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## Presentation and evaluation of the IPSL-CM6A-LR climate model



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#### **Key Points:**

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- The IPSL-CM6A-LR model climatology is much improved over the previous version although some systematic biases and shortcomings persist.
- A long pre-industrial control and a large number of historical and scenario simulations have been performed as part of CMIP6.
- The effective climate sensitivity of the IPSL model increases from 4.1 to 4.8 K between IPSL-CM5A-LR and IPSL-CM6A-LR.

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#### 49 Abstract

This study presents the global climate model IPSL-CM6A-LR developed at IPSL to study 50 natural climate variability and climate response to natural and anthropogenic forcings 51 as part of the 6<sup>th</sup> phase of the Coupled Model Intercomparison Project (CMIP6). This 52 article describes the different model components, their coupling, and the simulated cli-53 mate in comparison to previous model versions. We focus here on the representation of 54 the physical climate along with the main characteristics of the global carbon cycle. The 55 model's climatology, as assessed from a range of metrics (related in particular to radi-56 ation, temperature, precipitation, wind), is strongly improved in comparison to previ-57 ous model versions. Although they are reduced, a number of known biases and short-58 comings (e.g., double ITCZ, frequency of midlatitude wintertime blockings, and ENSO 59 dynamics) persist. The equilibrium climate sensitivity and transient climate response 60 have both increased from the previous climate model IPSL-CM5A-LR used in CMIP5. 61 A large ensemble of more than 30 members for the historical period (1850-2018) and a 62 smaller ensemble for a range of emissions scenarios (until 2100 and 2300) are also pre-63 sented and discussed. 64

#### 65 **Pla**

#### **Plain Language Summary**

Climate models are unique tools to investigate the characteristics and behaviour 66 of the climate system. While climate models and their components are developed grad-67 ually over the years, the 6<sup>th</sup> phase of the Coupled Model Intercomparison Project has 68 been the opportunity for the Institut Pierre-Simon Laplace to develop, test and evalu-69 ate a new configuration of its climate model called IPSL-CM6A-LR. The characteristics 70 and emerging properties of this new model are presented in this study. The model cli-71 matology, as assessed from a range of metrics, is strongly improved although a number 72 of biases common to many models do persist. The equilibrium climate sensitivity and 73 transient climate response have both increased from the previous climate model IPSL-74 CM5A-LR used in CMIP5. 75

#### 76 **1 Introduction**

The Institut Pierre-Simon Laplace Climate Modelling Centre (IPSL CMC, see https:// cmc.ipsl.fr) has set up a new version of its climate model in the runup of phase 6 of

the Coupled Model Intercomparison Project (known as CMIP6, see Eyring et al., 2016,

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for more information). Here we provide a brief description of the coupled model, document the model climatology and its performance against a range of observations and reanalyses, and present some key emerging properties of the model (internal variability and response to forcings). The implementation of the model boundary conditions (Lurton et al., 2019) and the development process for this new model configuration in preparation to CMIP6 are described in two companion papers.

IPSL CMC developed IPSL-CM5A-LR (CM stands for climate model and LR for 86 low resolution) as its main model for phase 5 of CMIP (Dufresne et al., 2013; Szopa et 87 al., 2013). IPSL-CM5A-LR also had two variants: a medium resolution configuration, 88 IPSL-CM5A-MR, and an experimental version, IPSL-CM5B-LR, based on a new ver-89 sion of the atmospheric physics (Hourdin et al., 2013). The resolution of the atmospheric 90 model was 96  $\times$  95 points in longitude and latitude in the LR configuration, and 144  $\times$  143 91 in the MR configuration. Both versions had 39 layers in the vertical. The nominal res-92 olution of the NEMO oceanic model was  $2^{\circ}$  for both configurations. Since then, many 93 improvements have been implemented in the various model components: LMDZ (atmo-0/ sphere), NEMO (ocean, sea ice, marine biogeochemistry) and ORCHIDEE (land surface, hydrology, land carbon cycle). In this article we describe only the coupled ocean-atmosphere 96 model and the carbon cycle in the terrestrial and marine model components as the full 97 Earth System version of IPSL-CM6 is still under development. The resolution of the at-98 mospheric model is now  $144 \times 143$  points in longitude and latitude, which corresponds 99 to an average resolution of  $\sqrt{(4 \pi R^2 / 144 / 142)} = 157$  km (R being the Earth's radius), 100 and 79 vertical layers (with a model top at  $\sim 80$  km). The low horizontal resolution (LR) 101 of IPSL-CM6 thus corresponds to the medium horizontal resolution (MR) of IPSL-CM5. 102 The nominal resolution of the ocean model has been increased to  $1^{\circ}$  and 75 layers in the 103 vertical. 104

This article provides an entry point to the IPSL-CM6A-LR model with a brief scientific and technical description of the model, a thorough evaluation of its climatology and some presentation of the DECK (Diagnostic, Evaluation, and Characterization of Klima) and ScenarioMIP simulations that were prepared for CMIP6. Further studies on the IPSL-CM6A-LR model emerging properties and model intercomparison studies are expected to be ongoing for the next few years.

#### <sup>111</sup> 2 Brief overview of the IPSL-CM6A-LR model

2.1 Introduction

IPSL-CM6A-LR is composed of the LMDZ atmospheric model version 6A-LR (Hourdin,
 Rio, Grandpeix, et al., 2020), the NEMO oceanic model version 3.6 (see references be low) and the ORCHIDEE land surface model version 2.0. We briefly describe below each
 of the three model components and the coupling procedure between them. Further de scription of the IPSL-CM6A-LR climate model is available on the ES-DOC interface (https://
 explore.es-doc.org/cmip6/models/ipsl/ipsl-cm6a-lr).

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#### 2.2 LMDZ6 atmospheric component

The atmospheric general circulation model LMDZ6A-LR is based on a finite-difference 120 formulation of the primitive equations of meteorology (Sadourny & Laval, 1984), on a 121 staggered and stretchable longitude-latitude grid (the Z of LMDZ standing for Zoom). 122 Water vapor, liquid and solid water and atmospheric trace species are advected with a 123 monotonic second order finite volume scheme (Van Leer, 1977; Hourdin & Armengaud, 124 1999). In the vertical, the model uses a classic so-called hybrid sigma-pressure coordi-125 nate. Regarding the physical parameterizations, IPSL participated to CMIP5 with two 126 versions: a "Standard Physics" version (atmospheric component LMDZ5A used in IPSL-127 CM5A, Hourdin et al., 2013) and a "New Physics" (NP) version (LMDZ5B used in IPSL-128 CM5B, Hourdin et al., 2013) based on a full rethinking of the parameterizations of tur-129 bulence, convection and clouds on which the 6A version is built. This NP package in-130 cludes in particular a turbulent scheme based on the prognostic equation for the turbu-131 lent kinetic energy that follows Yamada (1983), a mass flux representation of the orga-132 nized structures of the convective boundary layer called "Thermal Plume Model" (Hourdin 133 et al., 2002; Rio & Hourdin, 2008; Rio et al., 2010) and a parameterization of the cold 134 pools or wakes created below cumulonimbus by the evaporation of convective rainfall (Grandpeix 135 & Lafore, 2010; Grandpeix et al., 2010). The "episodic mixing and buoyancy sorting" 136 scheme originally developed by Emanuel (1991) used for deep convection was modified 137 to make the closure and triggering rely on the description of the sub-cloud vertical mo-138 tions by thermal plumes and wakes (Rio et al., 2009). Regarding convection, two impor-139 tant improvements were made from version 5B to 6A: a modification of the lateral de-140 trainment in the thermal plume model that allows to represent satisfactorily well the tran-141

sition from stratocumulus to cumulus clouds (Hourdin et al., 2019a) and the introduc-142 tion of a statistical triggering for deep convection (Rochetin, Couvreux, et al., 2014; Ro-143 chetin, Grandpeix, et al., 2014). The radiation scheme was inherited from the European 144 Centre for Medium-Range Weather Forecasts. In the LMDZ6A version, it includes the 145 Rapid Radiative Transfer Model (RRTM) code for thermal infrared radiation and an im-146 proved six-band version of the Fouquart and Bonnel (1980) scheme for solar radiation. 147 Cloud cover and cloud water content are computed using a statistical scheme using a log-148 normal function for deep convection (Bony & Emanuel, 2001) and a bigaussian function 149 for shallow cumulus (Jam et al., 2013). 150

The 6A-LR version is based on a regular horizontal grid with 144 points regularly 151 spaced in longitude and 142 in latitude, corresponding to a resolution of  $2.5^{\circ} \times 1.3^{\circ}$ . The 152 model has 79 vertical layers and extends up to 80 km, which makes it a "high-top" model. 153 It includes a representation of gravity waves generated by mountains as well as by con-154 vection (Lott & Guez, 2013) and fronts (de la Cámara & Lott, 2015; de la Cámara et 155 al., 2016). The model shows a self-generated quasi-biennal oscillation (QBO) whose pe-156 riod has been tuned to the observed one for the present-day climate. The source of wa-157 ter vapour in the stratosphere due to methane oxidation is not activated. 158

The readers are directed to Hourdin, Rio, Grandpeix, et al. (2020) for more details.

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#### 2.3 NEMO oceanic component

The ocean component used for IPSL-CM6A-LR is based on the version 3.6 stable 161 of NEMO (Nucleus for European Models of the Ocean), which includes three major com-162 ponents: the ocean physics NEMO-OPA (Madec et al., 2017), the sea-ice dynamics and 163 thermodynamics NEMO-LIM3 (Vancoppenolle et al., 2009; Rousset et al., 2015) and the 164 ocean biogeochemistry NEMO-PISCES (Aumont et al., 2015). The configuration used 165 is eORCA1 (with the e standing for extended), the quasi-isotropic global tripolar grid 166 with a  $1^{\circ}$  nominal resolution, and extended to the south so as to better represent the 167 contribution of Antarctic under-ice shelf seas to the Southern Ocean freshwater cycle (Mathiot 168 et al., 2017). The grid has a latitudinal grid refinement of  $1/3^{\circ}$  in the equatorial region. 169 Vertical discretization uses a partial step formulation (Barnier et al., 2006), which en-170 sures a better representation of bottom bathymetry, with 75 levels. The initial layer thick-171

nesses increase non-uniformly from 1 m at the surface to 10 m at 100 m depth, and reaches
200 m at the bottom; they are subsequently time-dependent (Levier et al., 2007).

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#### 2.3.1 Ocean Physics: NEMO-OPA

The eORCA1 configuration used has a non-linear free surface using the variable 175 volume layer formulation, which induces time-variability of all layer thicknesses (Levier 176 et al., 2007). It uses a polynomial representation of the equation of state TEOS-10 (Roquet 177 et al., 2015). The vertical mixing of tracers and momentum uses the turbulent kinetic 178 energy scheme (Gaspar et al., 1990; Blanke & Delecluse, 1993) and an energy-constrained 179 parameterization of mixing due to internal tides (de Lavergne, 2016; de Lavergne et al., 180 2019). There is no constant background diffusivity other than a floor at molecular lev-181 els:  $1.4 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$  for momentum and  $1.4 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$  for tracers. The mixing in-182 duced by submesoscale processes in the mixed layer is also parameterized (Fox-Kemper 183 et al., 2011). A quadratic bottom friction boundary condition is applied together with 184 a parameterization of a diffusive bottom boundary layer for the tracers with a coefficient 185 of  $1000 \text{ m}^2 \text{ s}^{-1}$ . The model uses an energy-enstrophy-conserving scheme for momentum 186 advection and a no-slip boundary condition is applied on the momentum equations. Lat-187 eral diffusion of momentum is performed on geopotential surfaces and uses a Laplacian 188 viscosity with a coefficient of  $20,000 \text{ m}^2 \text{ s}^{-1}$ . Lateral diffusion of tracers is performed along 189 isoneutral surfaces using Laplacian mixing with a spatially varying coefficient of 1000  $\mathrm{m^2 \, s^{-1}}$ 190 at the Equator decreasing with the reduction of the grid spacing with the latitude and 191 reaches a value less than 500  $m^2 s^{-1}$  poleward to 60°N and S. In addition, there is a pa-192 rameterization of adiabatic eddy mixing (Gent & Mcwilliams, 1990) varying spatially 193 as a function of Rossby radius and local growth rate of baroclinic instabilities. The con-194 figuration also includes representation of the interaction between incoming shortwave ra-195 diation into the ocean and the phytoplankton (Lengaigne et al., 2009). A spatially vary-196 ing geothermal heat flux is applied at the bottom of the ocean (Goutorbe et al., 2011), 197 with a global mean value of  $66 \text{ mW m}^{-2}$ . 198

