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Review of Tropical-Extratropical Teleconnections on Intraseasonal Time Scales

Cristiana Stan[1], David M. Straus[1], Jorgen S. Frederiksen[2], Hai Lin[3], Eric D. Maloney[4], and Courtney Schumacher[5]

[1]Department of Atmospheric, Oceanic and Earth Sciences and Center for Ocean-Land-Atmospheric Studies, George Mason University, Fairfax, VA, USA, [2]CSIRO Oceans and Atmosphere, Aspendale, Victoria, Australia, [3]Atmospheric Numerical Weather Prediction Research, Environment and Climate Change Canada, Dorval, Quebec, Canada, [4]Department of Atmospheric Science, Colorado State University, Fort Collins, CO, USA, [5]Department of Atmospheric Sciences, Texas A&M University, College Station, TX, USA

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- Teleconnections on intraseasonal time scales
- MJO, NAO, PNA, monsoon, and tropical convection
- Rossby wave breaking and high-impact weather

Correspondence to:
C. Stan,
cstan@gmu.edu

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Abstract The interactions and teleconnections between the tropical and midlatitude regions on intraseasonal time scales are an important modulator of tropical and extratropical circulation anomalies and their associated weather patterns. These interactions arise due to the impact of the tropics on the extratropics, the impact of the midlatitudes on the tropics, and two-way interactions between the regions. Observational evidence, as well as theoretical studies with models of complexity ranging from the linear barotropic framework to intricate Earth system models, suggest the involvement of a myriad of processes and mechanisms in generating and maintaining these interconnections. At this stage, our understanding of these teleconnections is primarily a collection of concepts; a comprehensive theoretical framework has yet to be established. These intraseasonal teleconnections are increasingly recognized as an untapped source of potential subseasonal predictability. However, the complexity and diversity of mechanisms associated with these teleconnections, along with the lack of a conceptual framework to relate them, prevent this potential predictability from being translated into realized forecast skill. This review synthesizes our progress in understanding the observed characteristics of intraseasonal tropical-extratropical interactions and their associated mechanisms, identifies the significant gaps in this understanding, and recommends new research endeavors to address the remaining challenges.

1. Introduction

The study of the general circulation of the atmosphere has been traditionally carried out by considering the regions of the tropics, midlatitudes, and high latitudes as separate components, with distinct dynamics and sources of variability. The mutual interactions between these components in the atmosphere and ocean on the seasonal and longer time scales have been well recognized (see the review of Liu & Alexander, 2007). Their perspective is that "teleconnections enable the atmosphere to act like a *bridge* between the different parts of the ocean and enable the ocean to act like a *tunnel* linking different atmospheric regions." These teleconnections act to restore the imbalances in the energy budget of the climate system caused by the meridional distribution of solar insolation, sea surface temperature (SST) anomalies associated with the El Niño–Southern Oscillation (ENSO), and internal variability of the climate system associated with midlatitude storms and gyre circulations. The interactions occur in well-defined spatial patterns known as climatic teleconnection patterns and manifest as simultaneous variations in weather and climate over remote locations on Earth.

In the Northern Hemisphere (NH) the climatic teleconnection patterns include the North Atlantic Oscillation (NAO) (Walker & Bliss, 1932); the circumpolar waveguide pattern, which projects onto NAO with a correlation coefficient of covariability larger than 0.5 (Branstator, 2002); the East Atlantic, West Pacific, East Pacific-North Pacific, Polar/Eurasian, and Pacific North-America (PNA) patterns (Wallace & Gutzler, 1981); Eurasia-1 and Eurasia-2 patterns (Barnston & Livezey, 1987); the Tropical/Northern Hemisphere pattern (Mo & Livezey, 1986); and the East Asia-Pacific pattern (Nitta, 1987). A comprehensive description of NH winter teleconnection patterns is given by Panagiotopoulos et al. (2002). In the Southern Hemisphere, only two climatic teleconnection patterns have been identified: the Pacific-South American and South Pacific Wave patterns (Mo, 2000).

Some of these climatic teleconnection patterns also exhibit variability on subseasonal time scales (Feldstein, 2003; Murakami, 1988; Wang, Wen, et al., 2016). Thus, in recent years there has been a greater appreciation of the importance of the two-way interactions between the tropics and the midlatitude and high latitude *on intraseasonal time scales* of 10 to 100 days. The higher-frequency intraseasonal teleconnections draw their energy from the atmospheric variability associated with organized convection of the tropics, extratropical cyclones of the midlatitudes, and their interactions. The long time scales in the "ocean tunnel" can only affect the mean state on which the intraseasonal teleconnections develop, so the ocean may not provide the first order driving mechanisms for intraseasonal variability. Examples of atmospheric mechanisms are upper level divergence in the tropics resulting from tropical convection, the interaction of the resulting divergent flow with the basic state vorticity gradient (known as the Rossby wave source), and midlatitude baroclinic and barotropic instabilities. All of these will be discussed in this paper.

The role of the ocean as a driver of tropical convective disturbances, and in particular those associated with the Madden Julian Oscillation (MJO) (Madden & Julian, 1971), is not well understood (DeMott et al., 2016). The role of atmosphere-ocean coupling in the life cycle of the Asian summer monsoon intraseasonal oscillations is also uncertain. While some studies find a minor influence (Belon et al., 2008), newer studies using state of the art climate models report a larger impact (Sharmila et al., 2013) of the atmosphere-ocean coupling on the simulation of boreal summer intraseasonal variability.

Many intraseasonal tropical-extratropical interactions were discussed in Frederiksen and Webster (1988). This review article will build on that review and will summarize and synthesize the existing work published since then that explores the interactions and teleconnections between the tropics and extratropics on time scales up to a season. These interactions and teleconnections are components of the subseasonal to seasonal (S2S) climate variability, whose forecast has important societal impacts (Vitart et al., 2017).

As described in Frederiksen and Webster (1988), pioneering observational studies of the influence of atmospheric processes in the tropics on the extratropics included those of Riehl (1950), Bjerknes (1966, 1969), and Weickmann (1983). Significant early modeling studies included those of Opsteegh and Van den Dool (1980), Hoskins and Karoly (1981), and Webster (1981, 1982). By analyzing the meanders on daily weather maps, Riehl (1950) noted the distinct communication between the tropics and extratropics that occurs as trains of vortices associated with tropical disturbances interleave themselves into the circulation at higher latitudes, helping to alter flow and form extended troughs in the extratropics. Bjerknes showed that the variability associated with tropical ocean temperature anomalies and associated convective activity has a global-scale influence and acts as a thermal source for barotropic and baroclinic circulation anomalies at higher latitudes. In the baroclinic model of Hoskins and Karoly (1981), the time mean midlatitude response to stationary tropical forcing consists of a wave train extending poleward and eastward from the source region, especially in the upper troposphere. The observational study of Weickmann (1983) confirmed the theoretical predictions of Hoskins and Karoly (1981) by uncovering a *close* relationship between the intraseasonal variability of the tropical diabatic heating and the midlatitude winter standing oscillations, which exhibit maximum variability in the region of midlatitude jets. However, simple linear steady state theory does not allow for the inherent instability of the atmospheric basic state and predicts that the Rossby wave train will change its location following the heat source as it moves eastward during equatorial warm episodes. This is contrary to the general circulation model (GCM) simulations of Keshavamurty (1982) and Geisler et al. (1985) and the observational study of Dole (1986), which showed that quite similar high-latitude responses occur for different locations of equatorial heating. The reason for this is that the response is highly influenced by large-scale atmospheric instabilities. The relevant instabilities have both baroclinic and barotropic natures (e.g., Frederiksen, 1982, 1983). Baroclinic instabilities lead to organized storm tracks, whose response to tropical forcing alters the time mean response (Held et al., 1989), while barotropic instabilities of wavy basic states can "flip" the flow from one equilibrium state to another (Wu, 1993) or grow through various mechanisms (Franzke, 2002).

A significant limitation of much of the early modeling work is the treatment of the midlatitude basic state, which is often considered zonally symmetric (i.e., independent of longitude). Lorenz (1972) showed in general that a zonally varying basic state can generate growing barotropic disturbances. The role played by such barotropic disturbances in modulating the midlatitude response to tropical forcing is addressed in a numerical study with a barotropic model by Simmons et al. (1983), who showed that the midlatitude response can grow significantly by extracting energy from the mean flow. A dependence of the tropics-midlatitude interactions on the climatology of the midlatitude flow was later found to exist in observations by Lau and Phillips (1986).

The primary purpose of this review article is to provide the reader with a comprehensive background for understanding current (i.e., post Frederiksen & Webster, 1988) knowledge on intraseasonal tropical-extratropical interactions, identify gaps in this body of knowledge, and highlight the need for new research on the S2S time scales. This review also builds on Roundy (2012) and Zhang (2013), who provided more recent reviews of intraseasonal tropical-extratropical interactions and global MJO impacts.

The remainder of this paper is organized as follows. Section 2 presents the observational and modeling perspectives of intraseasonal teleconnections, describing *how and where* these interactions occur. Throughout the paper "observations" refer to reanalysis data as well as traditional observations (satellite, weather stations, etc.) without making any distinction between them. Section 3 summarizes the sources of tropical forcing and follows the chronological timeline of theories developed to explain the mechanisms driving the interactions, teleconnections, and feedback between the tropics and midlatitudes. The goal is to understand *why* the teleconnections arise. Because the boundary between modeling and studies focused on physical mechanisms is not distinct, some overlap in the material discussed in sections 2 and 3 is inevitable. Section 4 encapsulates the predictability of intraseasonal teleconnections and their impact on S2S prediction skill. Section 5 poses a set of frontier research topics for overcoming the remaining challenges.

2. Observational and Modeling Perspectives

Statistical methods applied to observations led Ghil and Mo (1991a) to classify the intraseasonal variability (ISV) of the atmosphere into two broad categories: persistent, geographically fixed anomalies, such as blocking anticyclones, which tend to recur in preferred locations, and waves with broad spectral peaks in frequency that can propagate or exhibit oscillation in place. Intraseasonal variability exists in both the tropics and extratropics, and these variabilities interact through a variety of complex mechanisms.

The analysis of observations and modeling studies show that ISV of the tropics has both seasonal and regional dependencies. In observations, the boreal winter is dominated by the MJO, which is characterized by an envelope of organized convection confined to the equatorial region that propagates eastward around the globe at 5 m s^{-1} (Zhang, 2005) and the 30–50 day oscillation of the Australian summer monsoon (Hendon & Leibmann, 1990; Wheeler & McBride, 2005). During the boreal summer MJO activity weakens and the tropical ISV is dominated by the oscillatory modes associated with regional monsoonal systems (e.g., Annamalai & Slingo, 2001; Krishnamurti & Bhalme, 1976; Yasunari, 1979). Over most of the monsoonal regions, the boreal summer ISV exhibits northward propagating modes whereas in some regions westward-propagating modes occur (Chang et al., 1996; Hsu et al., 2004; Wang & Rui, 1990).

In observations, the ISV of the midlatitudes is mostly dominated by regional, persistent anomalies (e.g., Ghil & Mo, 1991a, 1991b; Krishnamurti & Gadgil, 1985; Kushnir, 1987). A similar variability was simulated in numerical experiments with GCMs (Branstator, 1987; Kushnir, 1987). While the study of the component of ISV related to oscillations in midlatitudes is almost uncharted territory, there has been some work in this area. This variability of the Northern Hemisphere's midlatitudes is influenced by two dominant oscillatory hemispheric wide modes with periods of 23 days (e.g., Branstator, 1987; Kushnir, 1987) and 45 or 48 days (Ghil & Mo, 1991a; Krishnamurti & Gadgil, 1985). The 23 day mode prevails in the western half of the Northern Hemisphere, consists of zonal wave numbers 1 and 2, and propagates westward. The 48 day mode has a standing component with the centers of action in the Pacific at about 150 W and in the Euro-Atlantic sector, with eastward and westward propagating components.

There are many features of these modes of variability that deserve detailed discussion, and they will be presented in the following subsections in a descriptive framework organized in chronological order.

The modeling work reviewed in this section includes studies designed to verify the robustness of teleconnections derived from a small sample of observations and test models' ability to reproduce the observations. The modeling work focused on understanding the mechanisms associated with intraseasonal teleconnections will be presented in section 3.

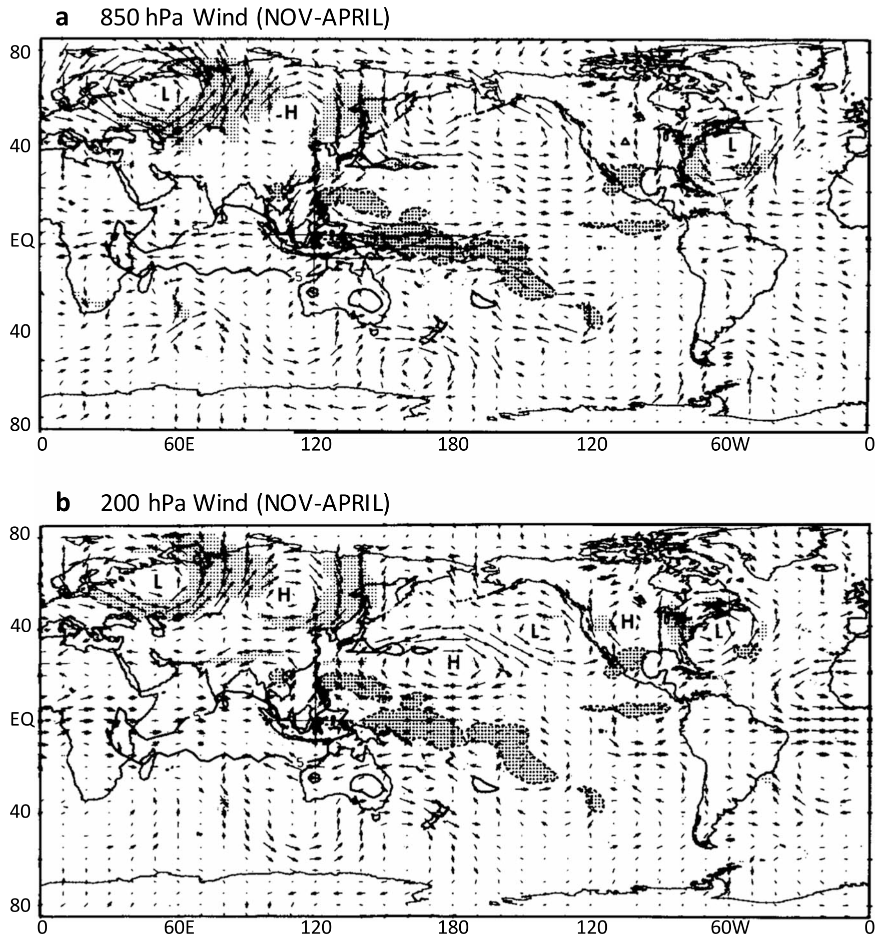

Figure 1. Composites of wind anomalies at (a) 850 hPa and (b) 200 hPa for days when the second principal component of the 30–60 day band-pass filtered 250 hPa velocity potential daily anomalies exceeds 0.85 standard deviation, indicating convection in the western Pacific Ocean. OLR composite anomalies of 5 W m^{-2} are superimposed in dashed contours with dark shading and they indicate decaying convection near the dateline as new convective anomalies are emerging over South America and Africa. The light stipple denotes regions of significant (99%) wind vectors in the extratropics (25–85 N). The cyclonic (L) and anticyclonic (H) circulation features suggest an extratropical barotropic wave train (reprinted from Knutson & Weickmann, 1987, ©American Meteorological Society. Used with permission.)

