



Impacts of the atmospheric mean state on austral early-winter Indian Ocean teleconnections with Chile in CMIP5/CMIP6 models

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ABSTRACT

This study investigates the impacts of atmospheric mean state biases on the Indian Ocean teleconnection with Chile, using ERA5 reanalysis and CMIP5/CMIP6 models. During austral early-winter (May–June), Indian Ocean heating anomalies trigger a dipole of increased (reduced) precipitation over central (southern) Chile. The dipole is driven by cyclonic circulation anomalies induced by a zonal wavenumber-3 Rossby wave train originating from South Australia. Models that fail to simulate this rainfall dipole consistently exhibit an equatorward shift of the Southern Hemispheric eddy-driven jet stream. The associated waveguide weakening displaces the refraction location of zonal wavenumber-3 waves westward, thus shifting anticyclonic anomalies from the Amundsen Sea toward Chile. Consequently, the dry signal associated with the Indian Ocean teleconnection moves from southern to central Chile. This waveguide bias likely stems from a reduced meridional gradient of surface temperature, linked to an underestimated Antarctic Sea ice cover. For this reason, accurately simulating Antarctic Sea ice cover is crucial for a better representation of the Indian Ocean teleconnection to Chilean rainfall.

1. Introduction

The Indian Ocean (IO) basin hosts two primary modes of variability: the Indian Ocean Basin-wide Mode (IOBM) (Chambers et al., 1999) and the Indian Ocean Dipole (IOD) mode (Saji et al., 1999). The IOBM refers to a basinwide warming of Sea Surface Temperatures (SSTs), typically in response to climate forcings such as the El Niño–Southern Oscillation (ENSO) (Schott et al., 2009). The IOBM is characterized by a spectral peak at interannual timescales with a seasonal phase locking peaking during the austral autumn following an ENSO event (Lau and Nath, 2003; Xie et al., 2009; Schott et al., 2009; Deser et al., 2010; Taschetto and Ambrizzi, 2012).

The positive phase of the IOD manifests as a dipole pattern of anomalously cold (warm) SSTs over the equatorial eastern (western) IO, while its negative phase is characterized by anomalously warm (cold) SSTs in the eastern (western) IO (Saji et al., 1999). The IOD occurs on interannual timescales, typically peaking during boreal autumn (SON), often coinciding with an ENSO event (e.g., Zhang et al., 2015; Molteni et al., 2015), although many IOD events appear to develop independently of ENSO (e.g., Wang et al., 2016; Zhang et al., 2021). Furthermore, recent studies suggest that the development of IOD events can be

strongly influenced by extratropical climate systems, including the Arctic Oscillation and Arctic Sea ice anomalies (Cheng et al., 2023; Cheng et al., 2024; Cheng et al., 2025).

While ENSO (Cai et al., 2020 for a review) and IOD (Saji et al., 2005; Chan et al., 2008; Sena and Magnusdottir, 2021; Kayano et al., 2021; Reboita et al., 2021; Gomes et al., 2021; Iacovone et al., 2024) impacts on Southern America (SA) have been extensively documented, the IOBM and austral autumn IO teleconnections with SA have attracted less attention. For instance, Taschetto and Ambrizzi (2012) investigated the influence of Indian Ocean SST variability on South American circulation using reanalysis data and General Circulation Model (GCM) experiments. Their findings indicate that, during austral autumn, the Indian Ocean affects SA through both tropical and extratropical pathways. The tropical connection occurs via a reorganization of the Walker circulation, leading to anomalous subsidence over northeastern SA. The extratropical teleconnection appears as a Rossby wave train originating from the southern IO, which, propagating southeastward, induces an anticyclonic circulation over SA. This circulation pattern, in turn, enhances moisture convergence over the La Plata region.

Kayano et al. (2021a) examined the relative role of the IOBM and ENSO in modulating interannual rainfall variability over SA, with a

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particular focus on the rainy season (i.e., austral summer). They found that the IOBM induces an anomalous wet condition over northeastern Brazil via an anticyclonic circulation over SA north of 20°S and a cyclonic circulation over the southwestern Atlantic between 20°S and 40°S, which enhances moisture flux over northeastern Brazil. Similarly, [Kayano et al. \(2021b\)](#) analyzed the IOBM influence on SA rainfall variability during austral winter and found that ENSO and the IOBM have contrasting impacts over the tropical and subtropical SA, with the IOBM which tends to enhance dry conditions over the 5°S–20°S band. [Souza et al. \(2023\)](#) investigated the modulation of the Interdecadal Pacific Oscillation (IPO) on the IOBM impacts during the decaying phase of ENSO extremes. During a positive IPO, a warm IOBM induces subsidence over SA via rearrangement of the Walker circulation. Conversely, during a negative IPO event, a cold IOBM supports upward motion over the Indian Ocean, which in turn alters the Walker circulation and enhances precipitation over SA.

Interannual modes of variability are often associated with severe socio-economic impacts over SA including crop-failure ([Poveda et al., 2001](#); [Anderson et al., 2017](#); [Hamed et al., 2023](#); [Aldunce Ide and González, 2009](#)), prolonging drought and wildfire events ([Aragão et al., 2018](#); [Jiménez-Muñoz et al., 2016](#); [Cordero et al., 2024](#); [Urrutia-Jalabert et al., 2018](#)), extreme flooding ([Grimm and Tedeschi, 2009](#); [Isla, 2018](#); [Tedeschi et al., 2016](#)), and an anomalous increase of vector-borne disease outbreaks ([Bouma et al., 1997](#); [Gagnon et al., 2002](#); [Mantilla et al., 2009](#)). For instance, during the austral winter preceding the 1997–1998 super El Niño, intense rainfall triggered severe flooding in Central Chile, leading to significant economic losses ([Aldunce Ide and González, 2009](#)). During the austral summers of 2017 ([Bowman et al., 2019](#); [De La Barrera et al., 2018](#)) and 2023 ([Cordero et al., 2024](#)), two record-breaking wildfires occurred, driven by prolonged “fire weather” conditions and coinciding with El Niño events.

Models’ systematic errors, such as mean state biases, can hinder the simulation of teleconnection patterns (e.g., [Di Carlo et al., 2022](#); [Sabatani and Gualdi, 2025](#)). [Di Carlo et al. \(2022\)](#) showed that a stronger Pacific jet stream can meridionally shift the geopotential response associated with Rossby waves forced by ENSO over North America during boreal winter. Using the historical simulations of the 5th and 6th phase Coupled Model Intercomparison Project (CMIP) models, [Sabatani and Gualdi \(2025\)](#) demonstrated that models which poorly simulate the wintertime ENSO teleconnection to the North Atlantic also share stronger Pacific and Atlantic subtropical jet streams. These stronger waveguides, in turn, constrain Rossby waves to lower latitudes, hence affecting the remote atmospheric response associated with ENSO.

To our knowledge, few studies have examined the impact of the IOBM on South America (i.e., [Taschetto and Ambrizzi, 2012](#); [Kayano et al., 2021a, 2021b](#); [Souza et al., 2023](#)), and none have specifically investigated its effects on Chile. This study fills that gap by providing new evidence on Indian Ocean teleconnections with Chile, focusing on austral early-winter (May–June, MJ), when this teleconnection is most active (see Section 3.1, [Fig. 2](#)). The analysis focuses on the extratropical pathway, which we investigate using a large ensemble of CMIP5 and CMIP6 models in what constitutes the first multi-model intercomparison of this teleconnection. Furthermore, we assess the role of the models’ atmospheric mean state in modulating the pathway’s strength and structure.

The next section describes the data and methodology implemented in this study. The Indian Ocean teleconnection with Chile in reanalysis is investigated in Section 3, along with the results regarding the impacts of the atmospheric mean state in CMIP5/6 models. The summary and discussions of the results are provided in Section 4.

2. Data and methodology

2.1. Reference and model data

The ERA5 reanalysis dataset ([Hersbach et al., 2020](#)) is used as the

reference dataset in this study. The analysis focuses on monthly means of precipitation, specific humidity, geopotential height, and zonal and meridional wind components over the period 1979–2024. As part of a sensitivity test, monthly precipitation data from the Global Precipitation Climatology Project version 2 ([Adler et al., 2003](#)) were also examined, yielding results consistent with those derived from ERA5 (not shown). Observed SST data is retrieved from the Hadley Centre Global Sea Surface Temperature (HadISST) dataset, covering 1979–2024 ([Rayner et al., 2003](#)).

A large ensemble of historical simulations from the CMIP5 ([Taylor et al., 2012](#)) and CMIP6 ([Eyring et al., 2016](#)) multi-model archives, comprising 37 CMIP5 models and 56 CMIP6 models, has been analyzed. The model data includes monthly geopotential height, zonal and meridional winds, precipitation, sea surface temperature, and sea ice concentration. To maintain equal weighting among models, only one ensemble member per model was selected. For CMIP6, the realization ‘r1i1p1f1’ was chosen where available, and for CMIP5, ‘r1i1p1’; alternative realizations were used when these were unavailable. The CMIP6 and CMIP5 models, comprising their associated modelling center, country, and ensemble member are listed in Tables S1 and S2, respectively. The analysis focuses on the final 64 years of each model’s historical simulation, specifically 1950–2014 for CMIP6 and 1940–2004 for CMIP5. Both reference and model data has been bi-linearly interpolated to a common 2.5°x2.5° grid before analysis, and to a common 1.0°x1.0° for data visualization.