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#### 2.3.2 Sea-ice: NEMO-LIM

IPSL-CM6A-LR utilises v3.6 of the Louvain-la-Neuve Ice Model (LIM), instead of
 version 2 for IPSL-CM5, hence many of the sea ice model features were revised since CMIP5.
 LIM3.6 is a multi-category halo-thermodynamic dynamic sea ice model embedded in the

NEMO environment (Vancoppenolle et al., 2009; Rousset et al., 2015), based on the Arc-203 tic Ice Dynamics Joint EXperiment (AIDJEX) framework (Coon et al., 1974). LIM3.6 204 combines the ice thickness distribution approach (Thorndike et al., 1975; Bitz et al., 2001; 205 Lipscomb, 2001), the conservation of horizontal momentum (Hibler, 1979), treating sea 206 ice as a 2D elastic-viscous plastic continuum (Hunke & Dukowicz, 1997; Bouillon et al., 207 2013), horizontal transport (Prather, 1986), and energy-conserving halo-thermodynamics 208 (Bitz & Lipscomb, 1999; Vancoppenolle et al., 2009). The multiple ice categories allow 209 resolving the enhanced growth of thin ice and solar radiation uptake through thin ice, 210 and the redistribution of thin onto thick ice through ridging and rafting. Sea ice salin-211 ity is an integral part of the model, evolving dynamically to resolve brine entrapment 212 and drainage; and influencing sea ice thermal properties and ice-ocean exchanges (Vancoppenolle 213 et al., 2009). Five thickness categories are used. Ice temperature and salinity fields are 214 further discretized onto two vertical layers of sea ice and one layer of snow. Horizontally, 215 the ice fields are resolved on the same grid as the ocean component. 216

The large-scale sea ice state was adjusted over atmosphere-forced, first, then at the 217 end of each tuning cycle step of the fully-coupled simulations, adding up to several thou-218 sands of simulated years. The sea ice tuning parameters include the cloud-sky albedo 219 nodal values (dry snow 0.87, wet snow 0.82, dry ice 0.65, wet ice 0.58) and the snow ther-220 mal conductivity (0.31  $\mathrm{W\,m^{-1}\,K^{-1}}$ ). The albedo values lie in the high end of the range, 221 to enhance sea ice formation and reduce melting, and compensate for the effects of high 222 air temperatures above sea ice, in particular in the Arctic winter. The albedo nodal val-223 ues were kept within observational uncertainty range, leaving a low Arctic sea ice bias 224 still. The ice strength parameter was set to  $P^* = 20,000 \text{ Nm}^{-2}$ . A maximum ice con-225 centration is imposed, which is equivalent to impose a minimum open water fraction, and 226 done specifically for each hemisphere (0.997 in the Northern Hemisphere, and 0.95 in the 227 Southern Hemisphere). This choice is justified by the difficulty of the model to main-228 tain open water within the pack, in particular in winter, and even more so for Antarc-229 tic sea ice. 230

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#### 2.3.3 Ocean biogeochemistry: NEMO-PISCES

The biogeochemical model is based on PISCES-v2 (Aumont et al., 2015) which simulates the lower trophic levels of marine ecosystem (phytoplankton, microzooplankton and mesozooplankton) and the biogeochemical cycles of carbon and of the main nutri-

ents (P, N, Fe, and Si). There are twenty-four prognostic variables (tracers) including 235 two phytoplankton compartments (diatoms and nanophytoplankton), two zooplankton 236 size-classes (microzooplankton and mesozooplankton) and a description of the carbon-237 ate chemistry. Formulations in PISCES-v2 are based on a mixed Monod/Quota formal-238 ism. On the one hand, stoichiometry of C/N/P is fixed and growth rate of phytoplank-239 ton is limited by the external availability in N, P and Si. On the other hand, the iron 240 and silicon quotas are variable and growth rate of phytoplankton is limited by the in-241 ternal availability in Fe. Nutrients and/or carbon are supplied to the ocean from three 242 different sources: atmospheric deposition, rivers, and sediment mobilization. PISCES is 243 used here to compute air-sea fluxes of carbon and also the effect of a biophysical cou-244 pling: the chlorophyll concentration produced by the biological component feedbacks on 245 the ocean heat budget by modulating the absorption of light as well as the oceanic heat-246 ing rate (Lengaigne et al., 2009). 247

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#### 2.4 ORCHIDEE land surface component

ORCHIDEE is a global process-based model of the land surface and the terrestrial biosphere, that calculates water, energy, and carbon fluxes between the surface and the atmosphere. The model, initially described in Krinner et al. (2005) for the version used in the IPSL-CM5 model, has been significantly improved in version 2.0 used in IPSL-CM6A-LR. We only summarize below the main characteristics of ORCHIDEE and key improvements from the CMIP5 version.

The vegetation heterogeneity is described using fractions of 15 different Plant Func-255 tional Types (PFTs, Prentice et al., 1992) for each grid cell. All PFTs share the same 256 equations but with different parameters, except for the leaf phenology. The annual evo-257 lution of the PFT maps (including a wood harvest product) is derived from the LUHv2 258 database (Lurton et al., 2019). In each grid cell, the PFTs are grouped into three soil 259 tiles according to their physiological behavior: high vegetation (forests) with eight PFTs, 260 low vegetation (grasses and crops) with 6 PFTs, and bare soil with one PFT. An inde-261 pendent hydrological budget is calculated for each soil tile, to prevent forests from ex-262 hausting all soil moisture. In contrast, only one energy budget (and snow budget) is cal-263 culated for the whole grid cell. Note that the energy budget is solved with an implicit 264 numerical scheme that couples the lower atmosphere to the surface, in order to increase 265 numerical stability. All components of the surface energy and water budgets, as well as 266

plant/soil carbon fluxes, are computed at the same time step as the atmospheric physics
(i.e. 15 min., Hourdin, Rio, Grandpeix, et al., 2020) using a standard "big leaf" approach,
but the "slow" processes (carbon allocation in the different plant reservoir and litter and
soil carbon dynamic) are computed on a daily time step. The routing scheme to transform runoff into river discharge to the ocean (Ngo-Duc et al., 2007) also proceeds at the
daily time step, and has not changed since IPSL-CM5.

A physically-based 11-layer soil hydrology scheme has replaced the 2-layer bucket 273 model used in IPSL-CM5. Vertical water fluxes are described using the Richard equa-274 tion discretized with 11 layers for a 2 m soil depth and a free drainage condition is im-275 posed at the bottom of the soil column (de Rosnay et al., 2002; d'Orgeval et al., 2008). 276 As detailed in F. Wang et al. (2016), the vertical discretization for heat diffusion is now 277 identical to that adopted for water up to 2 m. Furthermore the soil depth for heat dif-278 fusion is extended to 90 m, with a zero flux condition at the bottom and 18 calculation 279 nodes, extrapolating the moisture content of the deepest hydrological layer to the en-280 tire profile between 2 and 90 m. The soil thermal properties (heat capacity and conduc-281 tivity) of each layer now depend on soil moisture and soil texture, like the soil hydro-282 logical properties (hydraulic conductivity and diffusivity). Each model grid cell is char-283 acterized by the dominant soil texture, as derived from the map of Zobler (1986) (but 284 reduced to three classes: coarse / sandy loam, medium / loam, and fine / clay loam), 285 and controlling the constant soil parameters (porosity, Van Genuchten parameters, field 286 capacity and wilting point, dry and saturated thermal properties). All these changes have 287 a significant impact on the surface temperature and its high frequency variability in most 288 regions (Cheruy et al., 2017). 289

In contrast to IPSL-CM5, soil freezing is allowed and diagnosed in each soil layer 290 following a scheme proposed by Gouttevin et al. (2012), but the latent heat release/consumption 291 associated with water freezing/thawing is not accounted for. The freezing state of the 292 soil mainly impacts the computation of soil thermal and hydraulic properties, reducing 293 for instance the water infiltration capacity at soil surface. Finally, the 1-layer snow scheme 294 of IPSL-CM5 was replaced by a 3-layer scheme of intermediate complexity described in 295 T. Wang et al. (2013) and inspired by the scheme proposed in Boone and Etchevers (2001). 206 A revised parameterization of the vegetation and snow albedo has been also introduced 297 with optimized parameters based on remote sensing albedo data from MODIS sensor. 298

For the carbon cycle, photosynthesis depends on light availability, CO<sub>2</sub> concentra-299 tion, soil moisture and surface air temperature. It is parameterized based on Farquhar 300 et al. (1980) and Collatz et al. (1992) for C3 and C4 plants, respectively. We used the 301 implementation proposed by X. Yin and Struik (2009) that derives an analytical solu-302 tion of the three equations linking the net assimilation rate, the stomatal conductance, 303 and the intercellular  $CO_2$  partial pressure. In addition, the new version of ORCHIDEE 304 used in IPSL-CM6A-LR includes a "downregulation" capability which accounts for a re-305 duction of the maximum photosynthesis rates as the  $CO_2$  concentration increases in or-306 der to account for nutrient limitations. This downregulation mechanism is modelled as 307 a logarithmic function of the  $CO_2$  concentration relative to 380 ppm following Sellers et 308 al. (1996). Once the carbon is fixed by photosynthesis, we compute the autotrophic res-309 piration (growth and maintenance) and then allocate the remaining carbon into 8 plant 310 compartments (below and above ground sapwood and heartwood; leaves; fruit; roots; re-311 serves). Each compartment has a specific turnover depending on environmental stresses 312 and the living biomass is turned into a litter pool that is distributed in four compart-313 ments (metabolic or structural, both above or below ground). The litter is decomposed 314 following first order kinetics equations, modulated by upper soil moisture and temper-315 ature, with a fraction that is respired and a fraction that is distributed into 3 soil organic 316 carbon pools (active, slow and passive), following the CENTURY model (Parton et al., 317 1987). Each soil organic carbon pool is also decomposed following first order kinetic equa-318 tions modulated by soil moisture and temperature. Overall, the carbon respired from 319 the litter and soil carbon pools defines the heterotrophic respiration. 320

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#### 2.5 Coupling between the components

The LMDZ and ORCHIDEE models are coupled at every time step of the physics of the atmospheric model (i.e., 15 minutes) with the exception of the biogeochemical processes and the vegetation dynamics for which the coupling frequency is one day.

The coupling between LMDZ and NEMO in IPSL-CM5 is described in Marti et al. (2010). It is now performed with the OASIS3-MCT coupler. IPSL-CM6A-LR introduces some modifications and new features: models are coupled with a frequency of 90 minutes, which is both the timestep of the sea-ice model and of the radiation computation in the atmosphere. Atmospheric variables passed to the ocean model (heat, water and momentum fluxes) are averaged temporally over the six 15-minute timesteps of

the LMDZ physics. The flux of freshwater from rivers is passed to the ocean model at 331 river mouth locations with a frequency of one day, which is also the timestep of the river 332 routing in ORCHIDEE. On the ocean grid, the water coming from a river is smoothed 333 over about 200 km to avoid strong haloclines that may occasionally cause the ocean model 334 to crash. To ensure water conservation, the water flux into endorheic basins is globally 335 integrated and homogeneously redistributed over the ocean. Oceanic model variables sent 336 to the atmosphere every 90 minutes are sea surface temperature, sea-ice fraction, sea-337 ice surface temperature and albedo, averaged on two ocean dynamics time-steps. Albedo 338 for the open surface ocean is computed at every timestep in LMDZ following Séférian 339 et al. (2018), which represents a significant improvement over the parameterization used 340 in IPSL-CM5 models. Ocean albedo is a function of solar zenith angle, waveband and 341 surface wind speed; the optional dependence on chlorophyll content of the surface ocean 342 has not been activated. Separate albedoes are provided to the radiative transfer scheme 343 for direct and diffuse radiation. 344

The model includes a very simple scheme to represent the water budget of ice sheets. 345 Snow can accumulate on the land ice fraction of a gridbox, while water vapour can de-346 posit or sublimate depending on the surface relative humidity. The snowpack is capped 347 to a value of  $3000 \text{ kg m}^{-2}$  and any excess is sent to a buffer reservoir before returning 348 to the ocean. This buffering is achieved through a temporal smoothing of the freshwa-349 ter flux (with a 10 year e-folding time) to avoid any spurious low-frequency variability 350 in the freshwater input to the ocean. The flux is then integrated in three latitudinal bands 351  $(90^{\circ}N-40^{\circ}N, 40^{\circ}N-40^{\circ}S \text{ and } 40^{\circ}S-90^{\circ}S)$  and passed to the ocean. In the north and in 352 the tropical/subtropical bands, the flux is equally distributed over the ocean on the same 353 latitudinal bands. In the south band, it is split in two contributions of 50% each corre-354 sponding to ice shelf melting and iceberg melting. The ice shelf melting is geographically 355 and vertically distributed along Antarctica so as to mimic the observed distribution from 356 Depoorter et al. (2013) as described in Mathiot et al. (2017). The iceberg melting is spread 357 offshore following the observed geographical distribution of icebergs of Merino et al. (2016) 358 and distributed vertically over the top 150 m, similarly to river runoffs. 359

Insufficient information is available from the atmospheric and land surface models on the temperatures of freshwater inputs to the ocean so a number of simplifying assumptions are made: the temperatures of rain and snow reaching the ocean are assumed to be that of the SST or the ice surface temperature in the ice-covered areas, the temperature of the riverflow is assumed to be that of the SST at the river mouth (except if the latter is negative, in which case riverflow is assumed to be at  $0^{\circ}$ C). The freshwater flux from iceberg melting is treated as runoff hence the latent heat required to melt the ice is ignored and its temperature is set to the SST. In contrast the freshwater flux from iceshelf melting is treated as ice at  $0^{\circ}$ C and the latent heat required to melt it is accounted for.