2.1. Influence of the Tropics on the Extratropics

2.1.1. Modes of Climate Variability

During the boreal winter, the influence of the tropics on the midlatitudes in observations is made manifest as teleconnection patterns linking the MJO and the intraseasonal extensions and contractions of the midlatitude jets (e.g., Hsu, 1996; Knutson & Weickmann, 1987; Lau & Phillips, 1986; Weickmann, 1983; Weickmann et al., 1985).

A series of composite calculations performed by Yuan et al. (2011) showed that a configuration of the North Atlantic jet with enhanced westerlies centered around 40°N leads enhanced Indian Ocean precipitation and lag enhanced western Pacific Ocean precipitation. A jet configuration with two centers of enhanced westerlies (20°N and 60°N) lead enhanced western Pacific Ocean precipitation and lag enhanced Indian Ocean precipitation. These phase relationships manifest through the circumglobal waveguide pattern.

The response seen in numerical models of the midlatitude circulation (e.g., Frederiksen, 1982, 1983; Simmons et al., 1983) to the tropical forcing is a 20–100 day oscillation favored by those phases of the tropical MJO that have convective activity confined west of the dateline. The MJO phases are defined according to Wheeler and Hendon (2004). The midlatitude circulation anomalies precede and follow the tropical convective activity with about a 90° phase difference. The predominant features of the midlatitude wave train are a series of cyclonic and anticyclonic circulations over East Asia and the North Pacific, as seen (for example) by the observed composites of Knutson and Weickmann (1987) (Figure 1). These results are slightly different from

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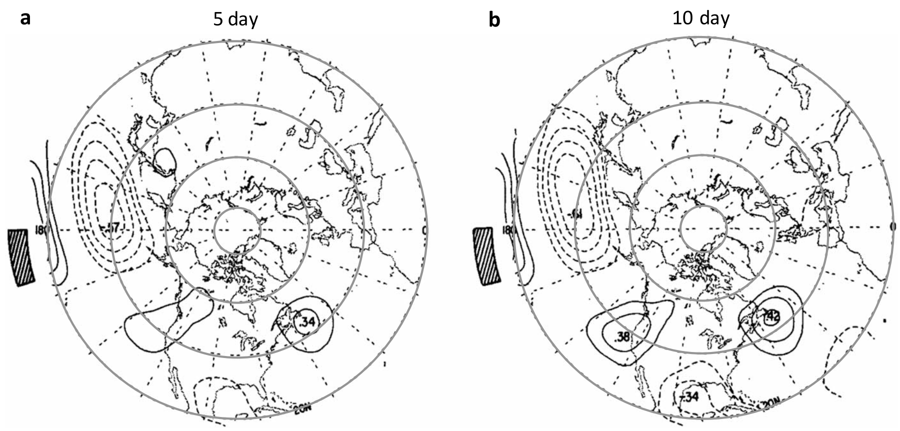

Figure 2. Simultaneous correlation between the area averaged (105 W–180, 10–15 N; stippled area) OLR and grid point 500 hPa geopotential height using (a) 5 day and (b) 10 day mean daily anomalies between 1974 and 1982. The anomalies are deviations from the low-pass filtered daily climatology, however no filtering is applied to the anomalies (reprinted from Liebmann & Hartmann, 1984, ©American Meteorological Society. Used with permission.)

Liebmann and Hartmann's (1984) findings, which suggest that the region of winter monsoon rainfall over the far western Pacific forces the midlatitude hemispheric flow (Figure 2). Most of these results are obtained from linear correlation analysis, which can be limited by the location of the domain chosen for analysis and by the lack of a clear link between cause and effect.

The variability of patterns associated with internal dynamics of the midlatitudes is also linked in observations to tropical convective activity located in the western and central Pacific (e.g., Ferranti et al., 1990; Ghil & Mo, 1991a; Higgins & Mo, 1997; Hsu & Lin, 1992; Schubert & Park, 1991). Ferranti et al. (1990) estimated that 25% of the intraseasonal variability of the geopotential height anomalies associated with the PNA teleconnection pattern is explained by the tropical forcing associated with the MJO and showed that the presence of a convective center east of the Philippines excites the PNA positive phase. The positive phase consists of above normal geopotential height over the western North America and below normal geopotential height over the eastern North America. The negative phase exhibits opposite features (Rodionov & Assel, 2001). The location of the tropical convection that acts as a source varies among various studies. For example, Hsu (1996) suggested that in observations, the PNA pattern is linked to the convective activity over the eastern Indian Ocean and Mori and Watanabe (2008) showed the existence of a phase locking between MJO and PNA. Johnson and Feldstein (2010) and Lukens et al. (2017) came to a similar conclusion. A slightly different conclusion was reached by Riddle et al. (2013), who showed that the MJO excites a PNA-like response rather than a canonical PNA pattern. The PNA pattern influences the subseasonal temperature and precipitation anomalies across North America (Barnston & Livezey, 1987).

Another leading pattern of atmospheric variability, the NAO, influences the weather in the Northern Hemisphere, especially in eastern North America and Europe (e.g., Barnston & Livezey, 1987; Wallace & Gutzler, 1981). The NAO pattern consists of an alternation of atmospheric mass between the North Atlantic regions of subtropical high pressure system (centered near the Azores) and subpolar low surface pressure (extending south and east of Greenland). The intensity of the pressure centers strengthens during the positive phase and weakens during the negative phase (Lamb & Peppler, 1987). Many previous model studies suggested that the primary mechanism for the NAO is the internal dynamics of the extratropical circulation (e.g., Frederiksen, 1982, 1983; Simmons et al., 1983). For example, Limpasuvan and Hartmann (1999) pointed out that the variability of the Northern Hemisphere annular mode, or the Arctic Oscillation (AO) (Thompson & Wallace, 1998), which is well correlated with the NAO (Thompson & Wallace, 1998), is associated with wave-mean flow interactions. Franzke et al. (2004) found that the variability of the NAO is related to midlatitude Rossby wave breaking. These mechanisms, which are related to extratropical atmospheric internal dynamics, imply a lack of predictability for NAO variability on S2S time scales.

However, recent observational studies reveal a robust lagged connection between the MJO and NAO. As discussed in Lin et al. (2009), about 2–3 pentads (a pentad is the average of five consecutive days) after the occurrence of MJO phase 3 (7), the probability of a positive (negative) NAO is increased significantly (Table 1). The phase 3 of MJO corresponds to enhanced convection in the tropical Indian Ocean and

Table 1
Lagged Percentage Probability of the NAO Index With Respect to Each MJO Phase and Lag

Phase	1	2	3	4	5	6	7	8
Lag-5		−35%	−40%			+49%	+49%	
Lag-4						+52%	+46%	
Lag-3		−40%					+46%	
Lag-2						+50%		
Lag-1								
Lag 0				+45%				−42%
Lag + 1			+47%	+45%				−46%
Lag + 2		+47%	+50%	+42%		−41%	−41%	−42%
Lag + 3		+48%				−41%	−48%	
Lag + 4						−39%	−48%	
Lag + 5				−41%				

Note. Lag *n* means that the NAO lags the MJO of the specific phase by *n* pentads, while lag −*n* represents that the NAO leads the MJO by *n* pentads. Positive values are for upper tercil, while negative for low tercil. Values shown are only for those that pass a 0.05 significance level according to a Monte Carlo test. Values greater than 45 are significant at the 0.01 level. Adapted from Lin et al. (2009).

reduced convection in the tropical western Pacific, while phase 7 corresponds to reduced convection over the Indian Ocean and enhanced convection over the western Pacific. Similar results of the MJO influence on the NAO were reported in Cassou (2008), and an analogous modulation of the AO by the MJO was found in Zhou and Miller (2005) and L'Heureux and Higgins (2008). The effect of the speed of propagation of MJO events on the NAO was addressed in a recent observational study of Yadav and Straus (2017), who find that the response of the positive phase of NAO (NAO+) is mainly due to the more slowly evolving episodes. For these episodes (characterized by convection taking over 20 days to propagate from the Indian Ocean to the western Pacific), the peak NAO+ response occurs about 15 days after the occurrence of MJO phase 4, somewhat later in the cycle than in studies which use all MJO episodes for the analysis.

Since the response to tropical heating in any phase of the MJO evolves over a 2 week time scale, the response to a propagating MJO episode will, in general, be the sum of responses to the tropical heating at different locations, with different time lags. To explore the potential role of destructive or constructive interference, Straus et al. (2015) performed mechanistic experiments to determine the effects of the MJO on the North Atlantic variability, using the Community Earth System Model (CESM). MJO-like tropical diabatic heating derived from the Tropical Rainfall Measuring Mission satellite data was *added* to the CESM as it runs: at each time step the four-dimensional (time-varying) MJO heating, encompassing a highly realistic sequence of MJO events, was added to the temperature tendencies produced by the model's dynamics and physical parameterization subroutines. Large ensembles of seasonal simulations (1 October to 31 March) were made with the identical sequence of heating added. Principal component analysis then yielded the midlatitude response to an evolving MJO heating. The leading component modes of total tropical diabatic heating, Rossby wave source, geopotential height, synoptic scale vorticity flux, and storm track kinetic energy show a coherent response to the MJO heating (Figure 3). The evolution of these modes is related to the daily evolution of the state of the North Atlantic circulation, which is assessed independently with a cluster analysis based on unfiltered daily data. In agreement with Lin et al. (2009) and Cassou (2008), the cluster corresponding to the positive phase of the NAO occurs more frequently 5–15 days after the MJO-related convection is over the Indian Ocean (phase 3). Roundy et al. (2010) and Roundy (2012) note that the nature of the MJO teleconnection to the Northern Hemisphere high latitudes is ENSO phase-dependent. In the context of the NAO, La Niña conditions appear to be when the MJO influence on the North Atlantic is strongest.

Unlike the boreal winter interactions, which are supported by a large body of literature, boreal summer teleconnection studies are relatively scarce. Earlier studies based on observations did not find a significant correlation between the extratropics and tropics in boreal summer (Ghil & Mo, 1991a; Knutson & Weickmann, 1987). More recent studies find that during the boreal summer at least 20% of the observed intraseasonal variability of the midlatitudes is driven by the tropical convective activity (Wang et al., 2013). Convection associated with the boreal summer monsoon systems forces an intraseasonal oscillation of the mid-Pacific troughs that propagates eastward at an average speed of 12 m s^{-1} (Chen & Yen, 1991) over the Pacific North-American sector (Moon et al., 2013). In addition, the response of the midlatitudes to boreal summer ISV consists of quasi-stationary patterns over Eurasia and the North American continent (Moon et al., 2013). Lin (2009) found different global teleconnection patterns associated with diabatic heating anomalies of the Indian summer monsoon and the western North Pacific summer monsoon. The Indian summer monsoon convective activity is associated with a global pattern that has a far-reaching impact in both hemispheres, while the western North Pacific summer monsoon variability is linked with a Southern Hemisphere wave train. The Indian summer monsoon intraseasonal variability tends to be connected to a circumglobal teleconnection pattern in the Northern Hemisphere midlatitudes (Ding & Wang, 2007). This variability also plays a substantial role on the initiation of the East Asia-Pacific pattern (Wang, Wen, et al., 2016), along with the convective systems over the South China Sea (Li et al., 2016), and affects the runoff formation of the Amudarya and Syrdarya rivers in central Asia (Schiemann et al., 2007).

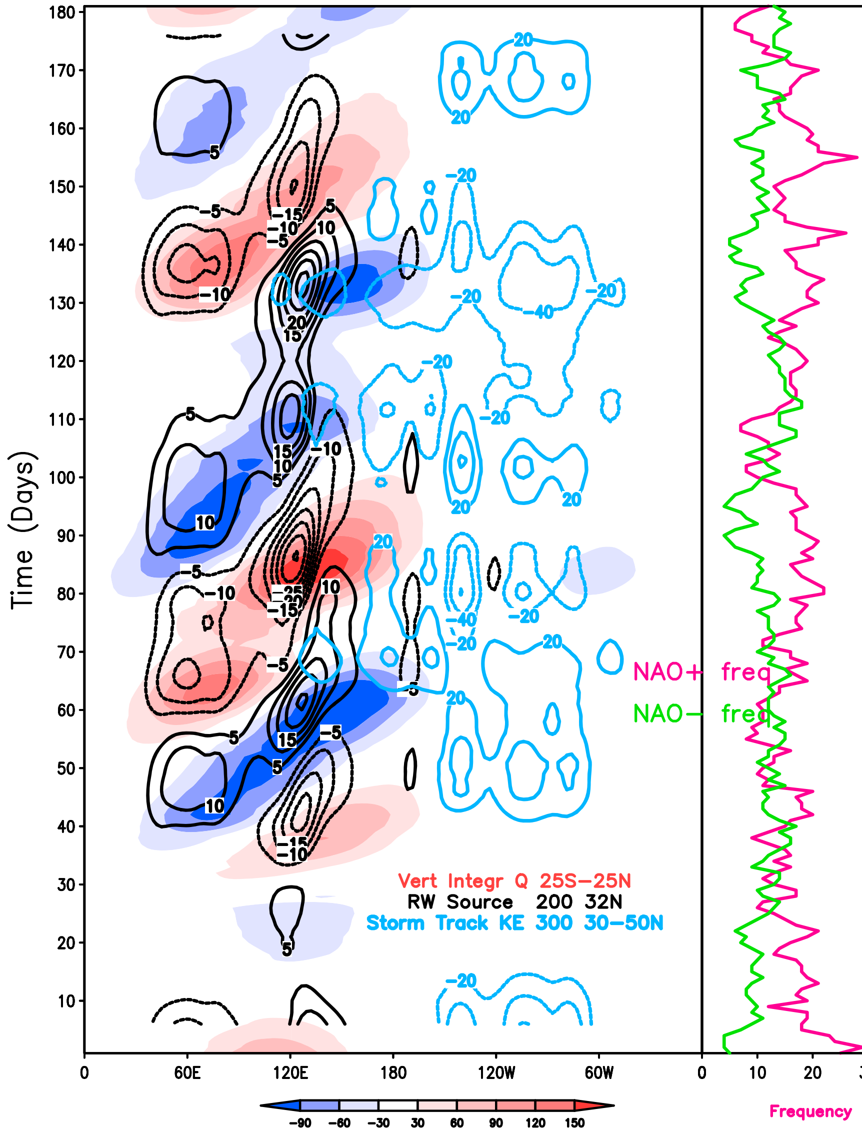

Figure 3. Summary of results of mechanistic experiments described in Straus et al. (2015), in which the identical seasonal evolution of realistic MJO-related diabatic heating was added to each of 50 seasonal simulations, all starting from 01 October initial conditions. The synthesis of the two leading most predictable modes (those with highest signal-to-noise ratio) is summarized. The shading gives the planetary wave component of the vertically integrated diabatic heating averaged from 25 S to 25 N, the black contours the Rossby wave source at 200 hPa and latitude 32 N (interval of 5.0×10^{-11} s^{-1}), and the light blue contours the high-frequency (periods less than 10 days) kinetic energy at 300 hPa, averaged over 30–50 N (interval of 20 m^2 s^{-2}). The red and green curves in the right-hand panel give the daily frequencies of occurrence of the NAO+ and NAO− clusters respectively, with frequencies given as the number of ensemble members (out of 50) for which the particular cluster occurred on that day.

