis data [Bozkurt et al. \(2016\)](#) used regional climate model (RCM) simulations to study the key local processes. Although the focus was on sensitivity experiments, their study is an excellent example for the benefit of using RCMs for investigating rainfall in an area with very limited ground observations, as is the case for the Atacama Desert.

The aim of this study is to contribute to a better understanding of the processes involved in rainfall events in the Atacama Desert. While most previous studies focus on large-scale processes, this study explores in more detail the synoptic conditions and the regional circulation patterns that trigger remarkable rainfall in different seasons and regions of this desert. Focus is placed on rainfall events occurring between 18° and 26°S and in altitudes below 3000 m MSL, a region that comprises the hyper-arid core of the Atacama Desert. We expect that the small precipitation amounts in this region are (to some extent) spillovers from the neighboring areas, and we purpose to uncover the processes facilitating these spillovers. In addition to reanalysis data we use, for the first time, simulated precipitation from a long-term (1982–2017) RCM run for our study. Similar to reanalysis datasets such an RCM simulation represents a physically consistent, complete, and homogeneous dataset. However, it has the advantage of a much higher spatial resolution, which is essential due to the orographic complexity of the study area ([Bozkurt et al. 2019](#)).

For the selection of the rainfall events analyzed in this study, a clustering of gridded annual rainfall cycles is done to identify winter and summer rainfall dominated domains ([section 3a](#)). Composite analyses for the selected top 10 events of each cluster domain are used to identify the relevant synoptic circulation patterns and moisture fluxes ([section 3b](#)). Further, individual cases are analyzed for detailed process studies using both reanalysis and RCM data ([section 3c](#)). The data and methods used in this study are described in [section 2](#), while a conclusion and discussion closes this paper in [section 4](#). A detailed evaluation of simulated rainfall is given in the supplemental material.

2. Data and methods

a. Reanalysis data

For composite analyses we use different variables and derived quantities of the ERA-Interim dataset

(Dee et al. 2011) of the European Centre for Medium-Range Weather Forecasts (ECMWF). It has a horizontal resolution of 0.75° and is available for the period from January 1979 to August 2019. Here, we use the time span January 1982 to December 2017. Composites are calculated for 500 and 200 hPa geopotential height fields (z_{500} and z_{200}), outgoing longwave radiation (OLR; as a measure for deep convection), sea surface temperature (SST), and integrated water vapor (IWV; also known as precipitable water vapor) as well as its horizontal flux (IWVF).

IWV is computed following formula in (1):

$$\text{IWV} = \frac{1}{\rho g} \int_{p_1}^{p_2} q dp, \quad (1)$$

where g is the acceleration of gravity, ρ is the density of water, q is the specific humidity, and p is the pressure. Here, we integrate the humidity between the surface (p_1) and 200 hPa (p_2).

IWVF consists of a zonal (IWVF_u) and a meridional (IWVF_v) component:

$$\left(\text{IWVF}_u = \frac{1}{\rho g} \int_{p_1}^{p_2} uq dp, \quad \text{IWVF}_v = \frac{1}{\rho g} \int_{p_1}^{p_2} vq dp \right), \quad (2)$$

where u is the zonal wind component and v the meridional wind component. To separate between low-level and upper-level moisture transport, IWVF is integrated between surface and 800 hPa, and between 800 and 200 hPa, respectively.

In this study, we also investigate the impact of mid-tropospheric cutoff lows (COLs; Reyes and Shao 2019) off the coast of northern Chile and of atmospheric rivers (ARs) that made landfall at the Atacama coast. COLs are upper-level closed low pressure systems that segregate from the large-scale flow during their life cycle, and COLs over the southeast Pacific are often associated with rainfall in northern Chile (Bozkurt et al. 2016; Pizarro and Montecinos 2000). The dates of COLs occurring in a domain ranging from 85° to 70°W and from 30° to 15°S are taken from Reyes and Shao (2019), who applied a COL detection method to z_{500} fields of ERA-Interim.

ARs are characterized by narrow and elongated bands of high water vapor content (Newell et al. 1992). They frequently produce precipitation when they hit land-masses, which in some cases may have extreme magnitudes thus leading to floods (e.g., Neiman et al. 2011). For central Chile (32° – 37°S) Viale et al. (2018) found evidence that ARs contribute to approximately half of the annual rainfall amount and that particularly in

austral winter ARs are active at the Chilean coast north of 30°S . A global database for the timing and location of ARs has been created by Guan and Waliser (2015). They applied their identification algorithm to different reanalysis datasets. Here, we use their AR Reanalysis Database based on ERA-Interim, which is provided on a horizontal $1.5^\circ \times 1.5^\circ$ grid at 6-hourly time steps. The dataset is available for the years 1979–2019. The algorithm identifies objects first based on IWVF intensity. This way objects are marked when IWVF is greater than the 85th percentile but at least greater than $100 \text{ kg m}^{-1} \text{ s}^{-1}$ at each grid cell. Second, the direction of the IWVF is considered (mean IWVF direction need to be within 45° of the AR shape orientation). Furthermore, the objects need to be longer than 2000 km and have a length to width ratio of 2 or more. For this study we filtered for ARs that made landfall at the Atacama coast between 15° and 30°S .

ERA-Interim is further used to identify back trajectory paths of air masses involved in the rainfall events over the Atacama Desert. Here, the HYSPLIT (Hybrid Single-Particle Lagrangian Integrated Trajectory) transport and dispersion model (<http://www.arl.noaa.gov/ready/hysplit4.html>, NOAA Air Resources Laboratory; Stein et al. 2015) is used to compute simple air parcel backward trajectories for 10-day periods, using 6-hourly meteorological 3D-fields as input.

b. The regional climate model WRF

The Weather Research and Forecasting Model (WRF; <http://www.wrf-model.org>; Skamarock et al. 2008), version 3.9, is used in this study. A simulation covering the period of 1982–2017 has been performed, using ERA-Interim as boundary conditions (Reyers 2018). Via a double one-way nesting (first from 0.75° to 30 km, then from 30 to 10 km horizontal resolution), a horizontal resolution of 10 km is achieved. The inner model domain (Fig. 1a) comprises 78 grid boxes in west–east direction, 111 grid boxes in south–north direction, and 44 vertical terrain-following eta-levels. The 6-hourly ERA-Interim data are used as initial and boundary conditions. Hence, this WRF simulation provides a physically consistent and dynamically downscaled dataset.