The lack of representation of the energy content of precipitation, riverflow and ice-370 bergs results in energy not being conserved exactly in the model. It should be noted that 371 these are not the only non-energy-conserving processes in the model. A number of subgrid-372 scale parameterizations in both the ocean (e.g., eddy-induced velocity, convection, hor-373 izontal momentum mixing, turbulent kinetic energy dissipation), in the atmosphere (e.g., 374 convection scheme) and at the ocean-atmosphere interface (e.g., wind stress interpola-375 tion) are not conserving energy exactly. A small lack of energy conservation is not a ma-376 jor issue as small energy sources and sinks do not prevent the model from equilibrating, 377 at least if the non-conserving terms are stationary. Achieving a more exact energy con-378 servation is an objective for the next version of our climate model. 379

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#### 2.6 Optional model components

Other model components can be activated in IPSL-CM6A-LR but are neither fur-381 ther described in this article nor used in the model experiments presented below. These 382 include atmospheric chemistry / aerosol microphysics models such as the INteractions 383 with Chemistry and Aerosols (INCA, Hauglustaine et al., 2014), the REactive Processes 384 Ruling the Ozone BUdget in the Stratosphere (REPROBUS, Marchand et al., 2012), and 385 the Sectional Stratospheric Sulfur Aerosol (S3A, Kleinschmitt et al., 2017) models. Ac-386 tivation of one of these model components (instead of specifying atmospheric chemical 387 composition and aerosol climatologies) requires a small re-tuning of either the LMDZ6A 388 model or, in the case of S3A, the background stratospheric sulfur budget in order to en-389 sure a similar baseline climate than in IPSL-CM6A-LR. The coupling of these chemistry 390 and aerosol models with the other model components will be described in forthcoming 391 publications. 392

#### 2.7 Testing, tuning and evaluation procedure

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The model was largely developed and tested under present-day climate using a setup we refer to as a *pdControl* setup, which corresponds to present-day climate forcings with an artificial sink of shortwave radiative energy reaching the ocean surface in order to compensate for the ongoing oceanic heat uptake of the current unequilibrated climate. A number of model features were tuned towards observations (see Hourdin et al. (2017) for the rationale).

The parameters considered during the tuning of the atmospheric model are given 400 in Table 3 of Hourdin, Rio, Grandpeix, et al. (2020). They concern in particular the con-401 trol of the deep convection scheme, the control of the conversion of clouds condensed wa-402 ter to rainfall, and the control of the vertical dependency of the width of the subrid-scale 403 water distribution for non convective clouds. A parameter was introduced as well in the 404 "thermal plume" model to control the representation of the transition from cumulus to 405 stratocumulus clouds (Hourdin et al., 2019b; Hourdin, Rio, Jam, et al., 2020). The threshold value for the conversion from liquid cloud water to rainfall as well as a parameter 407 that controls the indirect effect of clouds were used for the final tuning of the global ra-408 diative balance because they affect specifically the optical thickness of liquid (low) clouds, 409 thus modifying the total shortwave radiation much more than the longwave. 410

At some point during the development process, the main development stream switched 411 from a *pdControl* to a *piControl* setup, which corresponds to pre-industrial climate forc-412 ings. The spin up lasts several hundreds years but with some evolution of the model physics 413 as the tuning was being finalized. The final tuning process involved changes in param-414 eters associated with the sea ice physical properties (albedo and conductivity), subgrid-415 scale orography parametrization, and penetration of energy in the upper ocean with and 416 without sea-ice cover. Various options were envisaged as well concerning the control of 417 atmospheric deep convection and its competition with shallow convection. One impor-418 tant choice of the final configuration was to consider boundary-layer convective trans-419 port by the "thermal plume model" outside cold pools only. With this choice, thermal 420 plumes see a more unstable environment (since the thermal plume is more stable than 421 the mean column). Thermal plumes are therefore more active, which in turns favors shal-422 low convection compared to deep. The atmospheric deep convection activity over the ocean 423 was modified as well by using a different (larger) value of the horizontal density of cold 424

pools: 1 cold pool per (33 km)<sup>2</sup> over ocean versus 1 per (350 km)<sup>2</sup> over land (note that
the parameterization of this cold pool density is currently further tested in more recent
versions). The surface drag over the ocean was also modified by introducing a gustiness
term computed as a function of the vertical velocity associated with air lifting by the thermal plumes and by the gust fronts of cold pools.

The model code was then frozen (version 6.1.0) and subsequently altered only for correcting diagnostics or allowing further options and configurations. Versions 6.1.0 to 6.1.11 (the current version) are therefore bit-reproducible for a given domain decomposition, compiling options and supercomputer.

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A multi-centennial pre-industrial control was then simulated: 100 years (1750–1850) 434 as the *piControl-spinup* experiment and 2000 years (1850-3849) as the *piControl* exper-435 iment. It should be noted that the *piControl* experiment suffers from a small cooling drift 436 of  $\sim 0.2$  K in 2000 years. A shorter *piControl* experiment of 250 years labelled r1i2p1f1 437 was run on the Joliot-Curie supercomputer to check the consistency. A large ensemble 438 (32 members) of *historical* (1850–2014) simulations were performed following the CMIP6 439 protocol. Initial conditions for the historical simulations were sampled every 20 or 40 440 years of the *piControl* starting with year 1870 of the *piControl*. The r1i1p1f1 simula-441 tion was selected qualitatively among the first  $\sim 12$  available members at the time of se-442 lection on the basis of a few key observables of the historical period such as the evolu-443 tion of the global-mean surface air temperature, summer sea ice extent in the Arctic ocean, 444 and annual sea ice volume in the Arctic ocean. The rationale for highlighting a partic-445 ular member is that we expect many users to only consider r1i1p1f1 rather than the 446 whole ensemble. A more thorough ongoing analysis of our historical large ensemble shows 447 that other members appear to be closer to the observed record in many respects. Most 448 of the historical simulations were prolonged (outside the CMIP6 protocol) to 2059 us-449 ing SSP245 atmospheric, land use and solar forcings (except for the wood harvest and 450 ozone forcings, not available at the time, which have been kept constant to their 2014 451 values). An ensemble of scenario simulations for 2015-2100 with a few extensions to the 452 year 2300 were also performed following ScenarioMIP guidelines (O'Neill et al., 2016). 453

2.8 Infrastructure improvements

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The IPSL-CM6A-LR model can be extracted, installed and compiled on a specific machine using a suite of scripts called modipsl. Simulations are executed within the libIGCM running environment, which can be used to set up and run a simulation on a specific machine through a chain of computing and post-processing jobs. Metadata from the simulations are sent to the Hermes supervising tool which can be used to monitor progress in the simulations and key variables from different simulations can be intercompared using an intermonitoring tool on a dedicated web server (closed access).

The CMIP6 simulations were performed at the Très Grand Centre de Calcul (TGCC) on the Curie supercomputer with a switch during the CMIP6 production in October 2018 to the Joliot-Curie supercomputer. Both a *piControl* and a *historical* simulations initially performed on Curie were repeated on the Joliot-Curie supercomputer to ensure the climate statistics were comparable on both supercomputers. The throughput is about 13 and 16 simulated years per day on Curie and Joliot-Curie, respectively, on 960 processors with the full CMIP6 output.

Outputs from the IPSL-CM6A-LR model are managed by the XML Input/Output 469 Server (XIOS Meurdesoif et al., 2016). For the CMIP6 production, model output vari-470 ables in native format were kept to the minimum. Instead, and in sharp contrast to CMIP5, 471 the CMIP6-compliant model output has been produced on the fly using XIOS methods 472 from the code. XIOS is driven by XML files describing the whole netCDF file structure 473 (dimensions, attributes, etc.). Such XML files were produced using the dr2XML (https:// 474 github.com/rigoudyg/dr2xml) python library developed by our CNRM-CERFACS col-475 laborators, which translates the CMIP6 Data Request and Controlled Vocabulary (https:// 476 github.com/WCRP-CMIP/CMIP6\_CVs) into XML files for XIOS. We use dr2XML to gen-477 erate XML files for each simulated year of a given CMIP6 experiment and member. The 478 netCDF time series are created and filled by XIOS all along the simulation to avoid con-479 catenation during post-processing. 480

A quality assurance is applied at the end of each CMIP6 model simulation. The simulation is validated from a scientific point of view to make sure there is no critical issue or inconsistencies in the diagnostics (e.g., incorrect application of a forcing term, wrong sign, recurring patterns, etc). Each file then undergoes several checks against the CMIP6 controlled vocabulary to ensure its conformance with the CMIP6 Data Reference Syntax (http://pcmdi.llnl.gov/CMIP6) in terms of time axis and coverage, variable and global required metadata, filename syntax, etc.

The data are then published on the Earth System Grid Federation (ESGF), which guarantees a strong and effective data management. The espprep toolbox (http://esgf 489 .github.io/esgf-prepare) is a piece of software that eases data preparation accord-490 ing to CMIP6 conformance. Once a model simulation is validated and checked, the netCDF 491 files are migrated in the proper CMIP6 directory structure with the esgprep commands 492 in a shared space of the file system. The IPSL hosts an ESGF index with all datasets 493 from the French climate simulations and a data node to disseminate datasets from the 494 IPSL climate simulations. The IPSL CMIP6 data sets are published on the ESGF data 105 node using the usual *esqpublish* command-line provided by the node stack. During the 496 publication process, the Persistent IDentifier (PID) included in each netCDF file, is per-497 manently stored in a dedicated database at the German Climate Computing Centre (DKRZ) 498 allowing further data citation. 499

#### **3 Evaluation of present-day climatology**

In this section we evaluate the present-day climate of IPSL-CM6A-LR against our 501 CMIP5 flagship configurations, IPSL-CM5A-LR and IPSL-CM5A-MR, by considering 502 recent periods (i.e., the 1980–2005 period, when not mentioned otherwise) of our histor-503 *ical* simulations. The references (i.e., observations and/or reanalyses) used to evaluate 504 the models generally cover the same period, but may sometimes include some years af-505 ter (like ERA-Interim) or before (like WOA13-v2). We argue that not considering the 506 exact same periods for the simulations and the observations only has a minor impact on 507 the results given that i) the model internal variability is not synchronized with that of the observations and ii) large volcanic eruptions are included in the periods considered. 509 The list of evaluated model variables and datasets against which they are evaluated are 510 presented in Table 1. 511

Thirty-two members have been performed for the *historical* period. Most of the diagnostics are not qualitatively sensitive to the choice of the member (when looking essentially at the mean state). We thus use only the first member (r1i1p1f1) in the diagnostics and illustrate the spread within the ensemble for some of the diagnostics. We present the most common variables used in climatology (like sea surface temperature,

surface air temperature, precipitation etc.) and concentrate on variables that are affected 517 by the coupling between LMDZ, NEMO and ORCHIDEE. Thus we do not repeat the 518 evaluation of variables that are close to those presented in LMDZ6A AMIP paper (Hourdin, 519 Rio, Grandpeix, et al., 2020). There have been numerous developments in the different 520 components of the model; tracing back the evolution of the biases to particular devel-521 opments requires a well-defined experimental framework (Bodas-Salcedo et al., 2019) and 522 additional simulations which we have not performed. For this reason we focus this study 523 on the evolution of the biases between the IPSL-CM5 and IPSL-CM6 models. The eval-524 uation starts with surface temperatures (Sea Surface Temperature and Surface Air Tem-525 perature) and follows with results for the atmosphere, the ocean and the sea ice. 526