2.1.2. Blocking

Understanding the dynamics and assessing the predictability of blocking anticyclones in the midlatitudes are important for the prediction of extreme events such as cold air outbreaks, drought, and atmospheric river events (e.g., Buehler et al., 2011; Dole et al., 2011; Masato et al., 2012). MJO-induced heating has been demonstrated to significantly perturb geopotential height in the midlatitudes through Rossby wave teleconnections to the tropics in both modeling and observational studies (e.g., Gloeckler & Roundy, 2013; Hoskins & Sardeshmukh, 1987; Matthews et al., 2004). The extratropical impacts of the MJO have been shown to modulate a diverse set of phenomena in the midlatitudes including temperature extremes (Dole et al., 2014), the NAO (e.g., Cassou, 2008; Garfinkel et al., 2014; Lin et al., 2009), and West Coast precipitation events (e.g., Higgins & Mo, 1997), among other impacts. Blocking events are responsible for producing many of these phenomena. Given the ability of the MJO to force persistent flow anomalies in the midlatitudes, blocking events become more or less likely depending on the location of MJO-related tropical convection in the tropics, and hence so do the resulting MJO-related weather impacts.

Hoskins and Sardeshmukh (1987) documented the occurrence of a European blocking event during the winter of 1985 that was likely forced by an MJO event and produced a pronounced cold air outbreak. Cassou (2008) demonstrated that the MJO is associated with Scandinavian blocking during the phase when MJO convection is located just to the east of the Maritime Continent. These blocking anomalies are associated with complex interactions with the NAO in the North Atlantic basin. Moore et al. (2010) describe coherent variations in the extratropical North Pacific among MJO-induced geopotential height and wind anomalies, Rossby wave breaking, and atmospheric blocking, with the tendency for blocking anomalies to propagate eastward across the North Pacific as MJO convection anomalies transit the Maritime Continent region through the west Pacific.

Recent observational studies have provided more comprehensive analyses of the relationship of blocking to the MJO. Hamill and Kiladis (2014) used the 1D blocking index of Tibaldi and Molteni (1990) to show that European blocking frequency shifts from being anomalously low to being anomalously high as MJO convection shifts across the Maritime Continent, consistent with the results of Cassou (2008). Notable North Pacific blocking anomalies associated with the MJO were documented. More recently, Henderson et al. (2016) conducted a comprehensive analysis of Northern Hemisphere blocking as a function of MJO phase in observations using a two-dimensional blocking index based on the work of Masato et al. (2013). A significant decrease in east Pacific and Atlantic blocking activity was shown to accompany MJO phases with enhanced MJO convection in the Indian Ocean and suppressed MJO convection in the west Pacific, with the opposite relationship generally occurring for opposite signed heating anomalies. MJO-related blocking frequency anomalies in the North Atlantic as a function of MJO phase are quite substantial, approaching 100% of the winter mean blocking frequency. MJO-mediated variations in European blocking frequency are generally similar to those found by Cassou (2008). Enhanced European blocking during periods of enhanced MJO convection just east of the Maritime Continent is hypothesized by Henderson et al. (2016) to be associated with the preexisting conditions of an anomalous Atlantic anticyclone and a negative Pacific-North America pattern that is produced by the MJO. Significant central and western North Pacific blocking frequency anomalies occur during all MJO phases. However, Henderson et al. (2017) also showed that climate models have difficulty in simulating the teleconnected geopotential height patterns in midlatitudes that produce blocking, even if they can produce a good MJO. Biases in the Asian-Pacific jet are a prime factor in the inability of models to simulate these MJO teleconnections with fidelity.

2.1.3. Subseasonal Cariations of Precipitation and Near-Surface Air Temperature

The MJO is associated with modulation of temperature and precipitation on a global basis (e.g., Donald et al., 2006; Hoell et al., 2013; Matsueda & Takaya, 2015; Seo & Son, 2012), including extreme events. Observations show that during northern winter seasons the probability of precipitation over the Pacific Northwest and southwestern U.S. increases during the enhanced convective phases of MJO (Becker et al., 2011; Mo, 1999; Zhou et al., 2012) and that the frequency of extreme events is higher when convective anomalies are located over the Indian Ocean (Jones, 2000). The frequency of nor'easter events in the northeastern U.S. has been shown to be modulated by the tropical convection with higher values occurring during phases 7 and 8 of the MJO. During the northern summer and fall, the MJO impact on precipitation over is significant over Mexico and southeastern U.S. (Zhou et al., 2012). Lin, Brunet, and Fontecilla (2010) found that intraseasonal variability of precipitation in Canada is explained by the large-scale circulation anomalies induced by MJO diabatic heating. Precipitation anomalies begin to develop over the west coast of Canada one pentad (5 day average) after the peak convection over the eastern Indian Ocean and continue to extend over South Canada in the next two pentads. During strong MJO events the precipitation anomaly can increase by 20% over the mean seasonal variance.

Following the phase 3 of the MJO, surface air temperature over eastern North America in winter is found to be anomalously warm 10–20 days after (Lin & Brunet, 2009) and above-average surface air temperature appears along the eastern coast of the U.S. (Zhou et al., 2012). Central and northern Alaska experience warm surface air temperatures during MJO phases 2 and 7 and cold temperatures during phase 5 (Vecchi & Bond, 2004).

New England and the Great Lakes region experience below average surface air temperatures when convection is located over the equatorial Indian Ocean (Zhou et al., 2012). The warm anomalies in Alaska and the Arctic during phase 7 appear to be initiated by an enhancement of poleward propagating Rossby waves during previous phases (e.g., phase 5) that causes an enhanced eddy heat flux into the Arctic (Yoo et al., 2012). Seo et al. (2016) showed that warming of surface air temperature in the Northern Hemisphere winter is modulated by convection over the tropical oceans. Rossby waves with zonal wave numbers 2 and 3 forced by convective activity over the Indian and western Pacific oceans influence surface warming over North America, whereas the surface temperature over eastern Europe is modified by Rossby waves with wave number 4.

The intensity and frequency of precipitation over Southeast and East Asia are also modulated by MJO (Aldrian, 2008; Goswami, 2012; He et al., 2011; Hsu, 2012; Jeong et al., 2008; Jia et al., 2011; Wang et al., 2017; Zhang et al., 2009; Zhu et al., 2003) through the remote effects induced by the local tropical circulation in response to the MJO diabatic heating. When the centers of active convection are located over the Indian Ocean (phases 2 and 3), the precipitation rates show a statistically significant increase over most of the regions around 45 N in East Asia. When the MJO convection is active over the western Pacific (phases 6 and 7), these extratropical regions tend to experience a decrease in precipitation.

In addition to modulation of tropical South American rainfall and South Atlantic Convergence Zone (SACZ) by the MJO (e.g., Carvahlo et al., 2004; Souza & Ambrizzi, 2006), South American subtropical and extratropical precipitation also shows MJO modulation (e.g., Zhang, 2013). For example, Barrett et al. (2012) use surface gauge and satellite fields to show that teleconnected Rossby wave responses to Maritime Continent convection cause circulation and precipitation anomalies over Chile during austral winter. Rainier periods are favored when geopotential height is anomalously low to the west of Chile. A number of observational studies (e.g., Alvarez et al., 2016; Jacques-Coper et al., 2015; Juliá et al., 2012) showed a seasonal dependence on the effect of the MJO on South American precipitation, with austral summer characterized by enhanced rainfall in southeastern South America during phases 3 and 4, and June–November characterized by reduced rainfall in coastal regions influenced by the SACZ during phases 4 and 5.

The surface air temperature over the extratropical regions of the South American continent shows intraseasonal and seasonal variabilities consistent with the MJO phases (Alvarez et al., 2014, 2016). During the December–February season, warm anomalies are favored by MJO phases 6 to 1 whereas the other phases favor cold anomalies. During the June–August season, warm anomalies are associated with phases 4 to 7 and cold anomalies are favored by phases 8 to 3. These temperature anomalies are induced by the changes in the regional circulations over the SACZ and southeastern South America partially in response to local circulations driven by the MJO diabatic heating (Cerne, 2011).

The development of reanalysis data sets opened up the door for more extensive and systematic analyses designed to find the influence of the tropical intraseasonal variations on the Northern Hemisphere winter weather, especially extreme or high-impact weather events with regional scales (e.g., Bond & Vechhi, 2003; He et al., 2011; Jeong et al., 2005; Jones, 2000; Jones et al., 2004a; Mo & Higgins, 1998a, 1998b; Vecchi & Bond, 2004). MJO convection centers located over the Indian Ocean are associated with an increase in the frequency of occurrence of cold surges of air from southern Siberia over the north and middle of China (Jeong et al., 2005) and heavy snowfall over Korea (Park et al., 2010). The MJO has even been shown to modulate violent tornado outbreaks over the U.S. (Barrett & Gensini, 2013; Thompson & Roundy, 2013), as well as fire activity in Canada, Alaska, Siberia, and China (Mu et al., 2011; Zhang, 2013).

Extreme rainfall events in the extratropics are more common on a global basis when the MJO is active (Jones et al., 2004a). For example, Jones and Carvalho (2012) showed that in observations active MJO periods are associated with a doubling of U.S. extreme precipitation events relative to inactive periods, with extremes concentrated along the West and East Coasts. These extreme events may become more predictable when the MJO is active and its convective center is located in the Americas, Africa, or Western Indian Ocean (Jones et al., 2011). Moon et al. (2012) argued that the record breaking snowfall in the Eastern U.S. during the winter of 2009–2010 was due to the combined influence of the MJO and ENSO, primarily induced by teleconnections associated with enhanced Central Pacific convection.

Atmospheric rivers (ARs) represent a category of high-impact weather events since they are characterized by highly focused streams of precipitable water that are important climatically for extratropical vapor transport (e.g., Gimeno et al., 2014; Zhu & Newell, 1998). ARs are associated with extreme precipitation along the U.S. West Coast and other locations (Dettinger et al., 2011; Lavers & Villarini, 2013; Moore et al., 2012; Ralph et al., 2006; Warner et al., 2012). Atmospheric river development, at least in the eastern Pacific, shows close relationships with the jet location, Rossby wave propagation, and Rossby wave breaking (e.g., Payne & Magnusdottir, 2014; Ryoo et al., 2013). ARs can span the lifetimes of several individual extratropical cyclones. Because MJO teleconnections with high latitudes are an important modulator of extratropical flow fluctuations and the atmospheric jet on intraseasonal time scales, the MJO has also been demonstrated to be an important modulator of AR activity.

A coherent relationship has been demonstrated among the MJO, associated Rossby wave forcing, and "pineapple express" events, another name for ARs that are associated with U.S. West Coast flooding (e.g., Higgins et al., 2000). Ralph et al. (2011) presented a case study during March 2005 that demonstrated the importance of the MJO and concurrent tropical Kelvin waves for production of an AR that impacted the U.S. Pacific Northwest. Guan et al. (2012) used 13 years of in situ snow water equivalent observations to show that when MJO convection is situated just to the east of the Maritime Continent, the number of California ARs and the snow water equivalent production per AR in the Sierra Nevada is augmented, although total snow accumulation in the Sierra Nevada produced by the MJO shows a slightly different relationship as a function of MJO phase. Guan et al. (2013) demonstrated similar results for the winter of 2010–2011. Payne and Magnusdottir (2014) showed that the MJO appears to favor the production of landfalling ARs along the U.S. West Coast during certain phases through teleconnections that favor a jet extension towards the U.S. West Coast and production of anticyclonic wave breaking. In particular, Zhang (2013) demonstrated that boreal winter flooding events along the U.S. West Coast is favored when the MJO is in phase 6, when enhanced convection is located in the west Pacific. A similar suppression of flood events occurs during MJO phase 1 when enhanced convection is over the Western Hemisphere.

While many studies on ARs related to the MJO have had an emphasis on the U.S. West Coast during boreal winter, MJO interactions with ARs are more pervasive geographically and seasonally. Guan and Waliser (2015) and Mundhenk et al. (2016) show that the MJO modulates boreal winter AR activity first near the coast of Asia, with anomalies moving progressively further east across the Pacific as MJO convection progresses eastward. As a result, a significant modulation of AR activity by the MJO can be detected in diverse regions such as Korea and Japan, Hawaii, and the U.S. West Coast (Mundhenk et al., 2016). For example, the frequency of ARs is decreased near Hawaii when MJO convection is over the Maritime Continent. The MJO also modulates AR anomalies near Alaska, Europe, and in the Southern Hemisphere (Guan & Waliser, 2015). Mundhenk et al. (2016) further demonstrate that the MJO modulation of AR activity depends sensitively on phase of ENSO and the seasonal cycle, with the effects of these phenomena sometimes adding constructively or destructively on a local basis to produce a given AR anomaly.

Extreme precipitation and drought over Middle East and Southwest Asia have been reportedly associated with MJO activity (Barlow et al., 2005; Barlow, 2012; Cannon et al., 2017; Hoell et al., 2012; Li et al., 2016; Nazemosadat & Ghaedamini, 2010; Pourasghar et al., 2015). However, MJO is only one of the many modulating factors in the area and no single extreme event can be attributed to MJO activity (Cannon et al., 2017). An analysis of observations shows a 23% increase relative to the mean of boreal winter precipitation over the region when the MJO suppressed convective phase is located in the eastern Indian Ocean, and a corresponding increase during the MJO active convective phase (Barlow, 2012; Barlow et al., 2005; Nazemosadat & Ghaedamini, 2010; Pourasghar et al., 2015). Barlow et al. (2005) explained the MJO influence on the regional hydroclimate through the changes in the vertical velocity over Southwest Asia induced by the advection of MJO temperature anomalies by the mean wind and advection of mean thermal gradient by the MJO wind anomalies. In other areas in the region MJO affects the dynamic and thermodynamic mechanisms that drive the winter westerly disturbances causing extreme precipitation over Karakoram and western Himalaya (Cannon et al., 2017). Westerly disturbances have a synoptic time scale between 2 and 7 days, but changes in their frequency and intensity can produce changes in seasonal mean precipitation within High Mountain Asia (Cannon et al., 2015). The snow cover over the central and eastern Tibetan Plateau can increase in the MJO phases 1–8 and decrease during phases 4–5 and shows no changes in the other phases (Li et al., 2016). These variations in the snow cover have been attributed to the large-scale circulation induced by the active and suppressed MJO convection over the Maritime Continent. This circulation results in moisture advection over the region enhancing the water vapor transport to the Tibetan Plateau.

