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# **Southern Hemisphere Circumpolar Wavenumber-4 Pattern Simulated in SINTEX-F2 Coupled Model**

**Balaji Senapati<sup>1,2</sup>, Yushi Morioka<sup>3</sup>, Swadhin K. Behera<sup>3</sup>, and Mihir K. Dash<sup>1</sup>**

<sup>1</sup> Centre for Ocean, River, Atmosphere and Land Sciences, Indian Institute of Technology Kharagpur, Kharagpur, West Bengal, India

<sup>2</sup> Department of Meteorology, University of Reading, Reading, UK

<sup>3</sup> Application Laboratory, VAIg, Japan Agency for Marine-Earth Science and Technology, Yokohama, Kanagawa, Japan

Corresponding author: Mihir K. Dash (mihir@coral.iitkgp.ac.in)

## Key Points:

- First attempt to successfully simulate the wavenumber-4 pattern of Southern Ocean SST using a coupled model, uncovering the underlying physical processes.
- Southwestern subtropical Pacific SST plays a crucial role in generating SST wavenumber-4 pattern via circumpolar atmospheric variability.
- The ocean mixed layer is found to be important for the growth and decay of the SST pattern.

**Keywords:** Wavenumber 4, SINTEX-F2 coupled model and sensitivity experiments, Mixed layer heat budget, air-sea interaction, SST, Southern Hemisphere subtropics and mid-latitude

ORCID

*Balaji Senapati*: <https://orcid.org/0000-0001-5029-9731>

*Yushi Morioka*: <https://orcid.org/0000-0002-2709-4911>

Swadhin K. Behera: <https://orcid.org/0000-0001-8692-2388>

*Mihir K. Dash:* <https://orcid.org/0000-0003-1426-7200>

27 **Abstract**

28 Interannual sea surface temperature (SST) variations in the subtropical-midlatitude  
29 Southern Hemisphere are often associated with a circumpolar wavenumber-4 (W4) pattern. This  
30 study is the first attempt to successfully simulate the SST-W4 pattern using a state-of-the-art  
31 coupled model called SINTEX-F2 and clarify the underlying physical processes. It is found that  
32 the SST variability in the southwestern subtropical Pacific plays a key role in triggering  
33 atmospheric variability and generating the SST-W4 pattern during austral summer (December-  
34 February). In contrast, the tropical SST variability has very limited effect. The anomalous  
35 convection and associated divergence over the southwestern subtropical Pacific induce  
36 atmospheric Rossby waves confined in the westerly jet. Then, the synoptic disturbances  
37 circumnavigate the subtropical Southern Hemisphere, establishing an atmospheric W4 pattern.  
38 The atmospheric W4 pattern has an equivalent barotropic structure in the troposphere, and it  
39 interacts with the upper ocean, causing variations in mixed layer depth due to latent heat flux  
40 anomalies. As incoming climatological solar radiation goes into a thinner (thicker) mixed layer,  
41 the shallower (deeper) mixed layer promotes surface warming (cooling). This leads to positive  
42 (negative) SST anomalies, developing the SST-W4 pattern during austral summer.  
43 Subsequently, anomalous entrainment due to the temperature difference between the mixed  
44 layer and the water below the mixed layer, anomalous latent heat flux, and disappearance of the  
45 overlying atmospheric W4 pattern cause the decay of the SST-W4 pattern during austral autumn.  
46 These results indicate that accurate simulation of the atmospheric forcing and the associated  
47 atmosphere-ocean interaction is essential to capture the SST-W4 pattern in coupled models.

48

49 **Plain Language Summary**

50 In the subtropical-midlatitude Southern Hemisphere, we often observe year-to-year  
51 fluctuations in sea surface temperature (SST) linked to a specific pattern known as wavenumber-  
52 4. This study represents the first successful attempt to simulate this temperature pattern using a  
53 climate emulator called SINTEX-F2, allowing us to uncover its physical processes. Our research  
54 reveals that SST variations in the southwestern subtropical Pacific (SWSP) play a pivotal role in  
55 generating the wavenumber-4 pattern in the atmosphere, subsequently influencing SST during  
56 austral summer. Interestingly, this pattern is almost independent of tropical SST variability.

57 The process starts with heating in the SWSP, causing atmospheric disturbances. This leads  
58 to an undulation in mid-latitude atmospheric flow, evolving into a well-established global Rossby  
59 wave with four positive (negative) loading centers, forming a wavenumber-4 pattern. This  
60 atmospheric wave interacts with the ocean's surface, leading to heat exchange between the  
61 atmosphere and the upper ocean. In turn, it influences the depth of the mixed layer in the upper  
62 ocean, which receives solar energy. When solar energy penetrates into a shallower (deeper) mixed  
63 layer, it warms (cools) the mixed layer effectively, resulting in higher (lower) SSTs. Afterward,  
64 the energy exchange between the mixed layer and the deep ocean contributes to the decay of the  
65 SST pattern.

66

67 **1 Introduction**

68 Oceans cover more than 80% of the surface area in the Southern Hemisphere and  
69 primarily affect the rainfall variability over the landmasses, hence impacts the regional  
70 hydrology, agriculture, drinking water, economy and livelihood of the society. Further, rainfall  
71 deficit over these regions is closely associated with the climate variability over subtropical-  
72 midlatitude oceans (Barros & Silvestri, 2002; Behera & Yamagata, 2001; Diaz et al., 1998;  
73 Paegle & Mo, 2002; C. Reason, 1999, 2001; Taschetto & Wainer, 2008). The linkage of these  
74 rainfall variabilities with tropical climate variability has been extensively studied in recent  
75 decades (Ashok et al., 2003; Grimm et al., 2000; Richard et al., 2000). However, the dynamics  
76 of the subtropical-midlatitude oceans and their influence on the subtropical climate have drawn  
77 little attention until the 21st century. One of the major factors affecting the subtropical climate is  
78 sea surface temperature (SST) due to its interaction with the overlying atmosphere. The SST  
79 variability in the subtropical-midlatitude Southern Hemisphere has the potential to impact global  
80 climate directly or indirectly through the dynamics of the ocean, atmosphere, and climate on  
81 interannual to decadal timescales (Liu & Alexander, 2007). Thus, understanding the physical  
82 mechanisms of the interannual SST variability in the subtropical-midlatitude Southern  
83 Hemisphere is crucial in improving the knowledge of the underlying dynamics of climate  
84 variability.

85 The subtropical-midlatitude SST in the Southern Hemisphere reaches its peak during  
86 austral summer (December-February), shows the largest interannual variability because of  
87 intense insolation and shallower mixed layer depth (MLD), which facilitates interaction with the  
88 overlying atmosphere and upper ocean (Suzuki et al., 2004). The major SST variabilities over  
89 the subtropical regions are closely associated with the subtropical highs, called Subtropical  
90 Dipole (Behera & Yamagata, 2001; Fauchereau et al., 2003; Hermes & Reason, 2005; Morioka,  
91 Ratnam, et al., 2013; Venegas et al., 1997). Possible sources of the variations in the subtropical  
92 high are internal or connected to El Niño/Southern Oscillation (ENSO), Indian Ocean Dipole  
93 (IOD) and Southern Annular Mode (SAM) (Anila & Gnanaseelan, 2023; Crétat et al., 2018;  
94 Morioka et al., 2012; Morioka, Tozuka, et al., 2013; Rodrigues et al., 2015). For example, ENSO-  
95 induced Pacific South American pattern (Mo & Paegle, 2001) shifts the St. Helena High  
96 meridionally and hence generates the Subtropical Dipole in the South Atlantic (Rodrigues et al.,  
97 2015). Similarly, the Subtropical Dipole in the southern Indian Ocean develops due to variation

98 in Mascarene High influenced by ENSO, IOD and SAM (Anila & Gnanaseelan, 2023; Crétat et  
99 al., 2018; Morioka, Tozuka, et al., 2013). Further, in the absence of tropical SST variability,  
100 Rossby waves related to the SAM can generate Subtropical Dipoles by modulating sea level  
101 pressure anomalies in the mid-latitude Southern Hemisphere (Morioka et al., 2014a).  
102 Synchronization of these Subtropical Dipoles in each ocean basin turns into a global  
103 wavenumber-3 pattern in the subtropical Southern Hemisphere (Wang, 2010).

104 Despite these efforts, few studies have discussed the interannual SST variability over the  
105 entire subtropical Southern Hemisphere (Wang, 2010; Senapati, Dash, et al., 2021). This might  
106 be due to the paucity of observations over the regions (see Fig. 1 in Morioka et al. (2013)). The  
107 substantial increase in the SST observations after 1980s helps enhance confidence in studying  
108 the SST variability over the regions. Recently, a circumpolar wavenumber-4 (W4) pattern has  
109 been reported in the atmospheric circulation (Senapati, Deb, et al., 2021), which can influence  
110 the ocean (Senapati, Dash, et al., 2021) as similar to a wavenumber-3 pattern (Wang, 2010).  
111 Wang (2010) discussed the covarying Subtropical Dipoles in each basin of the Southern  
112 Hemisphere and suggested it as the wavenumber-3 pattern. Nevertheless, Senapati et al. (2021a)  
113 investigated the interannual SST variability, focusing on the subtropical- midlatitude Southern  
114 Hemisphere, and reported an SST-W4 pattern for the first time. SST-W4 pattern is different  
115 from the wavenumber-3 pattern in many aspects, including location and the number of centres  
116 of actions, generation dynamics (Wang, 2010; Senapati, Dash, et al., 2021), impacts (Senapati,  
117 Dash, et al., 2021), and linkage of extra-subtropical regions (Morioka et al., 2014b; Senapati,  
118 Dash, et al., 2021). The wavenumber-3 pattern (1st EOF mode of the subtropical SST) is  
119 generated mainly due to variations in horizontal wind components independent of ENSO and SAM  
120 (Wang, 2010). In contrast, the SST-W4 pattern (2nd EOF mode) develops only due to  
121 meridional wind variability (Senapati, Dash, et al., 2021).

