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Deforestation Impacts on Amazon-Andes Hydroclimatic Connectivity

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Deforestation impacts on Amazon-Andes hydroclimatic connectivity

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Abstract

Amazonian deforestation has accelerated during the last decade, threatening an ecosystem where almost one third of the regional rainfall is transpired by the local rainforest. Due to the precipitation recycling, the southwestern Amazon, including the Amazon-Andes transition region, is particularly sensitive to forest loss. This study evaluates the impacts of Amazonian deforestation in the hydro-climatic connectivity between the Amazon and the eastern tropical Andes during the austral summer (December-January-February) in terms of hydrological and energetic balances. Using 10-year high-resolution simulations (2001-2011) with the Weather Research and Forecasting Model, we analyze control and deforestation scenario simulations. Regionally, deforestation leads to a reduction in the surface net radiation, evaporation, moisture convergence and precipitation (~20%) over the entire Amazon basin. In addition, during this season, deforestation increases the atmospheric subsidence over the southern Amazon and weakens the regional Hadley cell. Atmospheric stability increases over the western Amazon and the tropical Andes inhibiting convection in these areas. Consequently, major deforestation impacts are observed over the hydro-climate of the Amazon-Andes transition region. At local scale, nighttime precipitation decreases in Bolivian valleys (~20-30%) due to a strong reduction in the humidity transport from the Amazon plains toward Andes linked to the South American low-level jet. Over these valleys, a weakening of the daytime upslope winds is caused by local deforestation, which reduces the turbulent fluxes at lowlands. These alterations in rainfall and atmospheric circulation could impact the rich Andean ecosystems and its tropical glaciers.

Keywords

Deforestation, high-resolution climate modelling, Amazon basin, Amazon-Andes transition region, energy and water balances.

1. Introduction

The Amazon is the largest tropical rainforest in the world. It exchanges water, energy, and carbon with the

41 atmosphere and is considered as an essential dynamic component of the climate system (Bonan 2008; Malhi et al.
42 2008). The Amazon ecosystem has been recognized as one of the major sources of evapotranspiration (Fisher et
43 al. 2011), releasing large amounts of water vapor to the atmosphere and sustaining rainfall not only in the Amazon
44 itself, but also in remote regions of the South American continent (Eltahir and Bras 1994; Van Der Ent et al. 2010;
45 Martinez and Dominguez 2014; Zemp et al. 2014). In addition, the latent heat release linked to moisture recycling
46 over the Amazon basin engines global and regional atmospheric circulation (Gill 1980; Eltahir 1996; Gedney and
47 Valdes 2000). Therefore, understanding the impacts of continued deforestation on the regional and global
48 hydroclimate has become an urgent and extensively explored scientific question.

49
50 The Amazonian forest has been intensively cleared since the 1970's (Fearnside 1993; Nepstad et al. 2009). After
51 a slowdown of deforestation rates during the period 2004-2012, the small-scale forest loss rates increased again
52 over the last decade (Kalamandeen et al. 2018; Eiras-Barca et al. 2020). Deforestation, as every effect of human
53 activities on the climate system, disturbs the Earth's radiation budget and can be an important forcing for climate
54 change (Hartmann et al. 2013; Mahowald et al. 2017). Indeed, when the rainforest is replaced by cropland or
55 grassland, the energy balance is disturbed by: (i) a reduction in surface roughness and actual evapotranspiration,
56 decreasing the surface energy loss from turbulent heat removal and latent heat flux (Gash and Nobre 1997; Lee et
57 al. 2011; De Noblet-Ducoudré et al. 2012); (ii) an increase in the incoming solar radiation at the surface linked to
58 less clouds and less evapotranspiration rates (Salati and Vose 1984; Bala et al. 2007); (iii) a reduction in absorbed
59 solar radiation due to an increased surface albedo (Berbet and Costa 2003; von Randow et al. 2004); and finally
60 (iv) less incoming infrared radiation driven by drier air as a result of the reduced evapotranspiration (Eltahir 1996;
61 Claussen et al. 2001). *In-situ* measurements find a decrease in the surface net radiation over cleared areas (Bastable
62 et al. 1993; von Randow et al. 2004). Although at the annual time scales an air temperature warming of the layer
63 close to the surface follows the forest clearance (Gash and Nobre 1997), the reduction in the net surface radiation
64 generates a cooling of the upper atmosphere and a subsidence that weakens the regional circulation (Eltahir 1996).
65 In this sense, perturbations that initially occurs at the surface and within the lower part of the atmosphere modify
66 the vertical transport of heat and moisture, altering convection and subsidence over extended regions and affecting
67 large-scale features such as the Hadley and Walker cells (Zhang et al. 1996; Snyder 2010; Badger and Dirmeyer
68 2015).

69
70 Atmospheric circulation and Amazonian rainforest compound a coupled bio-geophysical balanced system.
71 Perturbations in the vegetation state through deforestation can drive very different changes in atmospheric
72 circulation depending on its spatial scale. Observations and theoretical works show small-scale deforestation
73 (<10km) triggering thermally driven land breeze resulting from differences in sensible heat fluxes between
74 pastured and forested areas (Pielke, R. A. 1984; Atkinson 1989; Wang et al. 1996; Souza et al. 2000; Pielke 2001;
75 D'Almeida et al. 2007; Lawrence and Vandecar 2015). However, thermally driven circulations seem to be scale-
76 dependent and can weaken as deforestation areas increase (Avisar and Schmidt 1998; Patton et al. 2005). At
77 spatial scales larger than 100 km, land surface heterogeneities cannot produce enough surface gradient to sustain
78 thermal circulation (Dalu and Pielke 1993). In fact, a recent study in Rondonia-Brazil demonstrates how when
79 cleared areas grow, a shift occurs from a thermally convective regime towards a dynamically driven state where
80 the roughness length reduction dominates and favors (inhibits) convection over areas in the downwind (upwind)

81 direction of the deforested zone (Khanna et al. 2017). Since deforestation impacts in hydrological and atmospheric
82 processes are complex and scale dependent, regional and global climate models stand out as useful tools to better
83 understand the response of the climate system to land-use perturbations.

84
85 Numerical climate models have been extensively used for analyzing the impacts of deforestation over regional
86 and global climate (Nobre et al. 2016 and references therein). Mesoscale models are able to represent the observed
87 changes in circulation and clouds that follow small-scale deforestation (Chen and Avissar 1994; Wang et al. 2000;
88 Saad et al. 2010). Furthermore, the thermal-to-dynamical shift regime caused by the rainforest loss reported over
89 south-western Brazil is also reproduced by these models (Khanna et al. 2017). Under different deforestation
90 scenarios, models project more frequent dry events and a lengthening of the Amazonian dry season (Nobre et al.
91 1991; Bagley et al. 2014; Alves et al. 2017; Ruiz-Vásquez et al. 2020; Costa and Pires 2010). Although recent
92 observational studies report similar changes in Amazonian dry season hydrology since 1979, the attribution of
93 these alterations to deforestation is not straightforward and evidence suggest an additional influence of the tropical
94 north Atlantic Ocean warming (Fu and Li 2004; Yoon and Zeng 2010; Butt et al. 2011; Marengo et al. 2011; Fu
95 et al. 2013; Debortoli et al. 2017; Espinoza et al. 2019a; Leite-Filho et al. 2020). It is important to remark that,
96 taking into account the dependency of climate response to the size of deforestation, the current spatial scale of
97 Amazonian cleared area is not enough for predictions of extreme deforestation scenarios to be manifested or
98 detected (D’Almeida et al. 2007; Nobre et al. 2016). On the other hand, under extensive deforestation scenarios
99 climate models predict a reduction in precipitation recycling and rainfall in places downstream of the cleared areas
100 caused by the shut off of forest transpiration (D’Almeida et al. 2007; Moore et al. 2007; Correia et al. 2008; Swann
101 et al. 2015). At the same time, empirical estimates have found that south-western Amazon is particularly
102 dependent on evapotranspiration from other parts of the basin (Zemp et al. 2017a; Staal et al. 2018). Hence, several
103 simplified and regional climate models present a decreasing in rainfall and moisture transport over this particular
104 area (Sun and Barros 2015a; Boers et al. 2017; Zemp et al. 2017b; Ruiz-Vásquez et al. 2020). However, an
105 opposite effect of intensified moisture transport towards south-western Amazon and south-eastern South America
106 has been suggested by some modelling studies as the result of deforestation induced decreasing in surface
107 roughness (Lean and Rowntree 1997; Eiras-Barca et al. 2020). These different responses to similar deforestation
108 scenarios suggest the need to further assess the impacts of Amazon forest clearance over the regional climate.

109
110 In the Amazon-Andes transition region lies the rainiest and biologically richest area of the entire Amazon basin
111 (Myers et al. 2000; Killeen et al. 2007; Hoorn et al. 2010; Espinoza et al. 2015). This region encompasses complex
112 Andean topography, ranging between almost sea level in the Amazon lowlands to more than 6000 m a.s.l. over
113 the Andean summits. This causes a strong orographic effect that interacts with regional atmospheric circulation
114 to produce an exceptional spatial rainfall contrast (Chavez and Takahashi 2017; Junquas et al. 2018). There, high
115 precipitation rates lead to large values of runoff per unit area (Moquet et al. 2011), providing most of the sediment
116 load to Amazon rivers (Vauchel et al. 2017; Armijos et al. 2020). Consequently, extreme hydrological events over
117 this region can impact the whole basin downstream (Espinoza et al. 2014). Particularly, the Amazon-Andes
118 transition region exhibits the largest precipitation recycling (portion of the precipitated water previously transpired
119 by the Amazonian rainforest) of the Amazon catchment (Eltahir and Bras 1994; Dirmeyer and Brubaker 2007;
120 Van Der Ent et al. 2010; Staal et al. 2018). Therefore, this region can potentially be highly sensitive to Amazon

121 deforestation effects. Nevertheless, the surface-atmosphere coupling over this area has been poorly studied and
122 most of the referent work has been focused over Brazilian Amazon or was based on coarse horizontal resolution
123 modelling experiments insufficient to capture the required details of the topographical influence on local
124 convection.

125

126 This study aims to better understand the deforestation effects on moisture transport and rainfall over the Amazon-
127 Andes transition region during the wet season (December-January-February, DJF). During this season nearly 50%
128 of total annual precipitation falls over the region (Espinoza et al. 2009). Since Amazonian groundwater recharge
129 during wet periods can sustain forest transpiration rates during dry months (Miguez-Macho and Fan 2012), DJF
130 alterations in rainfall patterns may have a profound impact in the watershed hydrology. The strong summer
131 convection in the region is controlled by atmospheric processes that are part of the development of the South
132 American Monsoon System (SAMS; Vera et al. 2006; see section 3.1 for details) and which may be sensitive to
133 impacts on surface energy/water fluxes induced by land use change (Li and Fu 2004).

134

135 Taking into account the potential sensitivity of the Amazon-Andes transition region to deforestation and the need
136 for high-resolution studies, the present work analyzes the deforestation land-use change impacts on regional
137 energy and water budgets, as well as rainfall alterations at a local scale through the use of high-resolution
138 simulations (up to 1km horizontal grid size) with the regional climate model Weather Research and Forecasting
139 Model (WRF). Model hourly outputs are compared with observational data from rain-gauges and satellites in
140 order to elucidate the possible changes caused by Amazonian deforestation in the diurnal and nocturnal
141 precipitation inside a typical Andean valley during 2001-2011 austral summer. We describe the configuration of
142 the model, the deforestation experiment and reference data used in Sect. 2. In Sect. 3, we validate the model skill
143 to reproduce the precipitation, general circulation and moisture transport at regional scale, as well as its ability to
144 properly represent the deforestation signal in comparison with observations from the LBA project. The impacts
145 of deforestation on moisture transport and rainfall at regional, meso and local scales are analyzed in Sect. 4.
146 Finally, Sect. 5 summarizes and discusses the main results.

147

148 **2. Data and Methods**

149 **2.1 Model Description**

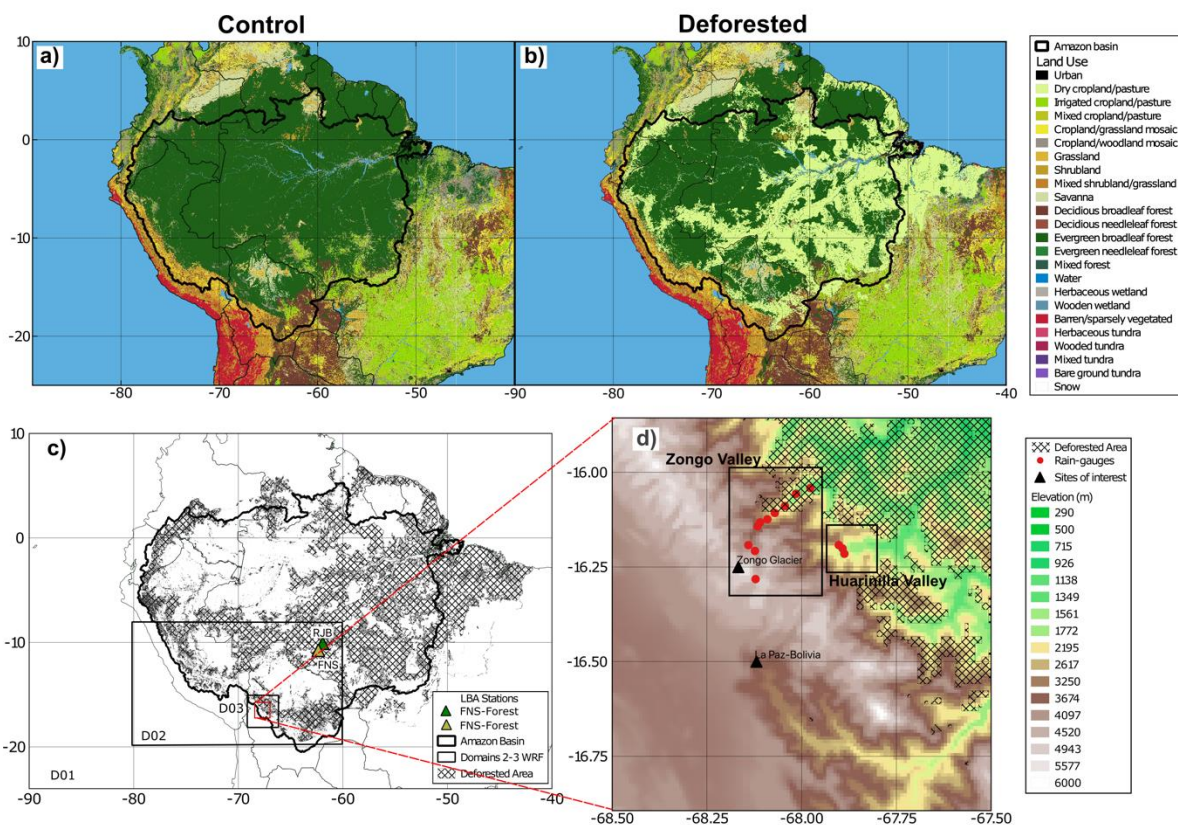
150 This study uses the regional climate model WRF version 4.1 (Skamarock et al. 2019). The model WRF was used
151 in previous works to study the climatology of tropical South America and the Amazon basin (Bagley et al. 2014;
152 Sun and Barros 2015a, b; Mourre et al 2016; Junquas et al 2018; Heredia et al. 2018; ; Spera et al. 2018; Trachte
153 2018; Yang and Dominguez 2019; Eiras-Barca et al. 2020; Saavedra et al. 2020). WRF is a nonhydrostatic model
154 with terrain-following vertical levels (we use 49 vertical levels with higher level density near the surface). We
155 employ three one-way nested modelling domains as shown in Figure 1c. Table 1 summarizes the information of
156 spatial and temporal resolution for each of the three domains used in this work. The first domain encompasses
157 tropical South America and has a horizontal resolution of 15 km x 15 km with a time step of 90 seconds. The
158 second and third domains have a spatial resolutions of 5 km x 5 km and 1 km x 1 km, respectively and cover the
159 south-western part of the basin. Time steps for these two domains are 30 seconds and 6 seconds, respectively.
160 Domain 3 comprises the instrumented Andean valleys of Zongo and Huarinilla used for model validation (Fig.

161 1d). In order to analyze the diurnal and nocturnal atmospheric conditions we use an output time resolution of 3
 162 hours for domains 1 and 2, and 1 hour for the finest domain.

163

164 Lateral boundary conditions from ERA5 reanalysis (Hersbach et al. 2020) are used. This reanalysis presents an
 165 enhanced horizontal resolution of $0.28^\circ \times 0.28^\circ$ (31 km) with data available every 6 hours
 166 (<https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5>). Significant improvements in the
 167 representation of several model processes in this reanalysis are reflected in a better fit in winds, temperature and
 168 humidity in the troposphere compared against radiosonde data (Hersbach et al. 2020).

169



170

171 **Fig 1** Land-use category distributions for: a) Control and b) Deforested scenarios. c) Deforested area and locations
 172 of selected flux towers from the Large-Scale Biosphere-Atmosphere Experiment in the Amazon. Black boxes
 173 show WRF domains. Delineated area represents the boundary of the Amazon River basin (black). d) D03
 174 subregion and locations of rain-gauges stations over two Andean mountain valleys

175

176 Preliminary tests for different model configurations (not shown) allow selecting the parameterizations shown in
 177 Table 2. Summarizing, we use the Yonsei University scheme for representation of the boundary layer (Hong et
 178 al. 2006) with the wind topographic correction activated (Jiménez and Dudhia 2012). This wind correction takes
 179 into account the momentum sink term generated by the unresolved terrain at sub-grid scale in the model. It has
 180 shown improvements in the simulated wind field over complex terrains (Jiménez and Dudhia 2012) and has
 181 already been tested in the tropical Andes (Junquas et al. 2018). Lin’s microphysics scheme is implemented (Lin
 182 et al. 1983) for phase change and precipitation. We also use Grell 3D parametrization for sub-grid scale convection
 183 over domains 1 and 2 (Grell and Dévényi 2002). Previous studies demonstrate convection parametrizations as one

184 of the major uncertainties in the representation of rainfall diurnal cycle (Brockhaus et al. 2008; Hohenegger et al.
 185 2008). Taking this into account and considering that a spatial resolution higher than 4 km is considered enough
 186 for non-hydrostatic models to explicitly solve convection (Weisman et al. 1997; Ramos da Silva and Avissar
 187 2006), we deactivate the cumulus parameterization over the finest domain.

188

189 In order to obtain a climatological state of the south-western Amazon wet season, our simulations cover 10 years
 190 of austral summers for the time period 2001-2011. Each simulation starts on November 1st and the first month is
 191 removed as a spin-up for the output analysis.

192

193 2.2 Description of the experiments

194 For each austral summer in the period 2001-2011, we conduct two simulations with WRF sharing the same
 195 atmospheric lateral boundary conditions, but with different land-cover states. The control simulation (“Control”
 196 or “Con”) corresponds to vegetation conditions for the year 2000 using the land cover map developed for South
 197 America by Eva et al. (2004) (Fig. 1a). This dataset was implemented in WRF by Saavedra et al. (2020) and
 198 exhibits a spatial resolution of 1 km x 1km. For the deforested simulation (“Deforested” or “Def”), we use the
 199 deforested scenario 2050 business-as-usual from Soares-Filho et al. (2006; Fig. 1b). This scenario reaches 45%
 200 of forest lost and assumes the continuation of the 1997-2002 registered deforestation rates, no definition of new
 201 protected areas in the basin and the building of all planned highways (Soares-Filho et al. 2006). This scenario
 202 does not include the forest impoverishment through logging and fires. The implemented deforestation polygon is
 203 shown in Fig. 1c. Amazon deforestation involves the replacement of native forest by mixed crops (mainly soybean
 204 and corn), as well as pasture for cattle feeding. For this reason, within the deforested polygon natural forest grid
 205 points are replaced by cropland/grassland mosaics.