3.1

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#### Sea Surface Temperatures and Surface Air Temperatures

We first evaluate sea surface temperatures (SST) simulated by the model keeping 528 in mind that the average SST between  $50^{\circ}$ S and  $50^{\circ}$ N was tuned to fit the observations 529 in the pdControl experiment. In this respect it should be noted that, towards the end 530 of the development process, a slightly negative bias in the model SST was deliberately 531 introduced, along with a tuning of some sea ice parameters, to partly compensate for a 532 negative bias in summertime sea ice volume. The overall biases in the SST (Figure 1) 533 have been significantly reduced between the IPSL-CM5A models and IPSL-CM6A-LR. 534 Part of the improvement is due to the fact that IPSL-CM5A-LR was inadvertently tuned 535 too cold. Nevertheless, the improvements are also clear when the mean bias is subtracted 536 (figures not shown). The North Atlantic negative anomaly around 45°N associated with 537 the position of the North Atlantic drift is slightly reduced with a value of -4.3 °C in IPSL-538 CM6A-LR, compared to  $-6.8^{\circ}$ C in IPSL-CM5A-LR (with the value taken as the min-539 imum temperature from the 60°W-15°W, 40°N-55°N box). For comparison this index 540 ranges from -6.8 to +2.0 °C (90% interval) with a median around -3.8°C in CMIP5 541 models, and ranges from -7.1 to  $+1.1^{\circ}$ C (90% interval) with a median around  $-3.7^{\circ}$ C 542 in CMIP6 models. We hypothesize that the increase in horizontal resolution (and the 543 better representation of the ocean topography that comes with it) together with the im-544 proved atmospheric circulation in LMDZ (notably the influence of the orography) have 545 contributed to improve the oceanic circulation and the resulting SST in the area. The 546 East Boundary warm biases have also been reduced in both extent and amplitude in IPSL-547 CM6A-LR subsequent to the improvements in the boundary layer humidity and stra-548

tocumulus clouds in those areas (Hourdin, Rio, Grandpeix, et al., 2020) and to a care-549 ful tuning of radiative fluxes. Yet this improvement is less clear in the tropical south At-550 lantic. In addition to global mean and latitudinal variation, the contrast between east-551 ern tropical basins and the rest of the tropical oceans was used as a target, considering 552 East Tropical Ocean Anomalies (ETOA) defined by Hourdin et al. (2015). A similar at-553 tention was given to the reduction of the classic latitudinal SST biases, which counter-554 acts a tendency of the model to produce too cold midlatitude SSTs and a warm bias close 555 to Antarctica (e.g., C. Wang et al., 2014). In the North Pacific, a warm bias (mostly dur-556 ing summer time) persists over the ocean in IPSL-CM6A-LR. This bias is much less vis-557 ible in Figure 1 (top and middle panels) because as indicated above, the CMIP5 versions 558 were globally too cold. Relative anomalies show that the North Pacific bias was already 559 present, although slightly weaker. This bias, robust to many tests which were conducted 560 during the tuning phase of the coupled model is also present in other CMIP5 and CMIP6 561 climate models (not shown). Its origin is still to be investigated. The cold bias over the 562 Equatorial Pacific, another classic deficiency of coupled models, is reinforced in IPSL-563 CM6A-LR. 564

Nevertheless, the SST biases against observations are altogether significantly reduced, even when comparing to IPSL-CM5A-MR which uses the same atmospheric horizontal grid as IPSL-CM6A-LR, with a reduction of the root mean square error (RMSE)
from 1.4 to 0.975 and an increase of the correlation coefficient from 0.986 to 0.988. It
is difficult to assess the statistical significance of such subtle changes. However it can be
noted that the correlation for IPSL-CM5A-MR falls outside of the range from the IPSLCM6A-LR ensemble members (0.9875 to 0.9885).

Consistent with the discussion above, there is a general reduction of the bias in sur-572 face air temperature (notably over the ocean) from IPSL-CM5A-LR to IPSL-CM6A-LR 573 (see Figure 2). Globally the increase in resolution has surely played a role as can be seen 574 by comparing IPSL-CM5A-LR and IPSL-CM6A-LR. However the difference between IPSL-575 CM5A-MR and IPSL-CM6A-LR is largely attributable to improvements in the model 576 physics and subsequent improvements in the radiative budget in LMDZ (Hourdin, Rio, 577 Grandpeix, et al., 2020) and a much more systematic and better tuning of the model key 578 parameters in IPSL-CM6A-LR in order to adjust the radiative fluxes. 579

Compared with IPSL-CM5A-MR, the warm bias over the Amazon basin and trop-580 ical Africa is reduced in IPSL-CM6A-LR. Cheruy et al. (2019) attribute the improve-581 ment to the reduction of the overestimation of the SW downward radiation at the sur-582 face. The cold bias over Asia (especially in winter) is stronger in IPSL-CM6A-LR than 583 in IPSL-CM5 models. The changes in the snow albedo in ORCHIDEE are likely to be 584 the cause of this amplification of the bias in comparison to IPSL-CM5A-MR. The new 585 snow scheme improves the realism of the physical properties of the snowpack (albedo, 586 density) in IPSL-CM6A-LR relative to IPSL-CM5A-MR. However, due to strong surface-587 atmosphere couplings, the larger value of the snow albedo appears to cancel out a pre-588 vious error compensation (Cheruy et al., 2019) and favours a too strong snow cover in 589 these continental areas. Part of this deficiency in summertime may be explained by the 590 fact that the parameterization of snow albedo does not account for shading effects in moun-591 tainous regions, a process which is thought to reduce the surface albedo on the scale of 592 a model gridbox. A strong negative bias is indeed observed over the Tibet including dur-593 ing summertime. A model development to account for the impact of orography on sur-594 face albedo is planned for a future model version. 595

In the Northern high latitudes the biases have also largely changed, due primar-596 ily to the revision of the boundary layer scheme which allows more decoupling in sta-597 ble situations. Modifications in the subgrid-scale orography parameters affecting the at-598 mospheric circulation, the sea ice model and the land surface scheme also contribute to 599 the change. The warm bias over the Northern part of Canada has been reduced in the 600 annual mean. There is actually some compensation of a warm bias in summer and a cold 601 bias in winter in IPSL-CM6A-LR that replaces a warm bias all year long in IPSL-CM5A-602 MR (Figure 2). The biases over the Arctic are linked to the position of the sea ice edge 603 and depend to some extent on the member being considered. However, a large warm bias is consistently simulated in winter over the Arctic. 605

Over inland Antarctica, a cold bias can be seen in IPSL-CM6A-LR surface air temperature in all seasons. This cold bias is likely to correspond to a warm bias diagnosed in the reanalysis surface temperature from a comparison with weather station data (Fréville et al., 2014; Jones & Lister, 2015) but the magnitude of this cold bias (up to 8 °C) exceeds the warm bias in ERA (up to 5 °C). In LMDZ6, the boundary layer scheme was indeed improved to match the temperatures observed at Dome C (Vignon et al., 2018). LMDZ5 tended to prevent the decoupling of the surface from the atmosphere in very sta-

ble conditions (Cheruy et al., 2019). To reach a good agreement with the observations, 613 the ice sheet albedo was also changed in IPSL-CM6A-LR following Grenfell et al. (1994). 614 Consistent with the reduction in SW radiation bias, the strong warm summer bias 615 in midlatitudes that was shared by many models participating in CMIP5 (Cheruy et al., 616 2014) is reduced in the CMIP6 version. However, it remains present in smaller areas, par-617 ticularly on the Southern Great Plains. In these regions the bias results from complex 618 interactions between the land surface and the atmosphere, especially through convec-619 tion (Koster et al., 2004). It is also likely that the lack of parameterization of propagat-620 ing mesoscale convective systems that are known to occur frequently in the region, con-621 tributes to this bias (Moncrieff, 2019). 622

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#### 3.2 Atmospheric variables

In this section we present the evaluation of a set of common atmospheric variables -namely surface precipitation and wind, temperature and atmospheric water on zonal mean diagnostics- and conclude with a set of evaluation metrics obtained with the PCMDI Metrics Package (PMP, Gleckler et al., 2016).

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#### 3.2.1 Atmospheric structure

The zonal mean temperature and zonal wind (Figure 3) show a decrease in the warm 629 bias over the Antarctic and of the cold bias at 200 mb in the polar vortex at both poles. 630 The cold bias at midlatitudes between 850 and 400 mb was present in IPSL-CM5A-LR, 631 vanished in IPSL-CM5A-MR but reappears in IPSL-CM6A-LR. The most striking im-632 provement is for the zonal atmospheric circulation (Figure 3, bottom row). The subtrop-633 ical jets used to be too close to the Equator in the two CMIP5 IPSL models. The dif-634 ference in resolution between IPSL-CM5A-LR and IPSL-CM5A-MR led to only slight 635 improvements. Despite the same horizontal resolution than IPSL-CM5A-MR, IPSL-CM6A-636 LR has a much better zonal circulation with jets moving polewards. This improvement 637 is mainly due to the changes in the physics of the atmospheric model and the increase 638 in vertical resolution from 39 to 79 layers. 639

The atmosphere is more humid than in previous models (Figure 4): the specific humidity in IPSL-CM5A (both LR and MR) used to be too low (i.e., corresponding to a dry bias) in the lower troposphere in the Tropics, and it is now slightly larger (i.e., cor-

responding to a wet bias) than in ERA Interim. In terms of relative humidity (RH), IPSL-643 CM6A-LR appears to be too saturated compared with ERA Interim between 30° and 644  $60^{\circ}$  in latitude (in both hemispheres). The wet RH bias in the free troposphere of the 645 midlatitudes was already present to some extent in the previous versions. This bias is 646 known to partly reduce with increasing horizontal resolution as illustrated by the com-647 parison of the IPSL-CM5A-LR and IPSL-CM5A-MR versions, as well as the compar-648 ison between the LR and HighResMIP horizontal grid in stand-alone atmospheric sim-649 ulations with the LMDZ6A version (Hourdin, Rio, Grandpeix, et al., 2020). The main 650 difference of the IPSL-CM6A-LR version compared to IPSL-CM5A-LR/MR is the much 651 wetter lower troposphere, at around 800 hPa. This change is related to the parametriza-652 tion of the boundary layer transport to the boundary layer top of the air evaporated at 653 the surface which is much more efficient with the thermal plume model in the IPSL-CM6A-654 LR version than with the old eddy diffusion scheme. This contributed to dry the near 655 surface air over the ocean, in better agreement with observation, but also resulted in a 656 moist bias in the lower troposphere when compared to ERA Interim. The subgrid-scale 657 distribution of total (vapour and condensed) water within a grid-box as a function of height 658 may also play a role in this. This wet bias should be put in relation with the increase 659 in equilibrium climate sensitivity (ECS) in IPSL-CM6A-LR relative to IPSL-CM5A-LR 660 and the diagnosed increased contribution of the water vapor feedback to the ECS (see 661 Section 6). 662

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#### 3.2.2 Surface precipitation

In terms of precipitation, biases are generally consistent between the three model versions (see Figure 5 for global maps), with the main changes concerning the Tropics (see Figure 6).