The review of Zhang (2013) provides a further summary of the extreme weather phenomena in the extratropics and elsewhere (e.g., Philippines) affected by MJO.

2.1.4. Storm tracks

A small number of observational studies report a relationship between the tropical convection and the Northern Hemisphere extratropical areas characterized by strong activity of synoptic-scale disturbances—the storm tracks (e.g., Deng & Jiang, 2011; Lee & Lim, 2012). The North Pacific storm track can have an anomalous dipole structure propagating northeastward or display an eastward shift of the maximum as the center of MJO convection moves from the Indian Ocean to the western Pacific. The former response is induced by the changes in the convergence of energy flux, baroclinic conversion, and energy generation due to interaction between synoptic eddies and intraseasonal flow anomalies, whereas the latter is attributed to a local Hadley circulation forced by MJO heating. Takahashi and Shirooka (2014) showed that this MJO influence on the storm track activity is ENSO-phase dependent.

Because the storm track variability projects strongly on the canonical teleconnection patterns (Wettstein & Wallace, 2010), the implied teleconnections between the tropical heating and storm track regions inferred from the observational studies may be an indirect response of the storm tracks induced by the modes of climate variability discussed in section 2.1.1.

2.1.5. Southern Hemisphere

The influence of tropics on the Southern Hemisphere extratropics is less studied because most of the land surface is located within the tropics and subtropics with the exception of the southern tip of South America.

The Southern Hemisphere is characterized by two regions with quasi-persistent convection, the South Pacific Convergence Zone (SPCZ) and the SACZ. The SPCZ extends from the tropical West Pacific southeastward into the Central Pacific subtropics (Trenberth, 1976) and is characterized by variability from intraseasonal to decadal time scales (Kidwell et al., 2016). The convective heating in the SPCZ is a significant Rossby wave source (Meehl et al., 2001; Shimizu & de Albuquerque Cavalcati, 2011), affecting the mean Pacific flow (van der Wiel et al., 2016). The subtropical portion of the SPCZ is governed by tropical-extratropical interactions between synoptic transients and the mean state (Allen et al., 2009; Kiladis & Weickmann, 1997; Niznik et al., 2015; Vincent, 1994) and exhibits variability on intraseasonal and longer time scales (Matthews, 2012). The convective band in the SACZ originates in the Amazon basin, extends eastward into the subtropics in the southeastern Atlantic Ocean, and is most pronounced during the austral summer (Kodama, 1992, 1993). The SACZ displays an intraseasonal seesaw pattern (Carvahlo et al., 2004, 2002; Gonzales & Vera, 2014; Nogués-Paegle & Mo, 1997) and is an important source region for the Euroasian and North Atlantic climatic patterns (Grimm & Silva Dias, 1995) and South American summer precipitation (Liebmann et al., 2004; Paegle et al., 2000).

Over the broader midlatitudes, Ghil and Mo (1991b) did not find significant correlations between the tropics and Southern Hemisphere during the boreal winter, while Jin and Hoskins (1995) showed that the Southern Hemisphere extratropics do respond to the tropical heating in the boreal winter. However, the Southern Hemisphere response is weaker than in the Northern Hemisphere and Hsu (1996) attributed the differences to the strength of the large-scale circulation in each hemisphere. Carvalho et al. (2005) argued that suppressed Indonesian convection associated with the boreal winter MJO could initiate the positive phase of the Southern Hemisphere annual mode, while central Pacific MJO convection favors the negative annual mode phase.

During the austral winter, the Southern Hemisphere is dominated by two modes of ISV, the Pacific South-American (PSA) (Mo & White, 1985) modes, which explain 5.4% and 5.1% of the total variance (Mo & Higgins, 1998c). The two PSA modes have the longitudinal wave number 3 structure, with phase in quadrature with each other and wave train patterns with maximum amplitudes in the Pacific-South American sector.

Mo and Higgins (1998c) demonstrated that PSA patterns are associated with tropical convection. The PSA 1 mode is linked to the enhanced convective activity of MJO in phases 5 and 6, whereas PSA 2 mode is related to MJO in phases 7 and 8.

Berbery and Nogués-Paegle (1993) found that in composite analysis of observations austral summer enhanced convection over Indonesia is associated with increases of the meridional divergent wind and westerly wind over Australia. These conditions favor the enhancement of the Rossby wave source in this region, and a wave train propagates poleward from this region into South America following the typical Rossby ray tracing theory.

Matthews and Meredith (2004) demonstrated that the austral winter southern hemisphere annular mode strengthens about 7 days after the peak in Indian Ocean MJO convection. A recent study by Chang and Johnson (2015) identified teleconnection patterns in the midlatitudes of Southern Hemisphere winter that exhibit oscillatory behavior on time scales of 20–30 days and with the frequency of occurrence modulated by the MJO phases. These patterns mix characteristics of the southern hemisphere annular modes (Limpasuvan & Hartmann, 1999; Thompson & Woodworth, 2014) and the asymmetric PSA modes.

2.2. Influence of the Extratropics on the Tropics

Early investigations of the extratropical forcing of the tropics preceded the availability of good analyses of the global circulation. One of the original motivations was the need to explain the energy source for the tropical waves observed by Yanai and Maruyama (1966) and Wallace and Kousky (1968) and characterized by Matsuno (1966) in his classic paper.

Observationally, extratropical influences on the tropical waves were found by Nitta (1970) using station data and the analyses of upper level tropical wind fields of Zangvil and Yanai (1980) and Yanai and Lu (1983). Clear equatorward energy propagation from the midlatitudes was seen up to the appropriate critical latitudes for waves propagating westward. Examples of cross-equatorial wave energy flux were also noted.

Extratropical forcing of tropical motions goes well beyond tropical waves. Morgan (1965) associated the hurricanes that formed in the eastern North Atlantic during 1962 and 1963 with the occurrence of the northward surge of the Antarctic anticyclone over the South Atlantic and hypothesized that intrusion of Southern Hemisphere air into the Intertropical Convergence Zone (ITCZ) erodes the static stability of the atmosphere and changes the winds, which eventually enhance the environmental conditions conductive to tropical cyclogenesis. The incursions of extratropical Northern Hemisphere upper tropospheric troughs have been observed to trigger tropical convection over the east and central Pacific as, for example, in Liebmann and Hartmann (1984) and Kiladis and Wickmann (1992), and to organize convection in the Indian Ocean (Hsu et al., 1990). In the western Pacific, lower tropospheric cold surge events over the east Asian coast (due to movement of the Siberian anticyclone) were observed to enhance air-sea interaction (Slingo, 1998) and were linked to the development of tropical cyclones and westerly wind bursts (e.g., Meehl et al., 1996).

2.2.1. Modes of Climate Variability

The MJO-NAO connection is not just a one-way influence from the tropics to the extratropics. Instead, it is a two-way interaction. As can be seen from Table 1, the occurrence of a strong positive (negative) NAO is likely followed by MJO phase 7, whereas a strong negative NAO tends to be followed by a MJO phase 3. Some earlier studies found coherent circulation anomalies across the tropical and extratropical regions (e.g., Lau & Phillips, 1986) and suggested a global view of intraseasonal variability (e.g., Hsu, 1996). The global view of intraseasonal variability is supported by the instability theory of Frederiksen and Frederiksen (1992, 1993, 1997), who found that some of the unstable modes in a global model couple the extratropics with a tropical 30–60 day disturbance, which is similar to the MJO. Using a dry atmospheric model, Lin et al. (2007) showed that a tropical MJO-like wave can be generated through tropical-extratropical interactions, and there is coherent circulation variability between the tropical and extratropical regions in the model atmosphere. Generation of MJO-like signals by extratropical forcing was also found by Ray and Zhang (2010) in reanalysis (Figure 4) and numerical experiments with a dry-channel model. As discussed in Lin et al. (2009), the organized convection of the MJO induces extratropical Rossby waves, which propagate into the North Atlantic and interact with synoptic-scale transients and lead to an NAO anomaly. The changed NAO,

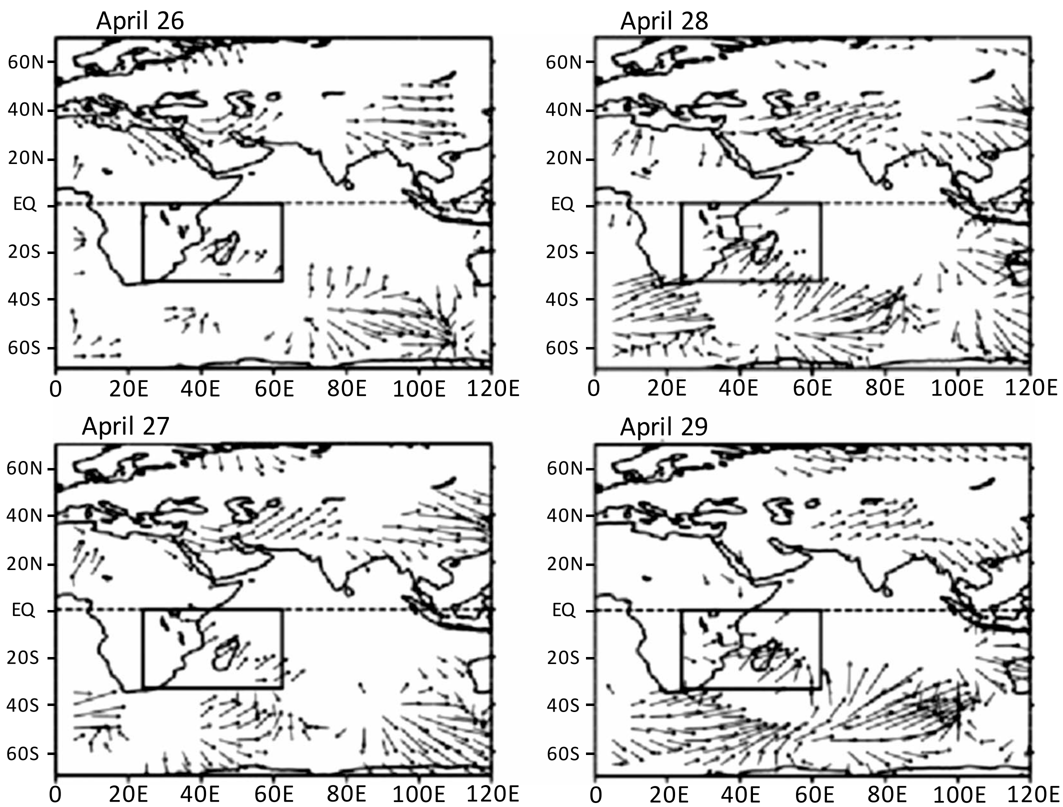

Figure 4. Wave activity flux vector (Takaya & Nakamura, 1997, 2001) at 200 hPa from 26 to 29 April 2002. This period coincides with the initiation of an MJO event in the region marked by the rectangular box (reprinted from Ray & Zhang, 2010, ©American Meteorological Society. Used with permission.)

through a southward wave activity flux in the North Atlantic, generates tropical zonal wind anomalies that help to trigger an MJO development in the Indian Ocean.

Randel (1992), using European Centre for Medium-Range Weather Forecasts (ECMWF) operational analyses, a showed a particularly striking example of a developing Southern Hemisphere anticyclone with associated poleward (negative) fluxes of zonal momentum exciting a mixed Rossby-gravity wave in the eastern tropical Pacific. Such bursts of momentum flux are associated with the decaying stage of baroclinic life cycles.

The MJO modeling study of Lin et al. (2000) utilized a quasi-equilibrium tropical circulation model with an intermediate level atmospheric model, which incorporated the effects of midlatitude disturbances. By replacing the temperature advection with its annual mean value, the authors were able to shut off the extratropical disturbances, which had the effect of killing the MJO. Excitation by extratropical variability is a major source of energy for the intraseasonal variability in this model.

Ray and Li (2013) also performed experiments in which the effects of extratropical disturbances impinging on the tropics were cut off, in this case by the imposition of wall-like boundary conditions in the subtropics, whose effect is to eliminate the MJO. Ma and Kuang (2016) point out, however, that such mechanism denial experiments lead to a change in the basic state of the atmosphere and that repeating these experiments with care taken to keep the basic state unchanged (with nudging on very long time scales) leads to the opposite conclusion: the MJO can exist quite well without extratropical forcing. Clearly, this is an area in which more study is needed.

2.2.2. Potential Vorticity Streamers

Extratropical transient upper level troughs that propagate eastward and equatorward during boreal winter have been linked to tropical plumes, extensive elongated cloud bands of length scale of 4000–16000 km and lifetimes of 3 to 9 days, as reviewed by Knippertz (2007). This interaction takes place in the active portions of the Intertropical Convergence Zone (ITCZ) where there is a westerly duct allowing the upper level extratropical disturbances to penetrate into the tropics. These disturbances can be described as tilted troughs or

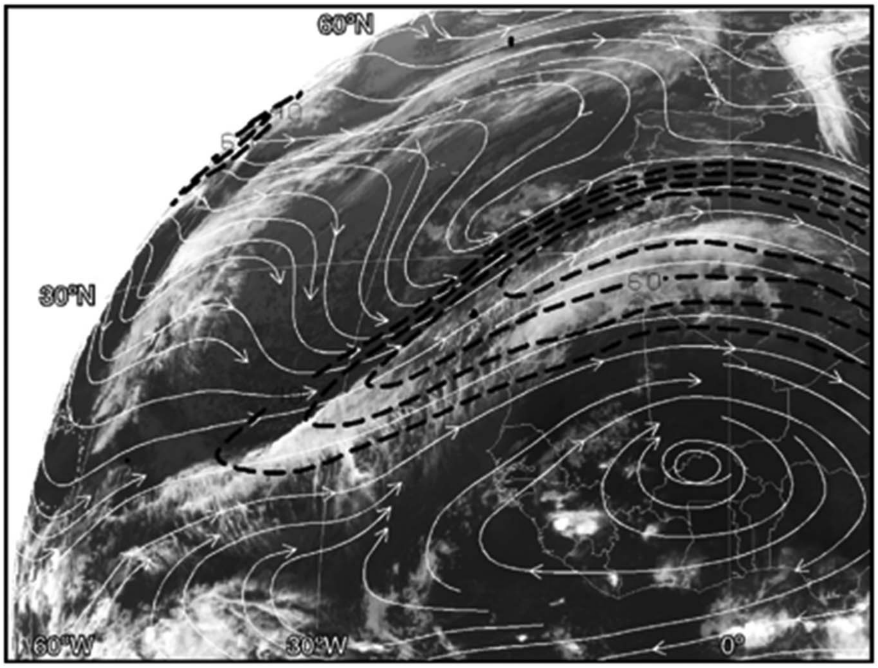

Figure 5. Meteosat infrared image of a tropical plume over northwest Africa at 00 UTC 31 March 2002. Superimposed are streamlines and isotachs on the 345-K isentropic level (dashed contours at 40, 50, 60, and 70 m s^{-1}) from the ECMWFTOGA analysis. The 345-K level is close to 200 hPa in the tropics streamlines that indicate extratropical wave incursion into the tropics. (reprinted from Knippertz, 2007, ©Elsevier. Used with permission).

equivalently as elongated potential vorticity (PV) "streamers" transporting stratospheric PV equatorward, leading to enhanced momentum flux and strong subtropical jet streaks. As pictured in Figure 5, they can be associated with Rossby wave breaking.