206

207 **Table 1.** Characteristics of the domains used in WRF simulations.

| | WRF-D01 | WRF-D02 | WRF-D03 |
|----------------------------|-----------------------------|-----------------------------|------------------------------|
| Horizontal resolution (km) | 15 | 5 | 1 |
| Domain | Tropical South America | South-western Amazon | Bolivian Andean valleys |
| Domain center coordinates | 7° 30' 00" S, 65° 00' 00" W | 13° 56' 12" S, 70° 4' 49" W | 16° 35' 49" S, 67° 38' 56" W |
| Configuration | Regional simulation | One-way nesting | One-way nesting |
| Forcing | ERA5 | WRF-D01 | WRF-D02 |
| Vertical resolution | 49 hybrid vertical levels | 49 hybrid vertical levels | 49 hybrid vertical levels |
| Run time step (s) | 90 | 30 | 6 |
| Output time resolution (h) | 3 | 1 | 1 |

208

209 **Table 2.** Physical parameterizations used in WRF simulations.

| | Parameterizations | References |
|--|--------------------------|-------------------|
| | | |

| | | |
|--------------------------|---|--------------------------|
| Radiation | Longwave: rapid radiative transfer model (RRTM) | (Mlawer et al. 1997) |
| | Shortwave: Dudhia scheme | (Dudhia 1989) |
| Cumulus | Grell 3D | (Grell and Dévényi 2002) |
| Planetary boundary layer | Yonsei University scheme | (Hong et al. 2006) |
| Land surface | Noah land surface model | (Chen and Dudhia 2001) |
| Cloud microphysics | Lin scheme | (Lin et al. 1983) |
| Surface layer | MM5 similarity | (Paulson 1970) |

210

211 2.3 Data

212 In order to validate the model's ability to represent the rainfall spatial distribution in the region, we use the product
213 Tropical Rainfall Measuring Mission 3B42 (TRMM; <https://disc.gsfc.nasa.gov>). This dataset, available from
214 1998, presents a $0.25^\circ \times 0.25^\circ$ resolution and includes precipitation rates every 3 hours (Huffman et al. 2007).
215 However, previous studies have identified a strong rainfall underestimation by TRMM in the Amazon-Andes
216 region (Espinoza et al. 2015; Zubieta et al. 2015; Chavez and Takahashi 2017). For this reason, we complement
217 the model validation with daily information of 10 rain-gauge stations (Fig. 1d) from the Bolivian Electric Power
218 Company (COBEE) and managed in collaboration with the Institute of Research for the Development (IRD) -
219 France. Additionally, three stations with hourly data were used from Schwabe et al. (2011). Table 3 presents the
220 general information of each rain-gauge station. The quality assessment of these rainfall stations was carried out
221 by previous works (Molina-Carpio et al. 2019 and Laura Ticona, personal communication). The use of *in-situ*
222 information also allows characterizing the model skill to capture the precipitation variability at small spatial scale
223 in an Andean valley where local processes are important (Egger et al. 2005; Trachte et al. 2010; Junquas et al.
224 2018). Additionally, the representation of the regional atmospheric circulation and moisture transport are validated
225 by a direct comparison between WRF outputs and ERA5 data.

226

227 To test whether WRF responds realistically to the changes induced by land-use alterations, we use observations
228 of energy surface fluxes, air temperature, precipitation and wind speed from the LBA project (Saleska et al. 2013).
229 This dataset is available at the website https://daac.ornl.gov/LBA/guides/CD32_Brazil_Flux_Network.html. The
230 Large-Scale Biosphere-Atmosphere (LBA) project was a large scientific effort carried between 1996 and 2006
231 for assessing the influence of tropical forest conversion in carbon storage, nutrient dynamics and trace gas fluxes
232 in the Amazon. Since the region of interest of this study is the Amazon-Andes transition region, we only use the
233 closest stations from the LBA project that fits inside our second modelling domain (Fig. 1c). Because of their
234 geographical location, these stations experience annual cycles of precipitation and temperature similar to the
235 south-western Amazon variability. The first station, *Fazenda Nossa Senhora de Aparecida* (yellow triangle
236 labeled as "FNS" in Fig.1c), is a cattle ranch located at the center of a deforested area of about 50 km in radius
237 cleared in 1977 for crop cultivation. The tower flux stands at 306 m a.s.l., 10.75°S , 62.35°W and presents daily
238 data for the period 1999-2002. The second station corresponds to *Rondonia Jaru Biological Reserve* and is located
239 at 191 m a.s.l., 10.08°S , 61.93°W (green triangle labeled as "RJB" in Fig.1c). This is a natural reserve of the
240 Brazilian Environmental Protection Agency covered by tropical dry forest with a mean tree height of 33 m that
241 was established in the 1960s and expanded in 2006. The proximity of these stations (around 87 km) allows to test

242 the impacts of land use change on surface fluxes and the atmospheric bottom conditions. For validation we only
 243 use data for DJF 2001-2002 when WRF simulations and LBA data availability periods overlap in time.

244

245 **Table 3.** Name, geographic location, elevation, temporal range and annual rainfall of rain-gauge stations.

| Name of station | Location (latitude, longitude) | Elevation (m a.s.l.) | Temporal period (years) | Annual (DJF) rainfall in mm/year(mm/month) |
|-----------------|--------------------------------------|-------------------------|----------------------------|---|
| Plataforma | 16.28° S, 68.12° W | 4719 | 1971-2018 | 691 (138) |
| Tiquimani | 16.20° S, 68.12° W | 3997 | 1971-2018 | 918 (196) |
| Botijlaca | 16.19° S, 68.14° W | 3557 | 1971-2018 | 983 (178) |
| Cuticucho | 16.14° S, 68.11° W | 2790 | 1971-2018 | 2053 (341) |
| Santa Rosa | 16.13° S, 68.10° W | 2659 | 1971-2018 | 1748 (317) |
| Sainani | 16.12° S, 68.09° W | 2224 | 1971-2018 | 1975 (358) |
| Chururaqui | 16.10° S, 68.07° W | 1867 | 1971-2018 | 2086 (334) |
| Harca | 16.08° S, 68.04° W | 1546 | 1971-2018 | 2061 (358) |
| Cahua | 16.05° S, 68.01° W | 1271 | 1981-2018 | 2595 (405) |
| Huaji | 16.04° S, 67.97° W | 1043 | 1999-2018 | 2676 (454) |
| Hornuni 1800 | 16.19° S, 67.90° W | 1800 | 2001-2002 | 2310 (311) |
| Hornuni 2600 | 16.20° S, 67.89° W | 2600 | 2001-2002 | 3970 (489) |
| Hornuni 3010 | 16.21° S, 67.88° W | 3010 | 2001-2002 | 5150 (651) |

246

247 2.4 Calculation of energy/water balances and atmospheric stability

248 The energy budget at the surface is described by (Duveiller et al. 2018):

249

$$250 N_R = (S_{in} - S_{out}) + (L_{in} - L_{out}) = Sens_H + Lat_H + Gr_H \quad (1)$$

251

252 Here, N_R corresponds to net radiation and is the summation of contributions from shortwave and longwave
 253 components: S_{in} is the incident shortwave or solar radiation, S_{out} is the reflected solar radiation, L_{in} is the
 254 incoming longwave component and L_{out} is the emitted longwave radiation flux. The net turbulent heat flux is
 255 later distributed in sensible heat flux ($Sens_H$), latent heat flux associated with evaporation from the surface (Lat_H)
 256 and molecular conduction flux from the ground (Gr_H). All the terms in Eq. (1) are in W/m^2 . Land use change
 257 disturbances in the energy budget are assessed computing the differences in DJF climatological means for each
 258 term of Eq. (1) between Deforested and Control scenarios.

259
 260 The atmospheric water budget establishes how precipitation, evapotranspiration and net moisture transport
 261 entering/leaving an atmospheric column controls the temporal changes in the vertically integrated water content
 262 (Peixoto and Oort 1992). At seasonal timescales the temporal changes in atmospheric moisture are negligible and
 263 precipitation, evapotranspiration and column moisture convergence nearly balances each other (Brubaker et al.
 264 1993; Li et al. 2013). Hence, the climatological changes induced by deforestation in the water balance can be
 265 expressed as follows:

$$266 \quad \Delta \bar{P}_{Def-Con} = \Delta \bar{E}_{Def-Con} - \frac{1}{g} \overline{\Delta \nabla \cdot \int_{1000hPa}^{200hPa} q \vec{V}_H dp}_{Def-Con} \quad (2)$$

267
 268 In Eq. (2), P is precipitation, E is evapotranspiration and $\nabla \cdot \int_{1000hPa}^{200hPa} q \vec{V}_H dp$ is the divergence field of vertically
 269 integrated moisture flux from 1000 hPa to 200 hPa. The bars denote climatological DJF means.

270
 271 Finally, we evaluate the deforestation impacts on the atmospheric stability in tropical South America. Over the
 272 Amazon, the moistening of mid-tropospheric levels is associated with the development of deep convection in the
 273 region, which is related to the decreasing of atmospheric stability between middle and upper levels of the
 274 troposphere (Scala et al. 1990; Zhuang et al. 2017). For this reason, recent works use the difference of equivalent
 275 potential temperature (θ_e) between middle (500 hPa) and upper (200 hPa) troposphere ($\theta_{e_{200-500}}$) as an
 276 atmospheric stability index (Segura et al. 2020). The equation of equivalent potential temperature can be written
 277 following Stull and Ahrens (2000):

$$278 \quad \theta_e = \theta + \left(\frac{L_V \theta}{C_p T} \right) r \quad (3)$$

279
 280 where, θ corresponds to potential temperature, L_V is the latent heat of vaporization, C_p is the specific heat for air
 281 at constant pressure, T corresponds to the air absolute temperature and r is the air mixing ratio. The first term of
 282 this equation is related with dry adiabatic processes while the second one includes the effect of moisture in the net
 283 energy of an air parcel. Thus, the changes in the difference between mid and upper tropospheric equivalent
 284 potential temperature ($\theta_{e_{200-500}}$) allows separating the individual effects of dry and wet processes in the regional
 285 atmospheric stability. Again, climatological mean differences in each term of Eq. (3) between Deforested and
 286 Control scenarios allow evaluating the impact of deforestation in the regional atmospheric stability.

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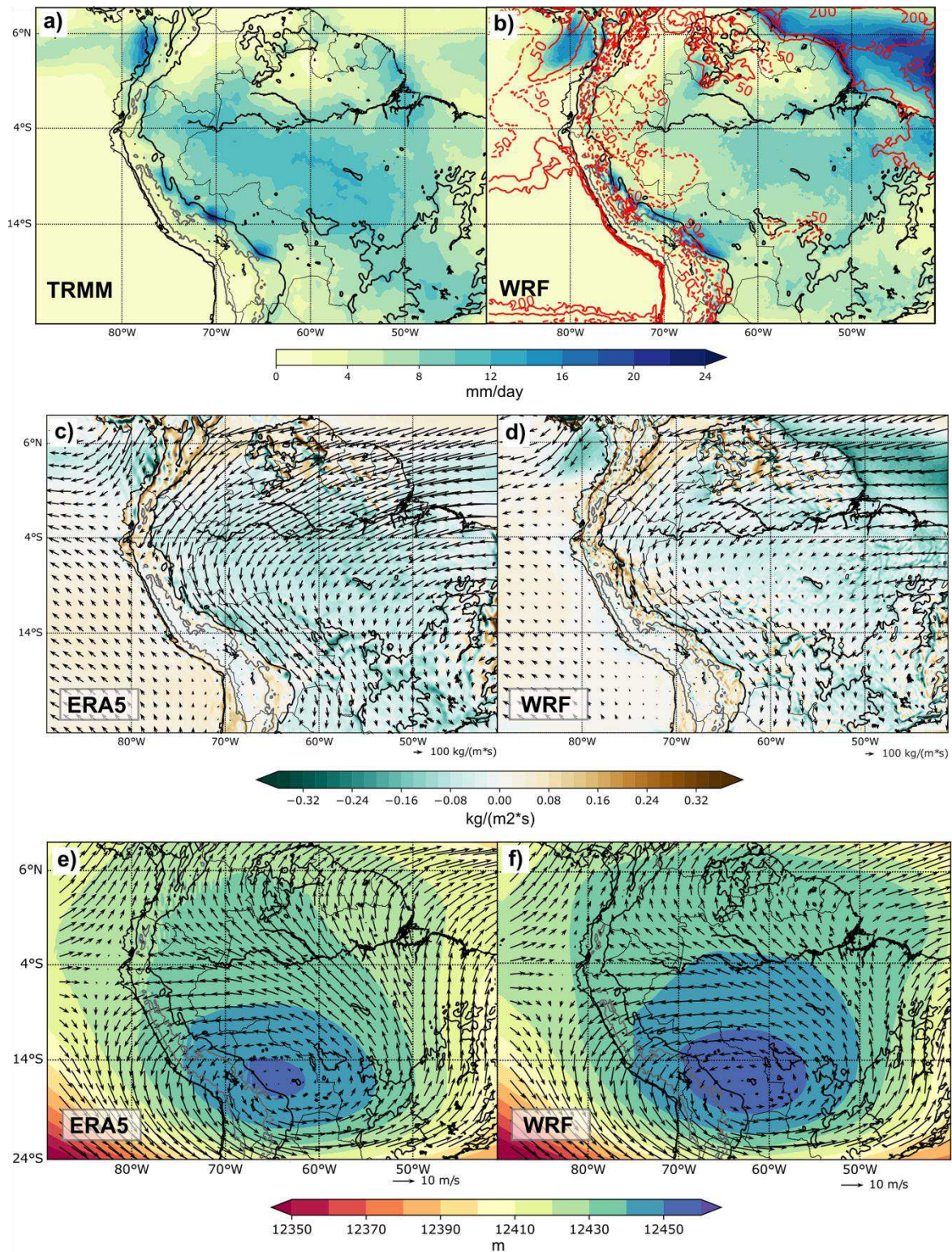
290 **3. Model validation**

291 3.1 Regional scale

292 The development of the SAMS establishes a northerly cross-equatorial flow (Wang and Fu 2002) and enhances
293 convection over central Amazon (Vera et al. 2006; Figs. 2a, c). A cloudiness/rainfall band known as the South
294 Atlantic Convergence Zone (SACZ) extends from southern Amazonia towards south-eastern Brazil and
295 accompanies the mature phase of the SAMS (Kodama 1992; Figueroa et al. 1995; Zhou and Lau 1998; Fig. 2a).
296 The diabatic heating released by the convective activity generates a large anticyclonic circulation at the upper
297 troposphere centered around 15°S - 65°W known as the Bolivian High (Lenters and Cook 1997; Fig. 2e). Latent
298 heat release over central Amazon also promotes a low-level continental-scale gyre that transports water vapor
299 from the Atlantic Ocean inland towards Amazon and La Plata basins (Rodwell and Hoskins 2001; Vera et al.
300 2006; Fig. 2c). Over the eastern flank of the Andes, this circulation reaches its maximum wind velocities forming
301 the South American Low-Level Jet (SALLJ). Over its trajectory, the transport of moisture through the SALLJ
302 encounters the topographical barrier of the Andes triggering forced convection in very localized places or rainy
303 hotspots mainly in the Bolivian and Peruvian piedmonts (Espinoza et al. 2009; Espinoza et al. 2015; Chavez and
304 Takahashi 2017; Fig. 2a).

305
306 In general terms, the model is able to reproduce the main precipitation features during the austral summer (Fig.
307 2b). WRF-D01 represents the strong precipitation over central Amazon but with its maximum shifted eastward.
308 The characteristic northwest-southeast rainfall band of the SACZ is present in WRF simulation. However, the
309 model shows a considerable negative bias over the northern Peruvian lowlands (until -50% of dry bias when
310 compared to TRMM; red contours in Fig. 2b). Rainfall Peruvian and Bolivian hotspots are also part of the model
311 climatology. Nevertheless, WRF underestimates the former and overestimates the latter. The Atlantic ITCZ is
312 strongly overestimated by WRF (up to +200% with respect to TRMM; red contours in Fig. 2b). This wet bias is
313 related to a too strong vertically integrated moisture convergence in the model output (Fig. 2d). Finally, upper and
314 lower atmospheric circulations are realistically represented by WRF (Fig. 2d, f). As in ERA5 outputs, WRF
315 easterly low-level winds carry moisture from the Atlantic Ocean toward the continent (Fig. 2d). The moisture
316 flow moves parallel to the Andes cordillera mechanically deviated exporting humidity toward southern latitudes
317 both for ERA5 and WRF. At the altitude of 200 hPa, the model develops the anticyclonic circulation of the
318 Bolivian High (Fig. 2f).

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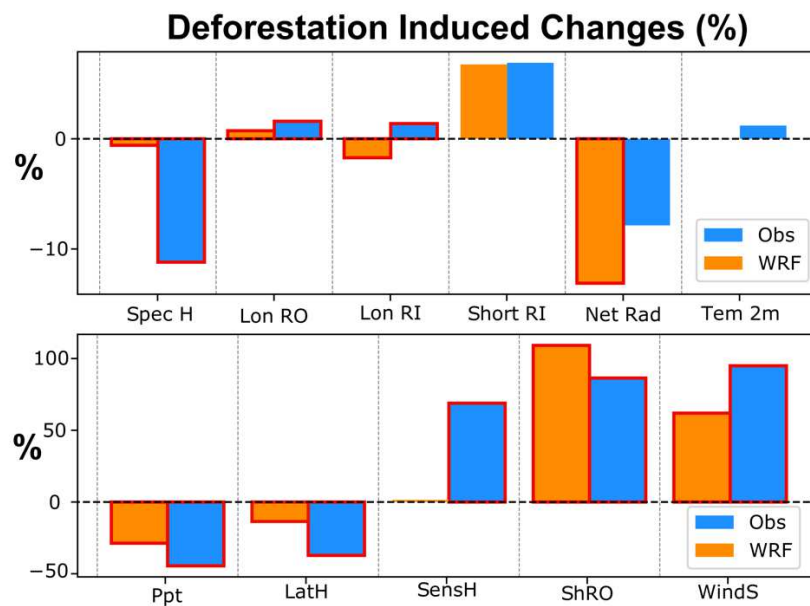
321 **Fig 2** DJF climatological means over the period 2001-2011 for: a) and b) precipitation from TRMM-3B42 (a) and
 322 WRF-Control (b) in mm/day. Red lines in b represent rainfall differences respect to TRMM in percentage. c) and
 323 d) vertically integrated moisture flux (vectors, $\text{kg}/\text{m}^2\cdot\text{s}$) and its divergence (shaded, $\text{kg}/\text{m}^2\cdot\text{s}$) from ERA5 (c) and
 324 WRF-Control (d). e) and f) horizontal winds at 200 hPa (vectors, m/s) and geopotential height (shaded, m) from
 325 ERA5 (e) and WRF-Control (f). Orography elevation is shown at 500 and 3500 m a.s.l. as black and grey contours,
 326 respectively

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3.2 Local scale

In order to validate if WRF reproduces the observed land use change impacts in surface fluxes and atmospheric variables at a local scale, we compare the selected LBA stations (section 2.3) against the WRF-D02 Control simulation for DJF 2001-2002. We take the paired pasture/rainforest sites from the LBA project and compute the observed changes in surface variables in percentage respect to the forest values. Then, similarly, we take the coordinates of the LBA pasture and rainforest stations and extract information of atmospheric variables from model outputs in order to compute the percentage changes. For WRF outputs, we average variables of interest around each location inside a surrounding area of 15 km x 15 km (one grid-point-radius) but selecting only the grids corresponding to cropland/grassland or forest respectively and ignoring grid cells with a different land use category. Between a 15km radius and a 50km radius, no significant changes are found in the WRF deforestation response (not shown).