The Equatorial Pacific is dryer in IPSL-CM6A-LR, reinforcing a classic bias of cou-667 pled model, associated with the above-mentioned negative SST bias. This dry bias, par-668 ticularly strong over the Warm Pool, is probably one of the most negative aspect of this 669 new model version. Preliminary analysis indicates that it may be associated with reduced 670 surface evaporation as a consequence of the modification of boundary-layer mixing by 671 the thermal plume model in this region. In contrast, rainfall over the Maritime Conti-672 nent is strongly overestimated. This strong overestimation, also present in stand-alone 673 atmospheric simulations, seems to be related to parameters of the deep convection schemes, 674

in particular those taking different values over ocean and over land, such as the verti-675 cal velocity at the basis of convective clouds and the density of cold pools. Preliminary 676 analysis suggests that the improvement of the SPCZ seems to be related to the activa-677 tion of the thermal plume model, and a better representation of the shallow versus the 678 deep convective regimes. Meanwhile, the so-called double ITCZ issue, with overestimated 679 rainfall South of the Equator over the East Pacific, is less pronounced in the new ver-680 sion. The double ITCZ issue is sometimes associated to entrainment in convective clouds 681 (Oueslati & Bellon, 2013, 2015). The rainfall is altogether reduced over the eastern part 682 of tropical oceans due to the modification of the parameterization of stratocumulus clouds 683 and a careful tuning of the parameters that control precipitation in these clouds. 684

Rainfall is generally increased over semi arid regions like North India, Sahel, Australia, or around the Mediterranean Sea, in better general agreement with observations. Another major improvement of the new version is the reduction of the strong dry biases over the Amazon basin (as stated above).

The global precipitation rate is overestimated in the last version of the model, more 689 than in the previous versions. For CMIP5, the global rainfall was considered as a tar-690 get of tuning, and a strong effort was done to reduce the mean rainfall, which otherwise 691 was generally overestimated by the IPSL model, as is the case in most global climate mod-692 els. This target was intentionally abandoned for the tuning of IPSL-CM6A-LR, which 693 explains for a large part the overestimation by  $0.3 \text{ mm day}^{-1}$  of the global precipitation 694 rate (about 10% of the observed value). This positive bias in the global mean precip-695 itation is common to many other models. It cannot be excluded that this overestima-696 tion is partly due to an underestimation of the observed precipitation rate attributable 697 to an underestimation of light rain over the tropical oceans (Berg et al., 2010; Stephens 698 et al., 2012; Hourdin, Rio, Grandpeix, et al., 2020). 699

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#### 3.2.3 PMP large-scale summary statistics

This section provides a general synthetic view of the evolution of the climatology of the atmosphere of the IPSL models between CMIP5 and CMIP6. We have used the PCMDI Metrics Package (PMP) to calculate a set of large-scale performance metrics (Gleckler et al., 2008), also called summary statistics, to summarize the agreement between the climate simulated by the model over the recent period and a set of references (observations and reanalysis, as listed in Table 1). Figure 7 shows the results for the most common atmospheric variables: 2-meter air temperature (tas), surface precipitation (pr), precipitable water (prw), pressure at sea level (psl), upwelling shortwave (rsut) and longwave (rlut) radiation at the top of the atmosphere, cloud radiative effect at top of atmosphere on longwave (rltcre) and shortwave (rstcre) radiation, temperature, zonal and
meridional wind at 850 mb (ta850, ua850, va850) and at 200 mb (ta200, ua200, va200),
and geopotential height at 500 mb (zg500).

We display the results of the metrics using parallel coordinates plots, which has the 713 advantage to display raw results and avoid the necessary normalization of the portrait 714 plot (Gleckler et al., 2008). For the sake of readability, the variables are sorted to dis-715 play the results by increasing order of performance for the IPSL-CM5A-MR model. The 716 individual members of IPSL-CM6A-LR (blue lines) are grouped together, with no sin-717 gle ensemble member coming out of the pack. For the large majority of the metrics, the 718 results for IPSL-CM5A-LR (red line) and IPSL-CM5A-MR (green line) are out of the 719 spread of the IPSL-CM6A-LR ensemble, demonstrating a difference in climatology that 720 cannot be explained by internal variability. The Root Mean Squared Error computed 721 over the globe over the 12 months of the climatological annual cycle (Figure 7, top panel) 722 has decreased for all variables except for ta\_850 and zg\_500. For many variables (va\_200, 723 ua\_200, psl, va\_850, ua\_850, rsut, rlut, rstcre and rltcre), the error has considerably de-724 creased compared to IPSL-CM5A-MR and is within or at the bottom of the CMIP5 model 725 range. The global bias has not necessarily decreased for all the variables. The colder at-726 mosphere of IPSL-CM6A-LR compared with IPSL-CM5A-MR shown in Figure 3 explains 727 the higher negative biases for ta\_850 and ta\_200. For tas, IPSL-CM6A-LR is a little colder 728 than IPSL-CM5A-MR but still shows the benefits of a better tuning compared with IPSL-729 CM5A-LR (which was much colder). The global bias for the meridional wind at 200 mb 730 (va\_200) has also increased (it is more positive), when it is only slightly more negative 731 at 850 mb (va\_850), and much closer to zero for the surface meridional wind (vas). For 732 zg\_500 the bias has increased (it is more negative) due to a general reduction of the al-733 titude of the geopotential at this standard level, over the whole globe except the Antarc-734 tic (not shown). The global bias for the upwelling shortwave and longwave radiation at 735 the top of the atmosphere has slightly increased in absolute value (actually close to IPSL-736 CM5A-LR). It has not really changed for prw, uas and va\_850. The increase in the bi-737 ases for ta\_850 and zg\_500 partly explains the relatively larger RMSE for these variables. 738

The improvement is also striking when looking at the correlation coefficients (Figure 7, bottom panel) with all the variables experiencing higher correlations in terms of their annual mean patterns.

3.3 Oceanic variables

This section evaluates the model in terms of Sea Surface Salinity (SSS), the global vertical temperature profile, the structure of the Atlantic Meridional Overturning Circulation (AMOC) and Mixed Layer Depth (MLD), the meridional heat transport and a set of mass transports through key transects (Figures 8–15).

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#### 3.3.1 Sea Surface Salinity

The climatology of SSS shows many evolutions since IPSL-CM5A-LR (see Figure 8). 748 Overall, the SSS is globally reduced. This corresponds to a relative increase of the pre-749 cipitation in subtropical basins. In the Atlantic Ocean, this translates into a reduction 750 of the positive bias in the subtropics and an increase of the fresh bias in the subpolar 751 latitudes. A SSS decrease is also visible in the South Pacific Ocean and Southern Ocean. 752 The negative bias around Indonesia is corrected in IPSL-CM6A-LR in spite of an over-753 estimation of precipitation locally. This may be due to enhanced exchanges between the 754 Pacific and the Indian Oceans (see transport in Indonesian Throughflow in Table 2). The 755 North and tropical parts of the Pacific Ocean are a little saltier in IPSL-CM6A-LR com-756 pared with IPSL-CM5A-MR, which is consistent with the reduction in precipitation in 757 the area 758

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#### 3.3.2 Vertical profile of temperature

The vertical profile of temperature as a function of latitude has strongly evolved 760 between IPSL-CM5A and IPSL-CM6A-LR (Figure 9). Overall, biases are larger in IPSL-761 CM6A-LR as compared to IPSL-CM5A, except to the north of 60°N, where warm anoma-762 lies are present in all versions. IPSL-CM6A-LR exhibit negative temperature anomalies 763 in the Southern Ocean and globally below 1500 m and positive ones above, except near 764 the surface (see the discussion on SST in Section 3.1). Changes in the Southern Ocean 765 are presumably associated with an increase in the ocean ventilation around Antarctica 766 (Figure 11) and local negative surface air temperature anomalies in winter (Figure 2i). 767

The cold ventilated water masses penetrate the deep ocean globally down to a depth of 768 2000 m. Above, temperature anomalies are positive, reflecting the fact that the model 769 is globally warmer in IPSL-CM6A-LR compared to IPSL-CM5A. Furthermore, the model 770 presumably forms too much mode water, as found in many other climate models (Stouffer 771 et al., 2017). A cold bias is also visible in the subtropical surface waters, reflecting the 772 relatively cold SST (see Figure 1). Altogether, this can be interpreted as a stronger (weaker) 773 thermocline (surface) stratification in midlatitudes in IPSL-CM6A-LR. To what extent 774 this stronger stratification is an outcome of the tuning of the eORCA1 configuration used 775 here, or a robust characteristic of the mean state in IPSL-CM6A-LR given the other com-776 ponents of the climate model, remains to be clarified. 777

### 3.3.3 Structure of the Atlantic Meridional Overturning Circulation and Mixed Layer Depth

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We now turn to the oceanic general circulation and to the AMOC in particular. 780 The above-mentioned excessive thermocline stratification at midlatitudes translates into 781 a pinching of the upper limb of the AMOC in the Atlantic Ocean (Figure 12). Indeed, 782 at 26°N, the RAPID-WATCH observations suggest a maximum overturning around 1000 m 783 depth, while it is reached at 700 m depth in IPSL-CM6A-LR. In this respect, the ver-784 tical profile was more realistic in IPSL-CM5A configurations, but with a lower magni-785 tude. Note that all versions of IPSL-CM exhibit an underestimation of the AMOC max-786 imum at  $26^{\circ}$ N (by about 25% in IPSL-CM6A-LR), a bias that is common to many coarse 787 resolution climate models in the absence of overflow parametrization (Danabasoglu et 788 al., 2014). This may in part be explained by the difference in time period used in this 789 comparison (2004–2017 for the observations as compared to 1980–2005 for the models). 790 However it is more likely to be due to biases in precipitation in the North Atlantic and/or 791 the representation of overflows and western boundary currents, which remains a chal-792 lenge in climate modelling. 793

The AMOC profile at 26°N also illustrates the excessive volume of cold deep water masses that is apparent in Figure 9: the streamfunction changes sign at a depth of around 2800 m in IPSL-CM6A-LR versus 4500 m in observations. Notwithstanding, the strength of the deep overturning cell is realistic, and the latitudinal extent of that cell compares well with previous versions of the model (Figure 13). In particular, the 0-contour of the AMOC around 2500 m depth, is very horizontal at all latitudes, a characteristic

of all three model configurations. Above that contour, the positive AMOC cell, is max-800 imum around 40°N in IPSL-CM6A-LR, as in previous versions of the model. This max-801 imum reaches roughly 14 Sv in IPSL-CM6A-LR, which is notably larger than before. This 802 may be related to the fact that dense water production in IPSL-CM6A-LR is different 803 than in previous versions (Figure 10). In IPSL-CM5A, deep mixed layers are found south 804 of Iceland and south of Greenland, which was unrealistic. This bias in IPSL-CM5A is 805 associated to an over extended winter sea ice in the Labrador and Nordic Seas. In IPSL-806 CM6A-LR, deep mixed layers are confined to the Labrador Sea and the Nordic Seas, which 807 is close to observed locations. Still, when looking at other members of the historical en-808 semble, it appears clearly that there is substantial variability in the North Atlantic deep 809 convection in this model (Figure 10, the three panels of the bottom row). 810

Deep convection in the Nordic Seas may be directly related to the strengthening of the upper limb of the AMOC in IPSL-CM6A-LR north of 60°N (Figure 13, although the streamfunction at these latitudes were integrated along distorted model grid lines). The northward transport through the Barents opening is also more intense in IPSL-CM6A-LR compared to previous versions, and so is the return flow through the Fram Strait (Table 2). This can be interpreted as more intense exchanges between the North Atlantic and the Arctic, which is likely to affect sea ice there (see below).

In IPSL-CM6A-LR, deep convection in the Southern Hemisphere is also very in-818 tense, much more than in IPSL-CM5A models. There are important observational un-819 certainties related to the mixed layer depth estimations (Pellichero et al., 2017). How-820 ever, this convection is possibly overestimated in IPSL-CM6A-LR (Figure 11). This pro-821 vides cold water masses that invade the deep ocean and strengthens the meridional den-822 sity gradients in the Southern ocean, inducing a very strong Antarctic Circumpolar Cur-823 rent (Table 2, Drake Passage). This constitutes a major difference in barotropic stream-824 function –and in the overall horizontal circulation– between IPSL-CM5A and IPSL-CM6A-825 LR (Figure 14). 826

One of the major influence of the ocean to the global climate is through the meridional heat transport. This quantity is closer to observations in IPSL-CM6A-LR compared to previous versions (Figure 15a), which is a substantial improvement. However a strong convergence of heat at 40°S remains, a feature which was already present in IPSL-CM5A model versions but is likely to be unrealistic. Direct observations at that latitude are not

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available, but the common view is that the global meridional heat transport is southward in the whole Southern Hemisphere (Trenberth & Caron, 2001), which is not the case in our models. This seems to be related to the strong meridional gradient in density at that latitude, in particular in the Atlantic Ocean, a consequence of excessive mode water formation to the north, as described above. We also see an anomalous northward heat transport at 50°S in IPSL-CM6A-LR, presumably linked to the excessive dense water formation to the south.