One particular region of the ITCZ whose variability on intraseasonal time scales is influenced by the midlatitude wave trains propagating equatorward is in the Atlantic (De Souza et al., 2005). During the austral summer, the Rossby wave train in the northern Hemisphere dominates and during the austral winter the influence of a Rossby wave in the Southern Hemisphere prevails (Tomaziello et al., 2016).

Similar interactions occur in the Southern Hemisphere with regard to the subtropical portion of the SPCZ and SACZ. Synoptic-scale waves travel along the subtropical jet over the southern Indian Ocean and south of Australia. These waves propagate equatorward into the westerly duct in the upper troposphere over the equatorial Pacific, triggering deep convection (van der Wiel et al., 2016). The influence on the SACZ is made manifest through the quasi-stationary wave pattern of the PSA (Castro Cunningham & De Albuquerque Cavalcanti, 2006).

2.2.3. Summer Monsoon Breaks

Studies evaluating the influence of the midlatitude circulation on the Indian monsoon rainfall date back to the case study of Ramaswamy (1962), who suggested that midlatitude circulation regimes with large-amplitude waves favor the troughs in the midlatitude westerlies to extend over India causing monsoon breaks, which can last between 3 and 5 days and sometimes between 17 and 20 days (Ramamurthy, 1969). Subsequent work based on longer time series confirmed the existence of a significant relationship between the midlatitude zonal circulation and summer monsoon rainfall (Chattopadhyay et al., 1994; Ding & Wang, 2007; Srivastava et al., 2014). Others (e.g., Kripalani et al., 1997; Raman & Rao, 1981; Samanta et al., 2016; Yasunari, 1986) found strong correlations between large-scale blockings over the Northern Caspian Sea and/or east Siberia and the monsoon breaks. Samanta et al. (2016) showed that in observations, blocking is not a climatological feature of the monsoon and is favored by Rossby wave breakings which allow the transport of low PV from the tropics toward north and the transport of high PV air toward the equator.

The west African monsoon system occasionally experiences dry spells that persist for 2 to 10 days over the eastern Sahel of Sudan and eastern Chad (Vizy & Cook, 2009, 2014), and their mechanisms can have tropical and/or extratropical origins. Midlatitude intrusions of cold, dry air from Europe and Mediterranean southward into northeast Africa (Chauvin et al., 2010; Roehring et al., 2011; Vizy & Cook, 2014) decrease the moisture supplied by the southwesterly flow of the monsoon system. The decrease in low-tropospheric latent energy results in a more stable environment that suppresses convective activity (Vizy & Cook, 2009, 2014).

2.2.4. Cold Air Surges

The regional and global characteristics of surges of cold midlatitude air reaching into the tropics have been reviewed by Garreaud (2001). Cold surges prevail over the Asia Pacific region (e.g., Chang et al., 1979; Chang & Lau, 1980; Mukarami & Sumi, 1981; Ramage, 1971), South Pacific and South America (e.g., Escobar et al., 2004; Garreaud, 1999; Garreaud & Wallace, 1998; Krishnamurti et al., 1999; Lupo et al., 2001; Müller & Ambizzi, 2007; Müller et al., 2005; Ricate et al., 2015; Sprenger et al., 2013); Central America and the Caribbean (e.g., Schultz et al., 1997, 1998), Africa (e.g., Jury & Parker, 1998; Lavaysse et al., 2010; Vizy & Cook, 2014), Indian Ocean and the Maritime Continent (e.g., Hsu et al., 1990; Tangang et al., 2008; Wang et al., 2012), along the east coast of Australia (e.g., Baines, 1980; McBride & McInnes, 1993; Perrin & Simmonds, 1995; Simmonds & Richter, 2000), and North America (Cole & Mass, 1995; Lin, 2015; Mecikalski & Tilley, 1992).

Although cold surges have regional dependencies, they all share common features such as a shallow layer of cold, dry air in the lower troposphere with a horizontal scale of 500–1000 km (Garreaud, 2001) and a time scale varying from 2 days to a few weeks. The short lived cold surges result from the interaction of synoptic-scale flow with the Earth's major mountain ranges (Garreaud, 2001; Mailler & Lott, 2010) and are accompanied by a hydrostatically induced ridge of surface pressure (Garreaud, 2001) and strong meridional low-level winds (Vizy & Cook, 2009). The longer time scale cold surge cases are associated with a number of mechanisms: (1) the development of a short wave train triggered by the expansion of the semipermanent core-pressure system, the Siberian high, over South East China (e.g., Ding, 1990; Lau et al., 1983; Martin et al., 1988; Ryoo et al., 2005; Wu & Chan, 1995; Yen & Chen, 2002); (2) a Rossby wave train emanating from the contractions/extensions of the westerly polar jet located over the northeastern Atlantic and propagating along the subtropical westerly jet waveguide across the Eurasia (e.g., Chauvin et al., 2010; Vizy & Cook, 2009); (3) variations in the position of the Pacific High (Schultz et al., 1998); and (4) Rossby waves emanating from the high latitudes of the Southern Ocean (e.g., Fukutomi & Yasunari, 2005; Perrin & Simmonds, 1995).

The short episodes of cold air intrusions are responsible for extreme or high-impact weather events. In South America, the outbreaks of cold air produce freezing temperatures from central Argentina to southern Brazil (e.g., Fortune & Kousky, 1983; Marengo, Cornejo, et al., 1997; Marengo, Nobre, et al., 1997; Marengo et al., 2002; Müller & Berri, 2012; Vera & Vigliarolo, 2000). In Central America and Mexico, cold surges can produce superstorms with strong wind gusts (~30 m/s) and a large temperature decrease (Schultz et al., 1997). Over North America, intrusions of midlatitude cold air associated with the high atmospheric pressure system over Alaska result in strong north wind events and temperatures below freezing in the southern Sacramento Valley of California (Grotjahn & Faure, 2008).

Likewise, persistent cold surges can also be associated with extreme weather events. In addition, they have been related to anomalous activity of tropical intraseasonal variability. In the Australian region, midlatitude cold surges associated with the wave trains from the Southern Ocean interact with the local cyclones on the west coast and produce explosive cyclogenesis (Ashcroft et al., 2009). The planetary scale circulation anomalies induced by the NAO occasionally force cold air southward over the southeastern U.S. and produce below freezing temperatures in this region. Cold surges in east-southeast Asia induced by cold-air outbreaks from the Siberian high produce a drastic increase of northerly and northeasterly winds (e.g., Boyle & Chen, 1987; Chen et al., 2002; Joung & Hitchman, 1982; Lau & Lau, 1984; Park et al., 2008; Takaya & Nakamura, 2005; Wu & Chan, 1995), heavy snowfall (Boyle & Chen, 1987; Ding, 1994) and rainfall (Chen et al., 2015; Koseki et al., 2014), and flooding (Tangang et al., 2008; Wangwongchai et al., 2005). The case study of Yokoi and Takayabu (2010) suggests that the development of tropical cyclone Nargis over the Bay of Bengal in 2008 was influenced by a cold surge from East Asia.

The long-lasting East Asian cold surges affect the local (~115°E) Hadley circulation (Chang & Lau, 1980), which exhibits a double-cell structure during the surge (Chu & Park, 1984), and the tropical upper level divergence and low-level convergence from western Pacific to the eastern Indian Ocean (Chang & Chen, 1992). The intensification of the local Hadley circulation strengthens the East Asian winter monsoon (Chen et al., 2002). Over the eastern Indian Ocean and Maritime Continent, these cold surges have been associated with tropical cyclogenesis (Chang et al., 2003; Takahasi et al., 2011) and the MJO initiation in the tropical western Indian Ocean (Wang et al., 2012).

In addition to the African monsoon breaks discussed above, cold surges during boreal winter are associated with below normal rainfall over equatorial Africa and along the Guinean coast (Vizy & Cook, 2014).

2.3. Coupled Tropical-Extratropical Variability and Feedback

There have been numerous observational and theoretical analyses of the tropical-extratropical interactions associated with the MJO (e.g., Ghil & Mo, 1991a; Hsu, 1996; Kiladis et al., 1994; Knutson & Weickmann, 1987; Lau & Phillips, 1986; Matthews & Kiladis, 1999; Meehl et al., 1996; Schubert & Park, 1991; Straus & Lindzen, 2000; Weickmann & Khalsa, 1990) and boreal summer intraseasonal oscillations (Fukutomi & Yasunari, 1999, 2002).

Based on correlations, empirical orthogonal functions (EOFs), and composite techniques applied to observations, Lau and Phillips (1986) suggested the existence of a coupling between the normal modes of the tropics and extratropics. Using a singular spectrum analysis applied to observations, Ghil and Mo (1991a) suggested that oscillations in the two regions are coupled by extratropical wave trains. In an observational study, Mechoso and Hartmann (1982) find some coherence between the Southern Hemisphere subtropics and midlatitudes for eastward moving wave number 1 during a single austral winter. Straus and Lindzen (2000) used reanalyses to examined the properties of eastward propagating fluctuations in the planetary wave components of the zonal wind in 39 boreal winters on time scales corresponding to both the MJO and to phase speeds in the 1–10 m/s range, the range of known baroclinic instability for meridional scales relevant to the observed jets. The two frequency ranges overlapped considerably. The extraordinarily high coherence found at zero phase lag between 200 hPa zonal wind fluctuations at the jet latitude (32^{o}N) and in the tropics (12^{o}N) for both frequency ranges considered is evidence for strong tropical-extratropical coupling. Their work supported the theory that planetary scale baroclinic instability and the tropical MJO are connected. Further support for this idea has also come from the studies of Pan and Li (2008). Matthews et al. (2004) and Roundy (2012) argue that penetration of extratropical energy into the tropics during certain MJO phases, for example, those phases with a retracted Asian-Pacific jet, are important for modulating MJO convection in the central and eastern Pacific. The inability of reduced complexity models (e.g., Frederiksen, 1982, 1983; Hoskins & Karoly, 1981; Simmons et al., 1983) to account for such eddy feedback may limit their ability to realistically simulate tropical-extratropical interactions with the MJO.

The subtropics and extratropics also play a role in cycling the MJO (Sakaeda & Roundy, 2015). The easterly wind in the upper troposphere of the Western Hemisphere previous to MJO convective initiation over the Indian Ocean was linked to equatorward-moving Rossby wave trains that are part of the extratropical circulation response to the convection associated with previous cycle of the MJO. Upward motion associated with this wave train induces deep cooling in the troposphere, which yields a trough in phase with easterly wind, producing the Kelvin wave structure. According to this study, the upper tropospheric intraseasonal wind field over Western Hemisphere does not consist only of a free Kelvin wave. Extratropical breaking waves couple northern and southern midlatitudes at the equator, and the circulation features in the dry regions of the MJO might influence its phase speed.

A two-way tropical-extratropical interaction has been shown to exist in observations over the western Pacific during June–July (Fukutomi & Yasunari, 1999, 2002). The influence of tropics on the midlatitudes comes from the convective activity over the South China Sea through a Rossby wave train with barotropic structure propagating over the North Pacific. The southern flank of the wave train modulates the circulation of the subtropics, which provides the means for the midlatitudes to influence the tropics. Lower tropospheric cyclonic and anticyclonic subtropical circulations propagate southwestward into the South China Sea where these circulations further modulate the convection over the ocean.

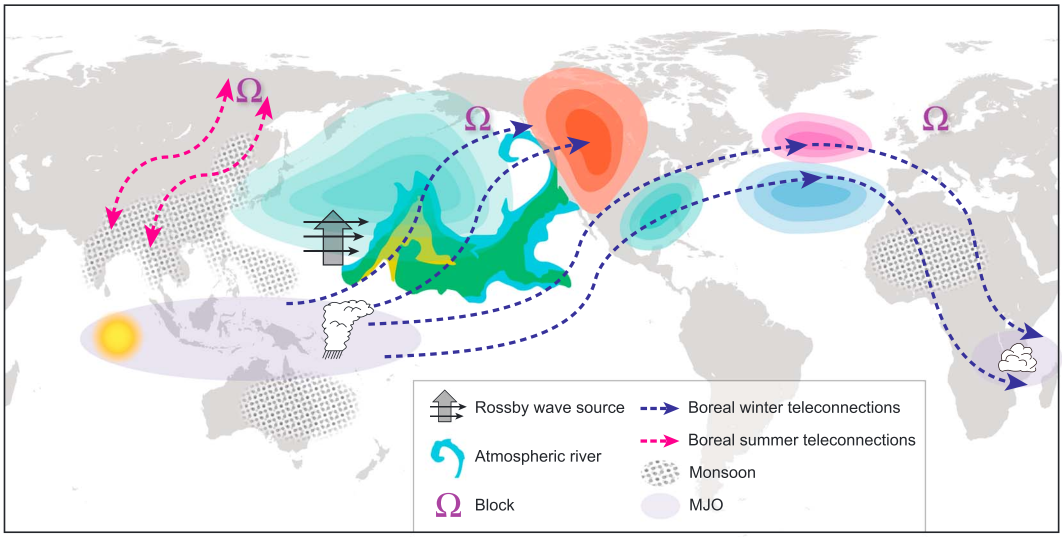

Figure 6. Schematic representation of tropical—Northern Hemisphere interactions and teleconnections inferred from observational and numerical studies. The convective activity of the tropics associated with MJO, indicated by the gray shading with the cloud symbol, creates a Rossby wave source, consisting of strong divergent flow (heavy arrow) oriented up the gradient of vorticity (light arrows). The resulting Rossby waves affect the probability of occurrence of the dominant modes of variability in midlatitudes, including the NAO (lue and magenta shading) and the PNA (light green and orange shading). The frequency of blocking (indicated by Ω) and atmospheric rivers (dark green and the yellow shading in the Pacific Ocean) are strongly affected by the MJO convectively driven heat source. The global monsoon systems (indicated by the dotted pattern) also interact with midlatiudes, in particular the Indian Monsoon system. Under certain conditions these circulation anomalies of the midlatitudes can feed back onto the subsequent evolution of the convective activity of the tropics (i.e., new MJO phases and breaks in the Indian summer monsoon precipitation), whereas at other times midlatitude nascent energy fluxes, without any relationship to a tropical forcing, can impinge into the tropics, forcing tropical circulations and organized convection. These potentially two-way interactions are indicated by the blue dotted lines and arrows (for boreal winter) and the magenta dotted lines and arrows (for boreal summer).