2.2. Indices

To assess the role of Indian Ocean heating anomalies in influencing the remote response over SA, this study utilizes a rainfall-based index, consistently with previous studies ([Molteni et al., 2015](#); [Abid et al., 2020](#); [Abid et al., 2021](#); [Molteni and Brookshaw, 2023](#); [Sabatani and Gualdi, 2025](#)). This is because large-scale teleconnection patterns are largely influenced by anomalous heating from strong tropical convection ([Gill, 1980](#); [Hoskins and Karoly, 1981](#)).

During austral early-winter, the ENSO-induced inter-basin connection with the Indian Ocean generates a dipole of precipitation anomalies, with increased rainfall over the south equatorial western Indian Ocean and reduced rainfall over the equatorial eastern Indian Ocean and Maritime Continent ([Fig. S1](#)). This precipitation pattern resembles the dipole structure associated with the IOD, although the western pole of enhanced precipitation is shifted southward by approximately 10°. Thereby, the rainfall index is defined as the difference between standardized detrended precipitation anomalies averaged over the western Indian Ocean (40°E–90°E, 0°–20°S; black dashed box in [Fig. S1](#)) and Maritime Continent (100°E–140°E, 10°N–10°S; black dashed box in [Fig. S1](#)), hereafter referred to as the Indian Ocean and Maritime Continent (IOMC) index.

SST anomalies in the Indian Ocean, particularly in the western basin, are often weakly coupled with atmospheric heating and rainfall responses ([Molteni et al., 2015](#)). Therefore, to directly capture large-scale atmospheric heating, we use the IOMC rainfall index rather than a conventional SST-based index. While distinct, our rainfall index is strongly linked to the IOBM. During May–June, the correlation between the IOMC rainfall index and the IOBM SST index, defined as the leading principal component of SST anomalies over the basin (40°E–100°E, 30°N–30°S), is 0.51. Sensitivity tests also suggest that the IOMC rainfall index is linked to a stronger mid-latitude upper-tropospheric response than the IOBM SST index is (not shown). This finding underscores the importance of using rainfall-based indices when studying extratropical teleconnections.

We also examined the potential influence of the IOD mode, particularly as recent studies show an increase in IOD events peaking during austral winter ([Sun et al., 2022](#)). The correlation between the IOMC rainfall index and the Dipole Mode Index (DMI; [Saji and Yamagata, 2003](#)) is 0.37 for May–June. A comparison of these relationships

indicates that the May–June rainfall variability captured by the IOMC index is more robustly associated with the IOBM ($r = 0.51$) than with the IOD mode ($r = 0.37$).

To assess the impact of heating anomalies over the tropical Pacific the Niño 3.4 precipitation index (hereafter Niño 3.4 index) is defined as the average of standardized monthly precipitation anomalies within the Niño 3.4 region (170°E–120°W, 5°N–5°S; outlined in black in Fig. S1), in agreement with previous studies (Molteni et al., 2015; Abid et al., 2020; Abid et al., 2021; Molteni and Brookshaw, 2023; Sabatani and Gualdi, 2025).

2.3. Statistical analysis and testing

Monthly detrended anomaly fields are regressed onto standardized indices. To isolate the contributions of heating anomalies in the Pacific basin from those in the Indian Ocean to regression patterns, we apply linear partial regression analysis. In this respect, the IOMC index can be decomposed as follows:

$$\alpha_{IOMC}(t) = \beta_{IOMC}(t) + p^* \alpha_{Ni\tilde{no} \ 3.4}(t), \quad (1)$$

where α_{IOMC} and $\alpha_{Ni\tilde{no} \ 3.4}$ denote the IOMC and Niño 3.4 indices, respectively, and β_{IOMC} represents the *pure* IOMC index (i.e., independent of the Niño 3.4 variability). The regression coefficient p^* is computed as follows:

$$p^* = \frac{\sum_{t=0}^N [\alpha_{IOMC}(t) \alpha_{Ni\tilde{no} \ 3.4}(t)]}{\sum_{t=0}^N \alpha_{Ni\tilde{no} \ 3.4}^2(t)} \quad (2)$$

From now on, we will refer to the pure IOMC regression as the pure IOMC teleconnection. In this way, the linear component of the May–June ENSO signal, which by construction accounts for variability associated with both El Niño and La Niña, is removed from the IOMC signal. The significance testing for ERA5 is assessed by implementing a Student's two-tailed t -test, while for the Multi-Model Ensemble (MME) mean a Student's one-sample t -test is implemented.

2.4. Atmospheric Rossby wave analysis

The Tropical Rossby Wave Source (TRWS) formula is a diagnostic tool used to examine anomalous vorticity tendency in the tropical upper troposphere (Sardeshmukh and Hoskins, 1988; Qin and Robinson, 1993). It captures how departures in the divergent wind field contribute to changes in absolute vorticity, and is expressed as the advection of climatological absolute vorticity by the anomalous irrotational component of the wind:

$$TRWS = -v'_x \nabla(\bar{\zeta} + f). \quad (3)$$

In eq. (3), v'_x indicates the anomalous irrotational wind, $\bar{\zeta}$ and f are the relative and planetary climatological vorticities, respectively. To obtain v'_x , the velocity potential (χ) is first computed from wind divergence following Mezzina et al. (2022). The irrotational wind field is then derived as the gradient of χ . The TRWS is calculated at the 200 hPa level in the upper troposphere, where Rossby wave sources typically reach their maximum intensity (Scaife et al., 2017).

The wave flux associated with Rossby waves propagation is investigated using the Wave Activity Flux (WAF) formula (Takaya and Nakamura, 2001), which provides a phase-independent estimation of waves' packets propagation. The formulation is based on the geostrophic approximation and has been widely applied in previous research (e.g., Ruggieri et al., 2021; Joshi et al., 2021). The zonal and meridional components of the WAF are expressed as follows:

$$WAF_x = \frac{p \cos \phi}{2\bar{U}} \left(\frac{\bar{u}}{a^2 \cos^2 \phi} [(\partial_x \psi')^2 - \psi' \partial_x^2 \psi'] + \frac{\bar{v}}{a^2 \cos \phi} (\partial_x \psi' \partial_y \psi' - \psi' \partial_{xy}^2 \psi') \right) \quad (4a)$$

$$WAF_y = \frac{p \cos \phi}{2\bar{U}} \left(\frac{\bar{u}}{a^2 \cos \phi} [\partial_x \psi' \partial_y \psi' - \psi' \partial_{xy}^2 \psi'] + \frac{\bar{v}}{a^2} [(\partial_y \psi')^2 - \psi' \partial_y^2 \psi'] \right) \quad (4b)$$

where p is normalized pressure (i.e., pressure divided by 1000 hPa), a is the Earth radius, ϕ is the latitude, $\bar{u}$ and $\bar{v}$ are the climatological zonal and meridional winds, $\bar{U}$ is the magnitude of the horizontal climatological wind vector $\vec{U} = (\bar{u}, \bar{v})$, and ∂_x and ∂_y are the partial derivatives along the zonal and meridional directions, respectively. The term ψ' represents the anomalous geostrophic streamfunction, defined as the anomalous geopotential φ' divided by the Coriolis parameter f .

The pathway of atmospheric Rossby wave is affected by the structure of the upper-tropospheric waveguide. As a proxy for the atmospheric waveguide, we analyze the meridional gradient of climatological absolute vorticity alongside the climatological zonal wind field, following the approach of Hoskins and Ambrizzi (1993). These diagnostics help identify regions where Rossby waves are likely to be trapped and guided by the mean flow. To further characterize the waveguide, we calculate the total stationary wavenumber, which determines the range of stationary Rossby wave modes that can propagate along it. This wavenumber is derived from the linear theory of barotropic Rossby waves and is given by the following equation:

$$K = \sqrt{(\beta^* / \bar{u})} = \sqrt{k^2 + l^2}, \quad (5)$$

where β^* is the meridional gradient of absolute vorticity, $\bar{u}$ is the climatological upper tropospheric zonal wind. k and l indicate the zonal and meridional wavenumbers, respectively.

2.5. Rossby wave ray tracing algorithm

An algorithm, based on linear theory and the stationary and barotropic assumptions, is employed to investigate the pathways of Rossby wave rays (Hoskins and Karoly, 1981; Hoskins and Ambrizzi, 1993). The algorithm involves computing the zonal and meridional components of the group velocity for stationary barotropic Rossby waves, which are expressed as:

$$c_{gx} = \frac{2[\bar{u}]^2 k^2}{(\beta - \partial_y^2 [\bar{u}])} \quad (6a)$$

$$c_{gy} = \frac{2[\bar{u}]^2 k \left(\frac{\beta - \partial_y^2 [\bar{u}]}{[\bar{u}]} - k^2 \right)^{1/2}}{(\beta - \partial_y^2 [\bar{u}])}, \quad (6b)$$

where $[\bar{u}]$ is the zonal average of the upper tropospheric climatological zonal wind, ∂_y is the partial derivative along the latitudinal axis, k is the zonal wavenumber, and β is the meridional gradient of the Coriolis parameter.