Figure 3 shows the simulated and observed (orange and blue bars, respectively) mean differences of surface variables due to deforestation. Significant differences are highlighted in red contour according to a t-test ($p < 0.05$). In general terms, WRF is able to properly represent the nature of most of the alterations in surface fluxes and atmospheric variables induced by the land use change. Land cover change is related to reduced precipitation by around -44% and -30% in observations and WRF, respectively. However, since precipitation is also related with non-local processes, it is difficult to establish a deforestation-rainfall causality relation. Consistently, significant reductions in surface specific humidity are observed but weakly simulated (around -11% and -1% in observations and WRF, respectively). Accordingly, there is a significant decrease in latent heat flux for observations (-37%) with a weaker response by WRF (-14%). This implies that WRF underestimates the evapotranspiration response induced by deforestation. There is a significant increase in sensible heat flux and surface air temperature in LBA stations (+68% and +1%, respectively). WRF strongly underestimates the sensible heat increase and even slightly invert the temperature response to land use change (+1.49% and -0.02% respectively). The effects of the increased sensible heat and temperature (only sensible heat in WRF) act to give an increase in the outgoing longwave radiation flux in LBA stations of +1.6% (+0.7% in WRF). Since pastures and croplands are brighter than forest (Oguntoyinbo 1970; Bonan 2008), an increase in the albedo rises the outgoing or reflected shortwave radiation in +86% for observations (Loarie et al. 2011). The albedo effect is overestimated by WRF, which presents an outgoing shortwave radiation pasture/forest difference of +104%. This overestimated energy loss can be related with the model underestimation of sensible heat flux increase. More incoming shortwave radiation reaches the surface in model and observations (+4% and +7% respectively). This difference in WRF simulation is a consequence of cloud reduction over deforested areas (not shown). There is a disagreement between the modelled and observed changes in the incoming longwave radiation. While WRF exhibits a decrease in this variable (-1.64%), observations report an increase of about +1.38%. This difference is possibly explained by biases in the model representation of clouds or because of the drier WRF mid-upper troposphere (not shown). At the end, driven by the incremented energy losses through longwave radiation fluxes and shortwave radiation reflection, there is a reduction in net radiation for WRF (-15%) and LBA stations (-8%). Finally, the enhanced surface wind speed at 2 m is a direct consequence of the deforestation induced roughness-length reduction in both model (+53%) and observations (+95%).



368

369 **Fig 3** Deforestation induced changes in surface atmospheric variables over LBA stations (blue bars) and WRF-370 D02 (orange bars) for DJF in the period 2001-2002. Significant differences are delineated in red (t-test, $p < 0.05$).

371 Abbreviations are as follows: Spec H - specific humidity at 2 m, Lon RO – longwave outgoing radiation, Lon RI

372 – longwave incoming radiation, Short RO – shortwave outgoing radiation, Short RI – shortwave incoming

373 radiation, Net-Rad – net radiation, Tem2 2m – air temperature at 2 m, LatH – latent heat flux, SensH – sensible

374 heat flux, Ppt – precipitation, WindS – wind speed at 10 m

375

376 **4. Results and discussion**

377 This section is composed of two subsections, in the first part we evaluate the effects of deforestation in the energy

378 and water balances, as well as the atmospheric stability at regional-scale (90° - 60° W, 24° S- 9° N). The second

379 subsection is focused on the alterations in the hydro climatology at meso and local scales, especially over the

380 Amazon-Andes transition region and Andean valleys of Bolivia.

381

382 **4.1 Regional changes in energy and water balances**

383 Climatological alterations in the surface energy balance (Eq. (1)) induced by deforestation are shown in Figure 4,

384 in which significant differences are highlighted with black slashes. The last term in Eq. 1 corresponding to soil

385 heat flux is omitted because it exhibits much smaller magnitudes with respect to the other terms ($\pm 1 \text{ W/m}^2$).

386 Indeed, in situ measurements report low soil heat fluxes below tropical forest (Shuttleworth et al. 1984), but

387 important values for crops (Oliver et al. 1987; Gash and Shuttleworth 1991).

388

389 Important changes in the incoming solar radiation are induced by the forest loss in the region (Fig. 4a). In general,

390 more (less) incident solar radiation reaches the ground over places where reduction (rise) in precipitation and

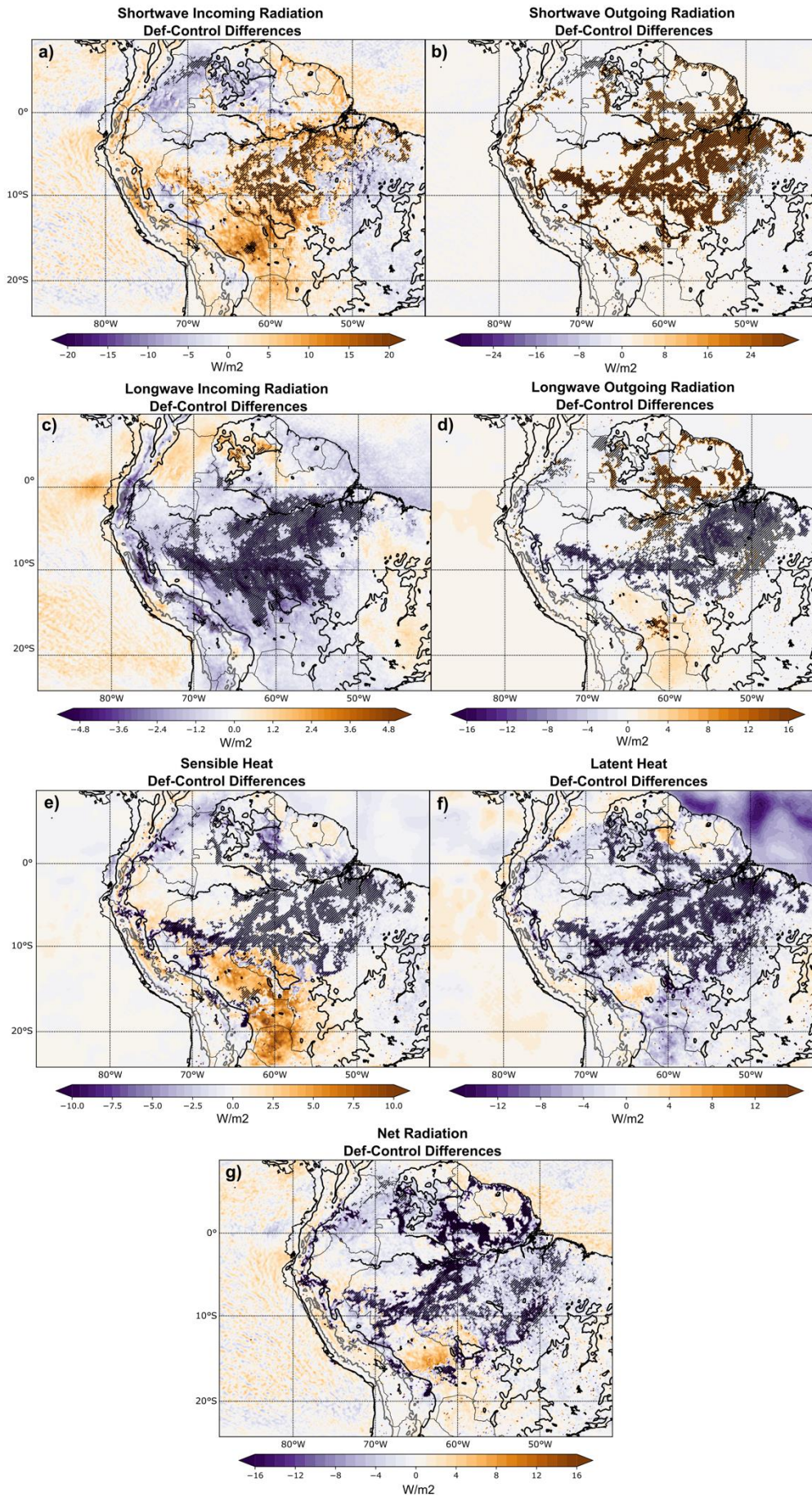
391 moisture convergence are located (see Fig. 5) such as in the central and south-western Amazon (north of South

392 America and eastern Brazil). Hence, disturbances in incident solar radiation are closely linked with changes in

393 clouds in WRF simulations as confirmed by analyzing top of the atmosphere (TOA) outgoing longwave flux and

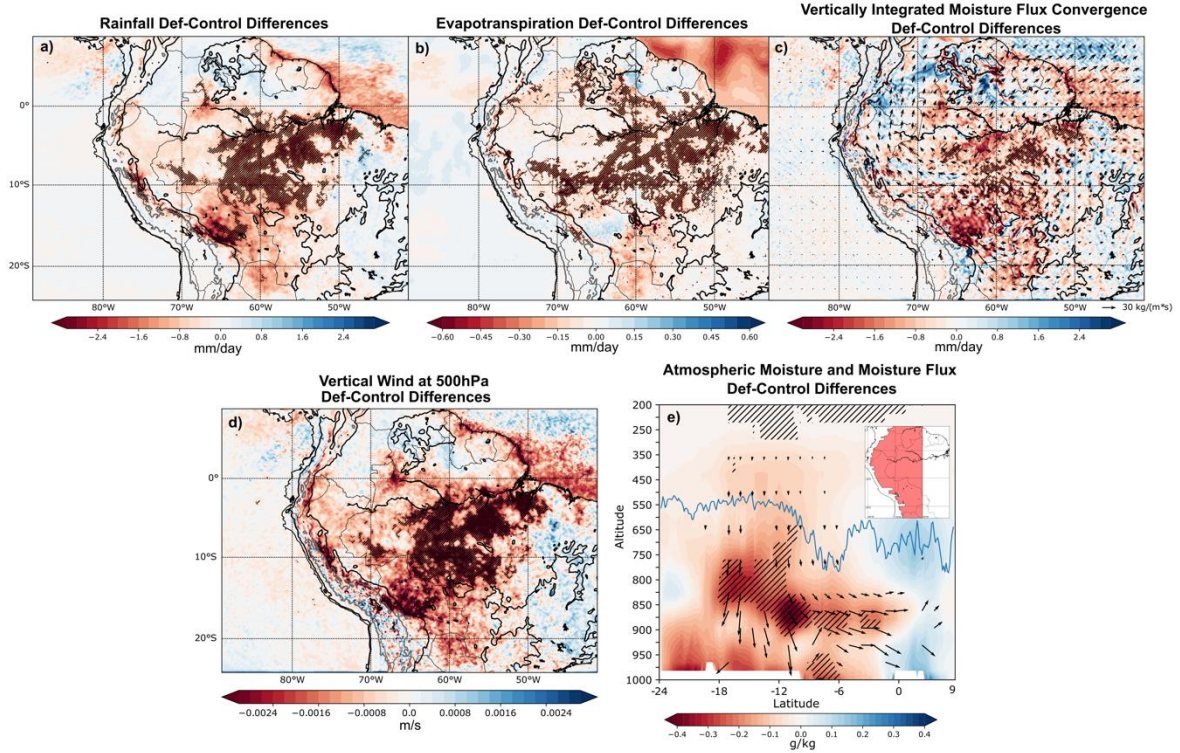
394 cloud water content (not shown). However, major changes in the energy balance occur in the outgoing shortwave

395 radiation (Fig. 4b). A significant increase in about $+20 \text{ W/m}^2$ occurs only over deforested areas which reflects the
396 very local response induced by the higher cropland/grassland albedo. With a minor response, reductions in the
397 longwave incoming radiation occur over most of the Amazonian watershed (Fig. 4c) as a result of a drier
398 atmosphere and lower cloud cover in the deforested scenario (not shown). On the other hand, there is a meridional
399 dipole in longwave outgoing radiation changes (Fig. 4d). Important disturbances in this variable take place mainly
400 over deforested areas which implies the influence of very local processes. North of the Amazon river ($<5^\circ\text{S}$) at
401 the East of the basin, upward longwave radiation increases as a consequence of an increase in surface emissivity
402 over this area (not shown). However, in the southern part of the basin, less outgoing longwave radiation is related
403 to reductions in sensible heat flux and air surface temperature. Sensible heat disturbances are linked to surface air
404 temperature changes (not shown). Thereby, reductions (increments) in this turbulent flux are located over
405 deforested areas (Fig. 4e) and are related with decreasing (increasing) surface temperature (not shown). Similarly,
406 latent heat flux decreases over regions with evapotranspiration reductions (Figs. 4b,5b). In the end, the enhanced
407 energy loss through reflected shortwave radiation as well as the reduced incoming longwave radiation flux
408 decrease the surface net radiation over the deforested and surrounding areas (Fig. 4g). This reduction in net
409 radiation is very important for driving alterations in the regional moisture convergence as pointed out by Zeng
410 and Neeling (1999). It is important to remark that the overestimation of the albedo effect and the weaker model
411 responses in sensible heat and longwave outgoing radiation fluxes, as well as roughness length and surface wind
412 velocity (see section 3.2) constrain our conclusions in the following sections.
413



415 **Fig 4** DJF seasonal climatological differences Deforested minus Control during 2001-2011 in surface energy
416 balance variables: a) Shortwave incoming radiation ($\Delta\overline{S_{inDef-Con}}$). b) Shortwave outgoing radiation
417 ($\Delta\overline{S_{outDef-Con}}$). c) Longwave incoming radiation ($\Delta\overline{L_{inDef-Con}}$). d) Longwave outgoing radiation
418 ($\Delta\overline{L_{outDef-Con}}$). e) Sensible heat flux ($\Delta\overline{Sens_HDef-Con}$). f) Latent heat flux ($\Delta\overline{Lat_HDef-Con}$). g) Net radiation
419 ($\Delta\overline{NRDef-Con}$). All figures are in W/m^2 . Significant changes (t-test, $p < 0.05$) are marked with black slashes.
420 Orography elevation is shown at 500 and 3500 m a.s.l. as black and grey contours, respectively
421

422 Figure 5 shows simulated changes in the water balance following Eq. (2). Significant reductions in the seasonal
423 rainfall are found over central and south-western Amazon closely matching the deforestation area (Fig. 5a). To
424 some extent, these precipitation changes are related to the local impact of the decreased evapotranspiration,
425 especially over the center of Brazilian Amazon (Fig. 5b). However, the drier conditions over Bolivian Amazon
426 are closely linked with changes in the regional atmospheric circulation, where increased moisture divergence is a
427 response to deforestation (Fig. 5c). The decrease in moisture convergence and rainfall over most of the Amazon
428 also implies a reduction in the vertical velocity (Fig. 5d), a rise in the atmospheric pressure at lower levels (not
429 shown), and a weakening of the regional Hadley cell (Fig. 5e). These changes are in agreement with the theoretical
430 frameworks laid out by Gill (1980) and Eltahir (1996), and with modelling studies (Zhang et al. 1996; Sun and
431 Barros 2015a; Badger and Dirmeyer 2016). Opposite to the forcing in Gill (1980), Amazon deforestation induces
432 a lack of diabatic heating as a result of a drier atmosphere and less convection in the region (Eltahir 1996; Badger
433 and Dirmeyer 2016). This feature is observed in Figure 5e as a drier troposphere below 650 hPa over $10^{\circ}S-18^{\circ}S$
434 and as an enhanced subsidence over the same latitudinal band. These changes are also associated with slightly
435 stronger westerlies at the upper troposphere and a small weakening of the Bolivian High (not shown). Similar
436 changes in the regional circulation and in the regional Hadley cell have been reported by previous modelling and
437 theoretical studies (Zhang et al. 1996; Eltahir 1996; Zeng and Neelin 1999; Badger and Dirmeyer 2016). In the
438 linear model developed by Eltahir (1996), the tropical subsidence is a consequence of a reduction in the boundary
439 layer entropy that follows large-scale deforestation. On the other hand, according to Zeng and Neelin (1999) and
440 in agreement with Figure 4, reductions in rainfall and moisture convergence over the Amazon basin induced by
441 tropical deforestation can be partially initiated by the enhanced energy loss due to the albedo effect.
442



443

444 **Fig 5** DJF seasonal climatological differences Deforested minus Control during 2001-2011 in water balance

445 variables: a) Rainfall ($\Delta\bar{P}_{Def-Con}$). b) Evapotranspiration ($\Delta\bar{E}_{Def-Con}$). c) Vertically integrated moisture flux

446 (vectors) and moisture flux convergence ($-\frac{1}{g}\Delta\nabla\cdot\int_{1000hPa}^{200hPa}q\vec{V}_H dp$; shaded). d) Vertical velocity at 500

447 hPa in m/s. e) Meridional moisture (shaded, in g/kg) and moisture flux (vectors in g/kg*m/s) zonally averaged

448 over the shaded region displayed at the upper right side. Only significant vectors are shown. Figures a-c are in

449 mm/day. Significant changes (t-test, $p<0.05$) are marked with black slashes. Orography elevation is shown in Figs.

450 5a-d at 500 and 3500 m a.s.l. as black and grey contours, respectively. The blue line in Fig. 5e represents the

451 Andes' profile calculated as the maximum topographic elevation inside the area displayed in red at the upper right

452 side

453

454 On the other hand, Figure 6 shows the differences between Deforested and Control scenarios for the DJF

455 climatology of the atmospheric stability index, as well as its adiabatic dry and diabatic moist process components

456 following Eq. (3). An increment in the equivalent potential temperature and the other terms of the equation (red

457 colors in Fig. 6) indicates an increase in the atmospheric stability and an inhibition tendency of the convective

458 activity. Conversely, a decrease in each term of Eq. (3) (blue colors in Fig. 6) implies a reduction trend in

459 atmospheric stability favoring convection development. A first evident result is that dry adiabatic processes

460 exhibit the tendency to reduce the atmospheric stability over the whole region (Fig. 6b). Although with

461 deforestation the troposphere cools at all levels (not shown), the upper levels exhibit a stronger cooling than

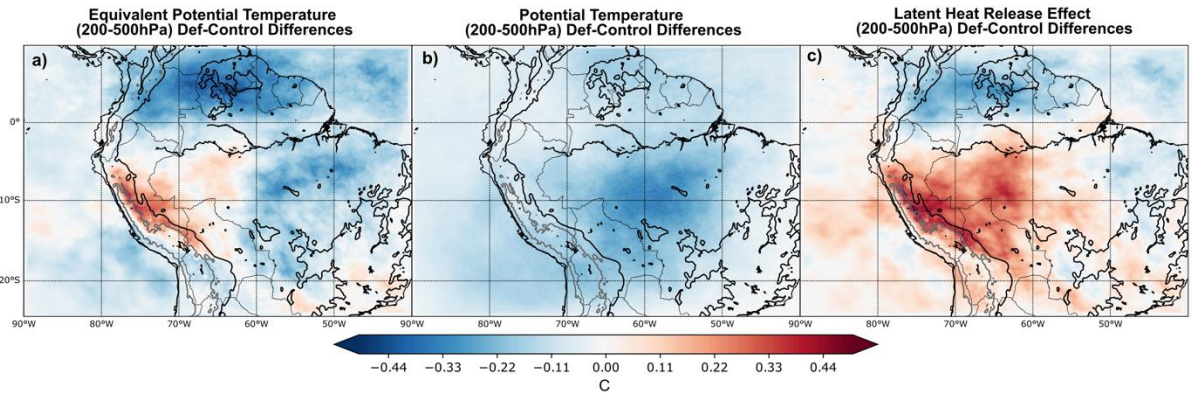
462 middle levels, generating the decrease in potential temperature shown in Figure 6b. This effect is more pronounced

463 over the center of the Brazilian Amazon, near the deforested area. Since the model response in sensible heat flux

464 and surface air temperature is strongly underestimated (section 3.2), the reduction in atmospheric stability due to

465 adiabatic processes could be even greater. On the contrary, changes in wet processes and latent heat release lead