In the Northern Hemisphere, the northward heat transport in IPSL-CM6A-LR is larger than in IPSL-CM5A versions. Notwithstanding, the simulated value remains slightly underestimated at the latitude where direct observations are available (24°N). Further north, it is very similar with observations, but this is due to an overestimated contribution from the Pacific Ocean (not shown). This might be partly responsible for the positive SST anomalies found in the north Pacific (Figure 1).

In the Atlantic Ocean, the meridional heat transport remains underestimated at all latitudes, particularly at tropical latitudes (Figure 15b). Still, this bias is much reduced in IPSL-CM6A-LR compared to IPSL-CM5A versions, which presumably contributes to reducing SAT biases over Europe and northern Africa (Figure 2).

<sup>849</sup> **3.4** Sea ice

Arctic sea ice was one of the targets considered during the tuning process (Hourdin, 850 Rio, Grandpeix, et al., 2020). We targeted around 20,000 km<sup>3</sup> of pre-industrial annual 851 mean Arctic sea ice volume and ultimately obtained slightly more, typically within a range 852 20,000-25,000 km<sup>3</sup>, much less than for IPSL-CM5A-LR, but a bit more than in IPSL-853 CM5A-MR. We also aimed for a seasonal cycle of ice coverage in our pdControl exper-854 iment that was broadly consistent with observations in both hemispheres. The Antarc-855 tic sea ice volume was not specifically considered during the tuning stage. Overall we 856 obtain a reasonably realistic simulation of sea ice, significantly improved, as compared 857 with IPSL-CM5A-LR. 858

In the Northern Hemisphere, there is less sea ice in IPSL-CM6A-LR than in both IPSL-CM5A models, which typically results in a better agreement with satellite data (Figure 16). The possible causes are a better tuning (Massonnet et al., 2018), higher model resolution, and more elaborated ice-ocean physics (Vancoppenolle et al., 2009; Uotila et al., 2017). Wintertime sea ice extent and area in IPSL-CM6A-LR slightly underestimate
satellite retrievals, but is still within observational uncertainty. Regionally, there is a lack
of winter sea ice in Okhotsk sea, associated with warm air temperatures, and less ice than
observed in Barents Sea. Summertime area and extent are generally lower than observed,
but are still within observational uncertainty. Excess summer ice decay occurs on the
Siberian Shelf.

Looking at both sea ice area and extent, the amplitude of the seasonal cycle appears to be on the high range. The annual mean volume and its seasonal cycle are within the rather wide observational range (Massonnet et al., 2018). There are noticeable simulated decadal fluctuations in sea ice volume.

In the Southern Hemisphere, IPSL-CM6A-LR overall improves over both previous 873 CMIP5 models, in particular in summer (Figure 17). Wintertime sea ice extent is over-874 estimated by 1-2 million square kilometers, sea ice area even more so. This points to the 875 classic high concentration bias of current sea ice models. Summertime extent and area 876 are within uncertainty range. As a result, the amplitude of the seasonal cycle of areal 877 sea ice coverage appears to be somewhat over-estimated. Sea ice volume varies between 878 5,000 to 25,000 km<sup>3</sup> with the season in the pre-industrial climate, which is much higher 879 than in our CMIP5 models. It is mostly wintertime sea ice that decreases in the 21<sup>st</sup> cen-880 tury. Summertime sea ice also decreases, but less clearly. 881

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#### 3.5 Model evaluation from a CMIP5 point of view

To complement the above evaluation of the model climatology, we now revisit the 883 recommendations for CMIP6 made by Stouffer et al. (2017) based on the results of a sur-884 vey made after the CMIP5 exercise. One of the main scientific challenges facing the cli-885 mate modelling community (first reported in Meehl et al. (2014)) is indeed to understand 886 "[...] the origins and consequences of systematic model biases". Stouffer et al. (2017) listed 887 six main long-lasting (across the various CMIP exercises) model biases from the survey 888 as major points for improvement: 1) the double ITCZ, 2) the Walker circulation, the dry 889 Amazon basin bias and tropical variability, 3) tropical and subtropical low clouds and 890 the East Boundary warm bias, 4) a too deep tropical thermocline, 5) too warm and too 891 dry continental surfaces during summertime, and 6) the position of the Southern Hemi-892 sphere subtropical jet. In this final subsection of the evaluation of the present-day cli-893

matology of the model we illustrate how some of these biases evolved between IPSL-CM5A
and IPSL-CM6A-LR in order to focus the evaluation of the model on identified problems for the CMIP community.

We make use of the diagnostics described in the previous sections along with a set 897 of large-scale evaluation metrics (Figures 18–21) presented in the context of other CMIP5 898 and CMIP6 climate models. Note that only CMIP6 models available on the ESGF at 899 the time of writing this study have been considered. All model outputs were regridded 900 to the same regular  $3^{\circ} \times 2^{\circ}$  resolution longitude-latitude grid before assessing global 901 mean biases and RMSE against observations. The model ranking shows a consistent but 902 varying improvement of IPSL-CM6A-LR over IPSL-CM5 for the metrics presented on 903 the Figures and discussed below. Some of these metrics have been considered during the 904 tuning process of the model, hence their improvement is expected. This is the case of 905 the radiative metrics (OLR, OSR, Figure 18, and the SW and LW cloud radiative effects, 906 Figure 19) even though the metrics used for the tuning are not exactly the same as those 907 presented here. Indeed it was not the RMSE on the seasonal cycle (considered as met-908 rics in the multi-model plots) which was used for tuning but rather its latitudinal de-909 pendency as well as the contrasts between East tropical oceans and the rest of the trop-910 ics (Hourdin, Rio, Grandpeix, et al., 2020). Most of the tuning procedure aimed at re-911 ducing the main regional SST biases. Thus the reduced RMSE on the SST is clearly an 912 outcome of the tuning process. 913

The rainfall and position of the jets were not directly considered as tuning targets 914 because the results were seen as reasonable enough from the beginning, but, if it would 915 not have been the case, some additional work or tuning would probably have been done 916 in this direction. Concerning the mean rainfall bias in version IPSL-CM6A-LR, we al-917 ready mentioned that it was abandoned as a target for tuning, explaining the increased 918 bias compared to IPSL-CM5A and IPSL-CM5B. However, the bias of IPSL-CM6A-LR 919 is only slightly larger than the averaged bias of CMIP6 models. The RMSE has slightly 920 decreased as a result of the combination of regional decreases and increases in the er-921 rors, as discussed above. 922

<sup>923</sup> Coming back to the six points listed by Stouffer et al. (2017), the following com-<sup>924</sup> ments can be made: 925

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Progress made on the double ITCZ, although not a target for tuning, is illustrated on the maps of annual mean precipitation climatologies on Figure 6 and on the scores shown on Figure 20. It can be seen that the southern branch of the double ITCZ in the eastern part of the Tropical Pacific basin (as well as in the Tropical Atlantic) has weakened in IPSL-CM6A-LR compared to the IPSL-CM5A models. The scores on Figure 20 (left panel) show the improvement of the Double ITCZ Pacific Index, with IPSL-CM6A-LR (in red) getting closer to the observed value. It must be noticed however that this improvement is accompanied by a reinforcement of another classic bias over the Pacific Ocean, consisting in a cold and dry tongue over the Equator that extends too far west toward the Maritime Continent. • Concerning the Walker circulation, the dry Amazon basin bias and tropical variability, we provide evidence for a reduction of the dry Amazon basin bias in Section 3.2.2 and Figure 20. We speculate that the improvement comes from a mix of better parameterizations of relevant local processes and a more realistic representation of the regional patterns of the radiative budget as teleconnections are known to influence precipitation in tropical South America (L. Yin et al., 2013). The East Boundary warm bias together with subtropical low cloud biases have also been reduced and are clearly one of the major improvements in IPSL-CM6A-LR (as noted in section 3.1). This reduction is linked for a large part to the improvement of the representation of cumulus and stratocumulus clouds in LMDZ (Hourdin et al., 2019b) and to a careful tuning of radiative and latent heat fluxes at the surface over tropical oceans.

Concerning the summertime warm bias over continents, IPSL-CM6A-LR shows almost no improvement against IPSL-CM5A-LR but a pronounced improvement against IPSL-CM5B-LR (see Figure 2, bottom row). More importantly the warm bias is much reduced in IPSL-CM6A-LR *amip* simulations for the two regions shown on Figure 21. It is thus clear that the lack of improvement between IPSL-CM5A-LR and IPSL-CM6A-LR is due to the general cold bias introduced by the tuning in the former version. The warm bias is reduced because of an improved –but still not perfect– shortwave radiative flux at the surface (Cheruy et al., 2014, 2019). Over the Southern Great Plains the complexity of land-atmosphere interactions together with the difficulty to represent the convective activity (Van Weverberg

et al., 2018) in relation with the absence of representation of the propagating convection in the present model (Klein et al., 2006) can explain the remaining bias. • The improvement to the position of the Southern Hemisphere subtropical jet is illustrated on Figure 21, with IPSL-CM6A-LR performing better than previous IPSL-CM5 model versions. The jets, which are located too close to the Equator in most CMIP models, are known to generally move polewards when the horizontal resolution is increased (Hourdin et al., 2013). This was clearly the case between the IPSL-CM5A-LR and IPSL-CM5A-MR model versions. However, IPSL-CM6A-LR –which has the same horizontal grid as IPSL-CM5A-MR– shows a much better location of those jets as seen in Figure 3 and, for the Southern Hemisphere, in Figure 21. We do not have a definite explanation so far, but it may be related to the much better tuning of the latitudinal dependency of radiative fluxes in IPSL-CM6A-LR, which in turn controls the thermal structure, itself tightly related to the zonal wind through the thermal wind balance.

#### **4 Modes of variability**

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We now turn to the main modes of variability of the model. We first present the 972 El Niño Southern Oscillation (ENSO) as it is the dominant coupled ocean-atmosphere 973 mode of variability, the new behaviour of the ocean multidecadal variability in IPSL-CM6A-974 LR and wintertime midlatitude variability and atmospheric blocking. We do not present 975 the atmospheric modes of variability defined with the leading Empirical Orthogonal Func-976 tions of dynamical variables like the North Atlantic Oscillation or the Pacific North Amer-977 ica pattern because they are presented in a separate study together with the role of orog-978 raphy parameterizations on those modes. 979

4.1 ENSO

The El Niño Southern Oscillation (ENSO) is the leading mode of interannual climate variability, emerging from air-sea interactions in the tropical Pacific, but with climate impacts worldwide due to atmospheric teleconnections (e.g., Timmermann et al., 2018). In particular, decadal modulation of ENSO results in decadal fluctuations of the global-mean surface temperature (GMST), a natural phenomenon that modulates anthropogenic climate change (e.g., Kosaka & Xie, 2013). It is thus a very important phe-

nomenon to represent in a global climate model. Previous studies have underlined char-987 acteristic biases in the representation of the tropical Pacific climate, that translate into 988 a misrepresentation of some key ENSO processes (e.g., Bayr et al., 2018). Such typical 980 biases include a too strong equatorial upwelling ("the cold tongue bias"), excessively dry 990 western equatorial Pacific, and the tendency to form a "double ITCZ". The cold tongue 991 and dry western Pacific biases have increased between IPSL-CM5A-LR and IPSL-CM6A-992 LR (Figure 1, top and bottom panels, Figure 5a top and bottom panels, Figure 22) but 993 the "double ITCZ" bias has been reduced, with a South Pacific Convergence Zone that 994 extends less into the eastern Pacific (Figure 6bd) 995

IPSL-CM5A-LR and IPSL-CM6A-LR reveal relatively similar ENSO pattern evo-996 lutions (Figure 22ef), with events that tend to start too early in spring and display west-997 ward phase propagation unlike in observations, and end too late the following year (Fig-998 ure 22def). The cold and dry equatorial biases in the mean climate result in SST, wind 999 and rainfall anomalies that are shifted west relative to those in observations, and too weak 1000 precipitation anomalies (e.g., Bayr et al., 2018). The amplitude of ENSO has increased 1001 by  $\sim 40\%$  between IPSL-CM5A-LR and IPSL-CM6A-LR (Figure 23), now being slightly 1002 above the observed value. One of the major issues of ENSO in IPSL-CM5A-LR was its 1003 seasonality (Bellenger et al., 2014). ENSO events indeed peak in boreal winter in obser-1004 vations, but IPSL-CM5A-LR tended to produce a maximum of equatorial Pacific SST 1005 variability in boreal spring (Figure 23a), due to its tendency to produce events peaking 1006 in spring in addition to those captured on Figure 22ef. This out of phase behavior has 1007 disappeared in IPSL-CM6A-LR, but the gap in amplitude between spring and winter 1008 ENSO signals remains too weak (Figure 23) and below the CMIP5 median. 1009

Overall, some mean-state biases thought to strongly influence ENSO representation have diminished (double ITCZ), while others have strengthened (cold tongue and dry equatorial biases). The major improvement in ENSO representation between IPSL-CM5A-LR and IPSL-CM6A-LR is a better representation of the ENSO seasonality, with other aspects of ENSO being quite similar in the two models.