122 The SST-W4 pattern is suggested to be phase-locked during austral summer and occurs  
123 even in the absence of other climate variabilities. It is noted that thermodynamic feedbacks  
124 between meridional wind, surface heat flux, and mixed layer mainly cause the SST-W4 pattern.  
125 The SST-W4 pattern also undergoes a decadal variability and impacts the precipitation over  
126 Southern continents (Senapati et al., 2022). The SST footprints of the decadal variability of the  
127 South Pacific Meridional Mode (SPMM; (Zhang et al., 2014)) creates an ideal environment for

128 the frequent occurrence of positive/negative type of the SST-W4 pattern, which results in the  
129 decadal modulation of the SST-W4 pattern after two years (Senapati et al., 2022). The SPMM,  
130 which drives the decadal variability of SST-W4 (Senapati et al., 2022), has been considered as  
131 a precursor to ENSO dynamics (Larson et al., 2018; You & Furtado, 2018). However, the  
132 tropical connection in the development of SST-W4 is lacking.

133 Nevertheless, the detailed air-sea interaction processes involved in the generation, growth,  
134 and decay of the SST-W4 pattern are not well investigated. In other words, how does the wind  
135 influence the MLD change and modulate the SST anomalies? Is the MLD change due to the  
136 wind stirring effect, thermal buoyancy effect, or both? If so, then what is their individual  
137 contribution to the MLD variability? The roles of wind, surface heat flux, and mixed layer  
138 variations in developing the SST-W4 pattern require thorough investigation. Also, Senapati,  
139 Dash, et al. (2021) suggested that the SST-W4 pattern decays because of the thermodynamic  
140 decoupling between the upper ocean and the overlying atmosphere and subsequent cut-off from  
141 the source. The ocean's role is not considered in the decaying phase and is essential to consider  
142 while investigating the SST variability, especially over the subtropics (Morioka et al., 2011,  
143 2012; Morioka, Ratnam, et al., 2013). Because of the very limited availability of observation  
144 datasets, a coupled general circulation model (CGCM) is required to explore the air-sea  
145 interaction processes in detail during the growth and decay phases of the SST-W4 pattern.

146 Furthermore, it is suggested that the convective activity over the southwestern subtropical  
147 Pacific (SWSP) is instrumental in generating atmospheric W4 (Senapati, Deb, et al., 2021),  
148 which later produces the SST-W4 pattern. The root cause of the convective activity over the  
149 SWSP, which acts as a Rossby wave source, remains unclear. Identifying the possible sources  
150 of the convective activity is of great importance in understanding and predicting the SST and  
151 atmospheric W4 patterns.

152 This research aims to investigate the seasonal evolution of the SST-W4 pattern both  
153 qualitatively and quantitatively using the output of 300-yr pre-industrial control (CTL)  
154 simulation from a CGCM. We also performed a series of 100-yr sensitivity experiments to  
155 clarify the possible sources of the convective activity over the SWSP in generating SST and  
156 atmospheric W4 patterns. The structure of this document is as follows. Section 2 briefly  
157 describes the model, the data used, and methodology employed in this investigation. Section 3

158 discusses the results comprising the following four subsections. Subsection 3.1 validates the  
159 performance of CGCM in simulating the SST-W4 pattern in the subtropical-midlatitude  
160 Southern Hemisphere. Subsection 3.2 examines the air-sea interaction processes that contribute  
161 to the growth and decay of the SST-W4 pattern. The connection between SST and the  
162 atmosphere, along with the associated generation mechanisms, is described in Subsection 3.3.  
163 Subsection 3.4 discusses the respective roles of the tropics and the SWSP in generating SST and  
164 atmospheric W4 patterns. Finally, the study is summarized and discussed in Section 4.

165 **2 Model, Data and Methods**

166 *2.1 CGCM and data*

167 The physical processes involved in the SST-W4 pattern are analyzed using a CGCM, namely  
168 the Scale Interaction Experiment-Frontier Research Center for Global Change 2 (SINTEX-F2;  
169 Masson et al., 2012), which is a subsequent version of the SINTEX-F1 (Luo et al., 2005). Previous  
170 studies demonstrated reasonable performance of the model in simulating the observed seasonal  
171 and interannual SST variability in the subtropical-midlatitude Southern Hemisphere (Doi et al.,  
172 2016; Morioka et al., 2011, 2012, 2014a, 2015; Morioka, Ratnam, et al., 2013). This CGCM  
173 comprises of (i) the fifth-generation atmospheric general circulation model 5 (ECHAM5;  
174 developed from the atmospheric model of the European Centre for Medium-Range Weather  
175 Forecasts (ECMWF) with a parameterization package designed at Hamburg) atmospheric model  
176 by Roeckner et al. (2003), (ii) the Nucleus for European Modeling of the Ocean 3 (NEMO3) ocean-  
177 sea ice model by Madec (2008) and (iii) the Ocean-Atmosphere-Sea-Ice-Soil 3 (OASIS3) coupler  
178 by Valcke (2006). The atmospheric grid comprises a T106 Gaussian horizontal grid (i.e.,  $1.125^\circ \times$   
179  $1.125^\circ$ ) and 31 vertical levels. NEMO3 includes the Louvain-la-Neuve Sea Ice Model 2 (LIM2;  
180 Fichefet & Maqueda, 1997) and features an ORCA05 tripolar horizontal grid (i.e.,  $0.5^\circ \times 0.5^\circ$ )  
181 with 31 vertical levels (Gurvan Madec & Imbard, 1996). NEMO3 employs a turbulent kinetic  
182 energy closure model to calculate vertical eddy viscosity and diffusivity in the ocean (Gurvan  
183 Madec et al., 1998). To represent the effects of diapycnal mixing on ocean salinity and  
184 temperature, NEMO3 utilizes the double-diffusive mixing model established by Merryfield et al.  
185 (1999). The OASIS3 coupler (Valcke, 2006) exchanges freshwater, heat, and air-sea momentum  
186 fluxes every 2 hours between NEMO3 and ECHAM5. Additional flux corrections using reanalysis  
187 products or observational data are omitted, allowing the ocean-atmospheric states in the model to

188 evolve freely. The ocean model is initialized with annual mean salinity and temperature data from  
189 the World Ocean Atlas 1998 (<https://psl.noaa.gov/data/gridded/data.nodc.woa98.html>). A 300-yr  
190 CTL simulation was performed with pre-industrial radiative forcings. The first 30 years of the  
191 simulations are excluded from the analysis to account for the time required for the ocean to adjust  
192 to the inter-annually varying atmospheric forcing (Morioka & Behera, 2021). Model results are  
193 validated using monthly mean SST data from Hadley Centre Global Sea Ice and Sea Surface  
194 Temperature (HadISST; Rayner et al., 2003) dataset with a  $1^\circ \times 1^\circ$  spatial resolution from 1979 to  
195 2018. Different climate indices such as the Oceanic Niño Index (ONI; Trenberth, 1997), Indian  
196 Ocean Dipole (IOD; Saji et al., 1999) index, South Atlantic Subtropical Dipole (SASD; Morioka  
197 et al., 2011) index, South Pacific Subtropical Dipole (SPSD; Morioka, Ratnam, et al., 2013) index  
198 and Indian Ocean Subtropical Dipole (IOSD; Behera & Yamagata, 2001) index are calculated in  
199 CGCM outputs to verify their connection with the SST-W4 pattern.

200 *2.2 Sensitivity experiments*

201 Two potential sources can create SST and atmospheric W4 patterns: (i) tropical SST  
202 variability and (ii) the SWSP SST variability. In order to identify the actual source, two SST  
203 nudging experiments are conducted using the SINTEX-F2 model. In these experiments:

204 (i) The model's SST in the tropical region ( $25^\circ\text{S}$ - $25^\circ\text{N}$ ) is nudged to the monthly  
205 climatological SST of the control simulation, referred to as the 'noTropics'  
206 experiment.

207 (ii) The model's SST in the SWSP region ( $150^\circ\text{E}$ - $160^\circ\text{W}$ ,  $50^\circ$ - $10^\circ\text{S}$ ) is nudged to the  
208 monthly climatological SST of the control simulation, referred to as the 'noSWSP'  
209 experiment.