466 to an increase in the atmospheric stability over the central and western Amazon, including the southern Andes and
 467 its eastern foothills (Fig. 6c). It is interesting to highlight that these areas have been recognized to be very sensitive
 468 to evapotranspiration upstream of the Amazon basin (Zemp et al. 2017; Staal et al. 2018). Over the northern and
 469 eastern parts of the continent, a decrease in atmospheric stability that is related to the enhancement in precipitation
 470 and moisture convergence (Fig. 5). Since WRF exhibits an underestimated response in evapotranspiration and
 471 latent heat flux to deforestation (section 3.2), the effects in atmospheric stability due to moist diabatic processes
 472 could be stronger. Similarly, earlier studies suggested two competing mechanisms as a response of tropical
 473 deforestation: (i) a dry atmosphere response linked to warmer surface that increases convergence; (ii) an enhanced
 474 divergence leaded by reductions in rainfall and latent heat release (Gill 1980; Lindzen and Nigam 1987; Eltahir
 475 and Bras 1993; Eltahir 1996). In summary, our results suggest that the combination of both dry adiabatic and
 476 moist (latent heat release) processes produces more stable atmospheric conditions and inhibits precipitation over
 477 the western part of the Amazonian basin and the eastern flank of the tropical Andes below the Equator (Fig. 6a).
 478 Colombia, Venezuela, Guyana and eastern Brazilian Amazon experience a net reduction in stability and more
 479 favorable conditions for convection in the deforested scenario. These alterations are more regional and not limited
 480 to deforested areas. Although changes in atmospheric stability are not statistically significant, the 10-year-period
 481 analyzed in this work might not be enough for identifying significant differences.
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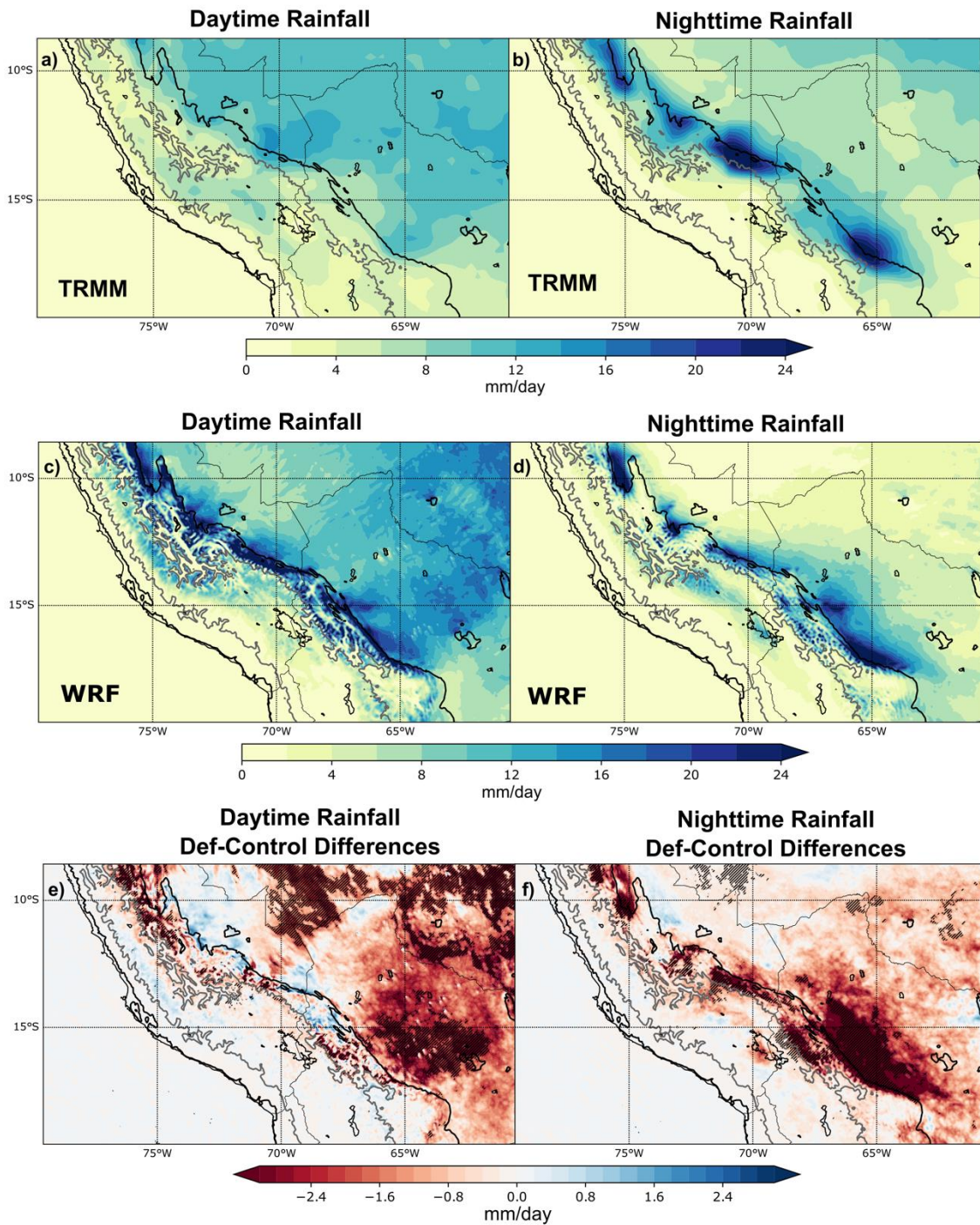


483
 484 **Fig 6** DJF seasonal climatological differences Deforested minus Control during 2001-2011 in atmospheric
 485 stability index $\theta_{e(200-500)}$: a) Equivalent potential temperature 200-500 hPa ($\Delta\overline{\theta}_{e200-500Def-Con}$). b) Potential
 486 temperature 200-500 hPa ($\Delta\overline{\theta}_{200-500Def-Con}$). c) Latent heat release effect 200-500 hPa ($\Delta\frac{Lv\theta}{CpT}r_{200-500Def-Con}$).
 487 All figures are in °C. Orography elevation is shown at 500 and 3500 m a.s.l. as black and grey contours,
 488 respectively
 489

490 4.2 Mesoscale and local changes in daily scale precipitation

491 The previous section displays how deforestation reduces precipitation and moisture transport, as well as increases
 492 the atmospheric stability and inhibits convective activity over the western Amazon and the eastern Andes flank.
 493 We analyze the deforestation effect at meso and local scales for daytime and nighttime conditions over the
 494 Amazon-Andes transition region, with a particular focus on the Andean valleys of Huarinilla and Zongo in Bolivia
 495 (see Figure 1 for precise location).
 496

497 Daytime and nighttime rainfall patterns are shown in Figure 7 for observations and WRF-D02. During the day,
498 TRMM exhibits high precipitation values over the western part of the Altiplano and Andes highlands (Fig. 7a).
499 Most tropical mountains display this daytime precipitation maximum over summits partially as a result of the
500 thermally driven upslope winds (Garreaud and Wallace 1997; Garreaud 1999; Kikuchi and Wang 2008; Zardi and
501 Whiteman 2013). There are also significant precipitation rates East of the Andes and over Amazonian lowlands
502 during the day as afternoon convection develops (Machado 2002; Romatschke and Houze 2010; Chavez and
503 Takahashi 2017; Fig. 7a). During the night, observed rainfall peaks are located over the eastern Andean piedmonts
504 associated with the so-called rainfall hotspots (Espinoza et al. 2015b; Killeen et al. 2007; Halladay et al. 2012;
505 Fig. 7b). This nocturnal convective activity is driven by the convergence of downslope winds from the top of the
506 Andes with the SALLJ nocturnal enhancement (Killeen et al. 2007; Romatschke and Houze 2010; Chavez and
507 Takahashi 2017; Junquas et al. 2018).
508



509

510 **Fig 7** DJF climatological means (2001-2011) for daytime and nighttime precipitation from: a-b) TRMM-3B42. c-
 511 d) WRF-D02. e-f) WRF-D02 Deforested-Control differences. All figures are in mm/day. Significant changes (t-
 512 test, $p < 0.05$) are marked with black slashes. Orography elevation is shown at 500 and 3500 m a.s.l. as black and
 513 grey contours, respectively

514

515 In general, WRF-D02 is able to represent the main features of diurnal and nocturnal rainfall variability in the
 516 region when compared to TRMM. Rainfall peaks above the 3500 m a.s.l. and over the central Peruvian Altiplano
 517 are simulated during the day by WRF as shown in Fig. 7c. Daytime convection over Amazon lowlands below 500
 518 m is also present in WRF outputs in agreement with TRMM. However, the model strongly overestimates daytime

519 precipitation over the Bolivian Amazon plains and over the eastern Andes slopes. Nevertheless, dry biases have
520 been identified by several studies in TRMM outputs over tropical Andes, which makes it difficult to properly
521 quantify the model biases (Condom et al. 2011; Scheel et al. 2011; Rasmussen et al. 2013; Espinoza et al. 2015).
522 During nighttime, WRF realistically represents the rainfall hotspots but exhibits a positive (negative) bias over
523 the Peruvian Andes highlands (Brazilian Amazon lowlands) (Fig. 7d).

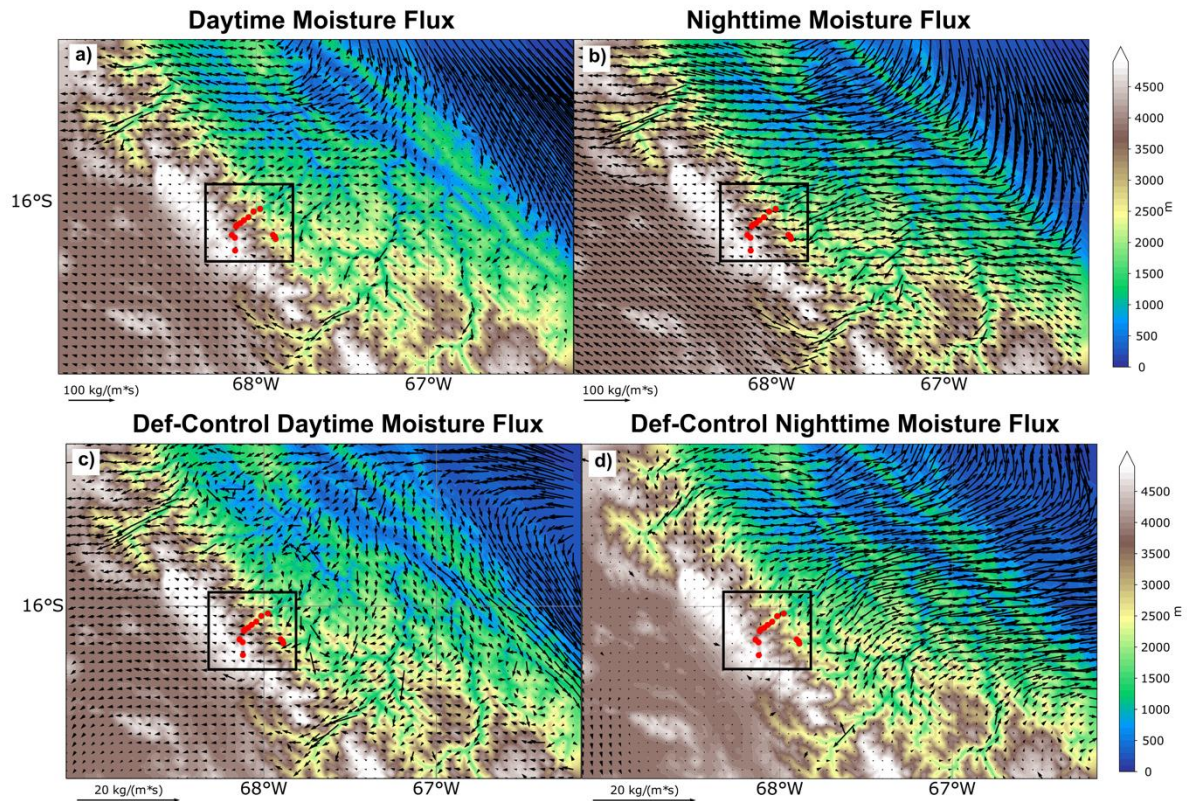
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525 Deforestation causes a general rainfall reduction in daytime and nighttime conditions but with different spatial
526 distributions (Figs. 7e-f). During the day, significant precipitation reductions occur over the Amazon plains
527 following the deforested area (between 10-20%), but no significant changes occurs over the tropical Andes above
528 3500m. On the eastern Andean slope between 500 m and 3500 m, significant changes are found, like in the central
529 Andean-Amazon valleys of Peru near the 10°S and 75°W (Fig. 7e). On the other hand, nighttime convection
530 decrease (between 20-30%) is concentrated over the eastern flank of the Andes and particularly over the Bolivian
531 piedmont (Fig. 7f). Taking into account that nighttime convection is responsible for nearly the 70%-80% of the
532 daily precipitation on the eastern tropical Andes slopes during austral summer (Perry et al. 2014; Chavez and
533 Takahashi 2017), this nocturnal rainfall reduction can have a profound impact in the hydrology of the region.
534 Since nocturnal precipitation over the eastern Andes slopes (including over rainfall hotspots) is closely related
535 to the presence of the SALLJ, in Figure 8 we evaluate the changes induced by deforestation in the vertically
536 integrated moisture flux over the Bolivian piedmont inside our third modelling domain (Figure 1). A northwesterly
537 moisture transport is present in both daytime and nighttime conditions during DJF in the Control scenario
538 associated with the SALLJ activity (Fig. 8; e.g. Marengo et al. 2004). This moisture transport is more parallel to
539 the Andes during the day and is stronger and more perpendicular to the slopes during the night in agreement with
540 previous works (Vernekar et al. 2003; Marengo et al. 2004; Junquas et al. 2018; Figs. 8a, b). The humidity flux
541 transported from Amazon toward the Andes is channelized both during day and night over the steepest valleys as
542 modelling and observational works reported in the region (Egger et al. 2005; Perry et al. 2013; Junquas et al. 2018;
543 Figs. 8a, b). The Andes act as a blocking barrier to this moisture laden winds coming from the Amazon lowlands,
544 triggering forced convection mainly during nighttime when the trajectory is more perpendicular to the topography.
545 This mechanism was hypothesized by Espinoza et al. (2015) based on *in-situ* stations and supported by modelling
546 studies (Junquas et al. 2018).

547

548 In the Deforestation scenario, the daytime upslope moisture flux is enhanced over the major valleys (Fig. 8c).
549 However, inside of our region of interest covering Zongo and Huarinilla valleys (black square in Fig. 8) this flux
550 exhibits little changes during the day. Conversely, during the night there is a strong reduction in the humidity
551 transport from the Amazon lowlands toward Andes highlands over the entire region (Fig. 8d) in agreement with
552 nocturnal rainfall reductions over eastern Andes slopes and Bolivian hotspot shown in Figure 7f. By suppressing
553 the condensation latent heating due to rainfall in the Peruvian and Bolivian Andean slopes in a modelling
554 experiment, Junquas et al. (2018) found a similar reduction in the entrance of the upslope moisture flux coming
555 from Amazon plains. A similar scope was assessed by Sun and Barros (2015a, b). Using high resolution climate
556 simulations, they prevent energy entry to the atmosphere from latent heat fluxes over the eastern flank of the
557 Andes in common weather conditions of austral summer and winter. They conclude that the energy available for
558 moist convection is strongly reduced, which is related to less moisture convergence, decreased rainfall and

559 cloudiness in the region. The upslope flow is also weakened as a consequence of increases in the static stability.
 560 Our results points towards the same direction: the crucial role of moist processes for sustaining the atmospheric
 561 circulation and hydrological condition over the Amazon-Andes transition region. We hypothesize that, as the
 562 atmosphere dries up in our Deforestation scenario and the northerly moisture flux weakens (Fig. 5), a decrease in
 563 precipitation starts over the wester Amazon. As rainfall decreases, less latent heat is released to the atmosphere
 564 over the region amplifying the precipitation reduction signal. Since latent heating is more important during the
 565 night (when sensible heating plays a minor role), major impacts on rainfall occurs.
 566



567
 568 **Fig 8** DJF climatological means (2001-2011) for daytime and nighttime vertically-integrated moisture flux from
 569 WRF-D03r: a-b) Control and c-d) Deforested minus Control differences. Terrain elevation (shaded) in m and rain-
 570 gauge stations (red dots). The black square shows the area where the Andean valleys of Huarinilla and Zongo are
 571 located
 572

573 In order to evaluate the potential impacts of Amazon forest loss in the hydroclimatology at local scales, our
 574 analysis is centered in two instrumented valleys located at the eastern flank of the Bolivian Andes: Huarinilla and
 575 Zongo Valleys (Fig. 9a). The spatial distribution of rain-gauges within the two valleys allows analyzing how
 576 precipitation varies with orography and altitude in two opposite directions with respect to the Andean Cordillera.
 577 Stations inside the Zongo Valley are located from north-east to south-west in a perpendicular direction to the
 578 Andes (Fig. 9a). Conversely, stations located in the Huarinilla Valley are disposed parallel to the mountain chain.
 579 Following the rain stations from highlands to lowlands, two cross-sections were defined as shown in black solid
 580 lines of Figure 9a. A third cross-section is analyzed in a perpendicular direction of the Zongo valley (northwest-
 581 southeast direction) in order to complement the understanding of local circulation alterations (black dashed line

582 in Fig. 9a). The climatological relationship between precipitation and elevation is shown in Figures 9b-c. There
583 is an opposite relation between rainfall and terrain elevation in the two transects. For Zongo cross-section,
584 precipitation generally decreases with elevation, reaching minimum values above 3000 m a.s.l. and rainfall peaks
585 near the 1000 m a.s.l. (Fig. 9b). This rainfall behavior was previously reported by Espinoza et al (2015). Chavez
586 and Takahashi (2017) demonstrated how the rainfall change with elevation along cross-sections transversal to the
587 Andes in the region follows the profile of the horizontal moisture flux carried by the SALLJ. In this sense, the
588 rainfall-elevation relation in the first transect seems to be more related with synoptic processes such as the SALLJ
589 and less with local upslope processes in accordance with previous studies (Romatschke and Houze 2010). An
590 opposite behavior is observed in Huarinilla Valley, where maximum rainfalls are located at high elevations around
591 3000 m a.s.l. (Fig. 9c). Here, daytime thermally-driven anabatic winds are responsible for transporting moisture
592 transpired inside the valley in the upslope direction and generates a rainfall peak in the afternoon (Molina-Carpio
593 et al. 2019).

594
595 WRF exhibits a realistic representation of the rainfall-elevation relationship for both cross-sections. The model
596 reproduces the observed rainfall increases over very local orographic elevations such as the rain-gauges of
597 Tiquimani, Cahua and Harca. However, a rainfall underestimation (around -30%) occurs in all the stations inside
598 the Zongo Valley and is stronger below the 3000 m a.s.l. Taking into account that precipitation is very related to
599 the SALLJ moisture transport in this valley, WRF rainfall underestimation could be related to the weaker SALLJ
600 moisture flux present in the model simulation (not shown). Daytime and nighttime rainfall conditions for both
601 transects are shown in Figures 9d-g according to WRF outputs. In Zongo Valley, diurnal precipitation is much
602 less than nocturnal values confirming that the dominant mechanism in this valley is the regional moisture transport
603 of the SALLJ (Fig. 9d-f). At this time, precipitation falls mainly over highlands for both transects as shown in
604 Figures 9d-e. Nighttime rainfall is concentrated mainly at lowlands in agreement with the mechanism described
605 before (Fig. 9-f). It is interesting to note that for Huarinilla Valley, similar precipitation magnitudes are simulated
606 for daytime and nighttime conditions (Fig. 9f-g). This suggests that local and synoptic mechanisms are equally
607 important in this valley as reported from previous works (Molina-Carpio et al. 2019).

608
609 Amazon deforestation induces significant reductions in the climatological rainfall mean but increases its
610 interannual variability over both Zongo and Huarinilla cross-sections (Fig. 9b-c). Major precipitation drop is
611 located over areas below 2000 m a.s.l. in the Zongo Valley (Fig. 9b). For this cross-section, daytime precipitation
612 presents little changes in its mean value with a slight reduction (increase) over highlands (lowlands). However,
613 daytime rainfall interannual variability increases in the deforestation scenario (Fig. 9d). Conversely, during the
614 night precipitation decreases, especially at lowlands, but no change in the variability occurs (Fig. 9f). This result
615 confirms our hypothesis that rainfall depletion in this valley is mainly related with the minor entry to SALLJ moist
616 winds from Amazon towards Andes highlands. On the other hand, Huarinilla cross-section does not present
617 significant changes in daytime rainfall (Fig. 9e). However, decreased rainfall and increased variability take place
618 during nighttime especially over highlands in agreement with the regional alterations in the SALLJ related
619 moisture flux (Fig. 9g).