#### 4.2 Multidecadal variability

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<sup>1016</sup> The climate variability at decadal to multidecadal timescales has strongly evolved <sup>1017</sup> in the latest version of the model (Figure 24 showing 500 years of the *piControl* simu-

lation of each of the model). The Atlantic Multidecadal Variability (AMV) index, de-1018 fined as the time evolution of the SST anomaly averaged between 0 and  $65^{\circ}$ N in the North 1019 Atlantic, seems to be dominated by a longer timescale in IPSL-CM6A-LR compared to 1020 both IPSL-CM5A versions (Figure 24, top panel). IPSL-CM5A-LR is indeed character-1021 ized by a marked bidecadal variability (Escudier et al., 2013; Ortega et al., 2015), also 1022 present, yet with weaker intensity, in IPSL-CM5A-MR (Wen et al., 2016). In the new 1023 model, the typical AMV timescale is much longer: successive peaks in the AMV index 1024 are separated by about 200 years. This bicentennial variability is very robust to small 1025 modifications in the oceanic code (not shown) but weakens towards the end of our 1200 1026 year-long piControl simulation. It should be noted that a similar feature is found in at 1027 least another CMIP6 model (CNRM-CM6, Voldoire et al., 2019) that shares the same 1028 ocean model as IPSL-CM6A-LR. Exact origin of this behavior is still under investiga-1029 tion. The spatial pattern of the AMV (Figure 25) exhibits a strong subpolar center of 1030 action and a relatively weaker tropical one as compared to observations. Note however 1031 that the AMV pattern in HadISST (Figure 25) was computed from the 1920–2016 pe-1032 riod and the global SST averaged between 60°S and 60°N was removed from all grid points 1033 before computing the North Atlantic average (0°N-60°N, 80°W-0°W) following (Trenberth 1034 & Shea, 2006). Hence this represents variability over a shorter and different period than 1035 the 500 years of the pre-industrial control and possibly still polluted by external forc-1036 ings in spite of the detrending. The AMV pattern in IPSL-CM6A-LR is also marked by 1037 a relatively clear teleconnection in the Pacific, with a pattern resembling a negative phase 1038 of the Interdecadal Pacific Oscillation (IPO) associated with a positive AMV phase, as 1039 in observations. Both IPSL-CM5A models failed to reproduce this teleconnection. 1040

The difference in the main timescale of variability is also found in the evolution of 1041 the AMOC maximum (Figure 24, middle panel). In IPSL-CM6A-LR, the AMOC has 1042 a predominant variability at centennial timescales, with peak-to-peak amplitude of al-1043 most 4 Sv. The same index has weaker variability, and predominantly over a shorter timescale, 1044 in IPSL-CM5A versions. The intensity of the Antarctic Circumpolar Current (ACC) mea-1045 sured at the Drake Passage is also different between IPSL-CM6A-LR and IPSL-CM5A 1046 versions (Figure 24, bottom panel). In IPSL-CM6A-LR, there is a marked periodicity 1047 with an 80-year timescale, with peak-to-peak amplitude of up to 15 Sv. Such a period-1048 icity is not visible in IPSL-CM5A models, although there seems to be a predominant vari-1049 ability at a similar timescale. The mechanisms leading to this variability are not yet fully 1050

understood. The AMOC centennial variability seems to be related to freshwater anomalies building up at very slow timescales in the Arctic Ocean and flushing into the North
Atlantic Ocean. The links between AMV, AMOC, and ACC variability in IPSL-CM6ALR remain to be investigated.

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#### 4.3 Wintertime midlatitude variability and atmospheric blocking

Figure 26a shows the frequency of wintertime blocked days in the IPSL models against 1056 observations. The envelope of blocking frequency from an ensemble of CMIP5 and CMIP6 1057 models is also reported. Blocking is defined estimating the reversal of the daily geopo-1058 tential height gradient at 500 hPa following D'Andrea et al. (1998). With respect to D'Andrea 1059 et al. (1998) here data is interpolated on a regular  $2.5^{\circ} \times 2.5^{\circ}$  grid, so that  $\Delta = 0^{\circ}, \pm 2.5^{\circ}, \pm 5^{\circ}$ 1060 and  $\Phi_n=80^{\circ}N$ ,  $\Phi_0=60^{\circ}N$ ,  $\Phi_s=40^{\circ}N$ . For a comprehensive review on blocking physics and 1061 climatology, the reader is referred to Woollings et al. (2018). It is particularly pertinent 1062 to analyse blocking frequency, because it has been a challenging phenomenon to repro-1063 duce for Numerical Weather Prediction (NWP) and global climate models alike for a long 1064 time. Davini and D'Andrea (2016) showed that there has been some improvement over 1065 generations of models, especially in the Pacific sector. In Europe, on the contrary, only 1066 a small number of models have blocking frequencies close to observed levels. This gen-1067 eral tendency of climate models is by and large confirmed for the CMIP6 generation (see 1068 the light orange and light blue bands of Figure 26a). IPSL-CM6A-LR simulates more 1069 blocked days than the two IPSL-CM5 models over Europe  $(0-30^{\circ}\text{E})$  in better agreement 1070 with observations, although the frequency of blocked days is still underestimated. In this 1071 region, IPSL-CM6A-LR remains in line with the average behavior of the other CMIP6 1072 models. There is a second maximum of blocking frequency at about  $70^{\circ}$ E, correspond-1073 ing to Ural blocking, that is largely overestimated with respect to observations and other 1074 CMIP5 and CMIP6 models. The Pacific sector is also slightly overestimated. 1075

In order to have a consistent understanding of the model behavior, Figure 26b–g gives an overview of the wintertime midlatitude variability of IPSL-CM6A-LR. Difference maps with IPSL-CM5A-LR are also shown Figure 26h–l. In the Atlantic sector, the midlatitude atmospheric jet is overestimated and too zonal, penetrating deeply into the European continent (Figure 26b and e) and carrying the Atlantic stormtrack along (Figure 26d and f). This brings about the underestimation of European blocking, and the overestimation of the Ural one. Over the Ural, excess low-frequency variability is con-
sistently found (Figure 26d and g). The tendency towards excessively zonal midlatitude
jets is linked to an underestimation of orographic drag (Pithan et al., 2016). In the Pacific sector the slight excess of blocking frequency is in agreement with a southward displacement of the jet (Figure 26b-c) and an excess cyclonic wave breaking (Rivière, 2009)
at high latitudes, as visible in the variability maps (Figure 26d-g).

Improvements with respect to IPSL-CM5A-LR are clearly visible. The overestima-1088 tion of the jet is much reduced in the new model (Figure 26h), which is consistent with 1089 the increase of blocking frequency in the Euro Atlantic sector. In IPSL-CM5 the jet is 1090 stronger and penetrates in the Eurasian continent slightly to the south with respect to 1091 IPSL-CM6A-LR. This causes larger low frequency variability in IPSL-CM5 (Figure 26j) 1092 in a region spanning the eastern Mediterranean to the low latitudes of the the Siberian 1093 region. At the same time the southward displacement of the jet explains the absence of 1094 an overestimation Ural blocking frequency. 1095

- 5 Simulations of the historical period
- 1097

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# 5.1 Simulation of global-mean surface temperature

As a reminder, the members of our ensemble of *historical* simulations have the same natural and anthropogenic forcings and differ only in their initial conditions which were sampled every 20 to 40 years in the *piControl* simulation. All *historical* simulations have been prolonged to 2030 using SSP245 forcings. Because of the large uncertainties in the observations before the 1880s, the analysis here is limited to the 1880–2018 period.

Figure 27 shows the time evolution of GMST (here computed from the surface air 1103 temperature), both in absolute terms and as an anomaly relative to the 1880-2018 pe-1104 riod. A large spread is present in both panels, with differences up to 0.75 K for a given 1105 year. The ensemble mean of the anomaly (Figure 27b) can be interpreted as the forced 1106 component of climate change (due to natural and anthropogenic forcings) with varia-1107 tions around it due to internal natural variability. We compare this anomaly to both the 1108 Cowtan and Way (2014) and Rohde, Muller, Jacobsen, Muller, et al. (2013); Rohde, Muller, 1109 Jacobsen, Perlmutter, et al. (2013) observational datasets. The observed GMST time 1110 series is within the spread of the ensemble simulations but the model ensemble mean qual-1111 itatively departs from observed changes around 1935–1945 and since 2005 (Figure 27b). 1112 The departure from observations for the recent period is slightly enhanced if the anomaly 1113

is computed from the 1850–1899 reference period (not shown). This large range of pos-1114 sibilities in the GMST evolution of the *historical* members is induced by their different 1115 initial conditions. It is not restricted to interannual variability as the long-term warm-1116 ing trends also depends on the *historical* member. The observed GMST response to the 1117 Pinatubo volcanic eruption is well represented by the model ensemble mean but there 1118 are large differences in the GMST evolution in the period around the Pinatubo eruption 1119 (i.e., 1990–1994) depending on the phasing of natural modes of variability in the sim-1120 ulations. 1121

Figure 28 shows the observed and simulated recent warming trends over the 1978– 1122 2018 period. Some members compares better to observations, e.g., with some degree of 1123 "warming hole" in the North Atlantic Ocean. There are however some discrepancies; in 1124 particular there is no member that reproduces the observed cooling trends in the South-1125 eastern Pacific and the Southern Ocean to their full extent. The model generally repro-1126 duces the land/sea contrast in warming, with an average ratio of 1.61 (ranging from 1.521127 to 1.79) between the global temperature over land and ocean over the 1978–2018 period 1128 compared to the observed ratio of 1.67 from the HadCRUT4 dataset (Morice et al., 2012) 1129 (the model data, SST over ocean and TAS over land, are regridded onto the observations 1130 temporally masked prior to the analysis). The Arctic amplification tends to be overes-1131 timated by the model, with an average trend of 0.88 K per decade over the  $70-90^{\circ}$ N re-1132 gion (ranging from 0.22 to 1.58 K per decade) relative to the global-mean trend of 0.26 K 1133 per decade (ranging from 0.16 to 0.36 K per decade), whereas the trend is about 0.79 K 1134 per decade over the 70–90°N region and about 0.19 K per decade for the global mean 1135 in the Cowtan and Way (2014) dataset. Further work is going on to assess the diversity 1136 of historical members of the IPSL-CM6A-LR model and their relevance against obser-1137 vations. 1138

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#### 5.2 Carbon fluxes

The historical simulations have prescribed  $CO_2$  atmospheric mixing ratio as per observations (Meinshausen et al., 2017). Global fluxes to the ocean and land can be estimated from the spatially resolved flux calculations of the NEMO-PISCES and ORCHIDEE models in response of atmospheric  $CO_2$  concentration and simulated climate (Figure 29 and Table 3). Compatible emissions are defined as the anthropogenic emissions that would be required to simulate the prescribed  $CO_2$  concentration if the carbon cycle were to be fully interactive in the model. These compatible emissions can be diagnosed from the following equation:

$$E_{\rm ff} = E_{\rm tot} - E_{\rm lcc} = G_{\rm atm} + S_{\rm ocean} + S_{\rm land} - E_{\rm lcc} \tag{1}$$

where  $E_{\rm ff}$  is the CO<sub>2</sub> emission flux from fossil fuel combustion and cement production,  $E_{\rm tot}$  is the total anthropogenic CO<sub>2</sub> emission flux,  $E_{\rm lcc}$  the CO<sub>2</sub> emission flux due to land cover changes (which is also estimated in the model),  $G_{\rm atm}$  the growth rate of atmospheric CO<sub>2</sub> concentration,  $S_{\rm ocean}$  the oceanic sink, and  $S_{\rm land}$  the terrestrial sink (not accounting for changes in land cover).