210 A large negative feedback ( $-2400 \text{ W m}^{-2} \text{ K}^{-1}$ ) is added to the surface heat flux to restore the  
211 upper 50m mixed layer temperature within one day (Morioka et al., 2014b). The gaussian method  
212 is applied at the boundaries within  $5^\circ$  of the SST nudging area for smoothing. We integrated these  
213 experiments for 100 years, from which the last 70 years are used for the analysis after removing  
214 the initial 30 years to account the ocean's adjustment to the atmospheric forcing (Morioka &  
215 Behera, 2021). It is important to note that the surface heat flux correction in the SST-nudging  
216 experiments tends to significantly reduce the coupled interactions and variability in the corrected  
217 region. In other words, we can scrutinize the impact of SST variability in each nudging area on the

218 SST and atmospheric W4 patterns by subtracting the results of the CTL and SST-nudging  
 219 experiments.

220 *2.3 Statistical analysis*

221 We calculated the monthly anomalies at all grid points for every variable by removing the  
 222 corresponding monthly climatology after detrending it using the least square fit. The cost  
 223 function of a variable ‘ $C = \sum_{i=1}^n (y_i - f(x_i))^2$ ’, is minimized using the least square approach  
 224 to generate the best-fit function, where,  $y_i$ ,  $x_i$ , and  $f(x_i)$  are dependent, independent variable,  
 225 and best-fit function, respectively. Following Senapati et al. (2021a), the SST-W4 pattern is  
 226 captured by employing the Empirical Orthogonal Function (EOF) analysis over the subtropical  
 227 and midlatitudes in the Southern Hemisphere (55°S–20°S). Further, North criteria (North et al.,  
 228 1982) was used to test the significance and independency of the EOF modes. In this method, the  
 229 standard error  $\left[ \Delta\lambda = \lambda \sqrt{\frac{2}{N}} \right]$  of the associated eigenvalues ( $\lambda$ ) was computed by using degrees of  
 230 freedom (N) present in the dataset. If the sampling error (i.e.,  $\Delta\lambda$ ) of an eigenvalue ‘ $\lambda$ ’ is shorter  
 231 than the gap between  $\lambda$  and the eigenvalue closest to it then ‘ $\lambda$ ’ is considered to be independent,  
 232 else otherwise. The role of atmosphere in generating the SST-W4 pattern is validated using  
 233 Maximum Covariance Analysis (MCA) over the region (Senapati, Dash, et al., 2021). The MCA  
 234 is a singular value decomposition approach that uses two variables' cross-covariance matrix.  
 235 Since the SST-W4 pattern peaks during austral summer (Senapati, Dash, et al., 2021), singular  
 236 value decomposition is applied to austral summer (December-February). Results from left  
 237 (right) singular vectors correspond to the covarying pattern of the first (second) variable during  
 238 austral summer. The temporal pattern is generated by projecting the corresponding pattern on  
 239 the original detrended anomalous data (consisting of all calendar months).

240 *2.4 Rossby wave source*

241 We calculated the source of Rossby waves associated with SST and atmospheric W4  
 242 patterns (Jianchun Qin & Robinson, 1993; Sardeshmukh & Hoskins, 1988; Senapati, Deb, et  
 243 al., 2021) using the following equation:

244 
$$\text{Rossby Wave Source (RWS)} = -(\zeta + f)\nabla \cdot V_\lambda - V_\lambda \cdot \nabla(\zeta + f) \quad (1),$$

245 where ‘ $f$ ’ represents planetary vorticity, ‘ $V_\lambda$ ’ denotes divergent wind, and ‘ $\zeta$ ’ represents  
 246 relative vorticity. The first term on the right-hand side of Eq. (1) indicates the vortex stretching  
 247 by eddies, while the second term corresponds to the advection of absolute vorticity by the  
 248 divergent wind.

249 *2.5 Rossby wave activity flux*

250 To illustrate the Rossby wave activities, two-dimensional Rossby wave activity flux (WAF;  
 251 Takaya & Nakamura, 1997, 2001) is estimated from the subtropics to the polar region using the  
 252 following equation:

$$253 \quad WAF = \frac{p}{2|\bar{U}|} \times \left\{ \begin{array}{l} \bar{u}(\Psi'^x_x - \Psi'\Psi'^{xx}) + \bar{v}(\Psi'^x_x\Psi'^y_y - \Psi'\Psi'^{xy}) \\ \bar{u}(\Psi'^x_x\Psi'^y_y - \Psi'\Psi'^{xy}) + \bar{v}(\Psi'^y_y - \Psi'\Psi'^{yy}) \end{array} \right\} \quad (2),$$

254 where  $p = \left( \frac{\text{pressure}}{1000 \text{ hPa}} \right)$ ,  $\Psi$  and  $|\bar{U}|$  represent the stream function and magnitude of horizontal  
 255 wind vector  $U(u,v)$ , respectively. Variables with bar, prime and subscript denote climatology,  
 256 perturbation, and partial derivatives, respectively. The regions of weak zonal wind or easterly  
 257 wind are omitted to diagnose the WAF (Takaya & Nakamura, 2001). The emergence (depletion)  
 258 of WAF aligns with the generation (dissipation) of Rossby waves. The WAF is defined within  
 259 the constraints of the quasi-geostrophic approximation, making it challenging to characterize the  
 260 flux near the equator. Given that the Rossby waves stem from vorticity disturbances related to  
 261 thermal or orographic influences (Brian J Hoskins & Karoly, 1981), we have opted to utilize the  
 262 divergence (convergence) of horizontal upper tropospheric winds instead to describe potential  
 263 generators of Rossby waves in this context.

264 *2.6 Mixed layer heat budget*

265 To investigate the air-sea interaction mechanisms that generate the SST-W4 pattern, we  
 266 conducted a budget analysis of the mixed layer temperature ( $T_m$ ; MLT) tendency over both  
 267 positive and negative SST anomaly poles in the subtropical-midlatitude Southern Hemisphere  
 268 (Morioka et al., 2011, 2012; Morioka, Ratnam, et al., 2013). This analysis involves the following  
 269 equation:

$$270 \quad \frac{\partial T_m}{\partial t} = \frac{Q_{net} - q_d}{\rho c_p H} - \mathbf{u}_m \cdot \nabla T_m - \frac{\Delta T}{H} w_e + Res. \quad (3).$$

271 The terms on the right-hand side of Eq. (3) represent various components, including the heat  
 272 flux at the air-sea interface (first term), horizontal advection (second term; both zonal and  
 273 meridional), entrainment (third term), and residual (fourth term). In the first term,  $Q_{net}$   
 274 represents the net heat flux at the ocean surface (NSHF; i.e., a sum total of net shortwave  
 275 radiation (NSW), sensible heat flux (SHF), net longwave radiation (NLW), and latent heat flux  
 276 (LHF)),  $q_d$  is the downward radiative heat flux at the bottom of the mixed layer (Paulson &  
 277 Simpson, 1977),  $\rho$  ( $=1027 \text{ kg/m}^3$ ) is the average density of ocean water, and  $c_p$  ( $=4187 \text{ JK}^{-1}\text{kg}^{-1}$ )  
 278 is specific heat capacity of the ocean.  $q_d$  is computed by following,

$$279 \quad q(z) = q(0) \left[ Re \left( \frac{z}{\gamma_1} \right) + (1 - R) e \left( \frac{z}{\gamma_2} \right) \right] \quad (4),$$

280 where,  $q(0)$  and  $q(z)$  represent the shortwave radiation at the surface and downward radiative  
 281 heat flux at depth 'z', respectively. In our study, we adopt water type 1B, with constants  $\gamma_1=1$ ,  
 282  $\gamma_2=17$ , and  $R=0.67$  (Dong et al., 2007). The MLD (H) is computed based on the  $0.5 \text{ }^{\circ}\text{C}$   
 283 temperature criteria, defined as the level where the temperature reduces by  $0.5 \text{ }^{\circ}\text{C}$  compared to  
 284 the SST. Similar results are obtained when MLD is calculated using a  $0.125 \text{ kg m}^{-3}$  density criteria  
 285 (the level where the potential density increases by  $0.125 \text{ kg m}^{-3}$  to that at the surface). Notably,  
 286  $q_d$  is only 5% of the shortwave radiation penetrating the surface, even under the thinnest MLD  
 287 conditions during the austral summer and is therefore negligible compared to  $Q_{net}$  (Morioka et  
 288 al., 2012). The average horizontal velocity in the mixed layer is indicated by  $u_m$  in the second  
 289 term. In the third term,  $\Delta T (= T_m - T_{-H-20m})$  denotes the temperature difference between MLD  
 290 and entrained water (i.e. water at 20 m below the mixed layer base; Yasuda et al., 2000).  
 291 Furthermore, the entrainment velocity ( $w_e$ ) is defined as (Qiu & Kelly, 1993; Kraus & Turner,  
 292 1967),

$$293 \quad w_e = \frac{\partial H}{\partial t} + \nabla \cdot (u_m H) \quad (5).$$

294 The "Res." term (i.e., residual) encompasses oceanic diffusion and additional processes with  
 295 high-frequency variability.

296

297 *2.7 Monin-Obukhov depth*

298 Generally, the mixed layer is shallower during austral summer compared to other seasons  
 299 in the subtropics. In this case, the change in the MLD is primarily attributed to wind stirring and  
 300 surface heat fluxes rather than subsurface ocean processes beneath the mixed layer (Morioka,  
 301 Ratnam, et al., 2013). To assess the contribution of each term to MLD variability, we estimate  
 302 the Monin-Obukhov depth ( $H_{MO}$ ),

$$303 \quad H_{MO} = \frac{m_0 u_*^3 + \frac{\alpha g}{\rho c_p} \int_{-H_{MO}}^0 q(z) dz}{\frac{\alpha g}{2 \rho c_p} (Q_{net} - q_d)} \quad (6)$$

304 (Qiu & Kelly, 1993; Kraus & Turner, 1967; Morioka et al., 2011, 2012; Morioka, Ratnam, et  
 305 al., 2013). Here,  $m_0$  (=0.5) is a coefficient of wind stirring (Davis et al., 1981), and  $u_*$  is the  
 306 frictional velocity  $\left[ u_* = \sqrt{\frac{\rho_a C_D u_{10}^2}{\rho}} \right]$ , where  $\rho_a$  (=1.3 kg/m<sup>3</sup>) is air density,  $C_D$  (=0.000125) is the  
 307 drag coefficient, and  $u_{10}$  is the 10 m wind speed. Additionally,  $g$  (=9.8 m s<sup>-1</sup>) represents the  
 308 acceleration due to gravity,  $\alpha$  (=0.00025 °C<sup>-1</sup>) denotes the coefficient of thermal expansion in  
 309 water and  $q(z)$  represents the downward solar insolation calculated using Eq. (5). Following the  
 310 approach of Morioka et al. (2011, 2012, 2013), the interannual variability of  $H_{MO}$  is evaluated  
 311 by

$$312 \quad \delta(H_{MO}) \left[ \equiv \delta \left( \frac{m_0 u_*^3 + q_*}{Q_*} \right) \right] = \frac{m_0 \delta(u_*^3)}{\overline{Q_*}} + \frac{\delta q_*}{\overline{Q_*}} - \frac{\delta Q_* (m_0 \overline{u_*^3} + \overline{q_*})}{\overline{Q_*^2}} + Res. \quad (7),$$

313 where  $q_* = (\rho c_p)^{-1} \alpha g \int_{-H_{MO}}^0 q(z) dz$  represents the effective penetrative radiation and  $Q_* =$   
 314  $(2 \rho c_p)^{-1} \alpha g (Q_{net} - q_d)$  is the effective buoyancy forcing. The overbar and  $\delta()$  represent the  
 315 climatology and their monthly anomaly, respectively. The first term and the sum of the second  
 316 and third terms on the right side of Eq. (7) depict the contribution of wind stirring anomaly and  
 317 net surface heat flux anomaly to the interannual changes in  $H_{MO}$ , respectively.