To find the optimal model setup we performed several test runs of individual rainfall events (e.g., the Atacama 2015 flood) using different physical configurations. This model setup includes the Thompson Scheme for microphysics (Thompson et al. 2008), the CAM scheme for radiation (Collins et al. 2004), the Eta Similarity scheme for surface layer physics (Janjić 1994), the Pleim–Xiu Land Surface Model (Xiu and Pleim 2001), the Mellor–Yamada–Janjić Scheme for the planetary boundary layer (Janjić 1994), and the modified Kain–Fritsch cumulus

scheme (Ma and Tan 2009). The WRF Model physics do not explicitly predict SSTs. However, for long-term simulations it is possible to prescribe SSTs by reading in time-varying data and update the SST fields at a given time interval (WRF option `sst_update=1`). Optionally, an appropriate diurnal signal can be calculated and added to the input SST (WRF option `sst_skin = 1`). For our simulation, time-varying SSTs from ERA-Interim are integrated and updated every 6 h. These SSTs are derived from different products, starting with the NCEP 2D-Var SST for the period prior to 2002 and using the Operational Sea Surface Temperature and Sea ice Analysis (OSTIA) from the ECMWF for the most recent period (Dee et al. 2011). OSTIA, for example, is produced on a daily basis from satellite data and in situ measurements from drifting and moored buoys (Donlon et al. 2012). Vazquez-Cuervo et al. (2013) demonstrated that the OSTIA SSTs off the Peruvian coast show clearly defined offshore maxima and large-scale structures associated with the Peruvian upwelling. It can thus be assumed that short- and long-term SST variations as well as coastal upwelling structures are captured realistically by the ERA-Interim SST products.

The WRF Model output is stored as hourly data. However, for our analysis hourly total precipitation (convective plus nonconvective rainfall) is accumulated to daily rainfall first. To overcome boundary effects, simulated rainfall at grid boxes at the edges of the model domain are disregarded.

c. Clustering

Two clustering methodologies are used in this study to systematically identify regions in the Atacama Desert that are dominated by winter/summer rainfall. While clustering procedures in climate science are often employed to find spatial cluster patterns, this technique can also be used to identify classes of time series with similar temporal characteristics. Here, we apply the *K*-means clustering (e.g., Hartigan and Wong 1979) and the agglomerative hierarchical clustering (e.g., Day 1984), using WRF-simulated mean annual cycles of rainfall at all model grid boxes located between 18° and 26°S over land and below 3000 m as input data.

K-means clustering is an iterative algorithm that finds an optimal partition of the input data (here, the gridded mean annual cycles of rainfall) into k clusters. The algorithm seeks so-called cluster centroids, and an optimal solution is obtained when the squared distances between the cluster members and their respective cluster centroids are minimized.

The agglomerative hierarchical clustering is a bottom-up approach. At the start, each of the n input data points (here, each gridded mean annual cycle of rainfall)

represents one cluster. Then, the two closest clusters (here, the two mean annual cycles with the most similar shapes) are consolidated, resulting in $n - 1$ clusters. This step is iteratively repeated, until one large cluster is established. Finally, a dendrogram representing the joint data points and their respective Euclidian distances is used to identify the optimal partition into k clusters.

3. Results

a. Rainfall cluster

The two clustering methodologies are applied to WRF-simulated rainfall. A comparison between simulated mean annual rainfall and precipitation from observations (see supplemental material) for the period 1982–2017 is depicted in Fig. 1b. A dry bias of the model is visible for the NE Atacama, which may either be associated with the underestimation of summer convection in the convective scheme used in the simulations, with the lack of moisture transport from the eastern side of the Altiplano, or with an underrepresentation of orographic gradients in our WRF Model domain. Simulated mean annual rainfall has a similar order of magnitude as the observations in the hyper-arid Central Valley south of 21°S (less than 10 mm yr⁻¹), along the coast, and also at stations near the western Cordillera (approximately 23.5°S, 68°W; >25 mm yr⁻¹). In contrast, in ERA-Interim, which is used to force the WRF simulation, mean precipitation is clearly overestimated, being partly up to two orders of magnitudes higher at near-coastal grid points (not shown). This clearly demonstrates the added value of dynamical downscaling for rainfall studies in this region. A detailed comparison of WRF-simulated precipitation with the observations (see supplemental material) indicates that simulated rainfall is suitable for our purposes, despite the dry bias in the eastern part of the focus area.

By using the Elbow Method for *K*-means and a dendrogram for the agglomerative hierarchical approach, both clustering techniques reveal that a partitioning into four classes is most appropriate to cluster the simulated mean annual cycles of rainfall in the Atacama Desert. The results of both methods are generally similar. However, as the outcome of the hierarchical clustering is somewhat smoother, we decided to finally use the results of this technique in our study. The four identified rainfall clusters and the respective cluster domains are shown in Fig. 2. Most WRF grid boxes of cluster 1 are located in the southern part of the hyper-arid Atacama Desert south of 23°S (Fig. 2a). This cluster is a winter-dominated rainfall zone, with strongest precipitation from May to August (Fig. 2e). The cluster 2 domain

FIG. 2. (a) Partitioning of the WRF Model grid points into four rainfall clusters, and (b)–(e) the simulated mean annual cycles of precipitation at all grid points belonging to clusters 1–4 as obtained by the hierarchical clustering. The lines in (b)–(e) are randomly shaded for visibility purposes. For more details, see main text.

comprises the southeast part of the target area toward the western Cordillera, and shows a rainfall peak in June (Fig. 2d). The driest conditions are found for cluster 3 in the northern part of the hyper-arid Central Valley, where monthly rainfall never exceeds 2 mm. Nevertheless, two “peak seasons” can be identified: January–February and June–August (Fig. 2c). The domain of cluster 4 corresponds to the northeastern part of the study area, with peak rainfall of up to 8 mm month^{-1} in summer (January–February, see Fig. 2b).

Based on these clusters, we selected the rainfall events to be analyzed in this study. To identify most common patterns, we focus on the peak seasons of the respective clusters. Further, we select 10 events for each case to obtain a reasonable number of events in a hyper-arid desert with rainless periods lasting for

several years. The selection is done by following three steps:

- (i) for each of the four cluster, simulated daily rainfall is aggregated over all grid boxes in the respective cluster domain;
- (ii) the aggregated daily rainfall amounts of the peak seasons are ranked;
- (iii) the 10 strongest events are selected for each peak season of each cluster.