The ray tracing algorithm is initiated from a chosen starting point, typically located near maxima or minima of upper-tropospheric Rossby wave sources or significant geopotential height anomalies. For a fixed zonal wavenumber k ranging between 2 and 4, values characteristic of tropical-extratropical teleconnection patterns (Hoskins and Ambrizzi,

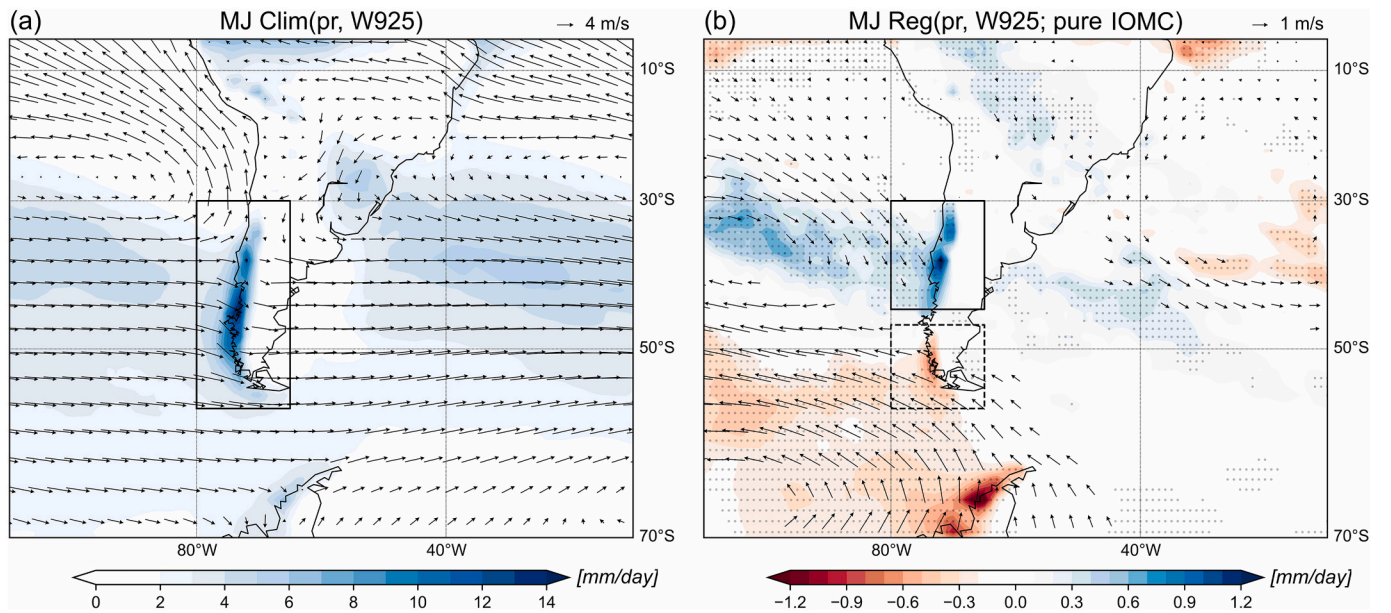


Fig. 1. (a) ERA5 May–June (MJ) climatology of precipitation (units mm/day) and lower tropospheric (925 hPa) wind (vectors; units m/s with unit vector of 4 m/s); (b) regression of MJ precipitation anomalies (units mm/day per standard deviation) and lower tropospheric wind (vectors; units m/s per standard deviation, unit vector of 1 m/s) onto the pure IOMC index. The light-gray stippling indicates the grid points in which the regression is significant at the 95 % level of confidence with respect to a two-tailed t -test. Vectors in (b) are displayed at the 90 % level of confidence. The black box in Fig. 1a indicates central and southern Chile. The black continuous and dashed boxes in Fig. 1b indicate the domains used to compute the DCR index.

1993), the zonal and meridional group velocities (from Eqs. 6a and 6b) are computed at the initial point. These velocities are then used to advance the ray trajectory forward in time and space over a 10-day period, using a 1-h integration timestep. This timescale is representative of the typical duration required for stationary Rossby waves to propagate from the tropics to midlatitudes (Scaife et al., 2017; Ayarza-güena et al., 2018; Sabatani and Gualdi, 2025). Sensitivity tests with longer propagation windows (e.g., 15–20 days, not shown) yield similar

results.

Waves with a constant zonal wavenumber k are refracted at critical latitudes, where the total stationary wavenumber K equals k , or equivalently, where the meridional wavenumber l becomes zero and ray propagation is purely zonal. To identify these turning latitudes, the algorithm computes K^2 (as defined in Eq. 5) from the zonally averaged zonal wind at each time step. For a given squared zonal wavenumber k^2 it then checks whether $l^2 = K^2 - k^2$ is negative. If so, the ray cannot

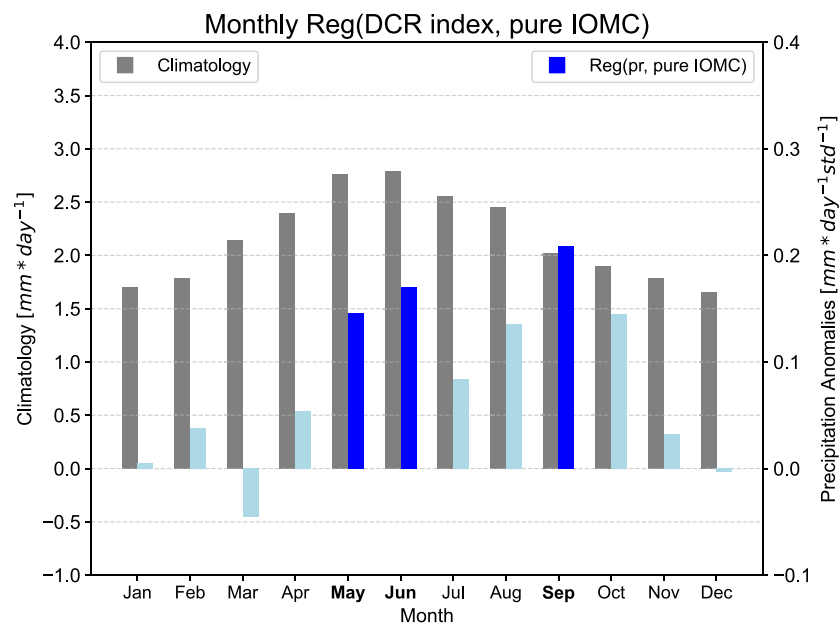


Fig. 2. Monthly precipitation climatology (gray bars; mm/day) averaged over central and southern Chile (80°W–65°W, 30°S–57°S; black box in Fig. 1a) from the ERA5 dataset for 1979–2024. Also shown are the monthly regressions of the Dipole Chile Rainfall (DCR) index (mm/day per standard deviation) onto the pure IOMC index (blue bars). The left y-axis represents climatology, while the right y-axis corresponds to regression values. Regression bars are colored blue when statistically significant at the 95 % confidence level based on a two-tailed t -test. Months are displayed in bold if the regression of the DCR index is significant at the 95 % level of confidence. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