### 318 3 Results

#### 319 3.1 Model validation

320 Before investigating the generation mechanism of the SST-W4 pattern in detail, we first assess  
 321 the ability of the SINTEX-F2 to capture this pattern by comparing it with the observed data.  
 322 Please note that we use the HadISST during 1979-2018 as observed data in our study. The results

323 from both HadISST and SINTEX-F2 are presented in the left and right panels of Fig. 1,  
324 respectively. The second EOF mode shows the SST-W4 pattern in the observed (Fig. 1a) and  
325 simulated (Fig. 1g) over the subtropical-midlatitude Southern Hemisphere. The model performs  
326 well in reproducing this pattern, exhibiting a high pattern correlation of 0.63 at  $0.5^\circ \times 0.5^\circ$  spatial  
327 resolution with that of observation (significant at a 99% confidence interval, de-correlation spatial  
328 scale  $\sim 20^\circ \times 20^\circ$ ). However, subtle differences are noticeable in the pattern over eastern New  
329 Zealand, southern Australia, and the eastern Indian Ocean. The second mode explains 6.8%  
330 (8.1%) of the total variances in the simulated (observed) data for the analysis period and are well  
331 separated from the 1st and 3rd EOF modes (North et al., 1982) as in the observation. The time  
332 series of the second EOF mode (PC-2) for simulated SST is independent of ENSO, the IOD, and  
333 the SASD, possessing an insignificant correlation coefficient of -0.07, -0.00, and 0.09,  
334 respectively, with their indices. The weak relationship with these climate modes is also found in  
335 the observational study by Senapati, Dash, et al., (2021). It should be mentioned that the first EOF  
336 mode (variance=9.68%) captures the Subtropical Dipole modes in the South Atlantic and Indian  
337 Oceans, having a correlation coefficient of 0.54 and 0.42 with the SASD and the IOSD indices,  
338 respectively. It agrees with the results of Morioka et al. (2011, 2012, 2014a). Also, the  
339 overlapping of the SST-W4 pattern in the Indian Ocean, as noticed by Senapati et al. (2021a),  
340 might lead to a low correlation of 0.33 with the IOSD index. The spatial map of the SST anomaly  
341 associated with the IOSD, SASD (Fauchereau et al., 2003; Senapati, Dash, et al., 2021), and SPSD  
342 indices are different from the SST-W4 pattern (Fig. S1). Note the covariability of IOSD and SASD  
343 is evident, in agreement with Fauchereau et al. (2003).

344 The seasonal dependence of the SST-W4 pattern is further examined by calculating the  
345 normalized monthly standard deviation of PC-2 (Figs. 1b, h). The variability of the SST-W4  
346 pattern from December to April is greater than one (Fig. 1h), indicating a seasonal phase locking  
347 in the simulated results, consistent with observations (Fig. 1b), although there is a slight  
348 difference in the amplitudes. This discrepancy may be attributed to variations in the length of  
349 the analysis periods, different mean state, and variability between the observation and the model  
350 simulation. We then defined the SST-W4 events as years when the normalized PC-2 exceeds  
351 one or minus one standard deviation during austral summer. This criteria resulted in eight  
352 positive years (1989-90, 1991-92, 1996-97, 1999-2000, 2004-05, 2006-07, 2010-11, and 2011-  
353 12) and six negative years (1979-80, 1984-85, 1986-87, 1987-88, 1997-98, and 2017-18) in the

observation. Similarly, we identified 48 positive and 44 negative years in the simulated outputs. Subsequently, we conducted a composite analysis of seasonal SST anomalies during positive SST-W4 events in both the observation data (Figs. 1c-f) and simulation outputs (Figs. 1i-l). The observation reveals that SST anomalies in the subtropics start to appear in the austral spring (September-November), reach their peak in austral summer (December-February), and then gradually decline (Figs. 1c-f). The simulation successfully captures this evolution (Figs. 1i-l). These results validate the model's capability to reproduce the SST-W4 pattern in the subtropical-midlatitude and provides a foundation for investigating the detailed mechanisms underlying the SST-W4 pattern.

### 363 *3.2 Growth and decay mechanisms of the SST-W4 pattern*

364 In this section, we diagnose the air-sea interaction processes on the growth and decay of the  
 365 SST-W4 pattern using the CTL simulation results. To do this, we performed a mixed layer heat  
 366 budget analysis using Eq. (3) at each grid to determine the processes on the tendency of the  
 367 mixed layer temperature. Since the PC-2 could not be useful to quantify each term in Eq. (3),  
 368 we define an index for the SST-W4 pattern (W4 index) as follows,

$$369 \quad [Var] \text{ W4 index} = [Var]_A + [Var]_C + [Var]_E + [Var]_G - [Var]_B - [Var]_D - \\ 370 \quad [Var]_F - [Var]_H \quad (8),$$

371  
 372 where  $[Var]_i$  is defined as the area-averaged anomaly for each variable ( $[Var]$ ) over the  
 373 box " i=A to H" (black boxes in Fig. 2a). The index represents the difference in area-averaged  
 374 anomaly of positive and negative poles over the boxes. The selection of the boxes is based on  
 375 the significant SST anomaly regions seen in composite years (Fig. 2a). It should be noted that  
 376 the overall results are not sensitive to variation in the area of the boxes within the anomalous  
 377 region. Each pole (Fig. 2a) in the SST-W4 pattern peaks during austral summer in both positive  
 378 (Fig. 2b) and negative (Fig. 2c) years. The defined SST-W4 index shows a very high  
 379 significant correlation, 0.88 (0.85), between the PC-2 time series of the simulated (observed)  
 380 data. Note that the regions selected for both simulated and observed analysis are the same. The  
 381 composite map generated using the SST-W4 index for positive and negative years matches  
 382 well with that produced using the PC-2 time series. Further, the budget of each pole for both  
 383 positive and negative years are similar with opposite signs (Fig. S2). Hence, the index is robust

384 for further analysis. Hereafter, the W4 index is used with other variables to quantify terms in  
 385 understanding the physical processes involving the air-sea interaction.

386 **3.2.1 Growth Phase**

387 The growth phase of the SST-W4 pattern is investigated by calculating the MLT ( $T_m$ )  
 388 tendency using Eq. (3) for positive (Fig. 3a) and negative (Fig. S3a) composite years. During  
 389 both the positive and negative years, the MLT anomaly grows from October to January and  
 390 decays afterward. Since the processes are mirror images for the positive and negative years  
 391 (Figs. 3, S3), here we discuss the mechanism for the positive years only. The growth (decay)  
 392 phase refers to the positive (negative) MLT tendency in positive years. The growth (decay) is  
 393 dominated by the anomalous net surface heat flux (entrainment) term (Fig. 3a). Here, the role  
 394 of the horizontal advection term is negligible for the growth and decay of the SST-W4 pattern.  
 395 With the increase in net surface heat flux term, the MLT begins to increase from October,  
 396 although the contribution from residual terms remains consistently negative. However, after  
 397 January, NSHF term begins to decrease, and entrainment term dominates (Fig. 3a). As a result,  
 398 the MLT begins to decrease, and eventually decays. To gain a deeper understanding of the  
 399 dynamics, we decomposed the net surface heat flux term into four components (i.e., NSW,  
 400 NLW, LHF, SHF), as illustrated in Fig. 3b. We find the dominant role of anomalous shortwave  
 401 radiation in the net surface heat flux term. Though anomalous LHF opposes the growth of the  
 402 SST-W4 pattern (Fig. 3b), the anomalous shortwave radiation generates the SST-W4 pattern.  
 403 To further dig into the physical mechanism, the contribution of surface heat fluxes during the  
 404 interannual variation in Eq. (3) is decomposed as

$$405 \delta \left( \frac{Q_{net} - q_d}{\rho c_p H} \right) \left[ \equiv \delta \left( \frac{Q}{\rho c_p H} \right) \right] = \frac{\delta Q}{\rho c_p \bar{H}} - \frac{\delta H \bar{Q}}{\rho c_p \bar{H}^2} + Res. \quad (9),$$

406 where overbar and  $\delta()$  represent the climatology and their monthly anomaly, respectively.  
 407 Different terms in the right hand side of Eq. (9) represents (i) the contribution of the surface heat  
 408 flux anomalies ( $\delta Q$ ) acting on a climatological MLD ( $\bar{H}$ ), (ii) the contribution of the MLD  
 409 anomalies ( $\delta H$ ) under climatological heating/cooling ( $\bar{Q}$ ), and (iii) residual terms. The  
 410 contribution of each term in Eq. (9) for anomalous shortwave radiation and LHF is computed  
 411 and shown in Figs. 3c,d. The second term in Eq. (9) plays a significant role in the anomalous  
 412 contribution of the shortwave radiation (Fig. 3c). It means that the warming of the mixed layer

413 is enhanced (suppressed) by the climatological shortwave radiation (Figs. 3c,f) with the thinner  
414 (thicker) MLD (Fig. 3e). On the other hand, the LHF is always negative, and the cooling  
415 (warming) of the MLT is enhanced by the thinner (thicker) mixed layer (Fig. 3d). However, the  
416 decreased (increased) LHF enhances the warming (cooling) as suggested in Fig. 3f, offsets the  
417 effect of MLT cooling (warming) by the thinner (thicker) mixed layer (Fig. 3d). The notable  
418 residual term, characterized by its significant influence, prevents the growth of the MLT (Fig.  
419 3a). This term encompasses oceanic diffusion and other high-frequency variability.