This results in the 10 strongest rainfall events occurring in May–August for cluster 1 (hereafter cl1_MJJA; the cluster 1 domain refers to South Atacama), in June for cluster 2 (cl2_JUNE; SE Atacama), and in June–August and January–February for cluster 3 (cl3_JJA and cl3_JF; North Atacama). A special case is cluster 4,

TABLE 1. Dates of the selected rainfall events of the cl1_MJJA, cl2_JUNE, cl3_JJA, cl3_JF, and cl4_JF composites. In parentheses “COL” and/or “AR” indicate that a cutoff low and/or an atmospheric river is identified on the day of the event (day 0), “COL-1” or “AR-1” indicates that a cutoff low or an atmospheric river is solely identified 1 day before the event (day 1), and the numbers after the semicolons show the maximum simulated rainfall occurring in a $10\text{ km} \times 10\text{ km}$ grid box during the individual events (in mm day^{-1}). The events marked in boldface are used for the case studies in [section 3c](#). The two events of cl1_MJJA marked with asterisk correspond to debris flow events in [Vargas et al. \(2006\)](#). For more details, see main text.

cl1_MJJA	cl2_JUNE	cl3_JJA	cl3_JF	cl4_JF
11 Jun 1983 (AR; 23.0)	3 Jun 1983 (AR-1; 17.0)	18 Jun 1982 (COL-1; 5.5)	4 Jan 1991 (5.5)	14 Jan 1993 (11.0)
1 Jul 1983 (COL, AR; 11.5)	8 Jun 1984 (AR-1; 6.3)	21 Jul 1987 (COL; 3.1)	2 Jan 1997 (4.0)	8 Feb 1997 (10.5)
8 Jun 1984 (AR-1; 28.8)	24 Jun 1984 (AR-1; 20.2)	10 Aug 1997 (COL; 3.1)	12 Jan 2001 (10.1)	3 Feb 1999 (19.1)
25 Jul 1987* (AR-1; 9.8)	11 Jun 1990 (COL-1; 3.9)	19 Jul 2003 (COL; 1.6)	2 Feb 2001 (2.8)	17 Jan 2000 (18.9)
18 Jun 1991* (48.9)	4 Jun 1992 (COL; 3.9)	2 Jul 2004 (COL; 4.7)	6 Feb 2004 (11.2)	17 Feb 2002 (8.1)
11 Aug 1993 (COL, AR; 60.0)	28 Jun 1999 (COL-1; 7.5)	24 Aug 2004 (COL; 3.5)	21 Jan 2005 (5.2)	10 Feb 2005 (10.0)
20 Aug 1993 (COL; 10.7)	2 Jun 2000 (COL-1; 7.8)	4 Jul 2011 (AR; 0.9)	12 Jan 2008 (7.6)	14 Jan 2006 (10.2)
2 Jul 2002 (AR; 13.1)	2 Jun 2006 (11.1)	24 Jun 2013 (11.5)	5 Jan 2010 (7.2)	22 Feb 2011 (8.8)
8 Jul 2011 (COL, AR; 17.8)	24 Jun 2007 (COL; 7.1)	20 Jul 2013 (AR; 4.1)	7 Feb 2011 (7.0)	30 Jan 2012 (12.2)
17 May 2013 (COL, AR; 24.5)	13 Jun 2013 (COL; 6.3)	24 Aug 2013 (COL; 5.1)	15 Feb 2012 (2.5)	13 Feb 2012 (9.9)

as it has six summer top events in common with cluster 3. This is due to the spillover of convective rainfall over the Altiplano into the northern part of the Atacama Desert. Hence, for cluster 4 steps (i) and (ii) are equal, whereas in step (iii) we removed those events from the top 10 list that are already included in cl3_JF, and replaced them by the next strongest events from the ranked list (hereafter cl4_JF; NE Atacama). Altogether, the rainfall database for our analysis thus comprises five composites (winter rainfall South Atacama, cl1_MJJA; June rainfall in SE Atacama, cl2_JUNE; winter rainfall in North Atacama, cl3_JJA; summer rainfall in North Atacama, cl3_JF; and summer rainfall in NE Atacama, cl4_JF) with 10 events, respectively. The dates of the selected top events are presented in [Table 1](#). Note that two dates of cl1_MJJA coincide with debris flow incidents along the coast of the Antofagasta region and associated heavy rainfall analyzed in [Vargas et al. \(2006\)](#). Further, for each individual event the maximum daily simulated precipitation occurring in a $10\text{ km} \times 10\text{ km}$ grid box is presented in [Table 1](#). In this regard, strongest winter rainfall ranges from 9.8 to 60.0 mm day^{-1} in South Atacama (cl1_MJJA), from 3.9 to 17.0 mm day^{-1} in SE Atacama (cl2_JUNE), and from 0.9 to 11.5 mm day^{-1} in North Atacama (cl3_JJA). Strongest summer rainfall ranges from 2.5 to 11.2 mm day^{-1} in North Atacama (cl3_JF), and from 8.1 to 19.1 mm day^{-1} in NE Atacama (cl4_JF).

b. Composite analysis

In this subsection, the five selected composites ([section 3a](#)) are analyzed with focus on (i) the spatial rainfall distributions [[section 3b\(1\)](#)], (ii) the different synoptic circulation patterns and the moisture fluxes associated with the selected rainfall events [[section 3b\(2\)](#)], and (iii) the

role of local and remote SST anomalies and the origin of the involved air masses via backward trajectories [[section 3b\(3\)](#)].

1) RAINFALL DISTRIBUTION

[Figure 3](#) shows the composite means of simulated rainfall for the five composites. Comparable strong precipitation of more than 8 mm per event on average is simulated for cl1_MJJA (winter rainfall in South Atacama), in particular in the southernmost part ([Fig. 3a](#)), and for much of the cluster 1 domain, rainfall is simulated for more than 7 events out of 10 ([Fig. 3f](#)). Moreover, [Fig. 3f](#) shows that winter rainfall in South Atacama is often accompanied by precipitation in higher altitudes toward the western slopes of the Andes (here, the domain of cluster 2). In contrast, the rainfall signal of cl2_JUNE is mostly restricted to SE Atacama ([Figs. 3b,g](#)). A similar behavior is found for cl3_JF (summer rainfall North Atacama) and cl4_JF (summer rainfall in NE Atacama), i.e., the summer rainfall events in NE Atacama selected for cl4_JF are often isolated in higher altitudes ([Figs. 3e,j](#)), while summer events of cl3_JF also affect higher elevated areas in the East ([Figs. 3d,i](#)). For both, summer rainfall in North Atacama and in NE Atacama, no rainfall is simulated in the Atacama Desert south of 22°S . Lowest rainfall amounts are found for the cl3_JJA composite (winter rainfall in North Atacama; [Figs. 3c and 3h](#)), spatially varying only between 0 and 1 mm per event on average.