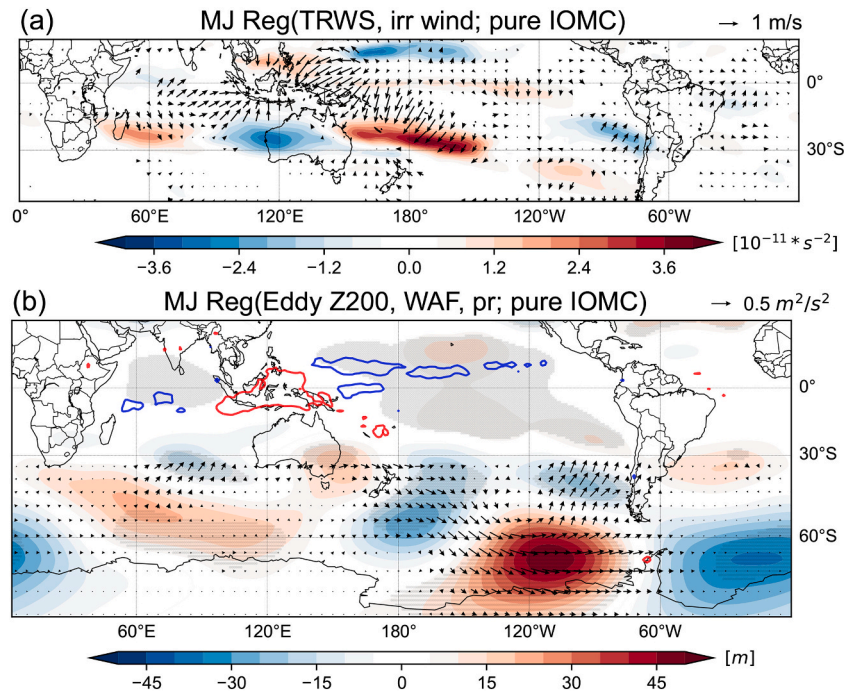


Fig. 3. (a) ERA5 MJ 200 hPa TRWS and 200 hPa irrotational wind anomalies regressed against the pure IOMC index. Shading indicates the regression of TRWS anomalies (units 10^{-11} s^{-2} per standard deviation), and vectors represent the irrotational wind (units m/s per standard deviation) with a unit vector of 1 m/s. Only TRWS that is significant at the 95 % level of confidence with respect to a two-tailed t -test is shown. (b) As in (a) but for the regression of 200 hPa eddy geopotential height (Eddy Z200), WAF, and precipitation anomalies. Eddy Z200 is defined as the deviation of geopotential height at 200 hPa from its zonal mean. Shading indicates the regression of eddy Z200 anomalies (units m), and vectors represent the WAF (units m^2/s^2 per standard deviation) with a unit vector of $0.5 \text{ m}^2/\text{s}^2$. Blue (red) contours in (b) represent significant precipitation anomalies at 0.5 (-0.5) mm/day per standard deviation. The light-gray stippling indicates the grid points in which the regression is significant at the 95 % level of confidence with respect to a two-tailed t -test. Vectors are displayed at the 90 % level of confidence. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

propagate meridionally and undergoes total refraction, with the angle of refraction equal to the angle of incidence. As a result, longer waves (smaller k) can propagate farther poleward, since the condition $K^2 > k^2$ is more easily satisfied. In contrast, shorter waves (larger k) are refracted closer to the equator, as they encounter turning latitudes sooner in regions where K is comparatively small.

We acknowledge that the ray tracing algorithm operates within a zonally symmetric framework and relies on the barotropic, stationary, and linear approximations, which imposes limitations on the robustness of the conclusions. Therefore, the results should be interpreted with caution (Scaife et al., 2017; Mezzina et al., 2023). It is also worth to note that barotropic Rossby waves propagate predominantly in the zonal direction, as c_{gx} increases at a faster rate of c_{gy} , thereby enhancing their effectiveness in driving the remote atmospheric responses (Hoskins and Ambrizzi, 1993; Hoskins and Karoly, 1981).

2.6. Moisture flux analysis

To identify sources and sinks of atmospheric moisture, we compute the vertically integrated moisture flux as:

$$Q = \frac{1}{g} \int_{300 \text{ hPa}}^{1000 \text{ hPa}} q \vec{U} dp, \quad (7)$$

where g is the gravitational acceleration, q is the specific humidity, and $\vec{U}$ is the total horizontal wind. This integral quantifies the total horizontal transport of moisture through the atmospheric column. Following Chen (1985) and Ratnam et al. (2014), the resulting moisture flux is further decomposed into its irrotational (divergent) and rotational (non-divergent) components, as follows:

$$Q = \frac{1}{g} \int_{300 \text{ hPa}}^{1000 \text{ hPa}} q \left(\vec{U}_{rot} + \vec{U}_{irr} \right) dp, \quad (8)$$

where $\vec{U}_{rot}$ and $\vec{U}_{irr}$ indicate the rotational and irrotational components of the total horizontal wind, respectively. This Helmholtz decomposition helps to isolate the divergent part of the flow, which is directly related to moisture convergence (sources) and divergence (sinks).

3. Results

3.1. IOMC teleconnection in reanalysis

Fig. 1a shows the austral early-winter climatology of precipitation and lower tropospheric circulation over SA. The circulation is dominated by an anticyclonic system over the subtropical southeastern Pacific Ocean, which drives westerly flow over central and southern Chile (80°W – 65°W , 30°S – 57°S ; black box in Fig. 1a). As a result, the anticyclonic structure enhances moisture flux over central and southern Chile, as shown by the irrotational (Fig. S2a) and rotational (Fig. S2b) components of the vertically integrated climatological moisture flux. Consequently, central and southern Chile receive most of their rainfall between April and August, as depicted by the vertical gray bars in Fig. 2.

The regression of the MJ precipitation and lower tropospheric circulation anomalies against the pure IOMC index is shown in Fig. 1b. The pure IOMC teleconnection is associated with a dipole pattern with increased (reduced) precipitation over central (southern) Chile. The lower tropospheric circulation related to the pure IOMC teleconnection features a cyclonic pattern over the subtropical southeastern Pacific, which enhances (reduces) the climatological moisture flux over central (southern) Chile.

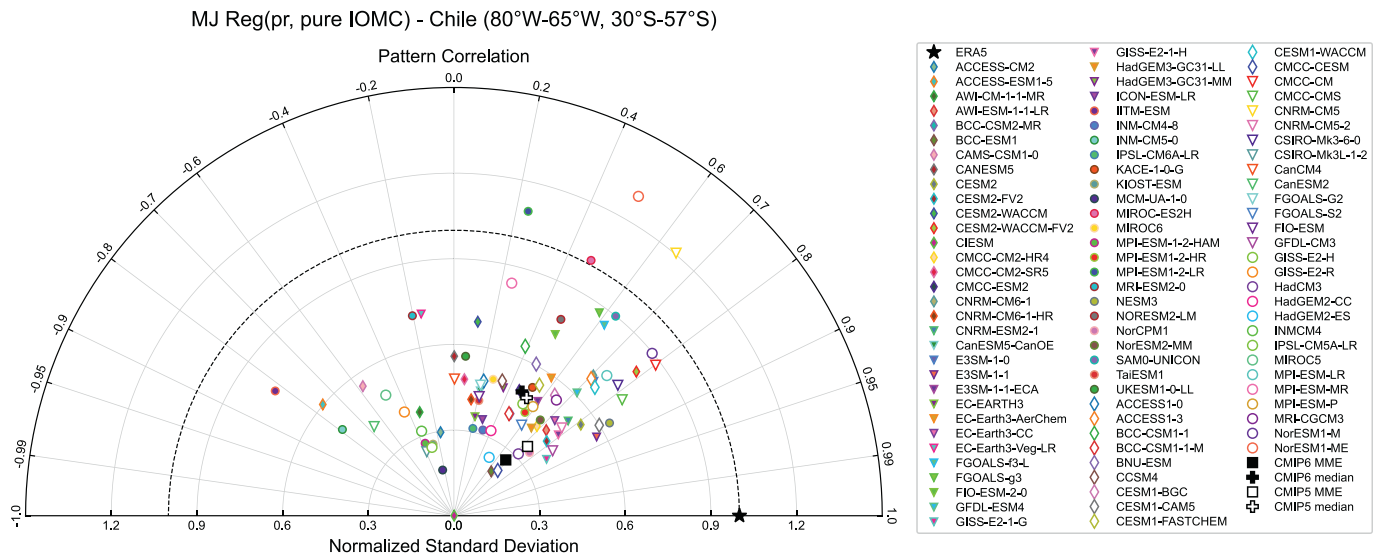


Fig. 4. Taylor diagram based on the MJ precipitation anomalies regressed against the pure IOMC index over central and southern Chile (80°W–65°W, 30°S–57°S) relative to ERA5. The black star represents the reference dataset, the black-filled square indicates the MME of CMIP6 models, and the white-filled square represents the MME of CMIP5 models. The black filled plus symbol indicates the median of CMIP6 models while the white filled plus symbol indicates the median of CMIP5 models. The color-filled symbols denote individual CMIP6 models, while the white-filled symbols denote individual CMIP5 models. The horizontal axis represents the standard deviation of the spatial pattern normalized over the reference standard deviation. The outer semicircle denotes the pattern correlation.

We define a Dipole Chile Rainfall (DCR) index as the difference between the average precipitation anomalies over central (80°W–65°W, 30°S–45°S; black continuous box in Fig. 1b) and southern (80°W–65°W, 47°S–57°S; black dashed box in Fig. 1b) Chile. Fig. 2 shows the monthly regression of the DCR index against the pure IOMC index, obtained after removing the ENSO influence.

The pure IOMC teleconnection has a positive and significant impact on Chilean rainfall in May, June, and September. Since the IOD mode is active in September, this month falls beyond the scope of this study, and thereby it is not considered in our analysis. The averaged DCR index regression for MJ is 0.16 mm/day per standard deviation. We now investigate how the extratropical IOMC teleconnection contributes to the maintenance of the positive Chilean rainfall anomalies during MJ.

First, we examine the characteristics of the atmospheric waveguide in the reference dataset (Fig. S3). The climatological 200 hPa zonal wind field (Fig. S3a) establishes both subtropical and mid-latitude waveguides, represented by the meridional gradient of absolute vorticity (Fig. S3b). The subtropical waveguide, characterized by total stationary wavenumbers ranging from 5 to 7 (Fig. S3c), is bounded poleward by a region of imaginary total stationary wavenumber K (Eq. 5), which can inhibit the meridional propagation of Rossby waves (McIntosh and Hendon, 2018). The bounding region is associated with the strong meridional gradient of absolute vorticity (Fig. S3b), related to a meridional curvature of the jet stream. In contrast, the mid-latitude waveguide supports total stationary wavenumbers between 2 and 4, facilitating the meridional propagation of Rossby waves.