420 The interannual MLD anomaly may be linked to the surface heat flux and wind stirring  
421 anomalies. To explore the relative contributions, we calculated  $H_{MO}$  anomaly and the  
422 contribution of each term using Eq. (7) during several seasons for both positive (Table 1) and  
423 negative years (Table S1). During positive years,  $H_{MO}$  anomaly increases from September-  
424 November to November-January, then gradually decreases (Table 1), similar to the MLD  
425 anomaly (Fig. 3e). With similar temporal evolution and significant months of the MLD anomaly,  
426  $H_{MO}$  anomaly successfully captures the dynamics of MLD variability and hence considered for  
427 analysis. The interannual variability of  $H_{MO}$  is mainly due to the buoyancy term caused by the  
428 surface heat flux anomalies (Table 1) in which the LHF anomaly plays a crucial role (Fig. 3f).  
429 The role of wind stirring anomaly in the  $H_{MO}$  variability is found to be negligible. Thus, the  
430 positive LHF (Fig. 3f) causes positive buoyancy (Table 1) in the ocean surface resulting in  
431 shallowing of the mixed layer (Fig. 3e) during positive years. In a similar manner, this dynamic  
432 behavior is opposite during the negative years (Table S1, Fig. S3).

433 Hence, anomalies in the mixed layer depth are primarily caused by buoyancy at the ocean's  
434 surface, which is mostly driven by the LHF. The interaction between incoming climatological  
435 solar energy and the mixed layer's thickness, therefore, leads to surface warming or cooling,  
436 subsequently aid in the formation of the SST-W4 pattern during the austral summer.

### 437 3.2.2 Decay Phase

438 After reaching its peak during December-January, the MLT anomaly starts decaying for  
439 positive and negative years mainly due to entrainment term (Figs. 3a, S3a). During early autumn,  
440 the entrainment term plays an essential role in the negative tendency of the MLT anomaly during  
441 positive years (Fig. 3a). Whereas, both entrainment and surface heat flux attributed to latent heat

442 terms, are vital during late autumn (Figs. 3a,b,f). It is obvious that decrease in the NSHF will  
 443 decrease the MLT. However the role of entrainment is not straightforward. To understand the  
 444 cause of entrainment, the entrainment term in Eq. (3) is decomposed into anomalous  
 445 contributions of four terms using the following equation,

$$446 -\delta \left( \frac{\Delta T}{H} w_e \right) = -\frac{\delta(\Delta T) \overline{w_e}}{\overline{H}} - \frac{\delta(w_e) \overline{\Delta T}}{\overline{H}} + \frac{\delta H \overline{\Delta T w_e}}{\overline{H}^2} + Res. \quad (10).$$

447 In this context, the first term on the right-hand side of Eq. (10) corresponds to the impact of  
 448 anomalous  $\Delta T$  (as described in the methods section) under climatological entrainment velocity and  
 449 the MLD on the variability of entrainment process. Likewise, the second term signifies the  
 450 influence of anomalous entrainment velocity with climatological  $\Delta T$  and MLD on entrainment  
 451 variability. The third term, in the presence of climatological  $\Delta T$  and entrainment velocity, accounts  
 452 for the effect of MLD changes on the entrainment process.

453 Fig. 4 displays each term of Eq. (10) for both positive and negative years. It is evident that the  
 454 primary driver of entrainment responsible for the decay of the pattern predominantly stems from  
 455 the first term in Eq. (10). The first term of Eq. (10) means that the variation in entrainment is due  
 456 to the changes in the difference in temperature of entrained water (i.e., thermocline) and the mixed  
 457 layer (Fig. 4). This temperature difference is largely attributed to the MLT anomaly because the  
 458 change in temperature of entrained water is usually minor (Morioka et al., 2011, 2012; Morioka,  
 459 Ratnam, et al., 2013). During late autumn, the anomalous contribution of surface heat flux is  
 460 mostly by the LHF (Figs. 3b,f). Fig. 3d represents the composite time series of anomalous terms  
 461 in Eq. (9) corresponding to LHF. The LHF anomaly in the late autumn is mostly explained by the  
 462 contribution of the MLD anomalies ( $\delta H$ ) under climatological heating/cooling (i.e. the second  
 463 term in eq. (9)). The first term plays a minor role in the LHF anomaly in May. Thus, the decay of  
 464 the MLT anomaly by the climatological LHF is intensified (weakened) by the thinner (thicker)  
 465 mixed layer during the positive (negative) years, as noticed in Fig. 3e (Fig. S3e). It can be  
 466 summarised from the analysis of different terms in Eqs. (9) and (10) that the positive (negative)  
 467 entrainment and the thickening (shallowing) of the mixed layer, decays the MLT in the negative  
 468 (positive) years, even though the climatological flux remains unchanged. During the decay period,  
 469 shortwave radiation term still contributes and maintains the SST-W4 pattern (Figs. 3b, S3b) in  
 470 early autumn, though the LHF term does not contribute much. The influence of shortwave  
 471 radiation term in early autumn (i.e., positive tendency) may be responsible for the 2-month lag  
 472 between the entrainment and LHF terms in the decay of the SST-W4 pattern.

473 To summarize, we have identified that LHF anomaly plays a pivotal role in positive buoyancy  
474 at the ocean surface, resisting mixing and thus causing a shallowing of the mixed layer depth. The  
475 interaction between incoming climatological solar radiation and the mixed layer's thickness  
476 results in surface warming (or cooling) and, consequently, positive (or negative) SST anomalies,  
477 which contribute to the development of the SST-W4 pattern during austral summer. Subsequently,  
478 the decay of the SST-W4 pattern during austral autumn is attributed to anomalous entrainment  
479 caused by the changes in temperature difference between the mixed layer and the water below it,  
480 as well as anomalous LHF. This underscores the critical importance of the upper ocean variability  
481 in both the growth and decay phases of the pattern. Now the question arises: How does the  
482 atmospheric W4 play a role in development of the SST-W4 pattern? The origins of the LHF  
483 anomaly linking the SST with atmosphere are elucidated in the subsequent section.

484 *3.3 Linkage of SST-W4 pattern with atmosphere in the subtropical-midlatitude Southern*  
485 *Hemisphere*

486 Following Senapati et al. (2021a), the MCA (see methods for details) is performed among  
487 various atmospheric and oceanic variables simulated by the SINTEX-F2 to examine and analyze  
488 the generation mechanism of the SST-W4 pattern linked with the atmospheric variability. The  
489 SST-W4 pattern emerges as the second mode of the SST anomaly with the meridional wind (V;  
490 variance= 13.43%), LHF (variance= 6.71%), and MLD (variance=8.42%) anomalies (Figs. 5a-  
491 c). It shoud be noted that, the role of zonal wind anomaly in the development of zonal SST  
492 pattern throughout the region is negligible (Fig. 5a) and hence omitted in this study. The SST  
493 pattern in the MCA (shaded; Fig. 5a) is similar to the earlier result of the second EOF mode  
494 (Fig. 1g). A strong correlation (0.85) is found between the expansion coefficient of SST pattern  
495 of second MCA mode and EOF method. The net surface heat flux anomaly is mostly dominated  
496 by the LHF anomaly compared to other surface fluxes in the generation of the SST-W4 pattern  
497 (Figs. S4, 3f) and hence considered here. Note that, all the 2nd modes obtained from the MCA  
498 are significantly independent from other modes (Table S2).

499 The difference in the centre of action between the SST and V is noticeable (Fig. 5a) and  
500 can be explained by their interactions with the MLD (Fig. 5b) and LHF (Fig. 5c) (Senapati,  
501 Dash, et al., 2021). Figs. 5d and 5e show the auto-correlation and cross-correlation of the  
502 variables to examine the evolution of the SST-W4 mode. The auto-correlation of the SST, MLD,

V, and LHF indicates their persistence up to 7, 4, 1, and 1 month respectively (Fig. 5d). Cross-correlations of SST-MLD, SST-V, and MLD-V bring out the origin of SST and MLD signals in response to the meridional wind. On the other hand, the LHF shows a significant link with the V (correlation coefficient of 0.4). Interestingly, the MLD, V, and LHF covary with no lag. The SST also exhibits maximum association with V at a one-month lag. Thus, cross-correlation analysis suggests that the meridional wind forces the ocean surface to generate the SST-W4 signal via MLD and LHF. The fading of atmospheric signals when the SST takes the lead is also noticeable (Fig. 2e). Nevertheless, the MLD significantly persists up to 4 months after SST takes a lead, which helps the SST-W4 pattern to remain for a while. The transfer of heat from the ocean to the atmosphere after the SST peak reverses the flux pattern and their cross-correlations after one month (Fig. 2e). The meridional wind could transport warm moist air (cool dry air) from tropical (polar) region to mid-latitude causing the humidity difference in the air-sea interface, and resulting in the LHF variability (Senapati, Dash, et al., 2021).