2) SYNOPTIC CIRCULATION PATTERNS AND ASSOCIATED MOISTURE FLUXES

In this subsection, composite fields of different quantities from ERA-Interim are presented. Focus will be first on upper-level circulation patterns and moisture fluxes ([Figs. 4–6](#)), while the moisture transport in

FIG. 3. (a)–(e) Composite means of WRF-simulated rainfall (in mm day^{-1}) and (f)–(j) number of days with rainfall (grid boxes with no rainy days are masked out) for (a),(f) cl1_MJJA; (b),(g) cl2_JUNE; (c),(h) cl3_JJA; (d),(i) cl3_JF; and (e),(j) cl4_JF. In (a)–(e) the respective cluster domains (see partitioning in Fig. 2a) are marked by the bold black lines. Note the different scaling of the color bars in (a)–(e).

lower levels is briefly discussed at the end of this section (Fig. 7).

The winter rainfall events in South Atacama (cl1_MJJA composite) are found to occur under the influence of a strong midtropospheric low pressure system over the southeast Pacific (Figs. 4a and 6a). During 5 out of the 10 events a COL occurs at the day of the rainfall events (day 0; Table 1). At the foreside of the trough visible in Fig. 4a a strong northwest–southeast-oriented upper-level IWVF (above 800 hPa) develops, which in 8 out of 10 cases fulfills all criteria of an AR (Table 1). Consequently, anomalous moist air is gathering over the entire Atacama Desert.

A slightly different picture is revealed for the winter rainfall events in SE Atacama (cl2_JUNE composite). The z500 composite mean also shows a trough over the southeast Pacific (Fig. 4b). However, this trough seems to be in the dissolving stage and is thus weaker when compared to the z500 composite mean of cl1_MJJA (cf. Figs. 6b to 6a). This result reflects the fact that three ARs and three out of six COLs are identified for the day prior the event (day 1), but not for the day of the event itself (day 0; Table 1). Consequently, the composite mean of

upper-level IWVF (Fig. 4b) has a stronger zonal component, and the moist IWV anomaly over the near-coastal Pacific is somewhat weaker when compared to the conditions during winter rainfall events in South Atacama.

The winter rainfall events in North Atacama (cl3_JJA composite) are related to a southeast Pacific trough as well (Fig. 4c). Only this time, the composite mean z500 field is anticyclonically deformed south and upstream of the trough. This is due to a strong high pressure wedge propagating southwestward (Fig. 6c), thus segregating the trough and often forming a COL. Accordingly, 6 out of 10 events are associated with a COL at day 0, and one with a COL at day 1. When compared to the conditions during winter rainfall events in South Atacama, the negative z500 composite anomaly is located farther north (cf. Figs. 6c to 6a). The upper-level IWVF at the foreside has thus a strong northerly component, and positive IWV anomalies are particularly moist over the northern part of the Atacama north of 25°S (Fig. 4c). Further, ARs seem to play only minor role (only two ARs are found for cl3_JJA; see Table 1).

The summer rainfall events in North Atacama (cl3_JF composite) occur under the influence of an anticyclone

FIG. 4. Composite means of $z500$ (green contours, every 20 gpm) and upper-level IWVF (black arrows, in $\text{kg m}^{-1} \text{s}^{-1}$), and composite anomalies of IWV (shading, in kg m^{-2}) from ERA-Interim for (a) cl1_MJJA, (b) cl2_JUNE, (c) cl3_JJA, (d) cl3_JF, and (e) cl4_JF. The IWV anomalies are given relative to the climatology over the respective season.

over the South American continent (Fig. 4d). This anticyclone corresponds to a southward displacement of the Bolivian high, an upper-level feature particularly visible in the $z200$ field (Figs. 5d and 6d). Coincidentally, negative OLR anomalies occur over and east of the Andes, reflecting high convective activity in this

region (Fig. 5d). At the northwestern periphery of the Bolivian high, the resulting upper-level atmospheric moisture is advected southwestward toward northern Chile (Fig. 4d). As a result, strong positive IWV anomalies are located over the Altiplano and northern Chile.

FIG. 5. Composite means of $z200$ (black contours, every 20 gpm) and composite anomalies of OLR (shading, in W m^{-2}) from ERA-Interim for (a) cl1_MJJA, (b) cl2_JUNE, (c) cl3_JJA, (d) cl3_JF, and (e) cl4_JF. The OLR anomalies are given relative to the climatology over the respective season.

FIG. 6. Composite anomalies of (a)–(c) $z500$ (in gpm) and (d), (e) $z200$ (in gpm) from ERA-Interim for (a) cl1_MJJA, (b) cl2_JUNE, (c) cl3_JJA, (d) cl3_JF, and (e) cl4_JF. The anomalies are given relative to the climatology over the respective season.

Similar results are found for the summer rainfall events in NE Atacama (cl4_JF composite; Figs. 4e and 5e). However, the Bolivian high for this composite is shifted eastward compared to the North Atacama cases (cf. Figs. 6e to 6d), and negative OLR anomalies only occur over the northwestern part of South America

(Fig. 5e). Hence, continental upper-level IWVF between 10° and 25°S is rather oriented from north to south, and strongest positive IWV anomalies are restricted to the Altiplano (Fig. 4e).

Based on the composite fields shown in Figs. 4–6 the most important results can be summarized as follows:

FIG. 7. Composite means of $z900$ (green contours, every 10 gpm) and low-level IWVF (below 800 hPa; black arrows, in $\text{kg m}^{-1} \text{s}^{-1}$), and composite anomalies of IWV (shading, in kg m^{-2}) from ERA-Interim for (a) cl1_MJJA, (b) cl2_JUNE, (c) cl3_JJA, (d) cl3_JF, and (e) cl4_JF. The IWV anomalies are given relative to the climatology over the respective season. Grid boxes above 500 m MSL are masked out.

strong winter rainfall in the Atacama Desert is generally related to low pressure systems over the southeast Pacific. Rainfall in South Atacama (south of 23°S) often comes along with ARs at the foreside of these systems, while strong events in North Atacama (18°–23°S) are mostly associated with COLs occurring anomalously far north. A southward displacement of the Bolivian high during summer accompanied by strong convective activity north and northeast of northern Chile may lead to strong rainfall in NE Atacama. When the Bolivian high is additionally shifted to the west, the enhanced easterly upper-level IWVF component may trigger rainfall in the northernmost Atacama (approximately 18°–20°S).