Fig. 3a shows the upper tropospheric TRWS and irrotational wind anomalies regressed onto the pure IOMC index. The IOMC heating anomalies form a tripole pattern, with increased rainfall over the off-equatorial South Indian Ocean and the off-equatorial western Pacific (blue contours in Fig. 3b) and decreased rainfall over the Maritime Continent (red contours in Fig. 3b). In the tropics, positive (negative) heating anomalies are balanced by upper tropospheric divergence (convergence). The resulting divergent winds interact with the atmospheric waveguide, generating vorticity tendency via vorticity advection and forming a TRWS. The TRWS pattern features three primary sources: one east of Madagascar and two over the subtropical western and eastern Australia. We note that the strong positive TRWS over eastern Australia can be interpreted as a response to the adjustment of the

Table 1

Statistics from the Taylor diagram analysis.

	MME corr	MME std	Median corr	Median std	Models with negative corr
CMIP5	0.73	0.35	0.52	0.49	16 %
CMIP6	0.69	0.27	0.49	0.50	19 %

Values of pattern correlation and normalized standard deviation of the MME and median of CMIP5/CMIP6 models relative to ERA5. The two rightmost columns indicate the percentage of CMIP5 and CMIP6 models with negative pattern correlation relative to ERA5.

Walker Circulation over the western tropical Pacific to IOMC rainfall anomalies.

Although in ERA5 the TRWS exhibits a tripolar structure, only the eastern Australian TRWS appears to effectively generate Rossby waves in the Southern Hemisphere. The limited impact of the other sources may be due to the presence of the bounding region of imaginary total wavenumber (Fig. S3c), which acts as a barrier for the meridional propagation of Rossby waves originating from the subtropical South Indian Ocean. The Southern Pacific Rossby wave propagates southeastward before being refracted toward the tropics around 110°W–65°S, over the Amundsen Sea, eventually reaching Chile and inducing cyclonic anomalies. At the refraction point, there is also a subtle indication of Rossby wave splitting, with part of the wave continuing its trajectory over the South Atlantic, where it forces cyclonic anomalies over the Weddell Sea.

The wave pattern is characterized by alternating geopotential height anomalies with a zonal wavelength of approximately 120°, corresponding to a zonal wavenumber of 3. The meridional scale of the wave spans the 30°N–65°N latitude band, indicating a meridional wavenumber of approximately 6. The zonal and meridional wavenumbers are consistent with the atmospheric mean state (Fig. S3c), which exhibits total stationary wavenumbers of 5–7 along the subtropical waveguide.

3.2. IOMC teleconnection in CMIP models

The ability of CMIP5 and CMIP6 models to simulate the pure IOMC teleconnection with Chile is assessed through a Taylor diagram (Taylor,

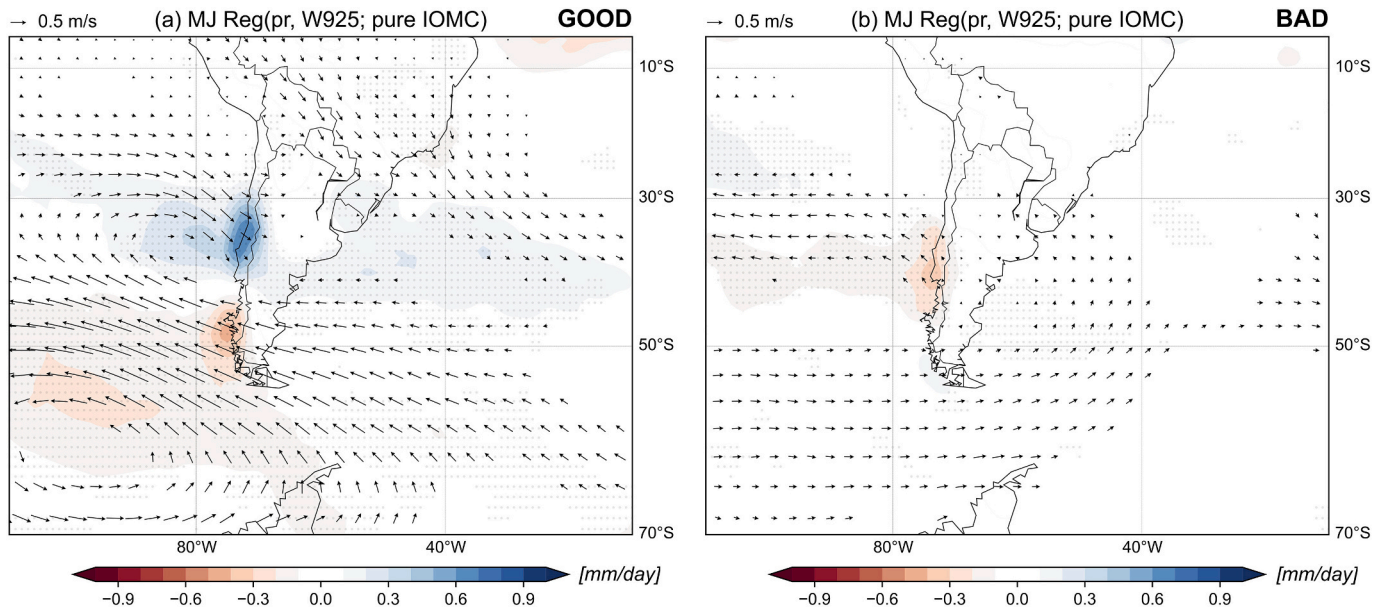


Fig. 5. Ensemble average of the MJ precipitation (units mm/day per standard deviation) and 925 hPa wind (vectors; units m/s per standard deviation, unit vector of 0.5 m/s) anomalies regressed against the pure IOMC index for (a) ensemble mean of good models (GOOD) and (b) ensemble mean of bad models (BAD). The light-gray stippling indicates the grid points in which the average is significant at the 95 % level of confidence with respect to a one-sample *t*-test. Vectors are displayed at the 90 % level of confidence.

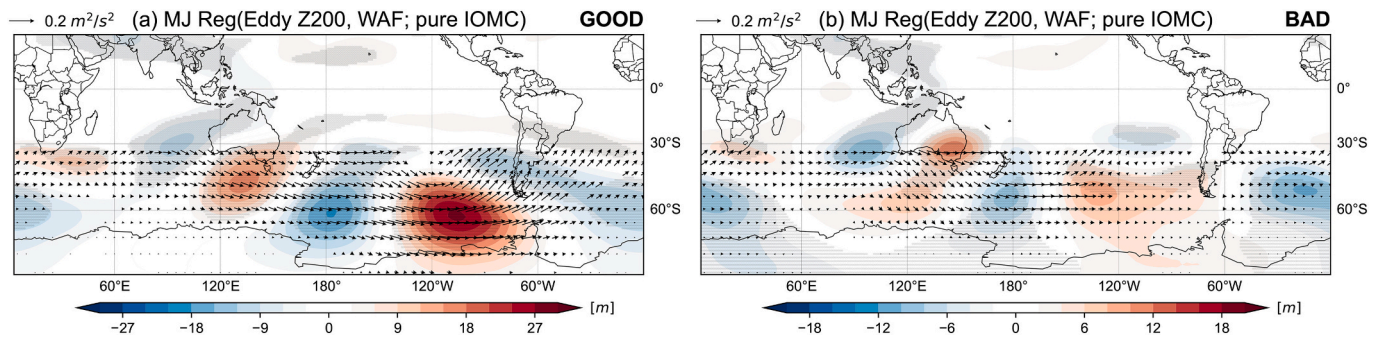


Fig. 6. Ensemble average of the MJ eddy 200 hPa geopotential height (Eddy Z200, units m per standard deviation) and WAF (vectors; units m^2/s^2 per standard deviation, unit vector of $0.2 m^2/s^2$) anomalies regressed against the pure IOMC index for (a) good models (GOOD) and (b) bad models (BAD). The light-gray stippling indicates the grid points in which the average is significant at the 95 % level of confidence with respect to a one-sample *t*-test. Vectors are displayed at the 90 % level of confidence.

2001), shown in Fig. 4, based on the regression of the MJ precipitation anomalies against the pure IOMC index over central and southern Chile (80°W–65°W, 30°S–57°S; black box in Fig. 1a).

Table 1 summarizes the statistics from the Taylor diagram analysis. Overall, CMIP models reasonably capture the pure IOMC teleconnection, with MME correlations of 0.73 for CMIP5 and 0.69 for CMIP6. However, no improvement is observed from CMIP5 to CMIP6 in simulating the teleconnection with Chile. The models tend to underestimate its amplitude, as indicated by the median of the normalized standard deviation, which is 0.52 for CMIP5 and 0.49 for CMIP6. On average, they simulate the teleconnection at approximately half the amplitude of the reference dataset.