The spatial correlation of PC-2 with meridional wind anomalies (shaded) and geopotential height anomalies (contours) at 250, 500, and 850 hPa depicts an equivalent barotropic structure of the W4 pattern in the troposphere (Fig. S5). The source and dynamics of the atmospheric wave that generates a similar pattern in the SST are investigated by using composite analysis (Fig. 6) and RWS analysis (Fig. 7). Fig. 6 shows the monthly composite maps of horizontal wind and geopotential height anomalies at 250 hPa for the months before the SST-W4 events (Figs. 6a–d). In September, three cells of geopotential height and wind anomalies are positioned in the South Pacific. One cell with a negative geopotential height is located over south of New Zealand. The other two cells are found with dipole-like structures oriented in a north-south direction in the eastern Pacific (Fig. 6a). A prominent cell of positive anomaly is also seen over the south of the African continent (Fig. 6a).

In October, the negative (positive) SST anomaly (Fig. 6e) over the western and eastern (central) South Pacific suppresses (enhances) the convective activity as represented by anomalous precipitation and Outgoing Longwave Radiation (OLR) (shading and contours; Fig. 6f) and hence the accompanied positive (negative) velocity potential anomaly (contours; Fig. 6e) and upper-level convergence (divergence) (vectors; Fig. 6e). As a result, the

negative geopotential height anomaly strengthens and expands over the off-shore of New Zealand (Fig. 6b). At the same time, the negative geopotential height moves further southeast in the eastern South Pacific (Fig. 6b). Also, a positive geopotential height anomaly in east Australia and the western Pacific is noticeable (Fig. 6b). In the following months, a well-developed atmospheric W4 pattern appears in the subtropical-midlatitude Southern Hemisphere (Figs. 6c-d). It is mainly due to the Rossby wave train emanating from the SWSP and confined in the westerly jet (solid blue contours in Fig. 6), as evident from the WAF (vectors; Fig. 6f).

This scenario is a mirror image during negative years (Figs. 6g-l). Climatologically, the precipitation band in 60°S-40°S (Fig. S6a) mainly corresponds to the relatively warm mean SST (Fig. S6b) and the strong meridional SST gradient. The region is also influenced by atmospheric circulations that promote moisture transport to the region from the tropical region. The SST anomaly over the SWSP with a climatological SST of around 25°C in 40°S-30°S (Fig. S6b) triggers the convective activity (as seen in OLR) and causes anomalous rainfall over there (Figs. 6e,f,k,l). Positive (negative) precipitation, negative (positive) OLR, negative (positive) velocity potential and divergence (convergence) during negative (positive) years suggest a diabatic source (sink) over the SWSP (Figs. 6e,f,k,l). However, the co-occurrence of precipitation, OLR, and velocity potential anomalies in other regions like the South Atlantic and central and eastern South Pacific creates an ambiguity regarding the possible sources that trigger the atmospheric W4 pattern. The RWS is calculated using Eq. (1), which consists mainly of the stretching of the vortex by eddies (Term-1) and the advection of absolute vorticity by divergent wind (Term-2). Since the peak of the SST-W4 pattern occurs with a 2-month lag of convective activity (i.e., October-November; Fig. 6), a correlation map is constructed between PC-2 of austral summer and anomalous RWS (Fig. 7a), Term-1 (Fig. 7b), and Term-2 (Fig. 7c) during October-November. The anomalous RWS shows three prominent areas in the subtropical Pacific (Fig. 7a) due mainly to the anomaly in Term-1 (Fig. 7b) as the possible source of the atmospheric W4 pattern. The contribution of anomalous Term-2 in RWS anomaly is negligible (Fig. 7c). Basically, Term-1 anomaly includes the anomalous RWS activity due to anomalous divergence under climatological absolute vorticity and hence the atmospheric W4 pattern. Since the Rossby wave travels eastward guided by the westerly jet in the Southern Hemisphere, the upstream

562 region in the SWSP is a potential source for the atmospheric W4 pattern. The WAF anomaly  
563 emitted from the SWSP (vectors, Figs. 6f,l) corresponding to precipitation-induced diabatic  
564 heating/cooling (shaded, Figs. 6f,l), OLR anomalies (contours, Figs. 6f,l), velocity potential  
565 anomaly (contours; Figs. 6e,k), and anomalous divergence (vectors, Figs. 6e,k) validates the  
566 RWS over the SWSP as the probable source in generating the atmospheric W4 pattern. Here, a  
567 natural question arises: What favors the formation of atmospheric W4 Rossby waves?

568 To identify such factors, we investigated the properties of the background atmospheric  
569 flow in the propagation of the Rossby wave during the development of the atmospheric W4  
570 pattern. The stationary Rossby wavenumber  $K_s [= (\beta_M / \bar{u})^{1/2}]$  is estimated for this purpose; where  
571  $\bar{u}$  denote the mean zonal wind. Here,  $\beta_M (= \beta - \bar{u}_{yy})$  comprising  $\beta$  and  $\bar{u}_{yy}$  represents the meridional  
572 gradient of planetary vorticity and the time mean of the meridional relative vorticity gradient,  
573 respectively (B. J. Hoskins & Ambrizzi, 1993). Fig. 8 shows the  $K_s$  and its components during  
574 the November-January months for the positive (left panel, Fig. 8) and negative (right panel, Fig.  
575 8) SST-W4 years. The existence of the stationary W4 in the midlatitude between  $30^{\circ}$ S to  $60^{\circ}$ S  
576 is clearly evident during both positive and negative SST-W4 years (red contour, Fig. 8a). This  
577 is due to the effective meridional gradient in planetary vorticity (Fig. 8b) guided by the curvature  
578 of the westerly jet (Fig. 8c). Therefore, the climatological background flow (during November-  
579 January) is necessary to establish the disturbance over the SWSP into an atmospheric  
580 wavenumber-4 pattern in the Southern Hemisphere.

### 581 3.4 Potential source of the SST and atmospheric W4 patterns

582 The above intriguing results indicate the role of the SWSP in generating the SST and  
583 atmospheric W4 patterns. To examine the root cause of the convective activity over the SWSP  
584 and its role in developing the SST and atmospheric W4 patterns, we conducted sensitivity  
585 experiments with the SINTEX-F2 (see Section 2.2). Prior to conducting any sensitivity  
586 experiments, it is imperative to acknowledge that the SINTEX-F2 demonstrates a high level of  
587 skill in capturing and predicting tropical and subtropical climate dynamics and their  
588 teleconnections up to seasonal scales (Doi et al., 2016). Therefore, the analysis of sensitivity  
589 experiments involving the nudging of the tropical and SWSP towards climatological value  
590 (noTropics and noSWSP) in the SINTEX-F2 is considered robust. The selection of the noSWSP  
591 sensitivity experiment area is based on the potential RWS (Fig. 7a), precipitation and OLR

592 patterns (Figs. 6f,l), and velocity potential and divergence patterns (Figs. 6e,k) associated with  
593 the SST and atmospheric W4 patterns. The spatial patterns of the second EOF mode of SST  
594 anomaly for all the simulations are shown in Fig. 9. It shows the SST-W4 pattern in the second  
595 mode for early 70 yr of CTL simulation (Fig. 9a), which shows similar features as of Fig. 1g. In  
596 the noTropics, the SST over the tropical region ( $25^{\circ}\text{S}$ - $25^{\circ}\text{N}$ ) is nudged to the climatological SST  
597 to suppress the influence of tropical region on the SWSP climate variability (Liess et al., 2014),  
598 so also on the generation of the W4 patterns. Interestingly, the EOF pattern remains unchanged  
599 (Fig. 9b), suggesting the generation mechanisms are not dependent on the variations of the SST  
600 in the tropical regions. Therefore, it can be hypothesized that the variability over the SWSP  
601 might be locally generated or influenced by the polar regions. To confirm this, we perform the  
602 noSWSP experiment by nudging the SST to its climatological value over the SWSP (black  
603 rectangle box, Fig. 9c). In the absence of air-sea interaction due to the SST anomalies over the  
604 SWSP suppresses the SST and atmospheric W4 patterns in the southern subtropics-midlatitudes  
605 (Fig. 9c), especially over the Indian Ocean and the ocean to south of Australia. The SST  
606 anomalies over the eastern Pacific and Atlantic Oceans are not sensitive to the SWSP SST  
607 variability, as seen in the Fig. 9c. Thus, the local air-sea interaction over the SWSP is crucial for  
608 the generation of the SST and atmospheric W4 patterns.