Interestingly, the onshore moisture transport in low levels (below 800 hPa) seems to play only a minor role for the analyzed rainfall events, both for the summer and winter (Fig. 7). For the winter composites in North Atacama and SE Atacama (cl3_JJA and cl2_JUNE), low-level IWVF is considerably weak along the Atacama coast, albeit a westerly component is visible (Figs. 7b,c). Only during the winter rainfall events in South Atacama (cl1_MJJA) a remarkable northwesterly IWVF occurs, as a result of a westerly shifted near-surface anticyclone and the formation of a low-level coastal low (Fig. 7a). However, also for this composite, low-level IWVF along the Atacama coast is rather weak when compared to upper-level IWVF (cf. Figs. 4a and 7a). Therefore, we conclude that the moisture transport for the analyzed rainfall events takes place predominantly above the marine boundary layer and is mostly associated with the midtropospheric circulation patterns.

Our results shown in Figs. 4 and 7 are confirmed by radiosonde data from Antofagasta, particularly for the winter rainfall events (see supplemental material). Positive IWV anomalies are found for almost all individual events in both, the radiosoundings and at the ERA-Interim grid box, which is nearest to the location of Antofagasta. Further, for all individual winter rainfall events a northwest–southeast upper-level moisture transport is observed in ERA-Interim as well as in radiosonde data, whereby the fluxes in ERA-Interim are clearly overestimated. Another common feature in both datasets is that the low-level IWFV is substantially weaker than the upper-level moisture fluxes.

3) BACKWARD TRAJECTORIES AND SST

We now analyze the origin of the air masses associated with the rainfall events. With this aim, 10-day backward trajectories ending at 23°S, 70°W (coastal range of the Atacama) and in four different altitudes [2000, 3000, 4000, and 5000 m above ground level (AGL)] are determined for the winter composites (cl1_MJJA, cl2_JUNE,

and cl3_JJA). For each composite, the trajectories are computed backward from the 10 dates of the selected rainfall events, thus yielding altogether 40 trajectories per composite. For the summer composites (cl3_JF and cl4_JF), backward trajectories arriving at 19°S, 69°W (NE Atacama) are analyzed. As the rainfall events for cl3_JF and cl4_JF are associated with the upper-tropospheric Bolivian high, we only consider the trajectories arriving at high altitudes (4000 and 5000 m), thus resulting in 20 backward trajectories per composite.

For the winter rainfall events in South Atacama (cl1_MJJA) two major source regions are identified. In some cases, the air masses involved in the rainfall events come from the South Pacific, as expected, but most of the trajectories originate from the South American tropics (Fig. 8a), with paths similar for all altitudes: air parcels pass over the Andes from east to west in tropical latitudes, and then follow the coast line toward northern Chile. The backward trajectories of the winter rainfall events in SE Atacama and in North Atacama (cl2_JUNE and cl3_JJA) are similar to that of cl1_MJJA, with air masses coming slightly more frequently from the open Pacific Ocean (Figs. 8c,e). SST composite anomalies indicate that the winter rainfall events in South Atacama are often accompanied by warmer SSTs (Fig. 8b). In particular at the tropical west coast of South America and in the adjacent Pacific warm SST anomalies occur prior to the majority of the analyzed rainfall events, such that the air masses have long distances to travel over warm ocean surfaces before they hit the coast of northern Chile. In contrast, SSTs only plays a minor role for the analyzed winter rainfall events in North Atacama and SE Atacama, as mostly cold SST anomalies are found for these composites (Figs. 8d,f). Lagged composites (see supplemental material) indicate that the primary moisture sources for these events are tropical continental areas and the tropical southeast Pacific, as well as the remote subtropical Pacific, which might explain why positive SST anomalies in the adjacent ocean are not necessarily required to trigger these rainfall events. Nevertheless, winter rainfall events in South Atacama, which are associated with enhanced SST, typically affect a broader area and yield higher precipitation amounts on the order of a magnitude compared to the other winter rainfall cluster cl2_June and cl3_JJA (Fig. 3), indicating the role of the regional SST.

Backward trajectories for the summer rainfall events in North Atacama (cl3_JF) are rather complex (Fig. 8g). The computed trajectory paths cover the midlatitude South Pacific, the coastal zone of central and northern Chile, Cape Horn, and the Amazon Basin. However, a common feature of the majority of the trajectories is that at their end they penetrate into the target area from the

FIG. 8. (left) ERA-Interim 10-day backward trajectories as obtained by the HYSPLIT model and (right) composite anomalies of SST from ERA-Interim (in K) for (a),(b) cl1_MJJA; (c),(d) cl2_JUNE; (e),(f) cl3_JJA; (g),(h) cl3_JF; and (i),(j) cl4_JF. In (a), (c), and (e) trajectories ending at 23°S, 70°W at altitudes of 2000 (red), 3000 (green), 4000 (light blue), and 5000 m (blue) AGL are shown, and in (g) and (i) trajectories ending at 19°S, 69°W at altitudes of 4000 (light blue) and 5000 m (blue) AGL. (b),(d),(f),(h),(j) The composite anomalies are computed for the SST averaged over the 7 days prior the events relative to the climatology over the respective month, and the black contours represent the isoline for 5 out of 10 events showing a positive SST anomaly larger than +0.5 K.

region east of the Altiplano. The air masses involved in the summer rainfall events in NE Atacama (cl4_JF), on contrary, mostly originate from regions east or northeast of the Altiplano, and in some cases even from the Atlantic (Fig. 8i). Further, while the SST composite anomalies of cl3_JF are negative (Fig. 8h), some of the cl4_JF events are associated with warmer than normal SSTs over the subtropical southeast Pacific (Fig. 8j).

c. Case studies

Regional processes responsible for rainfall events in the Atacama Desert, such as the local circulation, moisture anomalies, and the development of clouds, are investigated here. As these processes may strongly differ between and within the five composites, only some exemplary rainfall events are analyzed in detail (see Table 1). One event each is selected for winter rainfall in South Atacama (cl1_MJJA), in SE Atacama (cl2_JUNE), and in North Atacama (cl3_JJA), and for summer rainfall in North Atacama (cl3_JF). Two events for summer rainfall in NE Atacama (cl4_JF) are examined. The aim is to demonstrate the complexity of the events and the diversity of the synoptic and regional processes contributing to rainfall in different regions of the Atacama Desert.