By calculating the spatial correlation relative to ERA5, we performed a clustering process to categorize models based on their ability to simulate the pure IOMC teleconnection with Chile. Models that exhibit a pattern correlation greater than 0.80 are classified as “good” models. Good models include: E3SM-1-1, NESM3, GISS-E2-1-G, CMCC-CMS, CESM1-CAM5, CIESM, CESM2, and GFDL-CM3. On the other hand, models with a pattern correlation smaller than -0.40 are classified as “bad” models. Bad models include: IITM-ESM, INM-CM5-0, ACCESS-ESM1-5, CanESM2, CAMS-CSM1-0, CSIRO-Mk3L-1-2, GISS-E2-R, and

MIROC5.

Fig. 5 shows the regression of precipitation and lower tropospheric circulation anomalies onto the pure IOMC index averaged over good (Fig. 5a) and bad (Fig. 5b) models. In good models, the average regression of the DCR index onto the pure IOMC index is 0.16 mm/day per standard deviation, well in agreement with observations. The upper and lower quartiles are 0.19 and 0.13 mm/day per standard deviation, respectively. In contrast, bad models simulate a significant DCR regression value of -0.06 mm/day per standard deviation, with upper and lower quartiles of -0.01 and -0.09 mm/day, respectively. We note that the DCR value in bad models is about three times smaller compared to that in good models, suggesting a weaker teleconnection in the former cluster.

Good models effectively capture the cyclonic circulation anomalies associated with the pure IOMC teleconnection over Chile, albeit with reduced intensity compared to ERA5 (Fig. 3b). This cyclonic pattern, acting alongside the mid-latitude anticyclonic anomaly, enhances moisture transport over central Chile while weakening it over southern Chile, leading to positive and negative precipitation anomalies, respectively. In contrast, bad models simulate a weak but significant anticyclonic anomaly over the mid-latitude southeastern Pacific, which

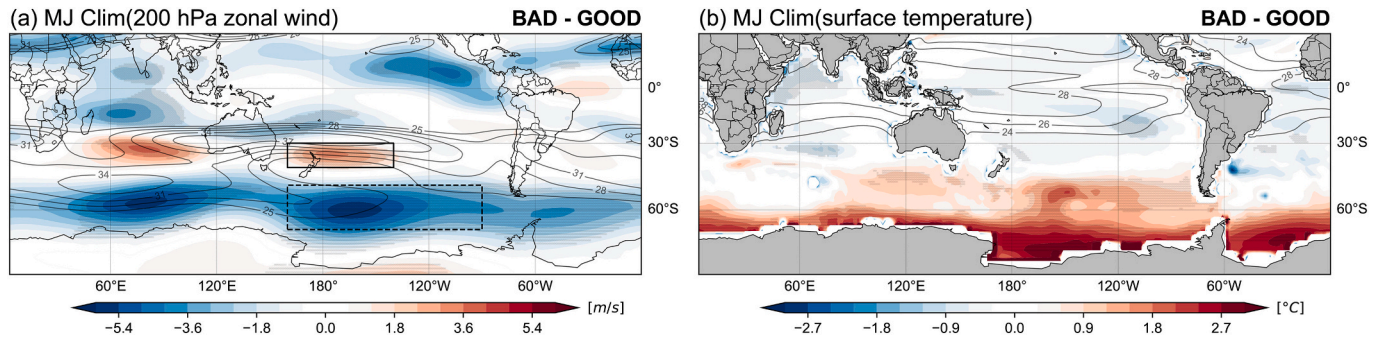


Fig. 7. Difference between the ensemble averages of bad and good models (BAD - GOOD) of the MJ (a) 200 hPa zonal wind (units m/s) and (b) surface temperature (units °C). The contour lines in (a) represent the 200 hPa zonal wind climatology for good models (units m/s), while the contour lines in (b) represent the surface temperature climatology for good models (units °C). The light-gray stippling indicates the grid points in which the difference is significant at the 95 % level of confidence with respect to one-sample *t*-test. The black continuous and dashed boxes in (a) indicate the domains (160°E-150°W, 30°S-42°S) and (160°E-90°W, 50°S-67°S), respectively, used to calculate the meridional gradient of the upper tropospheric zonal wind for Fig. 9.

suppresses moisture transport over central Chile, resulting in negative rainfall anomalies. This precipitation bias becomes more evident when taking the difference between the patterns shown in Fig. 5b and a (Fig. S4).

Fig. 6 shows the eddy (i.e., deviations from the zonal mean) 200 hPa geopotential height and WAF anomalies regressed onto the pure IOMC index for good (Fig. 6a) and bad (Fig. 6b) models. Good models capture key features of the zonal wavenumber-3 Southern Hemisphere Rossby wave train, consistently with ERA5 (Fig. 3b). However, they exhibit poleward wave propagation from the southeastern Indian Ocean and Australian sectors, differently from ERA5. This anomalous meridional propagation across the southern Australian sector may be linked to an underestimated Southern Hemisphere bounding region (Fig. S5e), which is, in turn, associated with a weaker meridional gradient of absolute vorticity (Fig. S5c) relative to that in ERA5 (Fig. S3b).

While bad models also simulate the wave train, albeit weaker, and show signs of poleward wave propagation over the southern Australian sector—likely due to the absence of the atmospheric bounding region (Fig. S5f) resulting from an overly weak meridional gradient of absolute vorticity (Fig. S5d)—they exhibit a weaker amplitude and a more equatorward-deflected trajectory compared to both good models and ERA5. As indicated by the wave flux field in Fig. 6b, the refraction point in bad models is westward and equatorward shifted, leading to a displaced upper tropospheric response. Specifically, the anticyclonic center typically positioned around 110°W-65°S (see Figs. 3b and 6a) is excessively deflected northwestward, thus shifting anomalous anticyclonic anomalies over central and southern Chile. As indicated by the wave flux, the wave train terminates over the eastern subtropical Pacific, where it induces an anomalous cyclonic anomaly.

3.3. The impact of the atmospheric mean state in CMIP models

In this section, we assess whether the anomalous wave behavior in bad models is associated with errors in the simulation of the atmospheric waveguide. Fig. 7a illustrates the difference in the climatological upper tropospheric jet stream between bad and good models. Compared to good models, bad models exhibit a weaker mid-latitude waveguide, with the jet stream being around 25 % weaker over the far southern Indian Ocean and its exit region (see also Fig. S5a and b for comparison). Furthermore, the subtropical jet is about 10 % stronger in bad models over the subtropical southern Indian and western Pacific Oceans.

The coupled weakening/strengthening of the mid-latitude/subtropical jet streams, indicating an equatorward shift of the mid-latitude jet, may be linked to an anomalously warm lower tropospheric Antarctic sector (Fig. 7b). Through thermal wind balance, this warming reduces the equator-to-pole temperature gradient and consequently weakens the upper tropospheric mid-latitude jet stream. Meanwhile, the warmer surface temperatures across the Southern Ocean and Antarctic sectors might be related to an underestimated Antarctic Sea ice cover (Fig. S6).

A previous study has shown that errors in the simulation of atmospheric waveguide might affect the ray path of stationary Rossby waves (Sabatani and Gualdi, 2025). Following a similar argument, we investigate the impact of an equatorward displacement of the mid-latitude jet stream on the trajectory of Rossby waves. The Rossby wave ray tracing algorithm is applied in a set of two experiments:

- using the averaged 200 hPa zonal wind MJ climatology across good models, hereafter referred to as $-u_{\text{GOOD}}$.
- using the averaged 200 hPa zonal wind MJ climatology across bad models, hereafter referred to as $-u_{\text{BAD}}$.

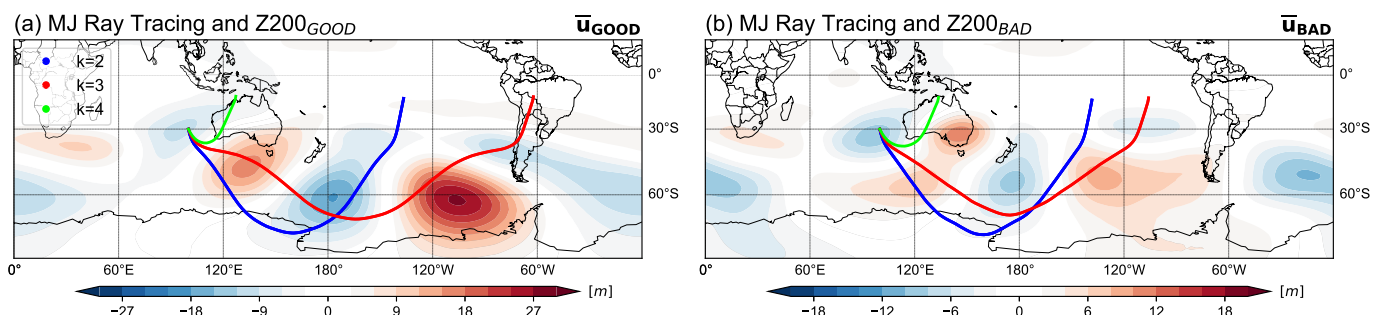


Fig. 8. (a) Ensemble average of good models of the MJ eddy 200 hPa geopotential height anomalies regressed against the pure IOMC index (Z200_GOOD, units m per standard deviation) and Rossby rays (colored lines) for different zonal wavenumber *k*. (b) As in (a) but for the ensemble average of bad models.