Thus, Rossby wave breaking is at least partially responsible for the two-way interactions between tropics and midlatitudes, in agreement with our discussion in the previous section. MacRitchie and Roundy (2016) suggest that anticyclonic wave breaking (Thorncroft et al., 1993), which occurs over the central Pacific Ocean and is modulated by the presence of convective activity over the Indian Ocean, enhances westerlies at low latitude over the central Pacific Ocean and leads to a higher amplitude MJO.

The tropical-Northern Hemisphere interactions and teleconnections inferred from these observational and numerical studies are summarized in the schematic representation depicted in Figure 6. The convective activity of the tropics associated with MJO, indicated by the gray shading with the cloud symbol, creates a Rossby wave source (Sardeshmukh & Hoskins, 1988; Tyrell et al., 1996), consisting of strong divergent flow (heavy arrow) oriented up the gradient of vorticity (light arrows). The resulting Rossby waves affect the probability of occurrence of the dominant modes of variability in midlatitudes, including the NAO (blue and magenta shading) and the PNA (light green and orange shading). The frequency of blocking (indicated by Ω) and atmospheric rivers (dark green and yellow shading in the Pacific Ocean) are strongly affected by the MJO convectively driven heat source. The global monsoon systems (indicated by the dotted pattern) also interact with the midlatiudes, in particular the Indian Monsoon system. Under certain conditions, these circulation anomalies of the midlatitudes can feed back onto the subsequent evolution of the convective activity of the tropics (i.e., new MJO phases and breaks in the Indian summer monsoon precipitation), whereas at other times midlatitude nascent energy fluxes, without any relationship to a tropical forcing, can impinge into the tropics, forcing tropical circulations, and organized convection. These potentially two-way interactions are indicated by the blue dotted lines and arrows (for boreal winter) and the magenta dotted lines and arrows (for boreal summer). Because the interactions and teleconnections of the Southern Hemisphere are not as well understood, they were not included in the current schematic.

3. Theory and Mechanisms of Tropical-Extratropical Interactions

The search for physical support for statistical correlations between the tropics and extratropics evolved in parallel to the analysis work. While the estimates of teleconnections from observations and modeling work discussed in section 2 take into account the cause and effect relationship, they do not establish the potential drivers of remote links. The picture that emerges from the available studies suggests a complex spatial and temporal signature of the interactions, which so far have been only partially explained.

3.1. Influence of the Tropics on the Extratropics

The first theoretical studies of the influence of the tropics on the extratropics were conducted with linear steady state models in which the atmosphere responds to a stationary, idealized source of heating localized in the tropics. By using a quasi-geostrophic two-level model on a β-plane, Opsteegh and Van den Dool (1980) showed that a tropical disturbance will excite stationary Rossby waves with the ability to influence the midlatitudes only if the zonal scale of the perturbation is large (corresponding to low zonal wave numbers), and the tropical heating is situated in a background state containing mean westerlies. Hoskins and Karoly (1981), also using a linearized nondivergent baroclinic model, showed that the response of the atmospheric circulation to a source of tropical heating located on the equatorial side of the basic westerly wind maximum is confined to the region of the source at lower levels and consists of a train of Rossby waves propagating poleward and eastward in the upper troposphere. To explain these results, Hoskins and Karoly (1981) developed an influential theory of Rossby ray tracing, applicable to stationary and nonstationary Rossby waves. Using the Rossby wave dispersion relationship with a fixed zonal wave number and fixed frequency, the direction and speed of wave trains is traced from a starting point assumed to be the location of the source. The meridional wave number of the wave changes with locations along the wave path so that the local dispersion relationship is always maintained. Stationary wave rays originating in the tropics propagate poleward and eastward until a critical latitude (dependent on the properties of the basic state) is reached at which point the wave, still propagating eastward, turns to propagate toward the equator. Longer waves (with smaller zonal wave numbers) can propagate further poleward before they turn. This simple theory has proven to be very useful in understanding many of the model and observed results on stationary wave propagation. Assuming that the total diabatic heating of the atmosphere is related to the SST anomalies, Webster (1981) found that the atmospheric response to a remote SST anomaly depends on the location of the SST anomaly. As in previous studies, strong teleconnections occur when the SST anomaly is located in a weak low-latitude basic flow. Both Hoskins and Karoly (1981) and Webster (1981) emphasize that poleward motion from the tropics is favored by a simple vorticity balance in which the zonal advection of high vorticity fluid is negligible and by the thermodynamic balance between the diabatic heating and vertical motion. Lim and Chang (1983) suggested that vertical shear of mean zonal wind also plays a role in the meridional propagation of disturbances generated by tropical heating, and Kasahara and da Silva Dias (1986) and Lim and Chang (1986) showed that by slightly increasing the complexity of a linear baroclinic model to account for the vertical shear of the mean zonal wind the efficiency of generating Rossby waves increases. In the presence of vertical shear the heating-induced equatorially trapped Rossby waves are slightly less equatorially confined (Wang & Xie, 1996). Majda and Biello (2003) and Kim et al. (2006) reached similar conclusions.

Another school of thought, pioneered by Simmons et al. (1983), Branstator (1985), and Sardeshmukh and Hoskins (1988) using linear barotropic models, associated the extratropical response to the fast-growing mode of tropical instability with a baroclinic structure whose dispersion on the climatological flow of the extratropics excites local anomalies through batrotropic instability and energy conversion. The midlatitude regions of preferred response tend to be located over the northeastern Pacific and Atlantic with their source regions located to their southwest (Branstator, 1985 and Simmons et al., 1983). Sardeshmukh and Hoskins (1988) also recognized the importance of the horizontal and vertical structure of the tropical heating for the divergent flow aloft, whose interaction with the mean vorticity gradient represents the source of Rossby waves.

While the observational studies were exploring the role of the MJO on the extratropical large-scale circulation anomalies, modeling studies were testing highly idealized sources of tropical heating. To fill the gap between the modeling and observational studies, Ferranti et al. (1990) designed an experiment with a linear barotropic model in which the tropical forcing was derived from the leading EOFs of the tropical convective activity. Despite the simplicity of the model, the simulated large scale circulation anomalies retain the coherence of the observed low-frequency modes of variability in the midlatitudes that project onto the PNA and NAO patterns. However, the amplitude of the oscillations was rather weak. A similar result was obtained by

Bladé (1994) using a nonlinear baroclinic model. An explanation for these differences was given by Bladé and Hartmann (1995) in a series of heuristic experiments which showed that the extratropical response is sensitive to the phase speed of the forcing.

A large-scale tropical diabatic heating can induce a Rossy wave train that propagates poleward and eastward (Li & Nathan, 1997; Magaña & Yanai, 1991; Tyrell et al., 1996). The extratropical response pattern is established in about two weeks (Jin & Hoskins, 1995; Matthews et al., 2004). With a linear model, Lin, Brunet, and Mo (2010) investigated the sensitivity of the Northern Hemisphere extratropical response to the location of tropical heating in boreal winter. Strong responses with a similar but out-of-phase structure were found when the equatorial heating is located in the Indian Ocean and western Pacific. A dipole tropical forcing associated with either above normal convection in the Indian Ocean and below normal convection in the western Pacific, corresponding to MJO phase 3, or below normal convection in the Indian Ocean and above normal convection in the western Pacific, corresponding to MJO phase 7, is most effective in exciting extratropical circulation anomalies.

The dynamical link between MJO and PNA suggested by analysis of observations was tested by Mori and Watanabe (2008) in forced numerical experiments with a linear barotropic model. Results of these experiments showed that PNA is only partly induced by tropical heating and cooling associated with MJO when only linear processes are considered. These results may explain the PNA-like response identified by Riddle et al. (2013) when the occurrence of the pattern was reconstructed solely from the MJO phases. The influence of tropical convection patterns on the PNA was investigated by Dai et al. (2017) in observations by contrasting PNA events associated and not associated with enhanced tropical convection. This comparison shows that convection-related PNA events are preceded by a stronger wave activity fluxes associated with the Eurasian wave train, have a higher frequency of occurrence (60% versus 40%), and persist longer (three versus two weeks) than the nonconvective PNA events. These results suggest that tropical convection is a modulator of the PNA events and not a trigger mechanism.

The development of more comprehensive GCMs led to simulations that reproduce some of the observed features of the extratropical response to tropical forcing (e.g., Higgins & Schubert, 1996; Hsu, 1996) and also allowed for sensitivity tests to be undertaken with primitive-equation models in which the varying basic state is driven by internal dynamics and external forcing. These studies not only confirmed the previous findings based on simple linear models but also revealed new characteristics of the midlatitude responses to tropical heating. In general, the extratropical anomalies tend to lag the tropical forcing by about 2 weeks (Branstator, 2014; Matthews et al., 2004). More specifically, the response depends on the history of tropical heating (Straus et al., 2015).

Using an atmospheric GCM, Seo and Lee (2017) investigated the debate regarding the location of tropical convection forcing the PNA-like pattern. Their results showed that both the Indian Ocean and western Pacific forcing contribute to the formation and maintenance of PNA-like pattern via Rossby wave propagation. The Indian Ocean heating induces two Rossby wave source regions—a negative region around Southern Asia and a positive region over the western North Pacific—and the wave propagation to the PNA region is accomplished by longwaves (zonal wave number 1 and 2) through a direct propagation and shorter waves that first needs to be displaced downstream by the jet waveguide effect before emanating at the Asian-Pacific jet exit. The western Pacific cooling induces a positive region of Rossy wave source, and the wave propagation to the PNA region occurs through a Rossby wave train.

In addition to Rossby wave mechanisms, many of the observational studies discussed in section 2.1 suggest that changes in tropical convection can affect the jet-level winds, vertical motion at the low and upper levels, storm tracks, and poleward moisture transport. All these mechanisms can influence the extratropical teleconnection patterns and their projections on the regional hydroclimate variability. Barlow et al. (2005) attributed this influence to the proximity of affected regions to the most active regions of the MJO convection and strong local wind response to the MJO tropical forcing.

Another possible pathway through which NAO can respond to MJO tropical forcing involves the stratosphere. This mechanism was proposed by Jiang et al. (2017) who found that in observations, the MJO-related negative NAO events are preceded by a deceleration of the stratospheric polar vortex caused by an increase in the vertical propagation of planetary-scale wave activity into the stratosphere. The MJO-related positive

NAO events are preceded by a strengthen stratospheric polar vortex due to a weaker vertical planetary-scale wave activity.

3.2. Influence of the Extratropics on the Tropics

One of the first papers to tackle the theory of extratropical forcing of the tropics was Charney (1969), who considered the problem from the extratropical point of view using a quasi-geostrophic system linearized about a zonally averaged basic flow. He concluded that propagation in the meridional direction in the presence of mean easterly flow was only possible if the phase velocity of the midlatitude wave is more easterly than the mean flow. Since such waves have small energy in the midlatitudes, the conclusion was that disturbances that propagate from the midlatitudes into the tropics will be in the upper troposphere and lower stratosphere where zonal winds are very weak. These results were consistent with those of Bennett and Young (1971), who used a linearized shallow water model relevant for the tropics with horizontal shear.

The effects of longitudinal variations in the basic state zonal wind (U) were considered by Webster and Holton (1982), who were motivated by observations of propagation of energy between the hemispheres in the eastern Pacific. Using a nonlinear shallow water model relevant for the upper level of the tropical atmosphere, they examined the response to stationary forcing at 20°N at various locations with respect to the westerly duct. They concluded that large-scale disturbances generated in the northern middle latitudes have little effect on low latitudes when a critical line ($U = 0$) lies between the source and the equator. They find, however, that these disturbances may have a significant influence on equatorial regions and the opposite hemisphere if a westerly duct ($U > 0$) is present in the equatorial zone and that the amplitude of the response increases with the strength of the westerlies. A generalization of these results can be obtained by replacing U with $U - c$, where c is the phase speed of the wave. These results are entirely consistent with many of the observations that place the extratropical effects in the eastern Pacific, where a westerly duct is observed at upper levels.

Recently, Francey and Frederiksen (2016) found that variations in interhemispheric transport, resulting in observed hemispheric differences in carbon dioxide and other gases, were correlated with the opening and closing of the upper tropospheric westerly duct. They attributed these finding to changes in Rossby wave propagation and changes in near-equatorial transient kinetic energy (see Figure 7 in Frederiksen & Webster, 1988), which depend on the existence and strength of the westerly zonal winds in the equatorial region.

Zhang and Webster (1989) and Zhang (1993) took a step back in considering only a zonal flow U independent of longitude but focused specifically on the forcing of the individual types of equatorially trapped waves. Using a linearized equatorial beta-plane model, they found that the amplitudes of westward propagating (Rossby and mixed Rossby-gravity) waves are larger in the presence of mean westerlies for low frequency forcing and are larger with mean easterlies for high-frequency forcing. The results were opposite for Kelvin waves.

The recognition that mean westerlies are *not* required for midlatitude excitation of tropical waves was expanded upon by Hoskins and Yang (2000), who focused on the direct excitation of equatorial waves in simple linear and primitive equation models by fluctuating vorticity and heat sources at 20°N. The equatorial response to extratropical forcing does not rely on the presence of westerly winds but is due to a direct projection of the forcing onto equatorially trapped waves. This response can be significant due to an approximate frequency match between particular wave numbers in the forcing and a particular equatorially trapped wave. In particular, higher-latitude forcing with an eastward phase speed is surprisingly effective in triggering the equatorial Kelvin wave.

Hall et al. (2016), conducted a numerical experiment with the Weather Research and Forecasting (WRF) model configured in the tropical channel mode to quantify the influence of the extratropics on simulated tropical intraseasonal variability. It was found that about half of the intraseasonal variance can be attributed to extratropical influences.

The mechanism through which the midlatitude circulation influences the Indian summer monsoon is probably the most understood extratropical-tropical interaction and is strongly supported by observations (Ramaswamy, 1962). In a nutshell, during situations with a strong meridional flow embedded with jets over the Tibetan Plateau and an existing large midlatitude trough in the 300 hPa westerlies north of 40°N, the trough extends into the Indian subcontinent and subsequently undergoes a slow eastward propagation.