The winter rainfall event in South Atacama (cl1_MJJA) is associated with both an AR and a COL. For SE Atacama, we selected a winter event, for which rainfall is mainly restricted to SE Atacama, in order to delimit it from the cl1_MJJA events. A winter rainfall event with a COL occurring extraordinarily far north is analyzed for North Atacama. To ensure a better comparability for the summer rainfall cases (cl3_JF and cl4_JF), we selected particular events from these composites that are both associated with nocturnal rainfall. Additionally, a second event is selected for NE Atacama to illustrate the impact of the typical afternoon circulation in the Atacama Desert on rainfall in this region. Aside from the synoptic conditions as obtained from ERA-Interim, the local circulation, moisture anomalies and the development of clouds as simulated by the WRF Model are examined. As a proxy for clouds, we use simulated cloud liquid water content at all individual vertical model levels (variable QCLOUD of the WRF Model output).

The strongest rainfall of the selected winter rainfall event in South Atacama (cl1_MJJA, 1 July 1983) is concentrated in the region south of 23°S (Fig. 9a). At day-1, a distinct COL is located over the southeast Pacific (Fig. 10a), which approaches northern Chile at day 0 (Fig. 10b). At the foreshore an AR is visible, which transports moisture in upper levels (above 800 hPa) toward northern Chile. As a consequence, anomalous moist air is lying over South Atacama in

heights up to 7 km (see longitude–height cross sections at 24°S in Figs. 10c–f). Embedded in the AR, clouds form already offshore over the Pacific (Fig. 10d), which cross the Atacama Desert from west to east due to the strong upper-level zonal wind component (Figs. 10e,f). This is an explanation for the rainfall pattern shown in Fig. 9a, which covers South Atacama from the coast to the SE Atacama. After the system crossed the Central Valley, dry air is advected from west (Fig. 10f).

For the winter event in SE Atacama (cl2_JUNE, 2 June 2006), the strongest rainfall is simulated for the eastern part of the cluster 2 domain (Fig. 9b). It is related to a trough approaching northern Chile (Figs. 11a,b). Upper-level IWVF and also the positive IWV anomaly are less strong over northern Chile when compared to the winter event in South Atacama. As the vertical movement associated with the trough is coincidentally rather weak, no clouds are simulated over the Pacific and the Atacama Desert west of 69.5°S (as shown for 25°S in Figs. 11c–e). However, toward the Andean slopes the moist air raises and clouds form over the SE Atacama in a height of 6–7 km MSL, leading to rainfall being restricted to the elevated easterly region (Fig. 9b). In contrast to the South Atacama event, orographic effects thus seem to be crucial for rainfall in this event.

A quasi-stationary COL is located off the coast of northern Chile at day 1 and day 0 of the winter rainfall event in North Atacama (cl3_JJA, 10 August 1997; Fig. 12). The COL is shifted to northeast when compared to the South Atacama event (cf. Figs. 12a,b to Figs. 10a,b), such that upper-level IWVF penetrates into the Atacama Desert only in the northernmost part, while the moisture transport is almost parallel to the coast south of 25°S (Fig. 12b). Consequently, positive IWV anomalies are strongest over North Atacama. Interestingly, moist IWV anomalies are restricted to the COL domain, encircled by anomalous dry air. The local circulation at 22°S is dominated by a very weak zonal component and unevenly distributed updrafts embedded in the COL (Figs. 12c–f), resulting in a rather patchy cloud pattern over the Pacific and the Central Valley (Fig. 12e). This is mirrored by the heterogeneous simulated rainfall pattern for this event, which is characterized by a rainband stretching in south-southeasterly direction and additional rainfall peaks west of it (Fig. 9c).

Prior to and during the summer event in North Atacama (cl3_JF, 2 January 1997), a pronounced anticyclone in 500 hPa is centered over northern Chile (Figs. 13a,b), which can be attributed to a southwestward displaced Bolivian high (not shown). As a result, a strong easterly IWVF is observed over North Atacama at day 0, resulting in moist IWV anomalies (Fig. 13b). The longitude–height cross section at 19.3°S (Figs. 13c–f)

FIG. 9. WRF-simulated daily rainfall (in mm day^{-1} ; grid boxes with no rain are masked out) for the rainfall events selected for the case studies. (a) 1 Jul 1983 of the cl1_MJJA composite, (b) 2 Jun 2006 of the cl2_JUNE composite, (c) 10 Aug 1997 of the cl3_JJA composite, (d) 2 Jan 1997 of the cl3_JF composite, (e) 3 Feb 1990 of the cl4_JF composite, and (f) 14 Jan 2006 of the cl4_JF composite. The respective cluster domains are marked by the bold black lines (see partitioning in Fig. 2a), and the black dashed lines represent the locations of the longitude–height cross sections shown in Figs. 10–15. Note the different scaling of the color bars.

FIG. 10. Composite means of $z500$ (green contours, every 20 gpm) and upper-level IWVF (black arrows, in $\text{kg m}^{-1} \text{s}^{-1}$) and composite anomalies of IWV (shading, in kg m^{-2}) from ERA-Interim for (a) day 1 and (b) day 0 of the 1 Jul 1983 event of the cl1_MJJ composite (South Atacama). The IWV anomalies are given relative to the climatology for July. (c)–(f) Longitude–height cross sections at 24°S (cf. Fig. 9a) of WRF-simulated u – w wind components (black arrows, in m s^{-1} ; w component has been multiplied by factor of 30), cloud liquid water content (contours, every 2 g kg^{-1}), and anomaly of specific humidity (shading, in g kg^{-1}) at 0800, 1400, 1800, and 2300 UTC of day 0. The specific humidity anomalies are given relative to the climatology over the respective hour of all July months. Height on the y axis is given in km.

reveals that the upper-level easterlies transport moist air westward, such that positive humidity anomalies are found in altitudes above 4 km, while the air masses below are anomalous dry (Fig. 13c). Over the Altiplano plateau, strong updrafts are simulated at 0000 UTC, which lead to the formation of deep convective clouds over this domain. Due to the strong upper-level easterly winds, fragments of these deep clouds are released and pushed westward (Fig. 13d), such that they traverse the North Atacama within a few hours (Figs. 13e and 13f for 0200 and 0300 UTC, respectively). Thus, the resulting rainfall pattern exhibits an east–west gradient, with strongest precipitation over NE Atacama and decreased rainfall amounts toward the coast (Fig. 9d).