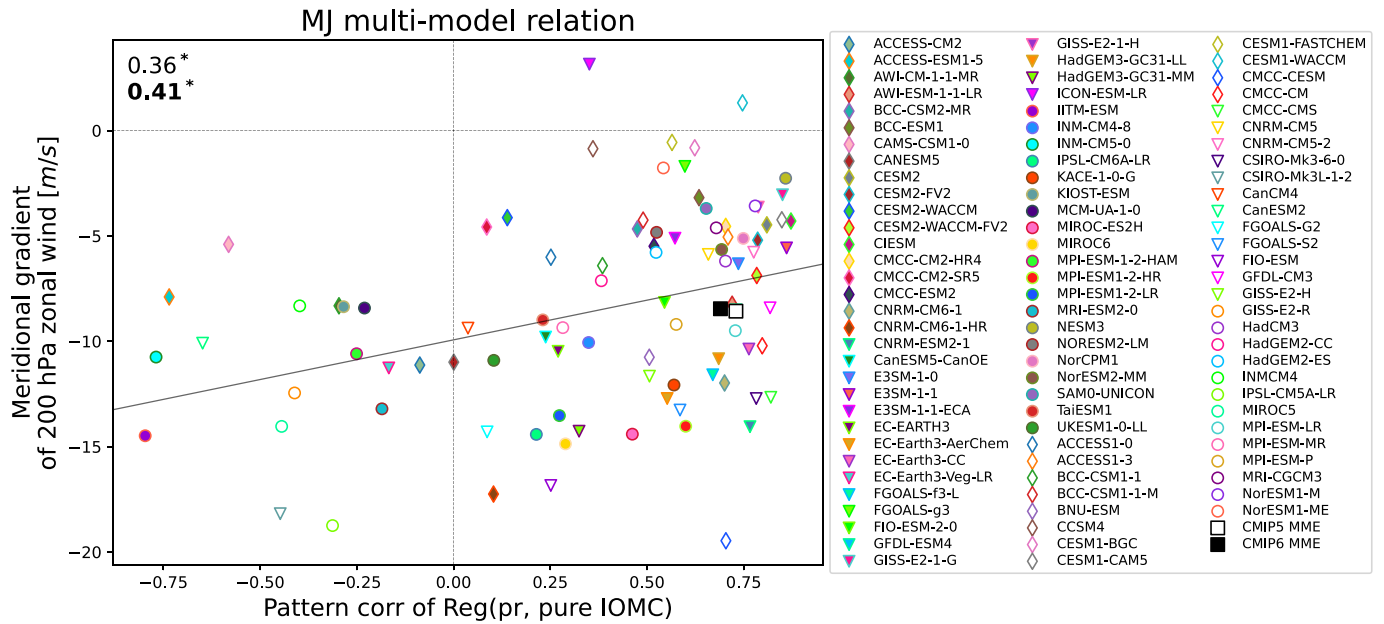


Fig. 9. MJ multi-model relation of the spatial correlation of the pure IOMC-related precipitation anomalies over Chile (see Fig. 4) and the meridional gradient of 200 hPa zonal wind over the Southern Hemisphere. The black-filled square indicates the MME of CMIP6 models, and the white-filled square represents the MME of CMIP5 models. The color-filled symbols denote individual CMIP6 models, while the white-filled symbols denote individual CMIP5 models. The black line indicates the best linear fit. Pearson correlation coefficients are shown in the top-left corner with asterisks indicating significance 95 % level of confidence. The bold value refers to the Pearson correlation after removing the outliers (CMCC-CSM and ICON-ESM-LR), identified using an interquartile range method.

Rossby wave rays are initiated from the cyclonic anomaly detected in both good (Fig. 6a) and bad (Fig. 6b) models west of subtropical Australia. Sensitivity tests demonstrate that analogous results are obtained when using local maxima/minima of TRWS as starting points for the algorithm (not shown). The ray positions are calculated for zonal wavenumbers spanning from 2 to 4, usually associated with tropical-extratropical teleconnection patterns (Hoskins and Ambrizzi, 1993). Fig. 8 shows the ray tracing results for good (Fig. 8a) and bad (Fig. 8b) models.

In good models, zonal wavenumber-3 Rossby waves (red lines) follow a ray path that eventually reaches Chile, well in agreement with the upper tropospheric circulation anomalies associated with IOMC heating anomalies (Fig. 8a). In bad models, the atmospheric waveguide is characterized by an equatorward shift of the mid-latitude jet stream, especially in its exit region (see Fig. 7a and Fig. S5b). As shown in Fig. 8b, the associated jet stream weakening affects the ray path of zonal wavenumber-3 stationary Rossby waves, which exhibit a too westward shifted refraction location. As a result, the ray path terminates over the subtropical southeastern Pacific, thereby displacing the cyclonic anomaly from the Amundsen Sea northeastward toward Chile. This northeastward shift, in turn, moves the precipitation response to IOMC heating anomalies from southern Chile to central Chile, where an anomalous dry signal emerges (see Fig. 5b).

Subsequently, we calculated for each model the difference between the averaged climatological 200 hPa zonal wind over the far southern Pacific Ocean (160°E-90°W, 50°S-67°S; black dashed box in Fig. 7a) and the southwestern Pacific Ocean (160°E-150°W, 30°S-42°S; black continuous box in Fig. 7a). This difference, constituting a meridional gradient of the upper tropospheric jet stream, is plotted in Fig. 9 against the pattern correlation coefficient of the pure IOMC-related precipitation anomalies over Chile (see Fig. 4). Sensitivity tests demonstrate that the relation is not dependent on the choice of the domains used to define meridional gradient values (not shown).

The multi-model relation shows a significant correlation of 0.36, indicating that models unable to reproduce the precipitation dipole over Chile also tend to simulate anomalously low meridional gradient values, indicative of an equatorward shift of the mid-latitude southern Pacific

jet stream. After removing outlier models (i.e., CMCC-CMS and ICON-ESM-LR), identified using the interquartile range method with a threshold of an interquartile range unit along the y-axis, the correlation increases to 0.41. These results further evidence that a source of uncertainty regarding the pure IOMC teleconnection with Chile stems from an equatorward shift of the mid-latitude southern Pacific jet stream.

While our atmospheric waveguide analysis explains the equatorward deflection of the teleconnection pattern in bad models, it does not account for its weaker amplitude (Figs. 5b and 6b). We therefore hypothesize that this weaker response is instead related to biases in the simulated spatial pattern of the IOMC heating anomalies. To test this, we analyzed the composite difference (bad minus good models) in precipitation, 200 hPa TRWS, and 200 hPa irrotational wind anomalies regressed onto the pure IOMC index (Fig. S7).

This analysis reveals that bad models overestimate IOMC-related rainfall over southwestern India and the Maritime Continent while underestimating it over the southwestern subtropical Pacific (Fig. S7a). We suggest this pattern is linked to a flawed adjustment of the Walker circulation in response to Indo-Pacific inter-basin coupling. Consequently, upper-tropospheric irrotational wind anomalies are excessively poleward over the southwestern Australian sector and equatorward over the southeastern sector. This anomalous circulation significantly weakens the TRWS in these key regions (Fig. S7b). Therefore, the weaker mid-latitude response in bad models appears to be a consequence of these upstream biases in IO heating patterns and the resultant weakening of the TRWS over the Australian sectors.

4. Discussion and conclusions

In this study, the impact of the atmospheric mean state on the simulation of the Indian Ocean impacts on Chile during austral early-winter (MJ) has been analyzed.