609  
610 The composite analysis of CTL, noTropics, and noSWSP simulation further provides an  
611 insight to the dynamical mechanisms involved. The left panel in Fig. 10 shows the SST anomaly  
612 (shaded) and 850 hPa wind anomaly (vectors) for positive composite years in the CTL,  
613 noTropics, and noSWSP simulations (Figs. 10a-c). Same for the negative years are shown in  
614 Figs. S7a-c. Similarly, the middle and right panels show the geopotential height anomalies  
615 (shaded) and WAF (vectors) at 250 hPa (Figs. 10d-f), and precipitation (shaded) and divergent  
616 wind anomalies (vectors) at 250 hPa (Figs. 10g-i), respectively. During positive years in the  
617 CTL simulation, wave activity flux emanating from the SWSP (vectors; Fig. 10d) in response  
618 to precipitation (i.e., diabatic heating) (shaded; Fig. 10g) and divergent wind (vectors; Fig. 10g)  
619 contributes to the development of the W4 patterns in the 250 hPa geopotential height (shaded;  
620 Fig. 10d), 850 hPa wind (vectors; Fig. 10a), and SST (shaded; Fig. 10a) anomalies. The Rossby  
621 wave train originates from the SWSP, travels towards Antarctica, where it gets trapped within  
622 the nearby westerly jet, subsequently following the great circle path to establish the atmospheric  
623 W4 structure in the mid-latitudes (Figs. 10d, g). In the noTropics experiment, the dynamics seen

624 to be similar (Figs. 10b,e,h) with the CTL simulation. The amplitude of SST variances do not  
625 change largely in noTropics as compared to CTL simulation (Figs. 9-10). Similar with  
626 atmospheric variables as well (Figs. 10e, h). Also, the frequency of occurrence (i.e., 19 extreme  
627 years in both CTL and noTropics) remains the same, suggesting the weak relation of the W4  
628 pattern with the tropics. In fact, the Subtropical Dipoles in each basin can be generated owing  
629 to the SAM in the absence of tropical climate variability (Morioka et al., 2014a). However, in  
630 the W4 case, the geopotential height does not vary over mid and high latitudes in the noTropics  
631 experiment, omitting the role of the SAM in SST and atmospheric W4 patterns (Fig. 10e). In  
632 addition, the convective activity over the SWSP is unaffected from the tropical region (Fig. 10h).  
633 Recently, the role of SAM is also found to be negligible in generation of zonal wavenumber-3  
634 in absence of perturbation in tropics or subtropics (Goyal et al., 2021). Thus, it can be concluded  
635 that, the generation mechanisms of SST and atmospheric W4 patterns are not affected by the  
636 tropical climate.

637 Conversely, in the reduction of convective activity over the SWSP, as demonstrated in the  
638 noSWSP experiment (Fig. 9c), both the circumglobal SST and atmospheric W4 patterns  
639 dissipate. The rainfall deficit amounts to -18.3, -13.1, and -7.3 mm month<sup>-1</sup> in the CTL,  
640 noTropics, and noSWSP experiments during positive years, respectively. This significant  
641 decrease in convection activity in the noSWSP experiment compared to the CTL experiment,  
642 implies much differences in the generation dynamics between the experiments (Figs. 10c, f, i).  
643 Due to less convective activity in the SWSP region, the Rossby wave train fails to propagate  
644 circumglobally and cannot form the characteristic W4 atmospheric structure. In this scenario,  
645 the Rossby waves emanating from the southeast of SWSP become trapped within a waveguide,  
646 propagating toward the equator over South America, not reaching the South Atlantic. As a  
647 consequence, the W4 pattern in SST and atmosphere could not be establish in the subtropical-  
648 midlatitude Southern Hemisphere. The scenario is the opposite during negative years (Fig. S7).  
649 It is interesting to observe the negative SAM in noSWSP experiment during positive years (Fig.  
650 10f). It could be due to the overwhelming role of the SAM, the dominant atmospheric intrinsic  
651 variability in the Southern Hemisphere, where the W4 patterns linked to SWSP are superposed.  
652 The SST anomalies, specifically over south Atlantic and Indian Ocean seems to be related to the  
653 negative SAM in noSWSP experiment (Ciasto & Thompson, 2008; Screen et al., 2010).

654        Despite convective activity in other regions in the noSWSP experiment, SST and  
655        atmospheric W4 pattern generation is absent, confirming the essential role of SWSP in this  
656        process. Hence the noSWSP experiment concludes that variability in the SWSP as a necessary  
657        condition for forming the circumglobal Rossby waves and, subsequently, the SST and  
658        atmospheric W4 patterns. The behavioral change in the Rossby wave propagation without the  
659        SST anomalies over the SWSP requires further investigation in future studies.

660        **4 Summary and Discussion**

661        Using the SINTEX-F2, we have demonstrated the W4 pattern of SST anomalies in the  
662        subtropical-midlatitude Southern Hemisphere. A realistic simulation of the W4 pattern in the  
663        CGCM has allowed us to conduct a comprehensive investigation of the generation mechanisms  
664        using the model experiments. The results are summarized using a schematic diagram (Fig. 11). (1)  
665        The convection activities and associated divergent wind anomaly over the SWSP stretch the vortex  
666        near the westerly jet due to diabatic heating and divergent wind during October-November. (2)  
667        This disturbance gets trapped in the westerly waveguide and circumnavigates the globe,  
668        establishing an atmospheric W4 pattern over the southern midlatitudes. The disturbance follows  
669        the meridional gradient of effective planetary vorticity and is guided by the curvature of the mean  
670        zonal flow, forming an atmospheric W4 pattern. The atmospheric W4 pattern has an equivalent  
671        barotropic structure in the troposphere and interacts with the upper ocean in the southern subtropics  
672        and midlatitudes. The air-sea interaction processes involved in the growth and decay of the SST-  
673        W4 pattern are investigated using mixed layer budget analysis. We also defined an index for the  
674        SST-W4 pattern to perform the budget analysis. (3, 4) The anomalous wind induces a variation in  
675        the MLD via a LHF anomaly over the region. (5) Because incoming climatological solar energy  
676        is absorbed in a thinner (thicker) mixed layer, the shallower (deeper) MLD supports surface  
677        warming (cooling). (6) In this way, the SST-W4 is generated in the southern subtropics and  
678        midlatitudes via a thermodynamic coupling between the upper ocean and atmosphere. Then, the  
679        SST-W4 pattern experiences the following three processes and starts decaying during austral  
680        autumn. Most dominantly, (7) the entrainment caused by the difference in temperature between  
681        the entrained water and the mixed layer causes the SST pattern to disappear in early autumn. The  
682        disappearance of the atmospheric wave is also an important cause of the pattern's decay during  
683        early autumn. Also, the anomalous LHF, which turns around after SST forcing, induces cooling

684 (warming) over the warm (cold) pole and hence contributes to the decay of the SST-W4 pattern in  
685 late autumn.

686  
687 The SINTEX-F2 has the ability to capture subtropical-midlatitude dynamics on a global  
688 scale. Although air-sea interaction processes are similar to those of Subtropical Dipoles in each  
689 basin (Morioka et al., 2011, 2012; Morioka, Ratnam, et al., 2013), this study highlights the role of  
690 ocean mixed layer in the life cycle of the SST-W4 pattern for the first time. The subsurface  
691 warming may have implications for the MLT, yet our investigation reveals no evidence of such  
692 warming in our study area, as confirmed by both observational data (Orsi, 1998; Orsi & Whitworth,  
693 2005) and SINTEX-F2 model outputs (Fig. S8).

694 Of particular interest is the weakening of the WAF over the South Atlantic and Indian  
695 Ocean during positive years (Figs. 6a-f). The influence of the SAM during positive years is  
696 apparent in the noSWSP experiment (Fig. 10f). SST anomalies, especially those over the South  
697 Atlantic and Indian Ocean, appear to correlate with the negative SAM in the noSWSP experiment  
698 (Ciasto & Thompson, 2008; Screen et al., 2010). However, the role of the SAM is not evident  
699 during positive years in the CTL experiment (Fig. 10d), though it becomes apparent during  
700 negative years (Fig. S7d). Thus, while the SAM may contribute to maintaining SST and  
701 atmospheric W4 over the South Atlantic and Indian Oceans, further investigation is required for  
702 clarification. Additionally, the presence of positive HGT over the southwestern Indian Ocean from  
703 September onwards, strengthening in subsequent months, is notable during positive years  
704 compared to negative years (Fig. 6). However, the role of the southwestern Indian Ocean in the  
705 development of SST and atmospheric W4 falls outside the scope of our study (Figs. 6a-f).

706 On the other hand, interaction of the SST-W4 pattern with the Subtropical Dipoles, South  
707 Atlantic-southern Indian ocean pattern (Lin, 2019), mid-tropospheric semi-permanent  
708 anticyclones (C. J. C. Reason, 2016), SAM, Pacific South American patterns (Grimm & Ambrizzi,  
709 2009), subtropical highs are intriguing but leave open questions for future work. Along with this,  
710 the contribution of SST-W4 variability apart from the Subtropical Dipoles in modulating rainfall  
711 over southern continents directly or by a local response (Morioka et al., 2015) needs further  
712 numerical modeling experiments. Furthermore, sensitivity experiments confirm the little  
713 involvement of the tropical forcings and SAM in the dynamics of SST and atmospheric W4  
714 patterns, which contrasts with the probable sources of Subtropical Dipoles (Morioka et al., 2014a).  
715 In these sensitivity experiments, the precipitation and divergent wind anomalies seen in the SST

716 nudged areas might be due to the high-frequency variabilities. The local air-sea interaction over  
717 the SWSP is found to be the necessary condition for the generation of SST and atmospheric W4  
718 patterns. However, it will be interesting to quantify the amount of precipitation/diabatic heating  
719 over the SWSP that necessary for the generation of W4 pattern. The role of the tropical climate in  
720 the W4 pattern on a decadal timescale (which is related to SPMM; Senapati et al., 2022), and the  
721 linkage of the Tasman Sea with tropical regions (Liess et al., 2014) cannot be neglected. In the  
722 absence of tropical climate variation, the W4 pattern can be generated but the tropical climate  
723 variability may play a role in modulating the W4 pattern through atmospheric teleconnection and  
724 changes in local SST over Tasman Sea, which will be studied in the future. The Rossby wave  
725 teleconnection from the SWSP in the absence of tropical influence provides new insights into  
726 subtropical climate variability. For example, studies suggest that the relationship between ENSO  
727 and the Antarctic sea ice has changed in recent decades because of the reduced response of the  
728 Tasman Sea to ENSO (Dou & Zhang, 2022). Therefore, a better understanding of the linkage  
729 between the Southern Hemisphere climate variability and the SWSP is required.