620

621 Aiming to complement the previous analysis, Figure 10 displays the atmospheric humidity and moisture flux in
622 the vertical cross-section along parallel (black solid line in Fig. 9a) and perpendicular (black dashed line in Fig.
623 9a) directions of the Zongo Valley. Similar results are found for Huarinilla Valley (not shown). During daytime,
624 upslope moisture flux (Fig. 10a) is causes the rainfall peaks over summits described in Figure 9d. It is interesting
625 to note that the strong ascent around Botijlaca station seems to be related to the maximum in precipitation around
626 this area (Fig. 9d). Similar anabatic winds are present in the perpendicular direction of the Zongo Valley (Fig.
627 10c). During nighttime, surface moisture flux weakens, and light katabatic winds develops in the Zongo Valley
628 (Fig. 10b-d). At the same time, the mid-tropospheric moisture coming from the Amazon plains is strengthened
629 mainly in the perpendicular direction of the Zongo Valley (Figs. 8a-b, 10b-d). Another interesting feature of the
630 nocturnal circulation in the perpendicular direction is the development of a moisture flux around 700hPa in the
631 opposite direction (northwest-southeast) of the main mid-tropospheric flow (Fig. 10d). This returning flow seems
632 to be a consequence of the ascension motion over lowlands summits to satisfy continuity.

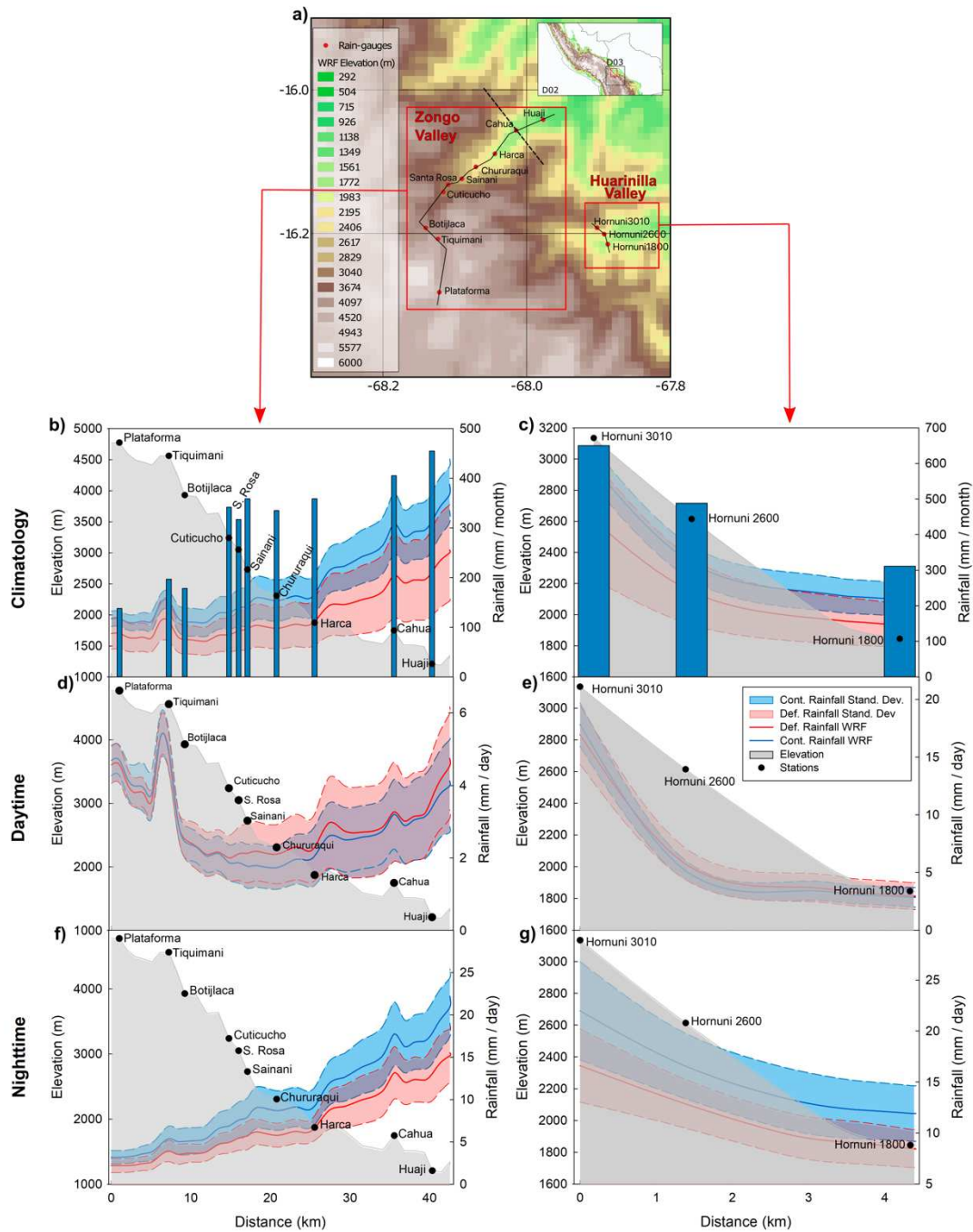
633

634 Deforestation induces a weakening of the daytime upslope flow over both directions of the Zongo Valley (Figs.
635 10e-g). A reduction in the intensity of the daytime circulation increases the atmospheric moisture near the surface
636 at the bottom of the valley and is related to the slight rainfall increase (decrease) over lowlands (highlands)
637 described before (Fig. 9d). Based on the thermodynamic of heat engines, Souza et al. (2000) explain the physical
638 mechanism behind the observed circulation over areas with different surface conditions in sloping terrains.
639 According to this work, the intensity of the circulation depends on the absorption of sensible and latent heat by a
640 parcel that moves along a closed streamline. They also recognize the hydrostatic pressure and nonadiabatic
641 expansion/compression as important effects. Following this argument, we assess the surface energy balance along
642 the Zongo Valley for the Control and Deforestation scenarios. We focus on daytime conditions, when local surface
643 heterogeneities play an important role (Fig. 11). However, nighttime changes were also analyzed and are not
644 significant (not shown). Small changes in the incoming shortwave radiation are induced by deforestations in the
645 Zongo Valley and are linked with differences in clouds (Fig. 11a). On the contrary, a strong increment in the
646 reflected shortwave radiation causes a lost of around -40W/m^2 at the bottom of the valley where deforestation
647 occurs (Fig. 11b). The longwave components of the energy balance remain almost unaltered (Figs. 11c-d). Due to
648 the increase in the outgoing shortwave energy loss, the surface net radiation decreases at lowlands below
649 Chururaqui rain-gauge (Fig. 11e). This net radiation reduction generates a decrease in the altitude of the planetary
650 boundary layer (PBL) (Fig.11f) and is distributed in reductions in latent heat but mainly in sensible heat flux over
651 lowlands (Fig. 11g-h). These fluxes are essential in the triggering and development of upslope flows (e.g., Junquas
652 et al 2018). The reductions in the sensible and latent heat fluxes explain the weakening of the upslope surface
653 circulation during daytime described in Figure 10.

654

655 On the other hand, nighttime differences in circulation mainly occur at regional scale through mid-tropospheric
656 moisture flux reduction in the deforested scenario (Figs. 8d, 10f-h). This moisture transport is very important for
657 developing the nocturnal convection and its decrease causes less rainfall over the entire region (Fig. 9f-g). The
658 nighttime conditions are also characterized by a strong reduction in atmospheric humidity above the 600hPa as a
659 consequence of deforestation (Fig. 10f-h).

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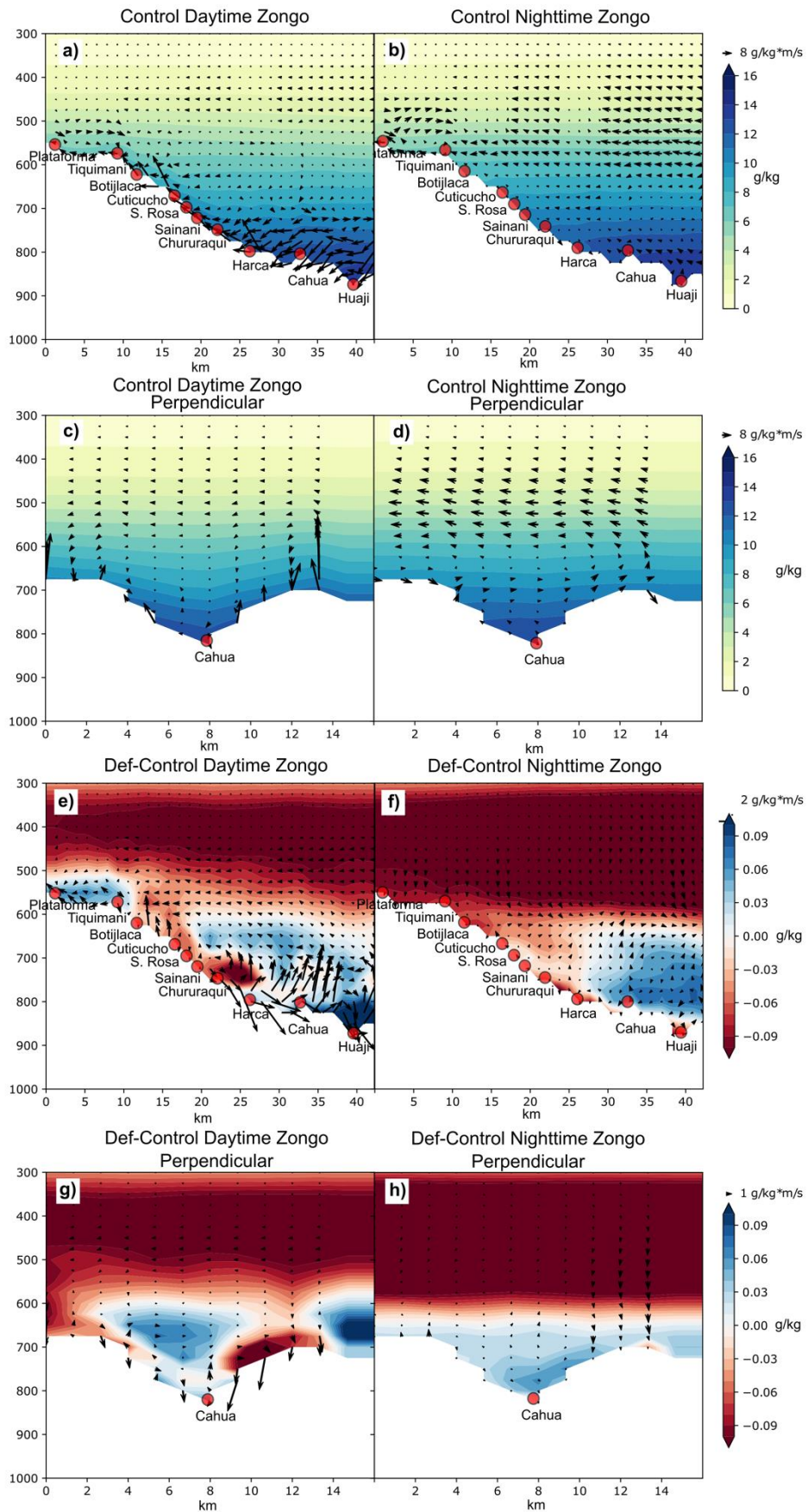


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Fig 9 a) Map of the subregion inside D03 with terrain elevation (shaded) in m and rain-gauge stations (red dots). Two selected cross-sections are indicated in black solid lines. A perpendicular cross-section in Zongo Valley shown in Figure 10 is indicated as a dashed black line. b-c) Cross-sections for b) Zongo Valley and c) Huarinilla Valley. Topographical profiles are displayed as a gray line (in m). Observed rain-gauge rainfall is presented as blue bars. WRF-D03 Control and Deforested rainfall climatological means are shown as blue and red lines and are labeled as Cont. and Def., respectively. Rainfall is in mm/month. d-g) Same as Fig. 9b-c but for daytime and nighttime conditions. Rainfall is in mm/day. The x axis represents the horizontal distance in km along the cross-sections. Blue and red envelopes represent model standard deviation for each scenario. Black dots represent rain-gauge stations

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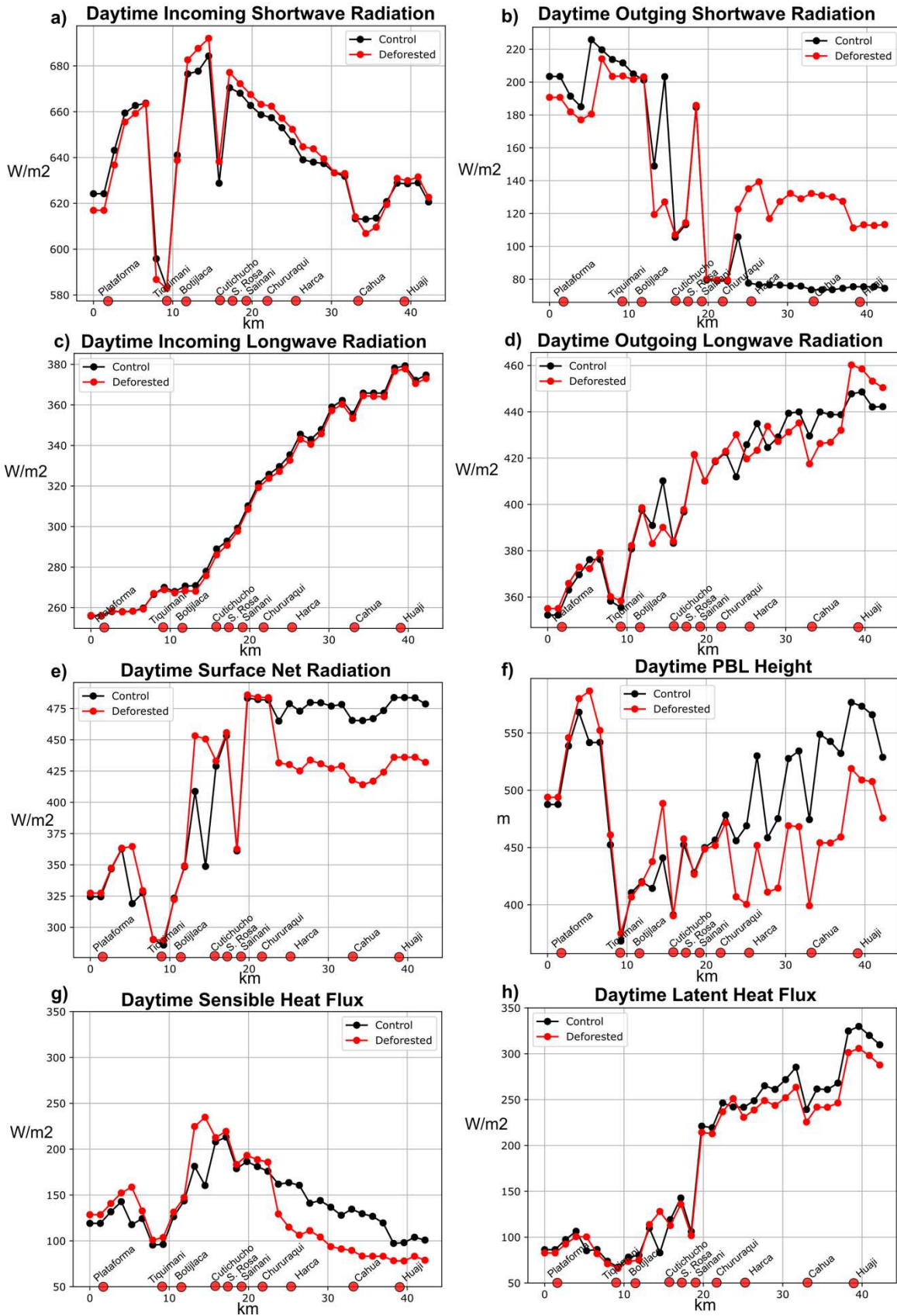
Rainfall reductions and increased rainfall variability over Huarinilla and Zongo Valleys can impact the wet ecosystems of local humid-montane forest and its endemic biological species (Ibisch et al. 2001; Mueller et al. 2002; Kessler and Kluge 2008). Additionally, the area encompasses part of the Cordillera Real-Bolivia and the Huayna Potosi Massif where several glaciers (e.g. Zongo Glacier) are located (Francou et al. 1995). The surface glacier mass balance variability of the tropical glaciers is strongly controlled by the precipitation amount and temporal distribution that determine the glacier surface albedo and therefore the surface energy balance (e.g. Sicart et al. 2005; Rabatel et al. 2013), as well as regional atmospheric circulation (Favier 2004; Sicart et al. 2016). Our results state that Amazonian large-scale deforestation can have an important effect in the regional circulation and precipitation. This means that retreat of the tropical Andes glacier documented since the early 1980s and linked to climate change (e.g. Ramírez et al. 2001; Francou et al. 2003; Rabatel et al. 2013) is likely to be further increased as extensive Amazonian deforestation can also have an impact on precipitation and therefore on the surface mass balance of tropical glaciers.