Results obtained from the ERA5 reanalysis and partial regression analysis show that the pure (i.e., independent of ENSO variability) IOMC teleconnection with Chile appear as a dipole of precipitation anomalies, with above-average rainfall over central Chile and below-average rainfall over southern Chile. This rainfall dipole is driven by a remote

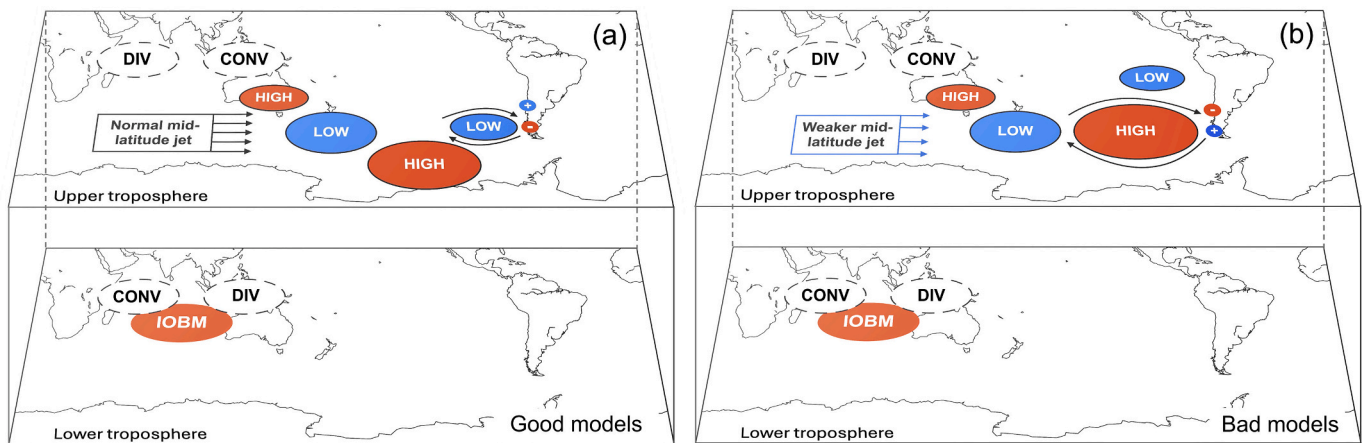


Fig. 10. Schematic illustration of the austral early-winter (MJ) IOMC teleconnection in (a) good and (b) bad models. CONV and DIV stand for anomalous convergence and divergence, respectively. The “+” and “-” symbols over Chile indicate enhanced and reduced precipitation anomalies, respectively.

atmospheric response associated with a zonal wavenumber-3 Rossby wave train originating from the eastern Australian sector and terminating over Chile, where it induces cyclonic tendency. The related circulation anomalies, in turn, strengthen (weaken) the climatological moisture flux over central (southern) Chile, thereby favoring the dipole of precipitation anomalies.

Decadal to multi-decadal variability may modulate the relationship between Indian Ocean heating and Chilean rainfall. To assess this potential non-stationarity, we performed a 40-year moving window analysis to ERA5 from 1940 to 2024, calculating the regression of the May–June DCR index onto the Niño 3.4, pure Niño 3.4, and pure IOMC indices (Fig. S8). The results reveal an increase in the IOMC contribution to Chilean rainfall after the 1990. This signal is driven by a simultaneous strengthening of the pure IOMC teleconnection and a weakening of the pure (i.e., independent of IOMC variability) ENSO teleconnection (Fig. S8, green and red scatters, respectively). However, this apparent non-stationary signal must be interpreted with caution. The reliability of tropical rainfall in reanalysis products is dependent on observational constraints, which improved dramatically with the advent of satellite data in the late 1970s. Before this era, precipitation estimates over the data-sparse tropical oceans are less certain. Consequently, substantial uncertainty remains in the characterization of multidecadal signals in reanalysis, and the above results should therefore be interpreted with caution.

Our analysis of CMIP5 and CMIP6 historical simulations reveals that CMIP6 models show no perceptible improvement over CMIP5 in simulating the pure IOMC teleconnection to Chile. Furthermore, we find that models from both generations tend to underestimate the strength of this extratropical teleconnection. This result is consistent with the findings of Fang et al. (2024), who identified a systematic underestimation of Southern Hemisphere extratropical teleconnection patterns in CMIP6 models.

Through a clustering process, we separated models able to simulate the Chilean dipole of precipitation related to the pure IOMC teleconnection (referred to as “good models”) from those that cannot (referred to as “bad models”). Results indicate that bad models do simulate the Southern Hemispheric zonal wavenumber-3 Rossby wave associated with the pure IOMC teleconnection to some extent. However, the wave train’s refraction location is shifted too far westward compared to good models and ERA5, causing anticyclonic anomalies to be displaced from the Amundsen Sea northeastward toward Chile. These anticyclonic anomalies, in turn, weaken the climatological moisture flux over central Chile, thereby shifting the IOMC-related dry signal from southern to central Chile.

An analysis of mean states across good and bad models reveals that bad models simulate a Southern Hemispheric mid-latitude jet stream

that is displaced too far equatorward. This shift may be linked to an overly warm Antarctic sector, which weakens the meridional temperature gradient. In turn, Antarctic warming may stem from an underestimated climatological Antarctic Sea ice cover. Indeed, variability in SSTs and sea ice extent can significantly influence the latitudinal position of the mid-latitude jet stream (Francis and Vavrus, 2015; Francis and Vavrus, 2012; Kidston et al., 2011; Brayshaw et al., 2008; Chen et al., 2010; Deser et al., 2004).

Perturbations in the meridional SST gradient can alter the position of the mid-latitude jet stream via thermal wind balance, shifting it toward regions of enhanced lower tropospheric baroclinicity. More specifically, in bad models, lower tropospheric warming is confined on the poleward flank of the mid-latitude jet, weakening the meridional temperature gradient and reducing baroclinic wave activity at high latitudes. As a result, the jet shifts equatorward. On the other hand, the underestimated Antarctic Sea ice cover in bad models might stem from model biases in the subsurface ocean and sea ice physics, warranting further investigation in future studies.

Using a Rossby ray tracing algorithm, we demonstrated that an equatorward shift and weakening of the mid-latitude jet leads to a westward displacement of the refraction location of zonal wavenumber-3 Rossby waves. According to Rossby wave theory, a weaker jet is associated with a slower eastward propagation of stationary waves, which tend to meander more. As a result, the wave train becomes excessively meridionally bent and westward shifted, disrupting the remote atmospheric response over Chile.

These results agree with previous research showing that the spatial structure of wave trains induced by atmospheric heating anomalies is highly dependent on the mean atmospheric flow (Ting and Sardeshmukh, 1993; Yun et al., 2010, 2011; Chen and Wu, 2017; Di Carlo et al., 2022; Chen et al., 2024; Sabatani and Gualdi, 2025). For instance, using a Linear Baroclinic Model, Yun et al. (2011) analyzed how sub-seasonal changes in the Asian jet alter Rossby wave propagation from tropical and mid-latitude forcings. Similarly, Chen and Wu (2017) used a barotropic model to demonstrate that the climatological waveguide contributes to changes in the Eurasian atmospheric circulation’s response to anomalous North Atlantic heating. A valuable next step, therefore, would be to use a Linear Baroclinic Model to formally investigate how the basic state modulates the teleconnection pathway from the Indian Ocean to South America.

More recent work highlighted, using an intermediate-complexity model, that the ENSO-induced atmospheric response over the Pacific–North American sector is affected by the position and strength of the subtropical Pacific jet stream (Di Carlo et al., 2022). In agreement, Sabatani and Gualdi (2025) suggested that a stronger basic flow leads to more zonally oriented stationary Rossby waves, therefore affecting the

remote atmospheric response to tropical heating forcings.

We also briefly investigated whether the link between the atmospheric mean state and the Indian Ocean–Chile teleconnection is evident in reanalysis data. Based on the analysis presented in Fig. S8, we defined two distinct epochs: a period of weak IOMC teleconnection from 1940 to 1982 and a period of strong teleconnection from 1983 to 2024. Specifically, the regression of the DCR index onto the pure IOMC index yields values of approximately 0.05 mm/day in the earlier period and 0.16 mm/day in the later one. An examination of the atmospheric mean state difference between these two periods reveals that the epoch of weaker teleconnection (1940–1982) was characterized by a weaker and more equatorward-shifted eddy-driven jet stream over the South Pacific (Fig. S9). This finding from reanalysis closely resembles our model results (Fig. 7a), providing further evidence that the Indian Ocean teleconnection to Chile is modulated by the mean atmospheric flow.

In conclusion, the analyzed CMIP models show that a key source of error in simulating the impact of the IOMC on Chile stems from an equatorward shift of the Southern Hemispheric mid-latitude eddy-driven jet stream, as schematized in Fig. 10. This displacement may, in turn, be linked to a reduced meridional SST gradient, potentially associated with an underestimated climatological Antarctic Sea ice cover. These findings underscore the need to address systematic errors that contribute to mean state biases. Identifying and mitigating these biases would enhance the models' ability to simulate extratropical atmospheric teleconnections, ultimately improving the skill of climate predictions in mid-latitudes.

CRedit authorship contribution statement

Davide Sabatani: Writing – original draft, Visualization, Methodology, Investigation, Formal analysis, Conceptualization. **Silvio Gualdi:** Writing – review & editing, Supervision, Methodology, Investigation, Funding acquisition, Conceptualization. **Swadhin Behera:** Writing – review & editing, Supervision, Methodology, Conceptualization. **Yushi Morioka:** Writing – review & editing, Supervision, Methodology, Conceptualization. **Ingo Richter:** Writing – review & editing, Supervision, Methodology, Conceptualization.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.atmosres.2025.108509>.

Data availability

The model data used in this study is freely available from the Earth System Grid Federation (ESGF) at <https://esgf-ui.ceda.ac.uk/>. The ERA5 reanalysis dataset was obtained from the Copernicus Climate Data Store (<https://cds.climate.copernicus.eu/>). The Global Precipitation Climatology Project version 2 (GPCP) dataset is available at <https://psl.noaa.gov/data/gridded/data.gpcp.html>. The Hadley Centre Global Sea Surface Temperature (HadISST) dataset is available at <https://www.metoffice.gov.uk/hadobs/hadisst/>.

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