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740 Operator, Python, and Matlab have been used for the analysis. Figures are plotted using  
741 PyFerret, Python, and Matlab.

742

### 743 **Open Research**

744 The observed SST data (Rayner et al., 2003) is available on the HadISST website  
745 (<https://www.metoffice.gov.uk/hadobs/hadisst/data/download.html>). SINTEX-F2 model output used to

746 make figures in this study will be available from the Zenodo repository by the time of manuscript  
747 publication.

748

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949

950 **Figures:**

951 Figure 1. (a) The second EOF mode of SST anomalies (in °C) from HadISST over the  
952 subtropical-midlatitude Southern Hemisphere. Values in the top right explain its variance. (b)  
953 The normalized monthly standard deviation of the second principal component (PC-2).  
954 Composite of the observed SST anomalies (in °C) during (c) SON(-1) (d) DJF(0) (e) MAM(0),  
955 and (f) JJA(0) of the positive SST-W4 pattern. "-1 or 0" in the brackets denotes the year before  
956 (during) the event year. (c-f) Values exceeding 90% confidence level using a two-tailed  
957 Student's *t*-test are colored. (g-l) Same as in (a-f), but for the SINTEX-F2 model results.

958

959 Figure 2. (a) Composite map of SST anomalies (in °C) in the CTL experiment during the austral  
960 summer of the positive years. The black rectangular boxes used for calculation of the SST-W4  
961 index correspond to the region A(120°E-165°E, 45°S-30°S), B(170°E-154°W, 55°S-30°S),  
962 C(142°W-105°W, 51°S-23°S), D(100°W-77°W, 50°S-26°S), E(52°W-27°W, 48°S-27°S),  
963 F(20°W-15°E, 45°S-23°S), G(31°E-68°E, 45°S-22°S), and H(70°E-110°E, 37°S-20°S). (b)  
964 Composite time series of SST anomaly averaged over each box during the positive years. (c)  
965 Same as in Fig. 2b, but for the negative years. To smooth the time series, a 3-month running  
966 mean is applied. Open circles show significant anomalies with 90% confidence levels using a  
967 two-tailed Student's *t*-test.

968 Figure 3. (a) Time series of composite anomalies of mixed layer heat budget terms in Eq. (3)  
969 ( $\times 10^{-7} \text{ }^{\circ}\text{C s}^{-1}$ ) for the positive years. (b) Time series of components of net surface heat flux  
970 terms in the right-hand side of Eq. (3) ( $\times 10^{-7} \text{ }^{\circ}\text{C s}^{-1}$ ) during positive years. Each term of Eq.  
971 (9) ( $\times 10^{-7} \text{ }^{\circ}\text{C s}^{-1}$ ) corresponding to  $\delta Q$  for (c) the short wave radiation and (d) the latent heat  
972 flux during the positive years. (e) Time series of the MLD anomalies (in m), defined as  $\delta H$  in  
973 Eq. (9). (f) Time series of NSHF, NLW, NSW, SHF and LHF anomalies in  $\text{W m}^{-2}$  during the  
974 positive years. In (a) MLT tend., NSHF term, Hor. adv., Ent., and Res., indicate the tendency of  
975 MLT, net surface heat flux, horizontal advection, entrainment, and residual terms, respectively.  
976 In (b,c,d,f) NSW, NLW, LHF, and SHF indicate net shortwave radiation, net longwave  
977 radiation, latent heat flux, and sensible heat flux, respectively. To smooth the time series, a 3-  
978 month running mean is applied. Open circles show 90% significant anomalies using a two-tailed  
979 Student's t-test.

980

981 Figure 4. (a, b) Time series of anomalous contributions from each entrainment term  
982 ( $\times 10^{-7} \text{ }^{\circ}\text{C s}^{-1}$ ) in Eq. (10) during the positive and negative years, respectively. To smooth the  
983 time series, we applied a 3-month running mean. Open circles show significant anomalies with  
984 90% confidence levels using a two-tailed Student's *t*-test.

985

986 Figure 5. The second MCA mode between (a) SST anomaly (shaded; in  $^{\circ}\text{C}$ ) and 850 hPa  
987 horizontal wind anomaly (vectors; in  $\text{m s}^{-1}$ ), meridional wind anomaly (contours, interval:  $0.2 \text{ m}$   
988  $\text{s}^{-1}$ ) (b) SST anomaly (contours, interval:  $0.05 \text{ }^{\circ}\text{C}$ ) and MLD anomaly (shaded; in m) and (c) LHF  
989 anomaly (shaded; in  $\text{W m}^{-2}$ ; positive downward) and SST anomaly (contours, interval:  $0.05 \text{ }^{\circ}\text{C}$ ).  
990 Variances are given on top-right of each panel (note: (a) is for variance between SST and  
991 meridional wind anomaly). Solid (dotted) contours represent positive (negative) values. (d) Auto-  
992 correlation and (e) cross-correlation of the second MCA time series among the SST anomaly,  
993 MLD anomaly, meridional wind anomaly (V), and LHF anomaly as indicated by the colors. The  
994 X-axis (Y-axis) represents the monthly lead/lag (correlation coefficients) and positive (negative)  
995 lag means the first variable is leading (lagging) the second. 99% statistical confidence level using  
996 Student's *t*-test are shown with dashed black lines.

997

998 Figure 6. Monthly composites of anomalous 250 hPa geopotential height (shaded; in m) and  
999 wind (vectors; in  $m s^{-1}$ ) for (a) September, (b) October, (c) November, and (d) December during  
1000 the positive years. The blue contours of the zonal wind highlight the westerly jet. (e) SST  
1001 anomaly (shaded) in October, velocity potential anomaly (contours, in  $10^4 m^2 s^{-1}$ ) and  
1002 anomalous 250 hPa divergent wind (vectors; in  $m s^{-1}$ ) during October-November, and (f)  
1003 anomalous precipitation during October-November (shaded; in  $mm month^{-1}$ ), Outgoing  
1004 Longwave Radiation (black contours, in  $W m^{-2}$ ) and WAF (vectors, in  $m^2 s^{-2}$ ). (g-l) Monthly  
1005 composites for the negative event years. Values exceeding 90% confidence levels using  
1006 Student's *t*-test are plotted.

1007

1008 Figure 7. Correlation maps between the PC-2 in austral summer (December-February) and  
1009 anomalous (a) RWS, (b) Term-1, and (c) Term-2 of Eq. (1) during October-November. Values  
1010 exceeding 99% confidence level using Student's *t*-test are colored. The contour of black-dotted  
1011 lines of zonal wind represents the westerly jet.

1012

1013 Figure 8. (a) Stationary Rossby wavenumber ( $K_s$ ), (b) meridional gradient of absolute vorticity  
1014 ( $\beta_M$ ) ( $\times 10^{-11} m^{-1} s^{-1}$ ), and (c) zonal velocity at 250 hPa during November- January of the positive  
1015 years. (d-f) Same as in (a-c), but for the negative years. The solid red (blue) contour in (a) and  
1016 (d) correspond to stationary Rossby wavenumber 4 (5).

1017

1018 Figure 9. Spatial patterns of the second EOF mode of SST anomaly over the subtropical-  
1019 midlatitude region ( $55^{\circ}S$ - $25^{\circ}S$ ) from (a) the CTL, (b) noTropics, and (c) noSWSP experiments.  
1020 Here we used initial 70-yr simulation in the CTL experiment after removing the first 30 yr to  
1021 have consistency with the noTropics and noSWSP experiments. Values in the top right explain  
1022 their variance. The black rectangle box in (c) shows the SST-nudging area in the noSWSP  
1023 experiment.

1024

1025      Figure 10. (a-c) Composite maps of SST (shaded; in  $^{\circ}\text{C}$ ) and 850 hPa wind anomalies (vectors; in  $\text{m s}^{-1}$ ) for the CTL, noTropics, and noSWSP experiments during positive years. The middle  
 1026      panel (d-f) is similar to the left panel but for anomalous geopotential height (shaded; in m) and  
 1027      wave activity flux (vectors; in  $\text{m}^2 \text{s}^{-2}$ ) at 250 hPa. Similarly, the right panel (g-i) is for  
 1028      precipitation (shaded; in  $\text{mm month}^{-1}$ ) and 250 hPa divergent wind anomalies (vectors; in  $\text{m s}^{-1}$ ). HGT and WAF indicate the geopotential height and wave activity flux. Values not satisfying  
 1029      90% confidence level in a two-tailed Student's *t*-test are masked out. The black rectangle box  
 1030      in (c) shows the SST-nudging area in the noSWSP experiment.  
 1031

1033

1034      Figure 11. Schematic diagram describing the generation mechanism of the SST-W4 pattern in  
 1035      the subtropical-midlatitude Southern Hemisphere.

1036

1037      **Table:**

1038

1039

	SON	OND	NDJ	DJF	JFM
Monin-Obukhov depth	-16.64	<b>-30.04</b>	<b>-35.29</b>	<b>-32.29</b>	-10.07
Wind stirring	-0.01	-0.19	<b>-0.28</b>	<b>-0.21</b>	0.5
Flux	-16.5	<b>-29.95</b>	<b>-35.22</b>	<b>-32.21</b>	-10.7
Res.	-0.13	0.1	0.21	0.13	0.13

1040

1041      Table 1. Monin-Obukhov depth anomaly (in m) and anomalous contributions from surface flux,  
 1042      wind stirring, and residual terms in Eq. (7) during September-November, October-December,  
 1043      November-January, December-February, and January-March of the positive years. Bold letters  
 1044      show significant anomalies with 90% confidence levels using a two-tailed Student's *t*-test.

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