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Fig 10 Vertical cross-section for daytime and nighttime DJF climatological means (2001-2011) in atmospheric moisture (shaded, g/kg) and moisture flux (vectors, g/kg*m/s) for the Zongo Valley in two opposite directions:

689 parallel to the transect (see black solid line in Fig. 9a) and perpendicular to the transect (see black dashed line in
690 Fig. 9a). Daytime (a) and nighttime (b) WRF-Control conditions for the transect parallel to the Zongo Valley.
691 Daytime (c) and nighttime (d) WRF-Control conditions for the transect perpendicular to the Zongo Valley.
692 Daytime (e) and nighttime (f) Deforested minus Control differences in the transect parallel to the Zongo Valley.
693 Daytime (g) and nighttime (h) Deforested minus Control differences in the transect perpendicular to the Zongo
694 Valley. Red dots represent rain-gauge stations. Vertical component of the moisture flux is exaggerated by a factor
695 of ten
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Fig 11 WRF-D03 daytime mean variation along the Zongo Valley transect in Control and Deforested scenarios (black and red lines, respectively) for: a) Shortwave incoming radiation. b) Shortwave outgoing radiation. c)

700 Longwave incoming radiation. d) Longwave outgoing radiation. e) Surface net radiation. f) Planetary boundary
701 layer (PBL) height. g) Sensible heat flux. h) Latent heat flux. Red dots represent rain-gauge stations

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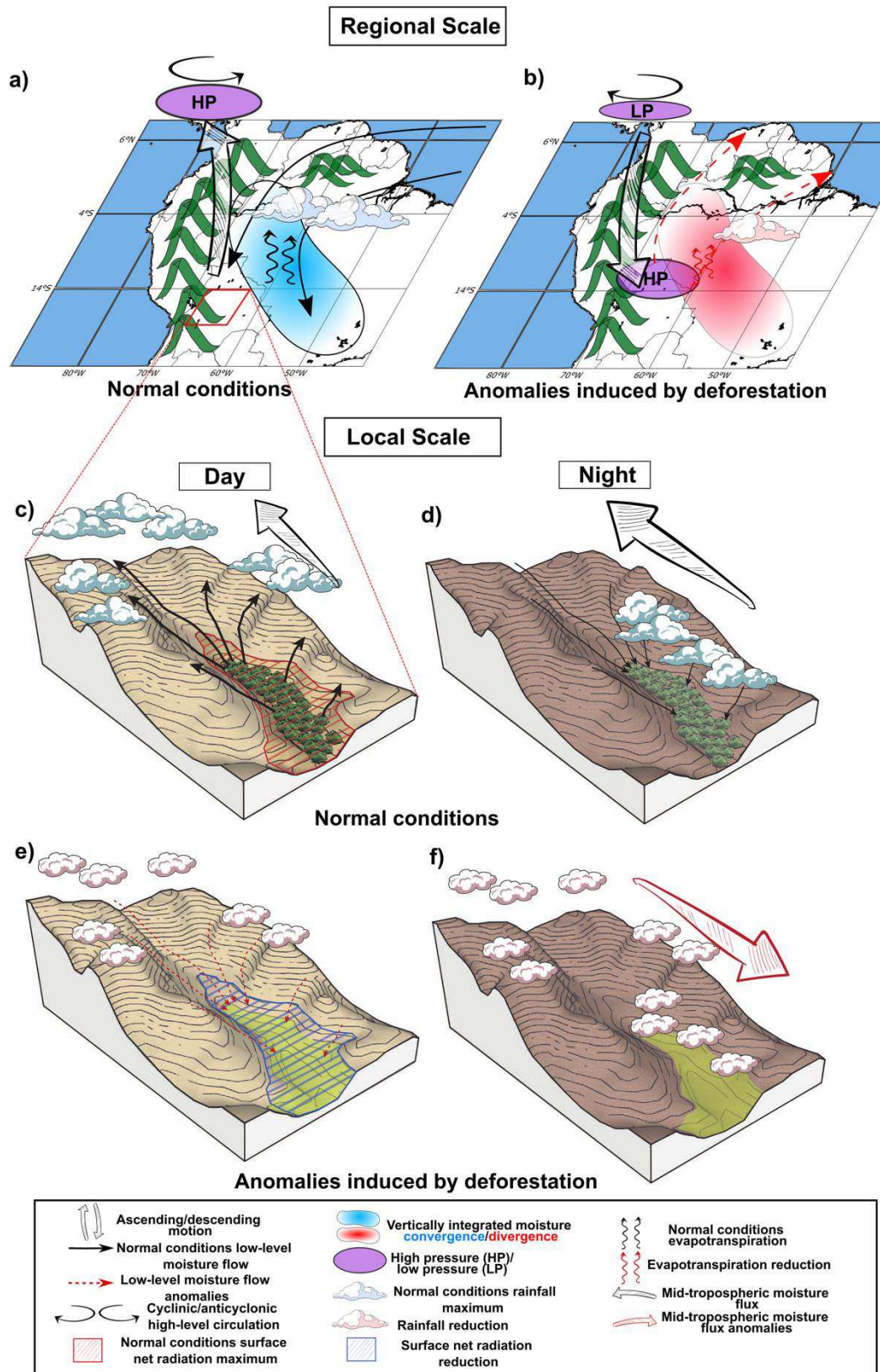
703 It is important to remark that spread deforestation can have strong nonlocal atmospheric responses (Avisar and
704 Werth 2005; Hasler et al. 2009). Thus, the use of a regional model where both control and deforested scenarios
705 present the same lateral boundary conditions generates a bias where these nonlocal effects are not included
706 (Medvigy et al. 2011). In this sense, the results discussed above do not include the possible remote responses of
707 deforestation in the boundary conditions of our analysis domain.

708

709 **5. Summary and conclusions**

710 This study aims to evaluate the impacts of Amazonian deforestation in the hydro-climatic connectivity between
711 Amazon and eastern tropical Andes during the austral summer in terms of hydrological and energetic balances.
712 To this end, we use 10-year December-January-February high-resolution simulations (2001-2011) with the
713 regional climate model WRF. Two scenarios are used: (i) a control simulation corresponding to vegetation of
714 conditions for the year 2000; (ii) a business-as-usual deforested scenario (45% of forest lost) designed by Soares-
715 Filho et al. (2006).

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Fig 12 Schematic representation of DJF climatological atmospheric conditions at regional (a-b) and local scales (c-f) for normal conditions (a-c-d) and deforestation induced anomalies (b-e-f) in WRF simulations. The red square in a indicates the approximated location of the Zongo Valley sketched in c-f. Black solid (red dashed) thin arrows represent surface moisture transport under normal conditions (deforestation anomalies). Black (red) thick

722 arrows represent mid-tropospheric moisture flux under normal conditions (deforestation anomalies). Vertical
723 thick arrows represent ascending and descending motions. Black (red) curved arrows represent evapotranspiration
724 under normal conditions (deforestation anomalies). Blue and red shaded areas indicate vertically integrated
725 moisture convergence and divergence, respectively. Purple areas indicate pressure changes. Blue and red clouds
726 indicate rainfall maximum and rainfall reductions under normal and deforested conditions, respectively. Red
727 (blue) hatched area in c (e) indicates surface net radiation maximum (reduction) under normal (deforestation)
728 conditions

729

730 By comparing with reanalysis (ERA5) and precipitation satellite outputs (TRMM), we find a realistic
731 representation of the main atmospheric circulation (e.g. SALLJ, the Bolivian High) and rainfall features (e.g.
732 SACZ, Amazon convection) during austral summer by WRF at a regional scale. The model is also able to
733 reproduce the daytime and nighttime rainfall conditions over the Amazon-Andes transition region and the rainfall-
734 elevation relation at a local-scales inside instrumented Andean valleys. However, at local scale, in regions with
735 high orography, day and night rainfall conditions were only compared with rain-gauge estimations for a very short
736 period (one year) in Huarinilla valley. The lack of observations at intra-diurnal time scale in the region is an
737 important limitation for assessing the high-resolution model representation of local atmospheric mechanisms and
738 its interaction with mesoscale features. We complement the validation by assessing the model skill to reproduce
739 the observed land use change impacts in surface fluxes and atmospheric variables at a local scale against LBA
740 stations. We find a realistic model response to deforestation for most of the surface atmospheric variables in
741 Amazon lowlands. However, WRF overestimates the effect of the albedo on the reflected shortwave radiation and
742 strongly underestimates the sensible heat flux and evaporation/latent heat response to forest loss. These model
743 biases constrain our conclusions about the impacts of deforestation on the regional climate. How these errors
744 interact with each other is something that remains unexplored in this work.

745

746 At a regional scale, the energy balance is profoundly disrupted by the forest loss. The combined effect of enhanced
747 energy losses through shortwave radiation reflection, longwave incoming and outgoing radiation fluxes leads to
748 a decrease in the surface net radiation over most of the basin. According to the deforestation theory developed by
749 Zeng and Neeling (1998), this reduction in surface net radiation may lead to alterations in the regional moisture
750 convergence. Accordingly, deforestation reduces precipitation almost over the entire Amazonian basin (~20%).
751 Rainfall depletion is related with the very local response of decreased evapotranspiration over cleared areas. The
752 drier atmosphere seems to present a decrease in latent heat release that weakens moisture convergence and causes
753 an anomalous winds subsidence in the region in agreement with early theoretical frameworks. In agreement with
754 these theoretical works, deforestation induces a slight weakening of the Bolivian High and decreases (increases)
755 the 200hPa geopotential height (surface pressure) over the southwestern Amazon. These alterations in the regional
756 precipitation and atmospheric circulation are summarized in Figures 12a-b. Regarding the regional atmospheric
757 stability, our analysis suggests the rise of two competing mechanisms for tropical deforestation: (i) an adiabatic
758 response decreases atmospheric stability and enhances convective activity over the continental area; (ii)
759 conversely, reductions in the latent heat release increase atmospheric stability and disfavor convection mainly
760 over western Amazon, and the eastern sides of tropical Andes. Although the effects of deforestation in atmospheric

761 stability are relatively small and non-significant, the underestimated WRF response in sensible and latent heat
762 fluxes and surface air temperature to forest loss implies that these effects could be stronger.

763

764 Over the Amazon-Andes transition region, deforestation causes a general rainfall reduction in both daytime and
765 nighttime conditions but with a clearly spatial differentiation. Daytime precipitation reductions occur over the
766 Amazon plains following the deforested area (between 10-20%). Conversely, nighttime convection decreases
767 between 20-30% mainly over the eastern flank of the Andes and over the Bolivian piedmont. Since nighttime
768 convection is responsible for nearly the 70%-80% of the daily precipitation on the eastern tropical Andes slopes
769 during austral summer, this rainfall reduction can seriously impact the hydrology of the region. The decreasing in
770 nocturnal rainfall over eastern Andean hillside is caused by a reduction in the humidity transport from the Amazon
771 lowlands toward Andes highlands during nighttime in agreement with previous modelling studies.

772

773 Finally, at a local scale, Amazon deforestation induces significant reductions in the climatological rainfall mean
774 over both Zongo and Huarinilla valleys (-25%) and increases its interannual variability. For the Zongo Valley,
775 major precipitation drop occurs over areas below 2000 m a.s.l. In this cross-section, daytime precipitation presents
776 little changes in its mean value, with a small increase (decrease) at lowlands (highlands) as a consequence of the
777 weakening of the thermally-driven anabatic circulation (Figs. 12c-e). However, during the night precipitation
778 decreases, especially at lowlands. The rainfall depletion in this valley seems to be mainly related with the minor
779 entry to SALLJ moist winds from Amazon towards Andes highlands (Figs. 12d-f). Similarly, Huarinilla cross-
780 section does not present significant changes in daytime rainfall but exhibits decreased precipitation especially
781 over highlands. However, as a result of numerical modelling these conclusions have to be taken with caution as
782 they are partially dependent on the model configuration. More efforts have to be done to understand the sensitivity
783 and robustness of these results to model parametrizations. Finally, this work and most of deforestation modelling
784 studies are focused on forest loss impacts on atmospheric processes at climatological time scales. How
785 deforestation can alter rainfall distribution and the atmospheric circulation patterns that drives moisture advection
786 at synoptic intra-seasonal time scale is a question that should be addressed to better understand the deforestation
787 disturbances at longer time scales.

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791

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797

798 **Declarations**

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803 Marcus Schawe for providing the rain-gauge data for Hornuni Valley.

804 **Conflicts of interest/Competing interests:** The authors declare no conflicts of interest or competing interests.

805 **Availability of data and material:** All dataset sources except WRF modelling results have been properly
806 referenced by showing the source (web site links). The datasets generated during the current study (WRF outputs)
807 are available from the corresponding author on reasonable request

808

809 **Author contributions:** All authors contributed to the study conception and design. Material preparation, data
810 collection and analysis were performed by Juan P. Sierra. The first draft of the manuscript was written by Juan P.
811 Sierra and all authors commented on previous versions of the manuscript. All authors read and approved the final
812 manuscript.

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Figures

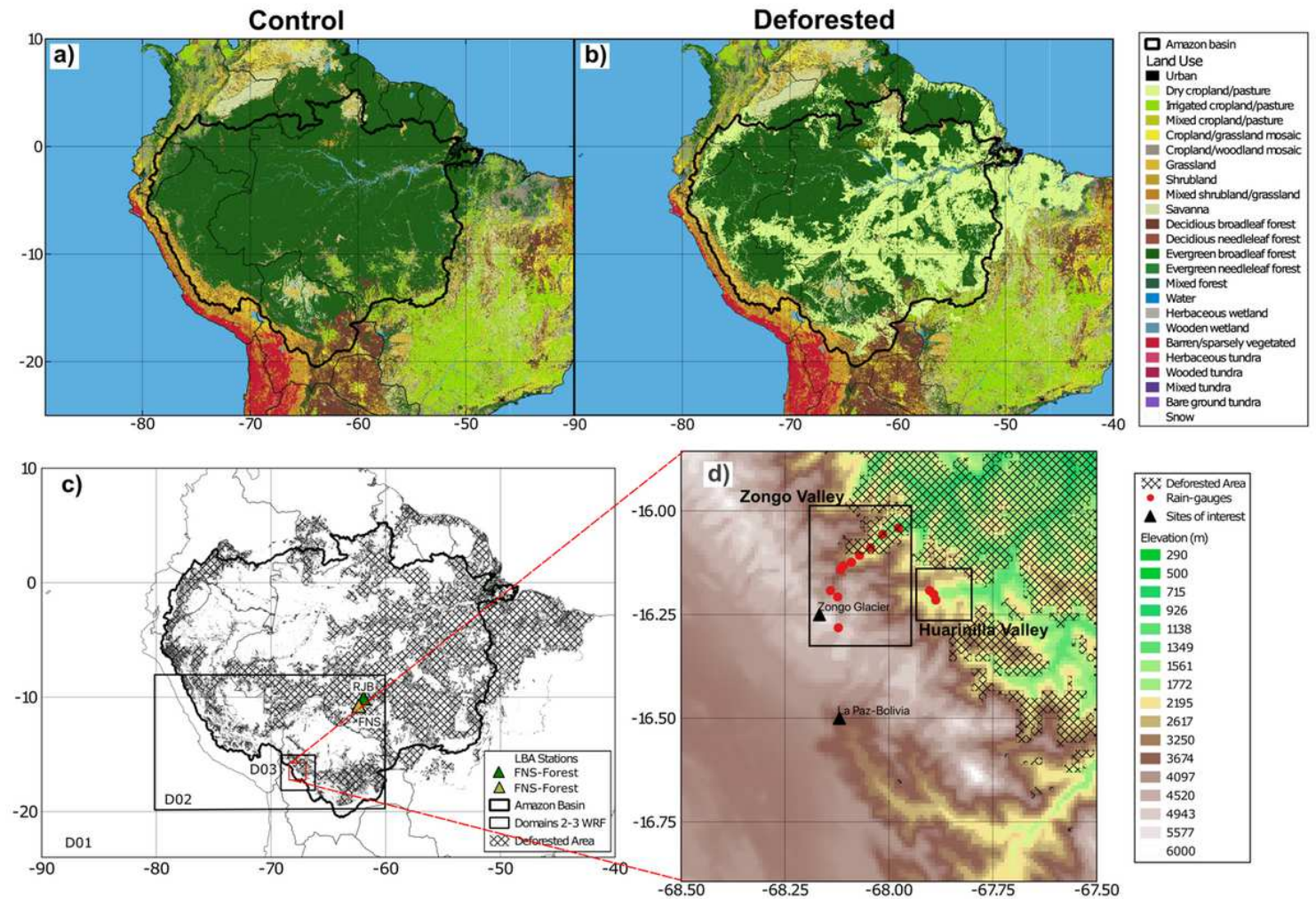


Figure 1

Land-use category distributions for: a) Control and b) Deforested scenarios. c) Deforested area and locations of selected flux towers from the Large-Scale Biosphere-Atmosphere Experiment in the Amazon. Black boxes show WRF domains. Delineated area represents the boundary of the Amazon River basin (black). d) D03 subregion and locations of rain-gauges stations over two Andean mountain valleys Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

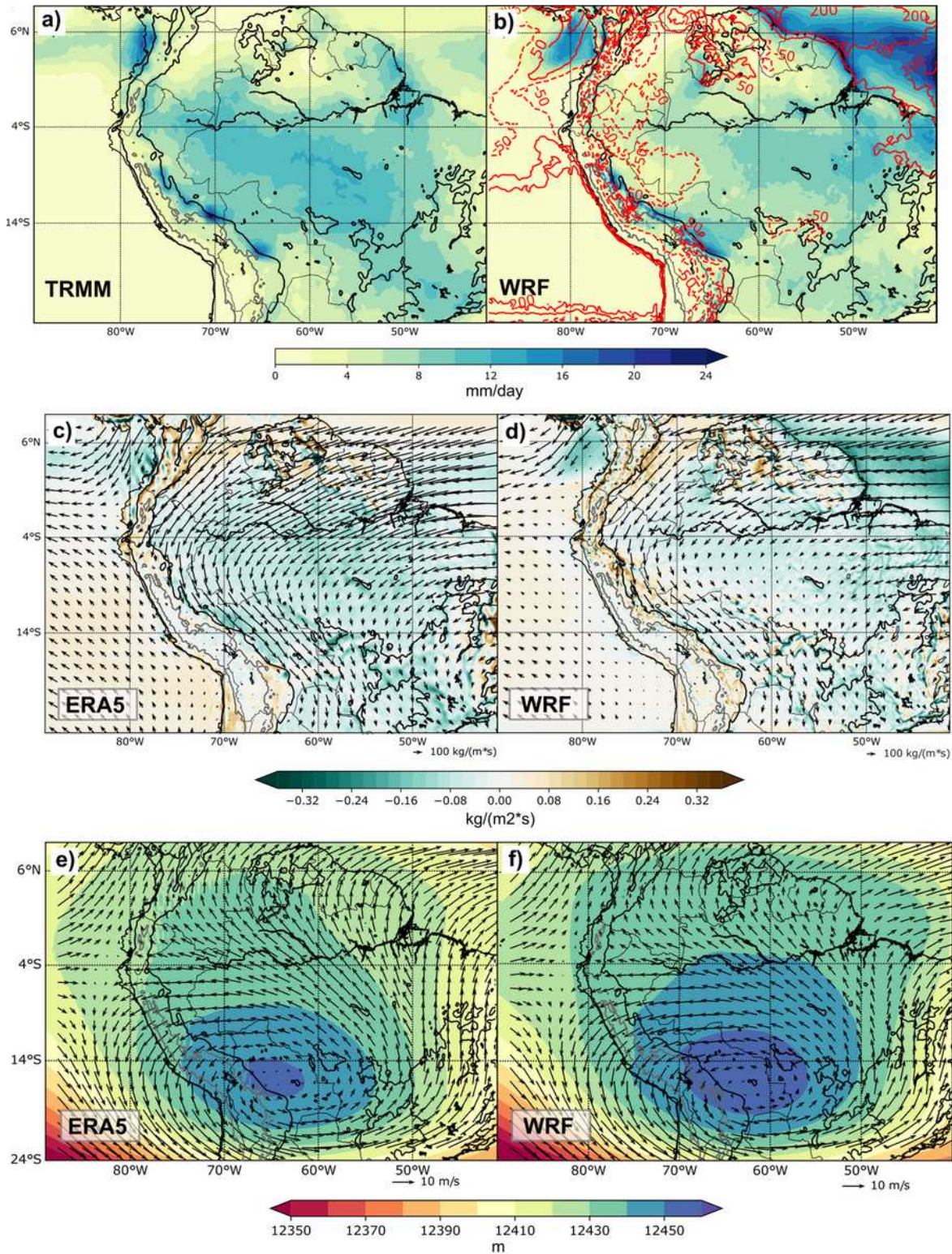


Figure 2

DJF climatological means over the period 2001-2011 for: a) and b) precipitation from TRMM-3B42 (a) and WRF-Control (b) in mm/day. Red lines in b represent rainfall differences respect to TRMM in percentage. c) and d) vertically integrated moisture flux (vectors, kg/m²s) and its divergence (shaded, kg/m²s) from ERA5 (c) and WRF-Control (d). e) and f) horizontal winds at 200 hPa (vectors, m/s) and geopotential height (shaded, m) from ERA5 (e) and WRF-Control (f). Orography elevation is shown at 500

and 3500 m a.s.l. as black and grey contours, respectively Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