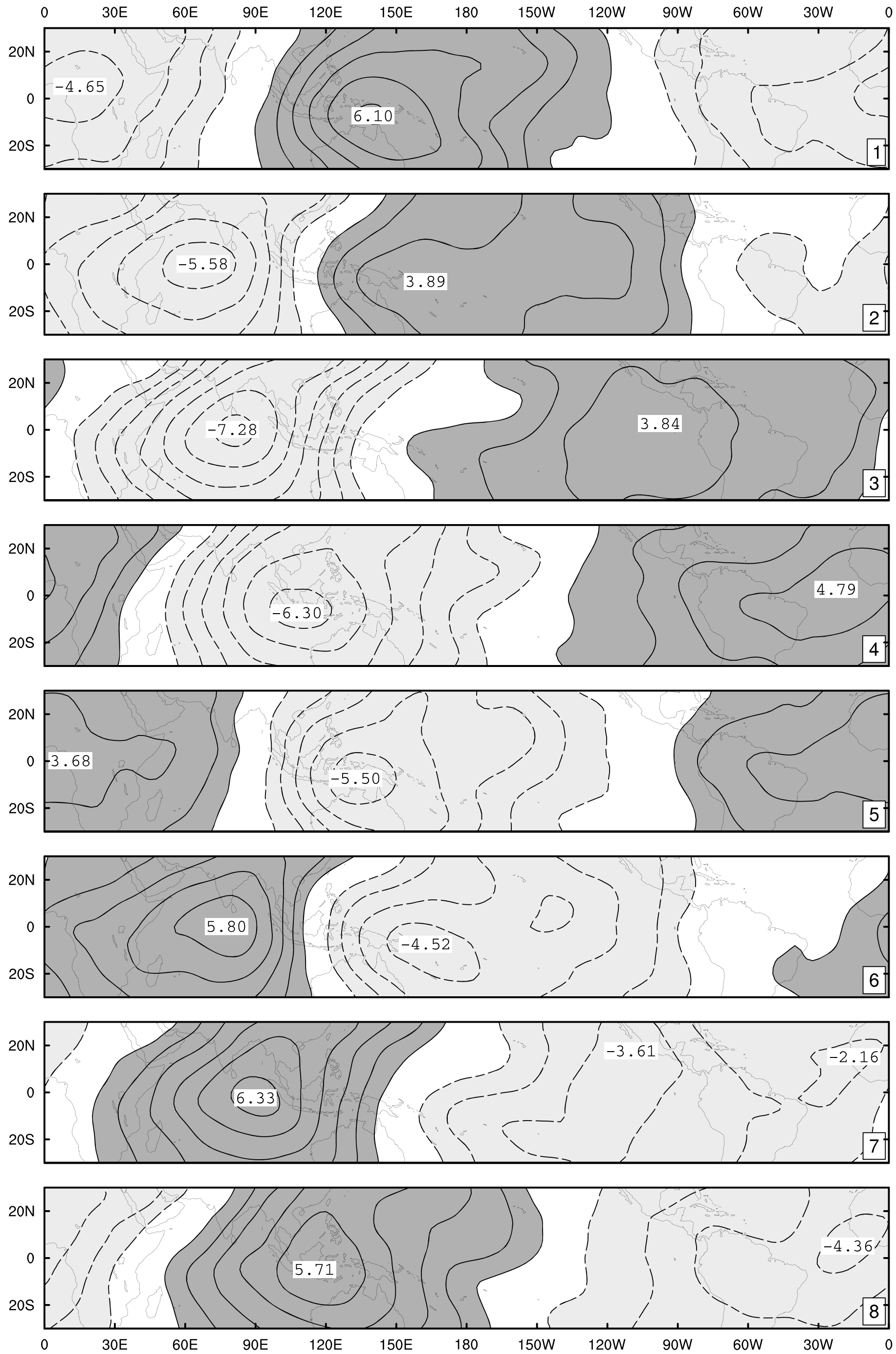

Figure 7. The evolution of the 200 hPa velocity potential (km^2 s^{-1}) at eight MJO phases based on 30 northern winters (December–January–February) between 1979/1980 and 2008/2009 and pentad data derived from the daily averaged NCEP/NCAR reanalysis data. Contour interval is 1 (reprinted from Frederiksen & Lin, 2013, ©American Meteorological Society. Used with permission.)

This eastward propagation of the trough weakens the easterlies at 500 mb associated with the Tibetan high and allow the Siberian High to extend southeastward, thus favoring dry conditions.

3.3. Coupled Tropical-Extratropical Variability and Feedback

The genesis mechanism of intraseasonal oscillations (ISOs) of the tropics was sought in coupled tropical-extratropical dynamics in the theoretical work of Frederiksen and Frederiksen (1993, 1997) and Frederiksen (2002). Their studies applied a linear two-level primitive equation model with a three-dimensional basic

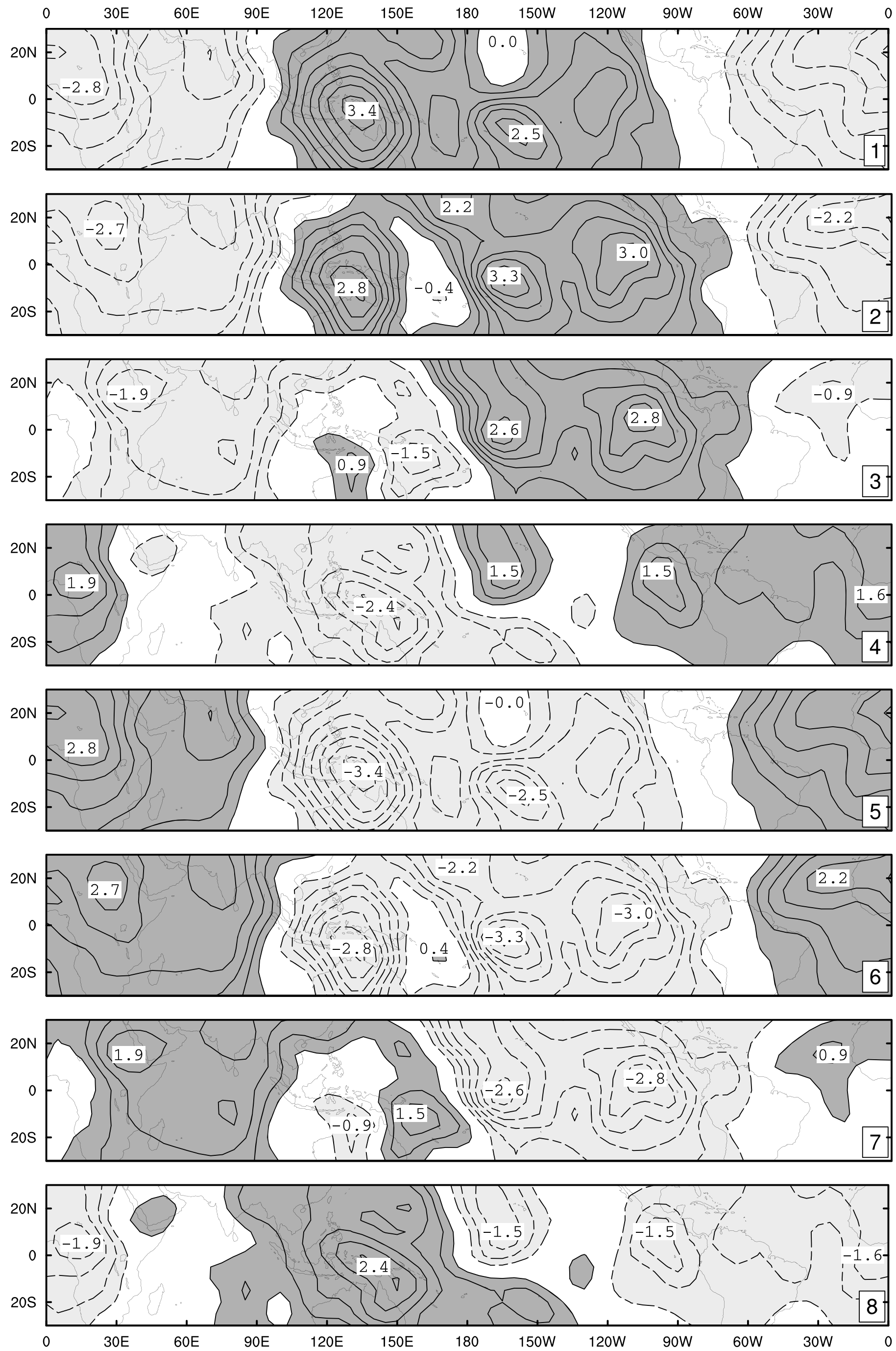

Figure 8. The evolution of the 300 hPa velocity potential at eight MJO phases for the leading theoretical intraseasonal mode for a basic state that include cumulus heating and evaporation-wind feedbacks (EVAP). Contour interval is 0.5 (reprinted from Frederiksen & Lin, 2013, ©American Meteorological Society. Used with permission.)

states characteristic of boreal winter and revealed a wide variety of disturbances, including growing modes that couple the extratropics with 30–60 day oscillations similar to the MJO. Their coupled ISO modes were found to derive their first internal (baroclinic) mode structure in the tropics due to convective interaction with dynamics (Neelin & Yu, 1994).

Frederiksen and Lin (2013) showed that evolution of the upper troposphere velocity potential for the observed MJO (Figure 7) and that for the theoretical intraseasonal oscillation mode of Frederiksen (2002) (Figure 8) were in broad agreement; importantly, the associated extratropical evolution of the upper

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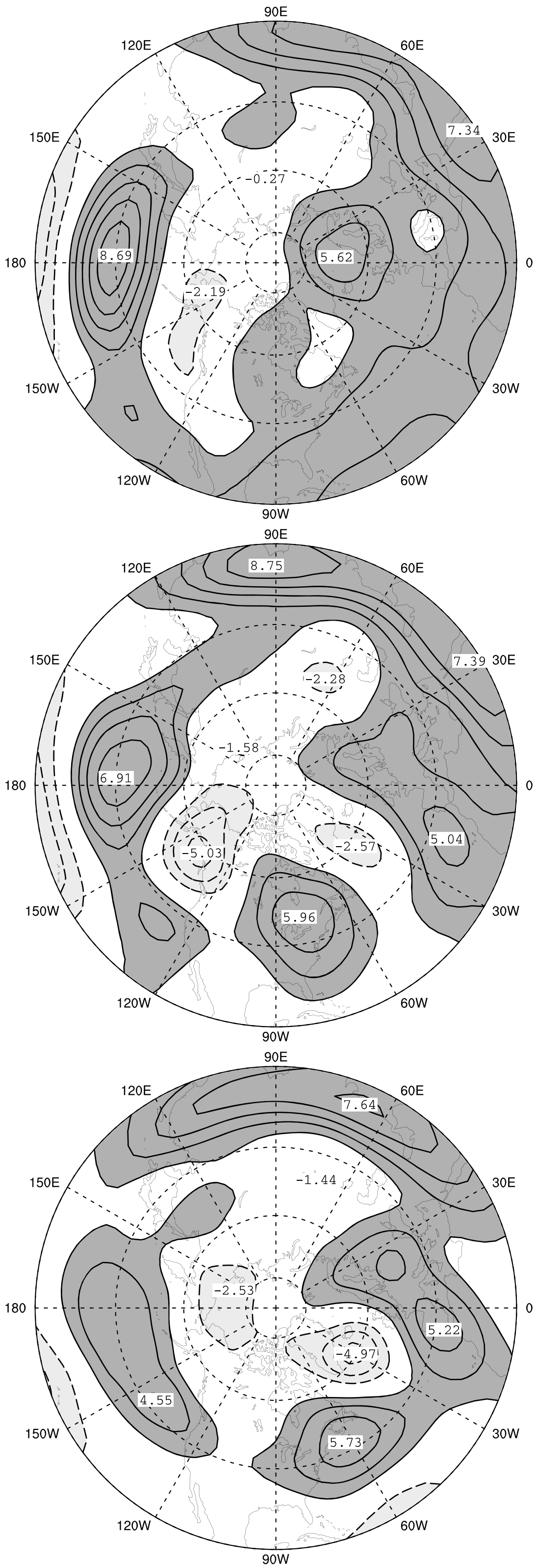

Figure 9. Extratropical NH 300 hPa streamfunction anomalies for reanalysis data, as in Figure 7 at MJO phases 3, 4 and 5. Contour interval is 1.5 (reprinted from Frederiksen & Lin, 2013, ©American Meteorological Society. Used with permission.)

tropospheric streamfunction from observations (Figure 9) and theory (Figure 10) exhibit PNA-like and NAO patterns at the same phases. They also found that fluxes of wave activity based on the upper-troposphere streamfunction of the theoretical ISO modes showed strong tropical-extratropical interactions and structures very similar to those of Lin et al. (2009) for observed extratropical anomalies associated with MJO convection. Frederiksen and Lin (2013) also studied theoretical ISOs for a variety of basic states and found similar links between the tropical ISO signal with PNA-like and NAO teleconnection patterns.

Recently, Whelan and Frederiksen (2017) examined the role of tropical-extratropical interactions on the intraseasonal time scale in the extreme rainfall and flooding over Australia during the 1974 and 2011 La Niña events. On the basis of observational analyses of OLR and model simulations, they found that the major extreme rainfall over northern Australia and related flooding during January 1974 and 2011 were associated with rapid growth of the MJO and the collision, or constructive interference, with Kelvin waves. Further, the leading intraseasonal oscillation and Kelvin waves modes associated with these events were found to be rapidly growing with extratropical responses over Australia. The modal structures were again obtained with the primitive equation instability model with Kuo convection (Kuo, 1974) and evaporation-wind feedback (EVAP) as in Frederiksen (2002).

Krishnan et al. (2009), using an AGCM, found a positive feedback between the break periods of the Indian summer monsoon and midlatitudes. The suppressed convection induces Rossby waves that have the ability to amplify the midlatitude circulation anomalies through the wave dispersion. These circulation anomalies induce cold air advection and favor the intrusion of dry air over the Indian subcontinent. The cooling reduces the meridional thermal contrast and the dry air decreases the convective instability. The overall effect is a weakening of the monsoon flow and prolongation of the break phase.

A complex tropical-extratropical interaction was suggested by Jia et al. (2011) to explain the intraseasonal variability of the East Asian winter monsoon. As mentioned in section 2.1.3 the MJO phases influence the rainfall probability in the area through the influence on the northward moisture transport coming from the Bay of Bengal and the South China Sea. Based on analysis of observations, Jia et al. (2011) find that circulation anomalies in midlatitudes accompany the eastward propagation of MJO convective anomalies and develop into the Siberian blocking high and low trough over Central Asia. This pattern favors the cold air surges from the west and north high latitudes.