The processes involved in the first analyzed summer event in NE Atacama (cl4_JF , 3 February 1990; Fig. 14) generally resemble those for North Atacama. For example, an anticyclone lies over northern Chile and strong nighttime convection is simulated over the Altiplano.

However, the midlevel anticyclone (Figs. 14a,b) and the corresponding Bolivian high (not shown) are shifted eastward when compared to the North Atacama event. Consequently, IWVF and particular its easterly component is much weaker at day 0 (cf. Figs. 14b to 13b), and only weak upper-level easterlies occur over the Atacama Desert (Figs. 14c–f). The westward transport of clouds, which is found for the North Atacama event, therefore fails to appear for here. Accordingly, notable rainfall is restricted mostly to higher altitudes toward SE Atacama during this event (Fig. 9e).

In the second summer event in NE Atacama (cl4_JF , 14 January 2006; Fig. 15), the local circulation as visible in the cross sections clearly differs from that of the 3 February 1990 event (cf. Figs. 15c–f to 14c–f). In the upper levels, moisture is transported from continental areas into the target area (Figs. 15c,d). At the western slopes of the Andes an upslope flow is revealed for the afternoon hours that transports anomalous moist air

FIG. 11. As in Fig. 10, but for the 2 Jun 2006 event of the *cl2_JUNE* composite (SE Atacama) at (c)–(f) 0200, 0600, 0800, and 1400 UTC of day 0 at 25°S.

masses from low levels toward the elevated NE Atacama (Figs. 15d,e). The upslope flow is the consequence of the strong heating of the western slopes of the Andes during afternoon, also known as Andean pumping (Rutllant et al. 2013). When these air masses interact with the moist air in high altitudes, clouds form solely over the elevated hillsides (Figs. 15d,e). Accordingly, notable rainfall is restricted exclusively to the SE Atacama in this event (Fig. 9f).

In summary, the location of rainfall in the Atacama Desert is mainly controlled by the geographical location of the relevant synoptic features (troughs and COLs in winter, Bolivian high in summer). These synoptic patterns not only steer the upper-level moisture flux into the Atacama Desert, but they also modify the local circulation in a manner that the ascent of moist air masses and thus condensation is facilitated. Note that the low-level IWVF is generally weak during the analyzed winter events (not shown), as already demonstrated for the composites in section 3b(2).

4. Conclusions and discussion

In this study rainfall events in the Atacama Desert are analyzed with focus on the region between 18° and 26°S

below 3000 m MSL (thus including the hyper-arid core). Synoptic and regional processes involved in such events are investigated. Aside from reanalysis data, we use, for the first time, a highly resolved long-term simulation with WRF for the analysis.

Some of the results found in this study confirm what is already known from previous studies: winter rainfall in the Central Valley and at coastal regions of the Atacama Desert are associated with low pressure systems (troughs and COLs) over the southeast Pacific (e.g., Montecinos and Aceituno 2003; Bozkurt et al. 2016), while summer rainfall in NE Atacama is controlled by a southward displaced Bolivian high and associated upper-level easterly winds (e.g., Garreaud et al. 2003; Meseguer-Ruiz et al. 2019).

Beyond these results our study also delivers new insights, which may be very helpful for various purposes:

- (i) Strong winter rainfall events in North Atacama (north of 23°S) are often associated with COLs occurring anomalously far north off the coast of the Atacama Desert. Further, anomalous warm SSTs are not necessary for triggering these events. In contrast, strong winter rainfall events in South Atacama

FIG. 12. As in Fig. 10, but for the 10 Aug 1997 event of the cl3_JJA composite (North Atacama) at (c)–(f) 0100, 0300, 0400, and 0700 UTC of day 0 at 22°S.

are linked to strong troughs and COLs occurring farther south over the subtropical southeast Pacific, and are often accompanied by warm SST anomalies over the tropical and subtropical near-coastal ocean.

- (ii) Case studies indicate that strong summer rainfall in North Atacama is highly sensitive to slight shifts in the geographical location of the Bolivian high. In case the northern periphery of the Bolivian high is located over northern Chile, anomalously strong upper-level easterly winds may push convective systems from the high Altiplano toward the coast, thus leading to precipitation over the northernmost hyper-arid core of the Atacama Desert. In contrast, when the center of the Bolivian high is located farther northeast, rainfall is often restricted to NE Atacama.
- (iii) The synoptic circulation patterns associated with strong rainfall not only control the large-scale moisture transport to the Atacama Desert, but they also disturb the predominant local circulation such that moist air is locally lifted and condensation is initiated. With respect to winter precipitation, orographic effects only seem to play a role for rainfall in the SE Atacama.
- (iv) Backward trajectories reveal some unexpected results for winter rainfall events associated with low pressure systems over the southeast Pacific: The involved air masses not necessarily come from the open Pacific Ocean, but in many cases they originate from the tropics in South America, passing over the Andes from east to west between the equator and 10°S, and then follow the coast line southward toward northern Chile. Some outliers of the backward trajectories of the winter composites (particularly those ending in lower levels) look similar to some trajectories identified for the summer events that are associated with the Bolivian high. In case of the winter composites these are the cases with a strong subtropical high extending deep into the troposphere, such that air masses at lower levels are steered counterclockwise around the anticyclone (not shown). In case of the summer composites the paths of the respective trajectories are determined by a Bolivian high shifted extraordinarily far to the west, thus following a similar route.
- (v) As for other regions of the world, atmospheric rivers also play a role in rainfall events in the Atacama Desert, particularly during winter and for

FIG. 13. As in Fig. 10, but for the 2 Jan 1997 event of the *cl3_JF* composite (North Atacama). (c)–(f) 0000, 0100, 0200, and 0300 UTC of day 0 at 19.3°S.

the region south of 23°S. They occur on the foreside of strong troughs or COLs over the subtropical southeast Pacific.