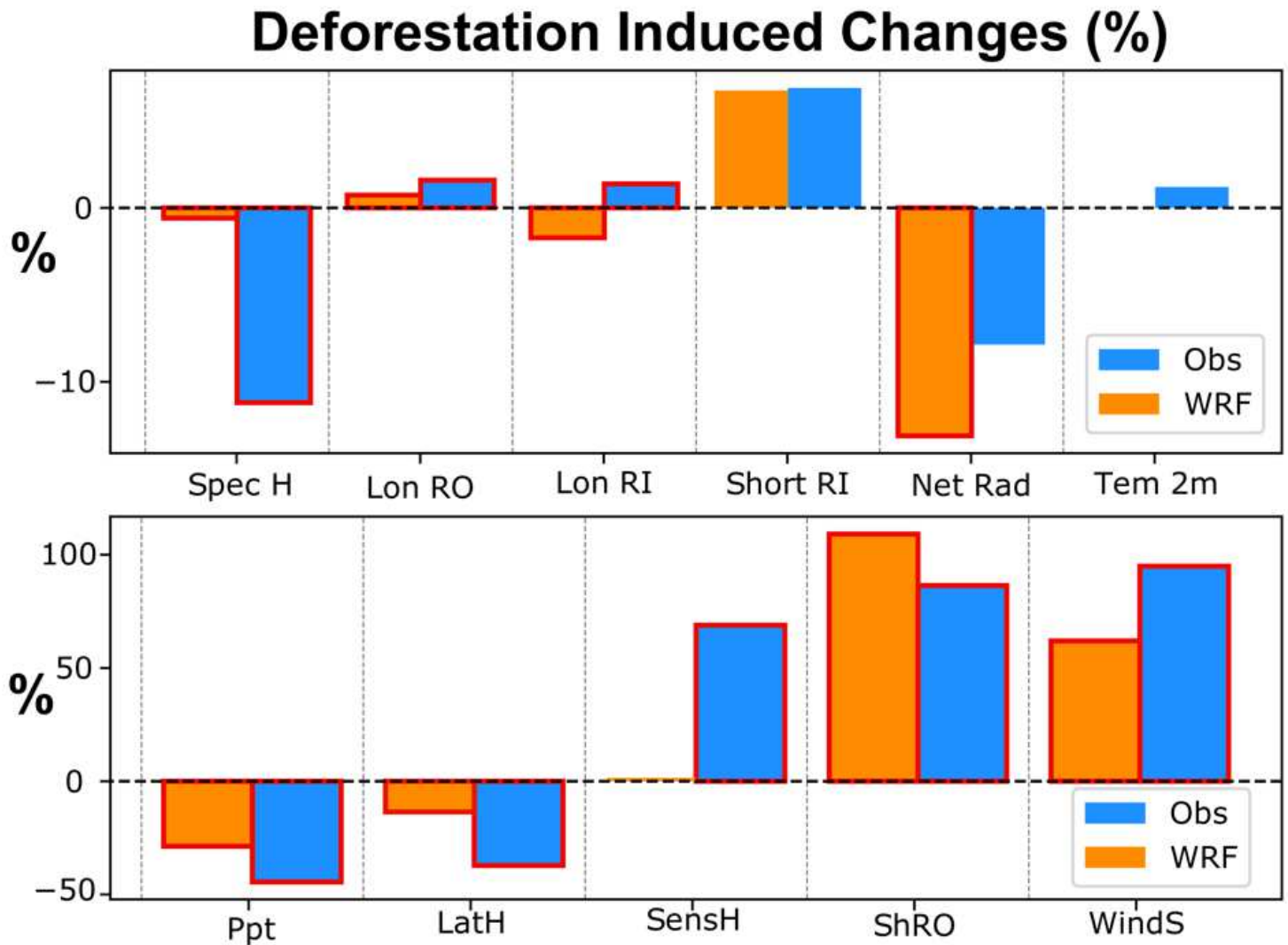


Figure 3

Deforestation induced changes in surface atmospheric variables over LBA stations (blue bars) and WRF-D02 (orange bars) for DJF in the period 2001-2002. Significant differences are delineated in red (t-test, $p < 0.05$). Abbreviations are as follows: Spec H - specific humidity at 2 m, Lon RO – longwave outgoing radiation, Lon RI – longwave incoming radiation, Short RO – shortwave outgoing radiation, Short RI – shortwave incoming radiation, Net-Rad – net radiation, Tem2 2m – air temperature at 2 m, LatH – latent heat flux, SensH – sensible heat flux, Ppt – precipitation, Winds – wind speed at 10 m

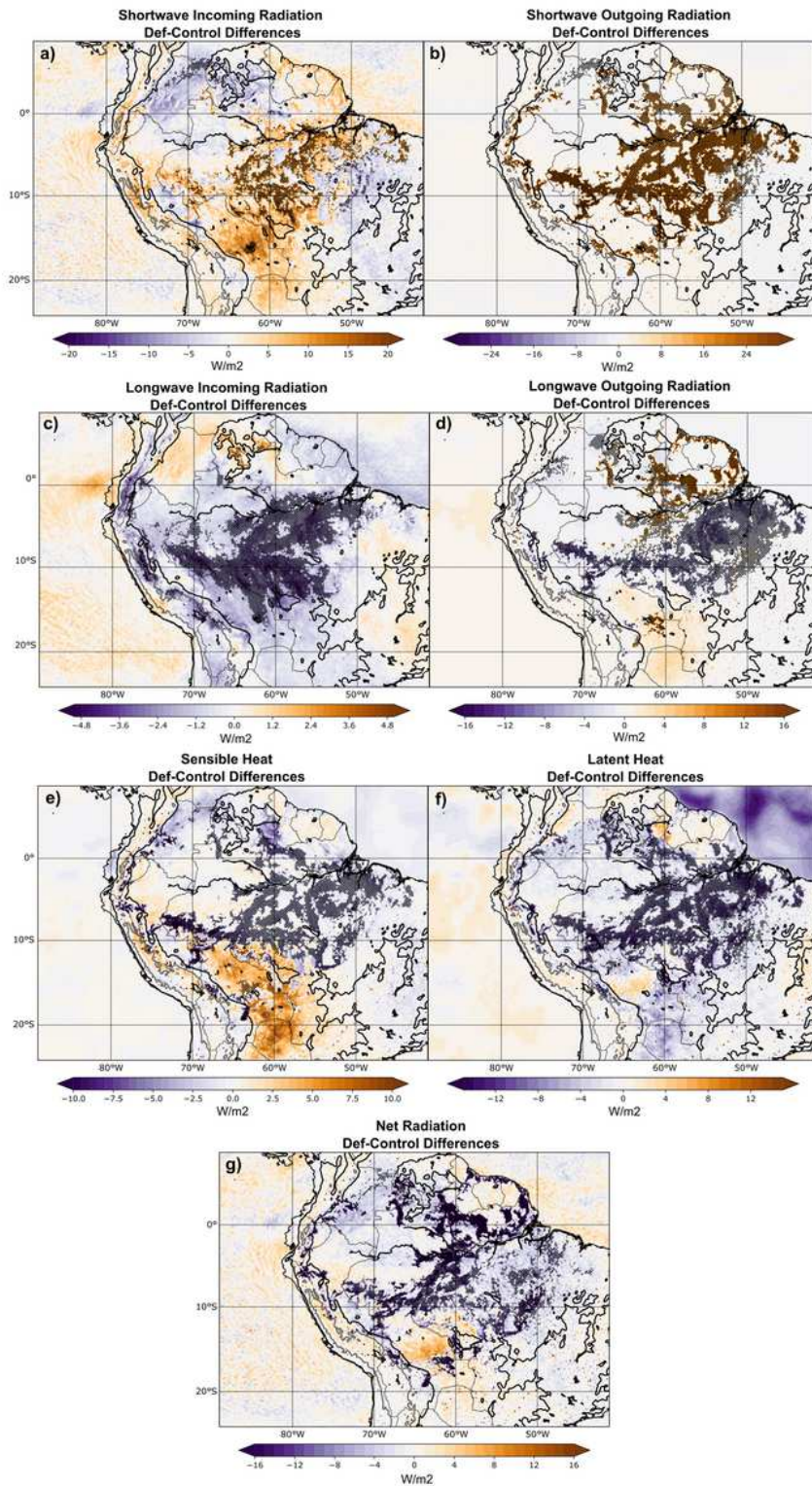


Figure 4

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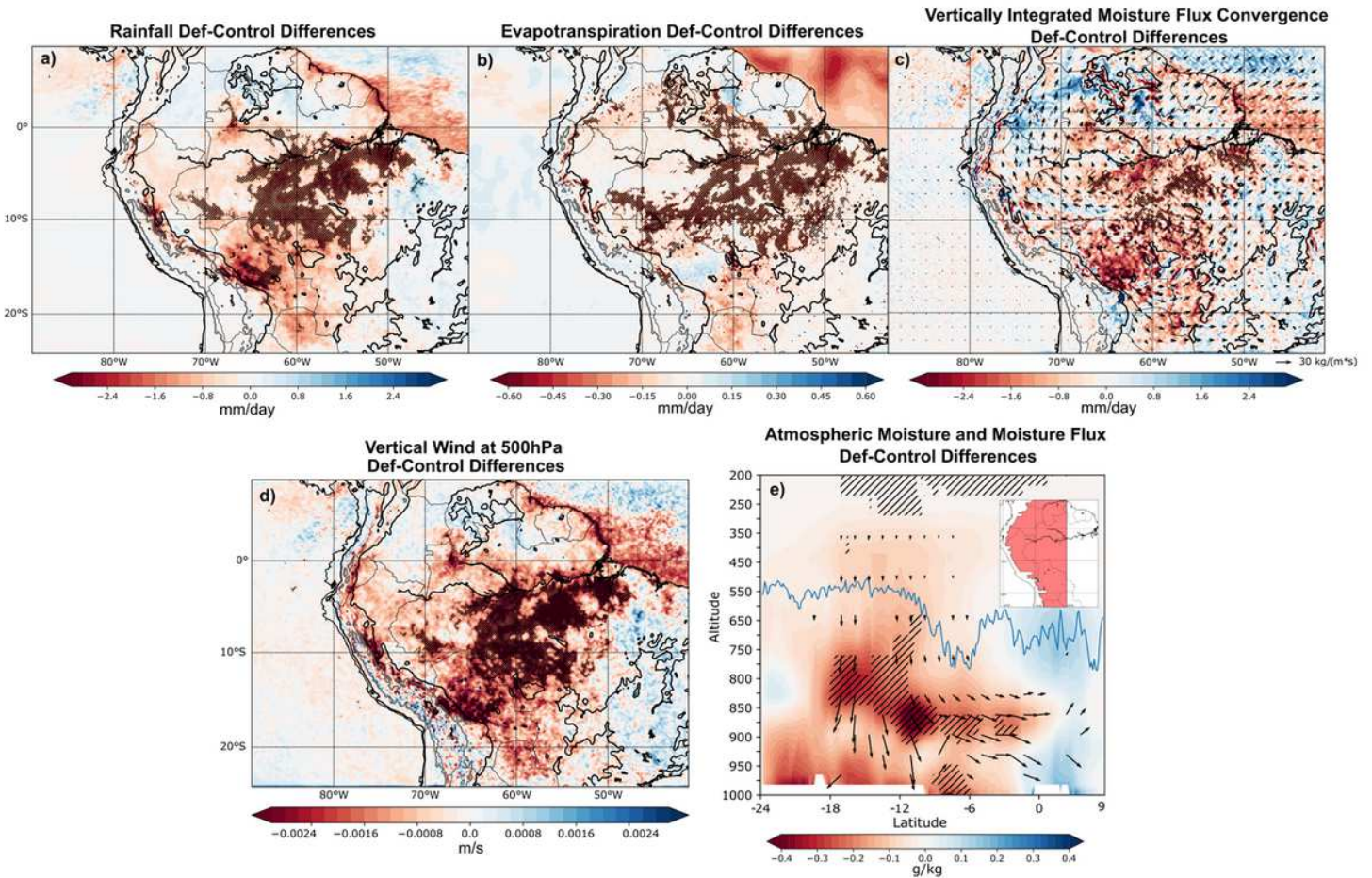


Figure 5

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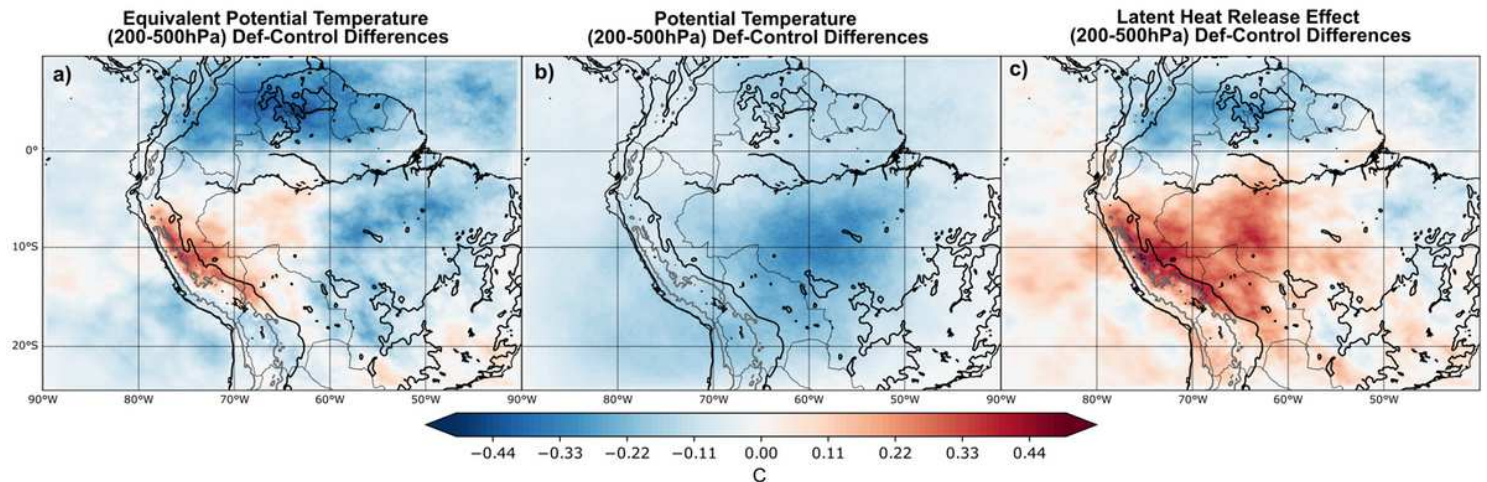


Figure 6

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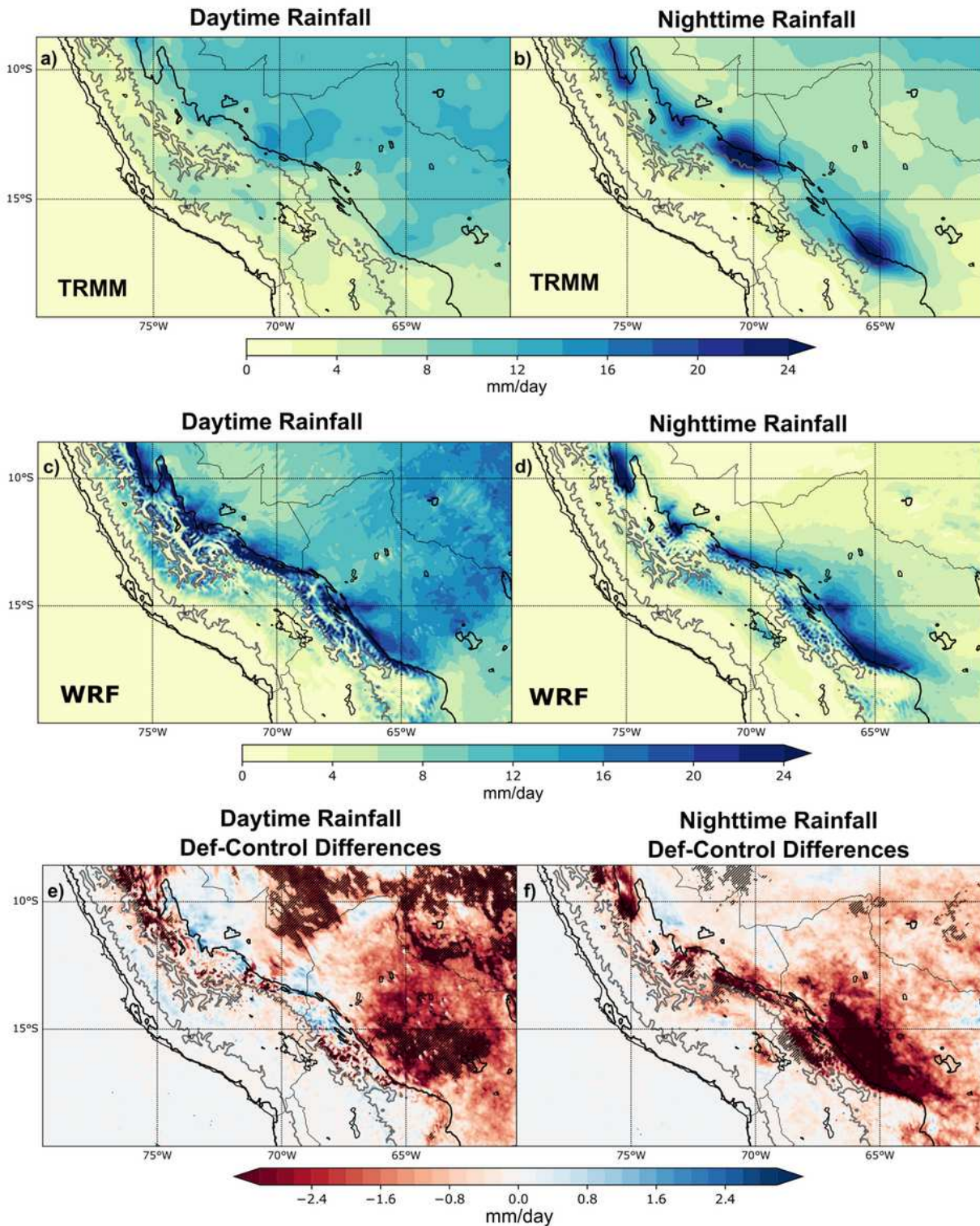


Figure 7

DJF climatological means (2001-2011) for daytime and nighttime precipitation from: a-b) TRMM-3B42. c-510 d) WRF-D02. e-f) WRF-D02 Deforested-Control differences. All figures are in mm/day. Significant changes (t-511 test, $p < 0.05$) are marked with black slashes. Orography elevation is shown at 500 and 3500 m a.s.l. as black and 512 grey contours, respectively Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

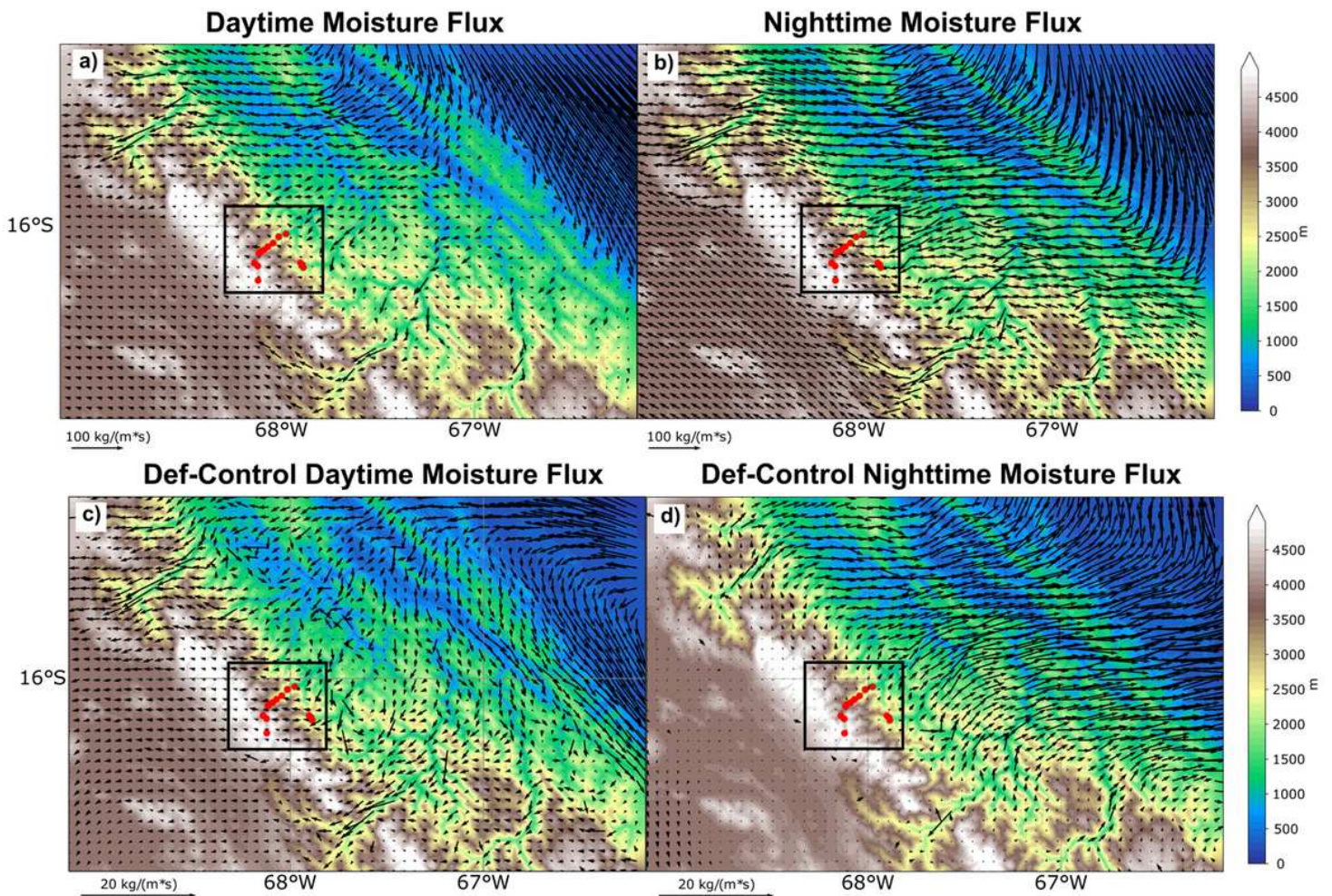


Figure 8

DJF climatological means (2001-2011) for daytime and nighttime vertically-integrated moisture flux from 568 WRF-D03r: a-b) Control and c-d) Deforested minus Control differences. Terrain elevation (shaded) in m and rain-569 gauge stations (red dots). The black square shows the area where the Andean valleys of Huarinilla and Zongo are 570 located Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