4. Forecasting Tropical-Extratropical Interactions

The time-lagged connection of the NAO to the MJO observed by Cassou (2008) and Lin et al. (2009) implies that part of the NAO subseasonal variability is generated by tropical forcing associated with MJO, so that knowing the initial state of the tropical MJO would help to predict the extratropical NAO. Lin and Brunet (2011) analyzed the reforecast experiments conducted with the Global Environmental Multiscale model and demonstrated that with a lead time up to about 1 month, the NAO forecast skill is significantly influenced by the existence of the MJO signal in the initial condition. (A reforecast is a forecast made from an initial condition in the past.) A strong MJO leads to a better

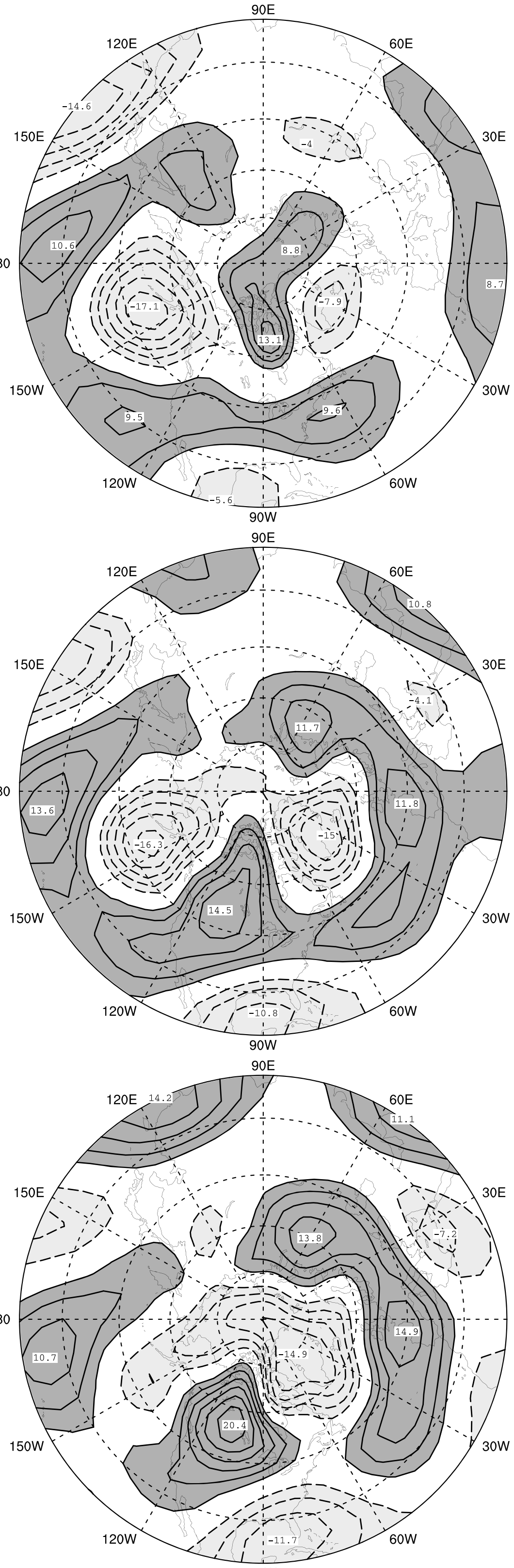

Figure 10. As in Figure 9 for the leading theoretical intraseasonal mode for the EVAP basic state (from contour interval is 3 (reprinted from Frederiksen & Lin, 2013, ©American Meteorological Society. Used with permission.)

NAO forecast skill than a weak MJO. An initial state with an MJO phase corresponding to a dipole tropical convection anomaly over the eastern Indian Ocean and western Pacific favors a more skillful NAO forecast than other MJO phases. These results indicate that it is possible to improve the skill of the NAO and the subseasonal forecast in the northern extratropics with an improved tropical initialization, a better prediction of the tropical MJO, and a better representation of the tropical-extratropical interaction in dynamical models.

Recent notable studies of the influence of tropical-extratropical interactions on forecast skill have used reforecasts made with full GCMs (see Hamill & Kiladis, 2014; Scaife et al., 2017; Vitart & Molteni, 2010; Vitart, 2014; Wu et al., 2016 for recent examples) and re-forecasts made with statistical models, in particular the Linear Inverse Model (LIM), (e.g., Newman et al., 2003; Pegion & Sardeshmukh, 2011; Winkler et al., 2001). Vitart and Molteni (2010) targeted the response of the extratropics to the MJO-related tropical diabatic heating in ECMWF forecasts out to 15 days and find that the response in the North Atlantic bears the same time-lagged relationship to the MJO as found by Cassou (2008) and Lin et al. (2009): positive NAO phase is preferentially excited about 10 days after the MJO heating is in the Indian Ocean (phase 3 of the MJO), while the MJO heating in the western Pacific (phase 6) is correlated with the appearance of the negative NAO phase 10 days later. However, the response in the reforecasts is somewhat too weak, a finding that Vitart and Molteni attribute to the bias in MJO phase speed in the reforecasts (the MJO propagates too slowly). Vitart (2014) assessed the change in the MJO–NAO interactions as the ECMWF forecast model has improved by examining a number of sets of reforecasts, each set made with a distinct version of the model. A clear gain in skill of the forecast of the NAO is seen with more recent versions of the model, and this gain can be unambiguously tied to the increase in MJO forecast skill. Vitart (2017) extended the evaluation of MJO-NAO analysis to the 10 S2S models (Vitart et al., 2017) and showed that in the forecasts the MJO teleconnections over the Euro-Atlantic sector are weaker than in observations.

The influence of the MJO is not limited to the North Atlantic, and general increases in the predictability of the extratropical atmosphere occur when the MJO is active (e.g., Jones et al., 2004b). For example, MJO influence is seen in forecasts of blocking highs, which occur in many extratropical locations. In particular, Hamill and Kiladis (2014) find (in reanalyses) an increase in blocking over western Europe associated with phase 6 of the MJO, a signal that is captured in short forecasts by the NOAA forecast model, but is damped by forecast day 8, likely due to errors in the forecast of the MJO, which propagates too slowly. They noted the inability of the NCEP Global Ensemble Forecast System to forecast blocking at long lead times even when a strong MJO was present. The association of western European blocking with MJO phase 6 is *consistent with* the lagged positive NAO phase response to MJO phase 3 if one takes into account the preferred regime transition path of positive NAO to Scandinavian Blocking to negative NAO phase (Cassou, 2008; Crommelin, 2004). The lag between MJO phase 3 and phase 6 can be more than 20 days, enough time for the positive NAO phase response to have developed and transitioned into the Scandinavian Blocking regime by the time MJO reaches phase 6. However, whether this is the correct physical explanation for the blocking signal seen by Hamill and Kiladis (2014) is not certain.

Surface air temperature over North America in winter is found to be anomalously warm 10–20 days following the MJO phase 3 (Lin & Brunet, 2009). Such a lagged connection implies predictability of North American temperature anomalies up to about 3 weeks given knowledge of the initial state of the MJO. Statistical models have shown that it is possible to extend the forecast range of North American temperature anomalies beyond 20 days, especially for strong MJO cases (e.g., Johnson et al., 2014; Rodney et al., 2013; Yao et al., 2011).

Insight into the importance of the tropical influence on forecasts in the extratropics has been obtained with the use of a class of statistical models, termed Linear Inverse Models (LIMs). A LIM is a model involving an empirical linear operator that evolves the low-frequency component of the atmospheric state and includes stochastic noise (Winkler et al., 2001). Both the operator and the noise process are derived from observational statistics of both extratropical and tropical fields, in particular tropical heating. The empirical-dynamical operator implicitly includes a linear representation of nonlinear processes such as the feedback of synoptic eddies on low-frequency flow. The LIM formulation allows tropical heating to coevolve with the extratropical circulation, so that the tropical-extratropical influences are fully captured. Winkler et al. (2001) find that the structure and magnitude of tropical diabatic heating are crucially important in forecasting the evolution of low-frequency circulation anomalies over time scales of 10 days and that neglecting diabatic forcing degrades forecast skill nearly everywhere. Newman et al. (2003) focus on the variability of potential forecast skill, which arises from the evolution of the tropical heating. Predictable variations are associated with those variations of the initial state projection on the growing singular vectors of the LIM's operator that have relatively large amplitude in the tropics. At times of strong projection, the Northern Hemispheric circulation is not only potentially more predictable but reforecasts at these times show higher skill. Pegion and Sardeshmukh (2011) emphasize the importance of identifying the small but significant fraction of forecast cases in which there is a potential for relatively high skill.

The two-way interaction of the MJO and NAO also implies that the forecast of the MJO would benefit from knowledge of the NAO state in the initial condition. In a hindcast experiment, it was found that a strong NAO leads to a better MJO forecast skill than a weak NAO (Lin & Brunet, 2011).

5. Summary and Remaining Challenges

Notable progress has been made in identifying intraseasonal tropical-extratropical interactions, including their centers of action and the statistical characteristics of the associated time series. A wide variety of physical mechanisms for these interactions have been identified, depending on the sense of cause and effect. The mechanisms of response of the extratropical circulation to organized deep tropical diabatic heating refer to such concepts as stationary wave theory, Rossby wave propagation characteristics, and feedback with the storm tracks. The response mechanisms by which tropical convection and circulation respond to extratropical incursions include local vertical velocity response to individual types of extratropical waves (on the very local level) and wave theory (on the more global level). Attempts at understanding true tropical-extratropical interactive coupling have generally invoked moist instability mechanisms encompassing both the tropics and extratropics, although in particular cases (e.g., the MJO cycling) individual feedback paths have been identified.

What is not clear is whether separate periods or instances of extratropical fluctuations forced by the tropics (and vice versa) exist in isolation or whether some degree of two-way coupling is always operating. Will focusing on individual pathways proves to be more valuable in extending the range of skillful forecasts of the atmosphere or is a truly interactive approach required?

Several specific challenges for tropical-extratropical interaction research are identified here, with brief suggestions for approaches to meet them. One challenge is to better understand the role of dynamical interactions that impact midtropospheric moisture. This appears to play a significant role in modulating the development of the MJO through its impacts on tropical convection, while the dynamical pathways of moisture transport into midlatitude and high latitude play a major role in atmospheric rivers and extreme precipitation events. A better handle on tropical/extratropical moisture transports and how they correlate with the MJO and associated teleconnections would be beneficial.

The background or basic state of the atmosphere is a feature that is of great importance to understanding and predicting tropical extratropical interactions. Errors in the basic state of many models include the incorrect simulation of the intratropical convergence zone, as described by Li and Xie (2014) in CMIP5 models. Such a fundamental error may negatively impact both the propagation of the MJO and its teleconnections. The mechanisms of the midlatitude response to the MJO involves the changing configurations of the Pacific jets, so errors in the climatology of the jet structure in models may degrade prediction of the MJO response. Finally, the basic state plays a very important role in the structure of global modes that arise from moist instability calculations that couple the tropics and extratropics.

It would be helpful for the community to identify specific hypotheses regarding the role of the basic state in the initiation and propagation of the MJO, in the formation of realistic MJO teleconnections, and in the influence on global instability modes. Key (common) basic state errors could be diagnosed from multimodel simulations (such as those of the CMIP and S2S projects), and controlled experiments could be conducted to understand the effect of the identified mean state errors.

One line of research that should be undertaken to better understand the evolution of the fully interactive system is the three-dimensional instability theory, which provides a natural extension of waves and instability theories. Expanding existing work to include linear moist primitive equation models with higher horizontal and vertical resolution, and a systematic understanding of the role of the basic state might be a fruitful approach. As with the dry primitive equation instability models, it might be valuable to extend such linearized calculations into the nonlinear regime to identify life cycles.

A better grasp of these fundamental issues should lead to improvements in the ability to predict the evolution of tropical and midlatitude circulation features for ranges beyond 2 weeks. Improvements in MJO forecasts, whether from more accurate representation of the physical processes involved or from the prediction of midlatitude precursors, should translate into better extratropical forecast skill. How much of an improvement is possible depends, among other things, on just how much extratropical variability is related to tropical forcing, which needs to be better quantified. Improved representations of the basic state in forecast models will positively impact the ability of midlatitude disturbances to propagate into the tropics and are essential for the forecasting of MJO downstream effects. As emphasized from the LIM approach, the tropics and midlatitudes evolve interactively.

The potential role of ocean variability in the subseasonal atmospheric variability is another line of research that is increasing its visibility with the advancement of high-resolution ocean modeling capabilities. A few recent studies suggest a significant relationship between the small-scale SST patterns and atmospheric circulation associated with the North-Pacific (Roundy & Verhagen, 2010; Ma et al., 2015) and Euro-Atlantic (Piazza et al., 2016; Wills et al., 2016) weather regimes. The results of Kilpatrick et al. (2016) obtained using an idealized configuration of the Weather Research and Forecasting (WRF) model also indicate that in the midlatitudes, the marine atmospheric boundary layer (MABL) dynamics is influenced by SST frontal zones and the atmospheric response depends on the alongfront background wind and the cross-front Rossby number. The value of the Rossby number determines the relative contribution of mechanisms involved in the MABL dynamics. Thus, physical processes associated with mesoscale ocean eddies can drive the intraseasonal teleconnections.

The research efforts to develop GCMs with cloud—and eddy—permitting resolutions or methods provide compelling evidence that improvements in the representation of tropical convection and large-scale climate features can be achieved (Kirtman et al., 2012; Stan & Xu, 2014; Wang, Sobel, et al., 2016) and highlight the importance of adopting these models into the forecast systems.

One aspect left out in this review is the role of timescale interactions on the links, teleconnections, and feedback between the tropics and extratropics on intraseasonal time scales. In particular, interannual and interdecadal variations of teleconnections may change the character of the intraseasonal teleconnections and their feedback. For example, summer rainfall over eastern China experiences decadal changes that manifest as a shift from a tripole pattern to a dipole pattern (Chen et al., 2012) and on interdecadal time scales the boreal summer intraseasonal oscillation increases the frequency of occurrence of phases (Kikuchi et al. 2012) with active convection over the Bay of Bengal and the eastern North Pacific along with pattern changes during the other phases (Wang et al., 2017). The associated changes in the atmospheric circulation can further affect the remote responses of this oscillation. On shorter time scales, the intraseasonal variability of the tropics is modulated by the ENSO (Bo et al., 2016; Fei et al., 2016), which can strengthen or weaken tropical intreaseasonal oscillations. Wang et al. (2017) explained the interdecadal variations of summer rainfall over southern China by changes in the frequency of individual phases of the oscillatory modes associated with regional monsoonal systems. Yuan et al. (2011) and Feldstein and Lee (2014) showed that the interdecadal poleward shift of the subtropical and eddy-driven jets in the Northern Hemisphere can be expressed in terms of the change in the frequency of occurrence of intraseasonal timescale teleconnection patterns.

Challenges for future studies of tropics-midlatitude interactions and teleconnections include a need for process-level diagnostics. Although it is possible to estimate teleconnections using linear analysis methods, new approaches designed to account for nonlinearities resulting from multiple space and time scale interactions are needed.

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