Our results may, for instance, contribute to a reasonable interpretation of precipitation archives for the Atacama Desert (e.g., Ritter et al. 2019; Diederich et al. 2020). Further, they are also relevant for social and economic aspects, as the Atacama Desert includes several mining companies. For the Atacama Salt Flat, lithium mining operations were estimated to have expanded from 20.54 to 80.53 km² in the period 1997–2017 (Liu et al. 2019), and extreme rainfall may have strong impacts on mining activities (Gonzalez et al. 2019).

In our study we have used traditional thresholded clustering methods to identify simulated rainfall classes with similar annual cycles. A shortcoming of these methods is that each predictor can only be assigned to one cluster, which may result in some inconsistencies. For example, cluster 3 not only contains grid boxes with peak rainfall in summer and winter, but also annual cycles with rainfall amounts close to zero throughout the year (cf. Fig. 2). Further, also some coastal grid boxes located south of 23°S are assigned to this cluster.

These issues could probably be overcome by using normalized annual cycles as predictors. However, as we are not only interested in the shape of the annual cycles but also in the rainfall amplitudes, we decided to use the original time series for the clustering. Based on the clustering, we have focused on the core winter and summer season in this study. Nevertheless, strong rainfall events may also occur in the transition months. Hence, future studies should also consider events occurring in other seasons, as it is not clear whether other processes than those detected here are relevant for these events.

As our focus was on synoptic and regional processes, the impact of the MJO or ENSO on WRF-simulated rainfall and the relevant processes controlling these impacts were not considered in this study. However, there is a connection between these Pacific-wide climate oscillations and the synoptic-scale circulation patterns identified in our study. For instance, Solman and Menendez (2002) found that SST anomalies associated with ENSO induce large-scale atmospheric circulation anomalies over large areas of the Southern Hemisphere, which in case of El Niño events result in an equatorward shift of the winter storm track over the subtropical Pacific. This shift favors the formation of troughs over

FIG. 14. As in Fig. 10, but for the 3 Feb 1990 event of the cl4_JF composite (NE Atacama) at (c)–(f) 0000, 0200, 0300, and 0400 UTC of day 0 at 19.3°S.

the subtropical southeast Pacific, which are associated with strong winter rainfall events in South Atacama. Recent studies reveal that synoptic-scale circulation patterns over the Pacific–South America region also vary by the different MJO phases. In phase 6, for example, strong positive geopotential height anomalies are found over the southeast Pacific (Barrett et al. 2012), which force upper-level low pressure systems anomalous far northward. In our study we found that the latter is connected with strong winter rainfall events in North Atacama. Further, intense El Niño events are often accompanied by anomalous warm coastal SST (e.g., Bozkurt et al. 2016), while the MJO modulates the position and the strength of the subtropical anticyclone (Barrett et al. 2012), thus affecting surface winds and the upwelling of cold water. As we found that strong rainfall events in South Atacama are favored by positive coastal SST anomalies, these connections are also important for the processes studied here.

Several recent studies highlighted the important role of northerly winds and meridional warm advection on the strengthening of the coastal El Niño, which in turn may contribute to strong rainfall events in the Atacama Desert (e.g., Bozkurt et al. 2016). For the winter rainfall

events in South Atacama (cl1_MJJA), we found that the midtropospheric troughs and COLs associated with these rainfall events may generate northerly 10 m wind anomalies off the Atacama coast (not shown), which might explain the positive SST anomalies in the adjacent ocean (see Fig. 8b). However, the findings from Takahashi and Martínez (2017) indicate that northerly wind anomalies must prevail for longer periods to lastingly relax coastal upwelling and increase coastal SSTs, and it is therefore likely that single synoptic disturbances such as the troughs and COLs analyzed here are insufficient to accumulate warm SST. This might also be an explanation for the fact that negative SST anomalies are found for the winter rainfall events in SE Atacama (cl2_JUNE), despite the large midtropospheric dynamic similarities to the South Atacama events. In future studies this scientific issue should be addressed by analyzing ocean dynamics in more detail. The differences in the composited SST anomalies between cl1_MJJA and cl2_JUNE, particularly along the Peruvian coast, are clearly reflected in the rainfall amounts of the two composites: composite means of WRF-simulated rainfall are much lower in cl2_JUNE when compared to cl1_MJJA (cf. Figs. 3a and 3b).

FIG. 15. As in Fig. 10, but for the 14 Jan 2006 event of the cl4_JF composite (NE Atacama) at (c)–(f) 1400, 1700, 2100, and 2300 UTC of day 0 at 19.3°S.

In this study orographic impacts of the Andes are only addressed for winter precipitation in SE Atacama (cl2_JUNE). Another orographic effect that might be important for rainfall in the Atacama Desert is the development of northerly barrier jets in 2–4 km MSL resulting from the blocking of westerly winds that are forced southward parallel to the Andes (Kalthoff et al. 2002). For Northern California it is demonstrated that such barrier jets associated with the Sierra Nevada may act as a virtual barrier, thus lifting moisture-laden air parcels within ARs and initiating or increasing precipitation west of the mountains (Kingsmill et al. 2013). It remains unclear whether the ARs, which are identified as important synoptic-scale features for winter rainfall in the Atacama Desert in our study, interact with the Andean barrier jet.

The long-term WRF simulation used in this study has a horizontal resolution of 10 km. While evidence is found that such a resolution is sufficient to capture important climate gradients imposed by complex topography in southwest South America (Bozkurt et al. 2019), convective storms over the Andes Cordillera, which may contribute to rainfall over the northern part of the Atacama Desert in summer, are not explicitly resolved.

For future studies it would thus be important to perform RCM experiments with a convective-permitting horizontal resolution (<3 km). Furthermore, we selected only one or two events per composite (altogether six cases) for the case studies, although it turned out that the individual events within a composite may strongly differ (not shown). Hence, detailed analyses of further rainfall events would be worthwhile. For additional studies dealing with rainfall in the Atacama Desert we made the WRF-simulated rainfall dataset freely available for the research community (<https://www.crc1211db.uni-koeln.de/data.php?dataID=38>; Reyers 2018).

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Data availability statement. ERA-Interim data can be downloaded on <https://apps.ecmwf.int/datasets/data/interim-full-daily/>. A registration is required. The global atmospheric river (AR) database created by Guan and Waliser (2015) is freely available via <https://ucla.app.box.com/v/arcatalog/folder/16397833893>. WRF-simulated precipitation with a horizontal resolution of 10 km (Reyers 2018) is freely available via the CRC1211 database (<https://www.crc1211db.uni-koeln.de/data.php?dataID=38>). Further output variables of this WRF simulation will be provided on request.

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