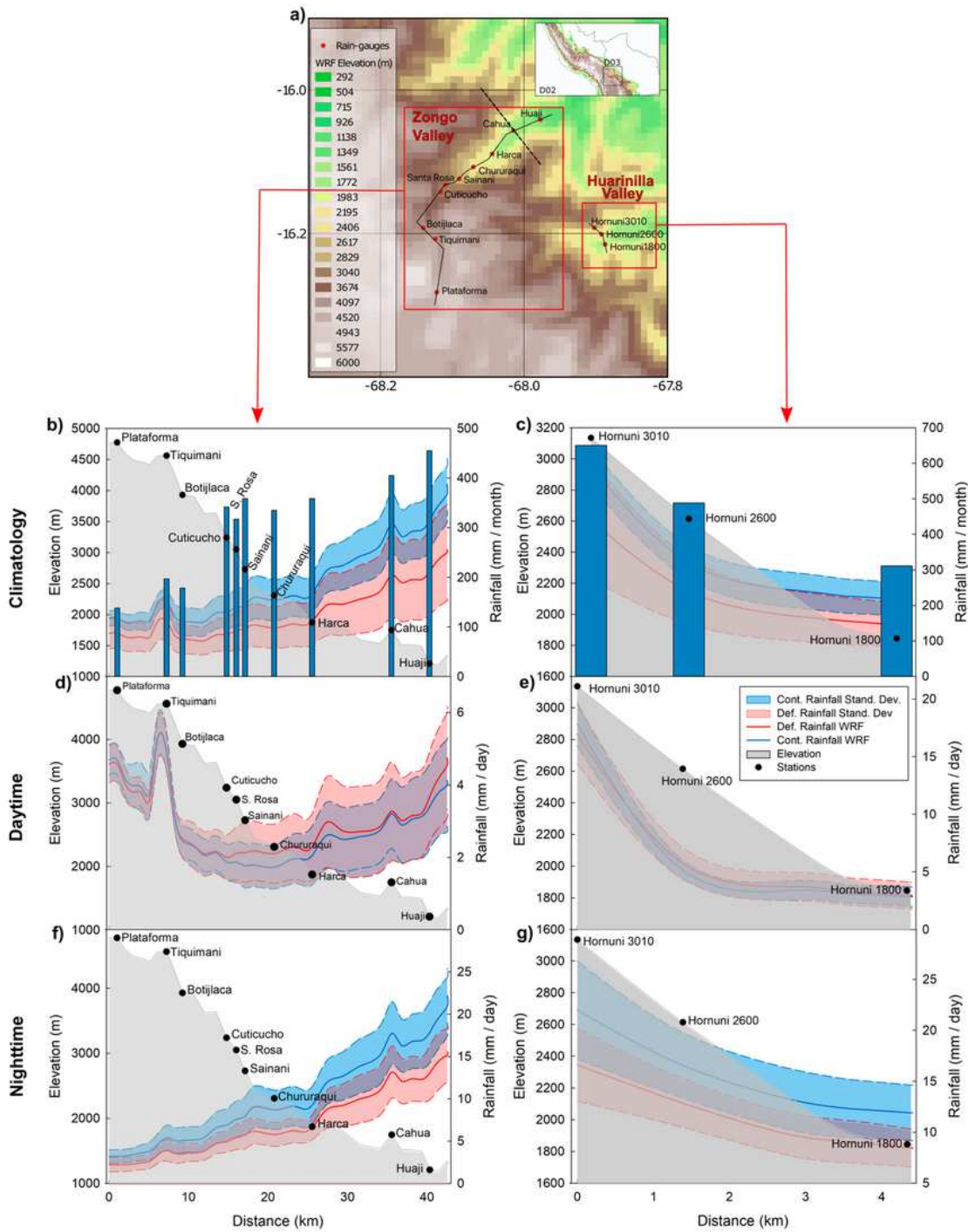


Figure 9

a) Map of the subregion inside D03 with terrain elevation (shaded) in m and rain-gauge stations (red dots). 662 Two selected cross-sections are indicated in black solid lines. A perpendicular cross-section in Zongo Valley 663 shown in Figure 10 is indicated as a dashed black line. b-c) Cross-sections for b) Zongo Valley and c) Huarinilla 664 Valley. Topographical profiles are displayed as a gray line (in m). Observed rain-gauge rainfall is presented as 665 blue bars. WRF-D03 Control and Deforested rainfall

climatological means are shown as blue and red lines and 666 are labeled as Cont. and Def., respectively. Rainfall is in mm/month. d-g) Same as Fig. 9b-c but for daytime and 667 nighttime conditions. Rainfall is in mm/day. The x axis represents the horizontal distance in km along the cross-668 sections. Blue and red envelopes represent model standard deviation for each scenario. Black dots represent rain-669 gauge stations Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

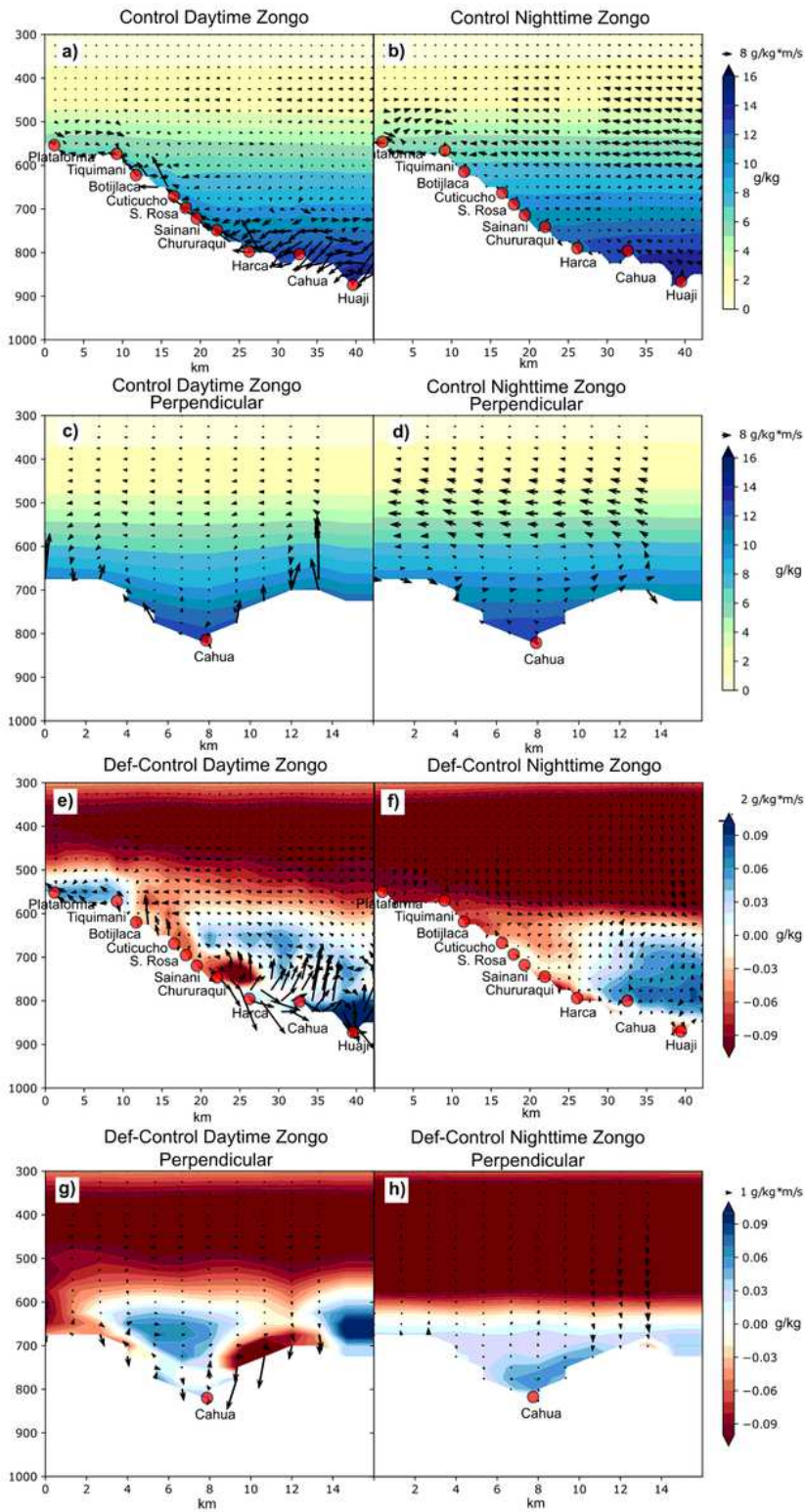


Figure 10

Vertical cross-section for daytime and nighttime DJF climatological means (2001-2011) in atmospheric 687 moisture (shaded, g/kg) and moisture flux (vectors, g/kg*m/s) for the Zongo Valley in two opposite directions: 688 28 parallel to the transect (see black solid line in Fig. 9a) and perpendicular to the transect (see black dashed line in 689 Fig. 9a). Daytime (a) and nighttime (b) WRF-Control conditions for the transect parallel to the Zongo Valley. 690 Daytime (c) and nighttime (d) WRF-Control conditions for the

transect perpendicular to the Zongo Valley. 691 Daytime (e) and nighttime (f) Deforested minus Control differences in the transect parallel to the Zongo Valley. 692 Daytime (g) and nighttime (h) Deforested minus Control differences in the transect perpendicular to the Zongo 693 Valley. Red dots represent rain-gauge stations. Vertical component of the moisture flux is exaggerated by a factor 694 of ten Note: The designations employed and the presentation of the material on this map do not imply the expression of any opinion whatsoever on the part of Research Square concerning the legal status of any country, territory, city or area or of its authorities, or concerning the delimitation of its frontiers or boundaries. This map has been provided by the authors.

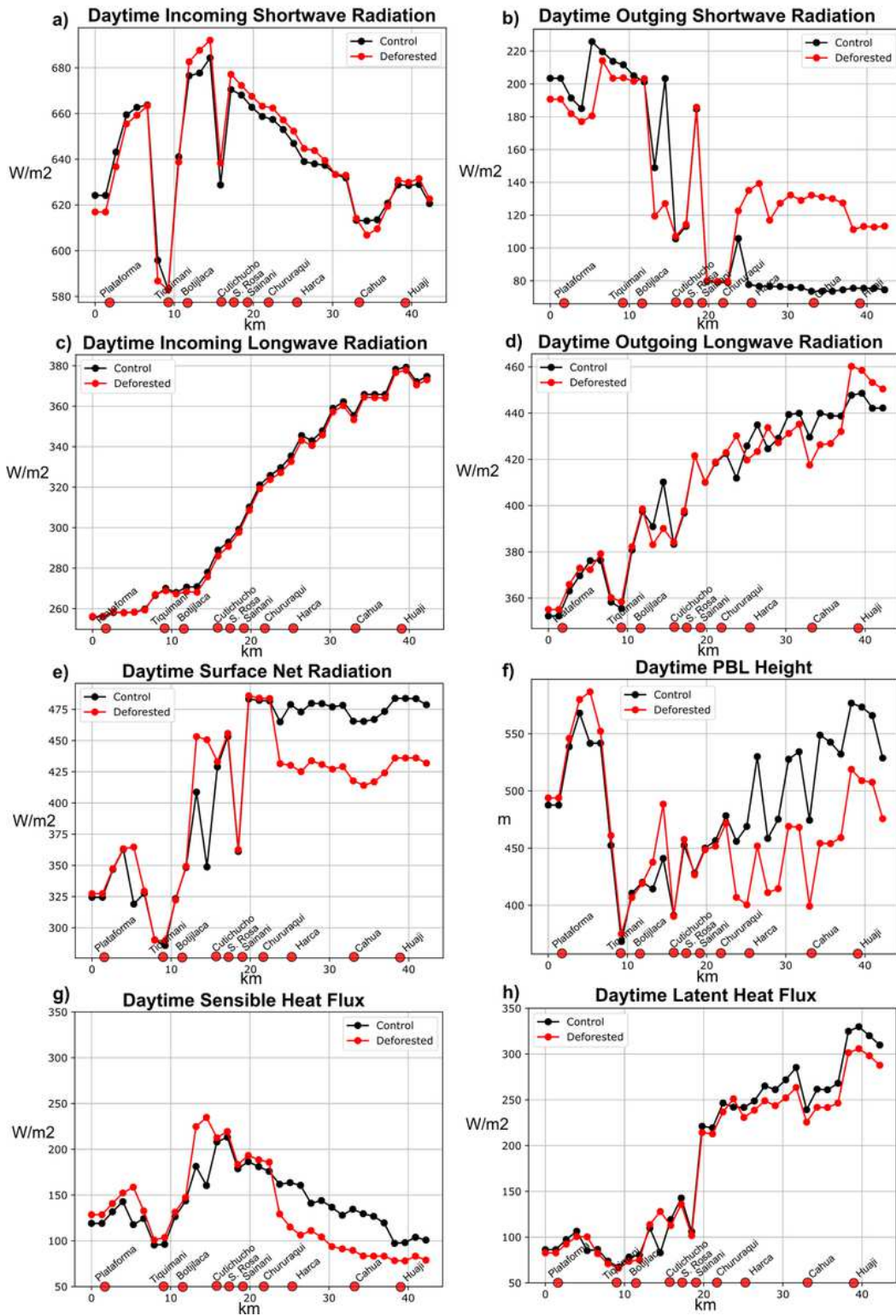


Figure 11

WRF-D03 daytime mean variation along the Zongo Valley transect in Control and Deforested scenarios 698 (black and red lines, respectively) for: a) Shortwave incoming radiation. b) Shortwave outgoing radiation. c) 699 30 Longwave incoming radiation. d) Longwave outgoing radiation. e) Surface net radiation. f) Planetary boundary 700 layer (PBL) height. g) Sensible heat flux. h) Latent heat flux. Red dots represent rain-gauge stations

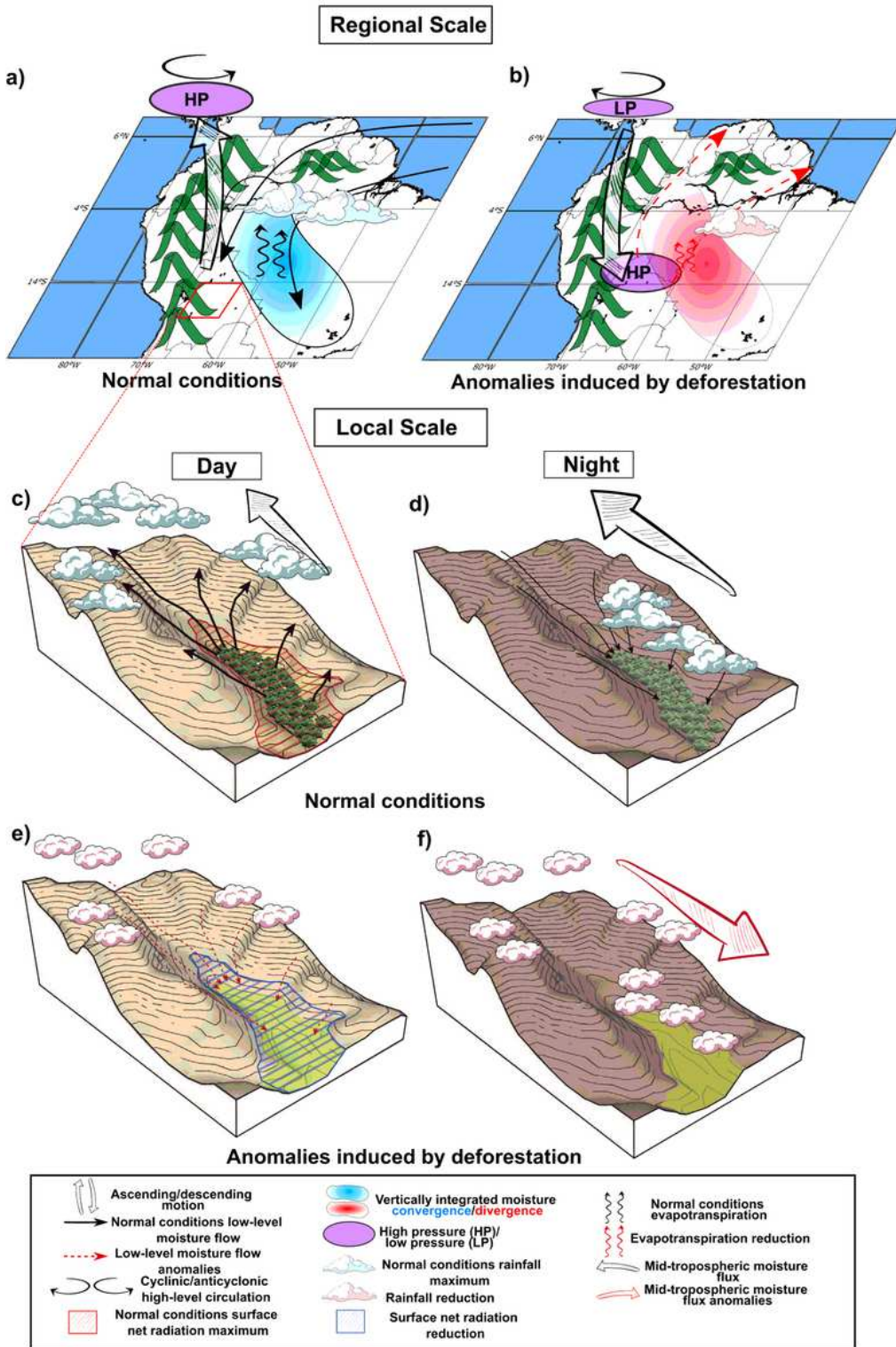


Figure 12

Schematic representation of DJF climatological atmospheric conditions at regional (a-b) and local scales 718 (c-f) for normal conditions (a-c-d) and deforestation induced anomalies (b-e-f) in WRF simulations. The red 719 square in a indicates the approximated location of the Zongo Valley sketched in c-f. Black solid (red dashed) thin 720 arrows represent surface moisture transport under normal conditions (deforestation anomalies). Black (red) thick 721 32 arrows represent mid-tropospheric moisture flux under

normal conditions (deforestation anomalies). Vertical 722 thick arrows represent ascending and descending motions. Black (red) curved arrows represent evapotranspiration 723 under normal conditions (deforestation anomalies). Blue and red shaded areas indicate vertically integrated 724 moisture convergence and divergence, respectively. Purple areas indicate pressure changes. Blue and red clouds 725 indicate rainfall maximum and rainfall reductions under normal and deforested conditions, respectively. Red 726 (blue) hatched area in c (e) indicates surface net radiation maximum (reduction) under normal (deforestation) 727 conditions