The ocean is a net sink of  $CO_2$  and that sink increases from near-zero in 1850 to  $\sim 2.9 \text{ PgC yr}^{-1}$  in 2018 with very little variability among the 32 *historical* members (standard deviation of  $\pm 0.07 \text{ PgC yr}^{-1}$ ). This simulated oceanic sink is consistent with the  $2.5\pm 0.6 \text{ PgC yr}^{-1}$  estimate from the Global Carbon Project for the 2009–2018 decade (Friedlingstein et al., 2019). Similarly, the simulated oceanic sink over the 1990–1999 decade  $(2.1\pm 0.04 \text{ PgC yr}^{-1})$  is very similar to the 2.2 PgC yr<sup>-1</sup> flux diagnosed in IPSL-CM5A-LR.

The net terrestrial flux remains broadly negative (i.e., a source to the atmosphere) 1152 until approximately 1970 due to land cover change effects. The flux then increases and 1153 the land becomes a sink due primarily to the increasing  $CO_2$  fertilisation effect that dom-1154 inates the land cover change effect. The net sink reaches  $1.5 \text{ PgC yr}^{-1}$  in the last decade 1155 of the historical period. It should be noted that, over the 1990–1999 decade, the simu-1156 lated sink is very close to that of the IPSL-CM5A-LR model  $(1.3\pm0.13 \text{ PgC} \text{ yr}^{-1} \text{ com-}$ 1157 pared to  $1.28\pm0.1$  PgC yr<sup>-1</sup>). The simulated net terrestrial sink is however less than the 1158 net flux of  $2.1\pm0.7$  PgC yr<sup>-1</sup> estimated by (Friedlingstein et al., 2019) for the 2009–2018 1159 decade. There is also a fairly large year-to-year variability due to climate variability at 1160 the regional scale (e.g., Schaefer et al., 2002), and a correspondingly large variability 1161 between the 32 ensemble members. The net terrestrial carbon fluxes,  $S_{\text{land}}$  -  $E_{\text{lcc}}$ , are 1162 consistent with the Global Carbon Project (Friedlingstein et al., 2019) estimates even 1163 though the  $E_{lcc}$  emissions are underestimated. This leads to simulated compatible emis-1164 sions within the range of estimated fossil fuel emissions and cement production,  $E_{\rm ff}$ , of 1165 the Global Carbon Project (Friedlingstein et al., 2019). 1166

# 6 Transient climate response and equilibrium climate sensitivity

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## 6.1 Estimates

Transient climate response (TCR) and equilibrium climate sensitivity (ECS) are two important quantities that characterize the model response to the  $CO_2$  radiative forcing. There is increasing awareness however that these quantities are not intrinsic properties to the climate system (or to a given climate model) but may depend on the climate state (Mauritsen et al., 2019; Rugenstein et al., 2020). Furthermore estimates of these quantities depend on the details of how they are estimated in a particular model.

The equilibrium climate sensitivity (ECS) is traditionally defined as the equilibrium global-mean surface temperature change for a CO<sub>2</sub> doubling. We follow Gregory (2004) to estimate an effective ECS by assuming a linear –or quasi-linear– forcing-feedback relationship between the anomalies of the net downward radiative flux at the top of the atmosphere  $\Delta R$  and the global mean surface air temperature  $\Delta T$  to a giving forcing F, and extrapolating  $\Delta T$  to its value for  $\Delta R = F + \lambda \Delta T = 0$ , where  $\lambda$  is the feedback parameter.

To calculate these radiative and temperature anomalies, we subtract the pre-industrial 1182 global-mean value of the net downward radiative imbalance and near-surface air tem-1183 perature ('rtmt' and 'tas', respectively, from the first 500 years of the pre-industrial con-1184 trol run, 1850-2350) from the respective radiative and temperature values from the *abrupt2xCO2* 1185 and abrupt-4xCO2 rliplf1 experiments. Results are sensitive to the length of the sim-1186 ulation because of the spatial and temporal dependence of the feedback parameters (Andrews 1187 et al., 2012, 2015; Knutti et al., 2017; Rugenstein et al., 2020). An ordinary least squares 1188 regression of the radiative imbalance on temperature anomalies results in equilibrium 1189  $\Delta T$  for CO<sub>2</sub> quadrupling of 9.05, 9.49 and 10.02 K depending on whether 150, 300 or 1190 900 years of simulation are considered (Table 4). This translates into an ECS of 4.75 K 1191 (95% confidence interval: [4.32, 5.20]) when the fit is performed on 300 years and a fac-1192 tor two is used to scale a quadrupling to a doubling  $CO_2$  as per the usual assumption 1193 of a logarithmic dependence of the  $CO_2$  radiative forcing on its atmospheric mixing ra-1194 tio. This corresponds to a 17% increase from the value of 4.06 K in IPSL-CM5A-LR (95% 1195 confidence interval: [3.73, 4.41]), estimated by the same method (Table 4). 1196

The factor two scaling from  $4 \times CO_2$  to  $2 \times CO_2$  can be questioned because: i) the 1197  $CO_2$  forcing is supra-logarithmic in the atmospheric  $CO_2$  concentration as shown by Zhong 1198 and Haigh (2013) and Etminan et al. (2016) for detailed radiative transfer models and 1199 Lurton et al. (2019) for our climate model; and ii) feedback parameters may depend on 1200 the forcing magnitude. In our model the effective radiative forcings (ERF) for  $CO_2$  dou-1201 bling and quadrupling are 3.46 and 7.53  $\mathrm{W m^{-2}}$ , respectively (Lurton et al., 2019). Renor-1202 malizing the surface temperature change extrapolated in the  $CO_2$  quadrupling with the 1203 estimated ERF values leads to a reduced ECS for a  $CO_2$  doubling of 4.35 K. This is to 1204 be compared to an ECS of 3.83 K if the regression is performed directly on the 300-year 1205 abrupt2xCO2 experiment. This last value may correspond better to the original defini-1206 tion of the ECS. 1207

The transient climate response (TCR) is defined as the temperature change at the 1208 time of  $CO_2$  doubling in an experiment where the  $CO_2$  atmospheric concentration in-1209 creases by 1% per year. More specifically it is computed as the twenty-year global av-1210 erage in the 2-meter surface temperature around the time of  $CO_2$  doubling (years 61 to 1211 80) in the 1pctCO2 experiment relative to the same quantity in the corresponding pe-1212 riod of the *piControl*. TCR amounts to 2.45 K in IPSL-CM6A-LR against published val-1213 ues of 2.09, 2.05 and 1.52 K in IPSL-CM5A-LR, IPSL-CM5B-LR and IPSL-CM5B-MR, 1214 respectively (Dufresne et al., 2013). Thus the larger ECS in IPSL-CM6A-LR also trans-1215 lates into a larger TCR in comparison to our previous generation of models. 1216

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#### 6.2 Differences in ECS between IPSL-CM5A-LR and IPSL-CM6A-LR

As discussed above, the effective ECS increases from 4.1 to 4.8 K between IPSL-1218 CM5A-LR and IPSL-CM6A-LR. The relative contributions to ECS are calculated fol-1219 lowing Dufresne and Bony (2008) and Vial et al. (2013) and illustrated in the bar plots. 1220 This method decomposes the contributions to ECS into i) rapid tropospheric and strato-1221 spheric adjustments to carbon dioxide and ii) temperature-mediated feedbacks operat-1222 ing on longer timescales. More specifically the rapid tropospheric adjustment includes 1223 the climate response associated with all tropospheric adjustments (temperature, water 1224 vapor, and clouds), surface albedo change, and the small land surface warming due to 1225 the  $CO_2$  forcing (Vial et al., 2013). The method also quantifies the relative contributions 1226 of the water vapor and temperature lapse rate, surface albedo, and cloud feedbacks. In-1227 dividual feedbacks are calculated by the radiative kernel method (Bony et al., 2006; So-1228

den et al., 2008; Shell et al., 2008). A radiative kernel acts as a partial derivative, rep-1229 resenting the sensitivity of the radiative flux to changes in a climate variable, such as wa-1230 ter vapor, temperature, and surface albedo. The radiative kernel is multiplied by the change 1231 in the climate variable of interest (i.e., water vapor) diagnosed from a model simulation 1232 and then normalized by the GMST change to produce the feedback value. We employ 1233 the same kernels as in Shell et al. (2008) for water vapor, temperature, and surface albedo. 1234 The cloud feedback is calculated as a corrected residual term, correcting for a cloud-masking 1235 term (Vial et al., 2013), which adds a consistent offset to net cloud feedback value es-1236 timated from the cloud radiative effect method (Andrews et al., 2012). A small resid-1237 ual term reflects nonlinearities in the relationship between radiative perturbation and 1238 the temperature response. 1239

The main drivers of this larger ECS in IPSL-CM6A-LR are more positive rapid tro-1240 pospheric adjustment to CO<sub>2</sub>, and a stronger combined lapse rate and water vapor feed-1241 back (Figure 30a). We diagnose the strong tropospheric adjustment from aqua-4xCO21242 and amip-4xCO2 simulations, as well as the abrupt-4xCO2 simulations, and find that 1243 the stronger adjustments come from clear-sky regimes (not shown). The stronger wa-1244 ter vapor feedback primarily results from strong moistening tendencies in weak ascent 1245 regimes around 500 hPa (Figure 30c). We diagnose this moistening tendency in weak 1246 ascent regimes by projecting the relative humidity anomalies, defined as the difference 1247 between relative humidity after 150 years of the *abrupt-4xCO2* simulation and the pi-1248 *Control*, into a circulation regime basis, wherein  $\omega_{500}$ , the vertical pressure velocity at 1249 500 hPa, acts as a proxy for the large-scale tropical circulation (Bony et al., 2004). This 1250 framework introduced by Bony et al. (2004) allows for attribution of changes in a cli-1251 mate variable to a given tropical circulation regime, ranging from strong ascent to strong 1252 subsidence regimes with increasing  $\omega_{500}$  values. Relative humidity anomalies reach up 1253 to 15% in these weak ascent regimes. However, it has also been shown that the IPSL-1254 CM6A-LR model is too moist in the tropical atmosphere compared with ERA-Interim 1255 data (see Figure 4) which suggests this moistening might be exaggerated as well. 1256

The net cloud feedback, in contrast, is less positive in IPSL-CM6A-LR than in the previous model version. Compensating positive and negative feedbacks in the tropics give rise to a less positive tropical cloud feedback. Plotted in Figure 30b is the spatial distribution of the net global cloud feedback, calculated from the kernel method, in  $W m^{-2}$ per K of GMST. A positive, warming feedback is in red, while a negative, stabilizing feed-

back is in blue. This feedback map demonstrates that the cloud response in IPSL-CM6A-1262 LR is spatially heterogeneous, with large swathes of the tropical ocean covered in pos-1263 itive or negative cloud feedbacks. To interpret the spatial discontinuity between the re-1264 gions of positive and negative cloud feedbacks, we project the net cloud feedback in the 1265 Tropics onto the  $\omega_{500}$  basis, analogous to what was done for relative humidity anoma-1266 lies. Based on the decomposition, the regions of positive cloud feedback can be linked 1267 to weak ascent regimes  $[-20, 0 \text{ hPa day}^{-1}]$  and moderate to strong subsidence regimes 1268  $[25, 100 \text{ hPa day}^{-1}]$  (Figure 30d). By contrast, negative net cloud feedbacks arise in deep 1269 convective regimes and a portion of weak subsidence regimes. Moreover, we divide the 1270 net cloud feedback into SW and LW components to see whether the SW or LW compo-1271 nent drives the net cloud feedback in particular regimes. In convecting regimes, the net 1272 cloud feedback more closely tracks the negative, LW cloud feedback, while in subsiding 1273 regimes, the net cloud feedback more closely follows the positive, SW cloud feedback (Fig-1274 ure 30d). The cloud feedback map shows that, geographically, positive values are found 1275 in regions of large-scale subsidence, which cover large parts of the tropical ocean and are 1276 associated with marine boundary-layer cloud such as stratocumulus and shallow cumu-1277 lus (Bony & Dufresne, 2005). By contrast, negative cloud feedback values occur in re-1278 gions of deep convection, such as the Western Pacific Warm Pool. A negative feedback 1279 also occurs over the Southern Ocean, which could result from phase changes or thermo-1280 dynamic changes with warming (Ceppi et al., 2016). 1281

- 1282 **7 Future scenarios**
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#### 7.1 Change in surface temperature

We now briefly present and discuss some results from the scenario simulations. The 1284 time evolution of the global-mean surface air temperatures are shown on Figure 31. The 1285 temperature change in 2100 relative to 1850-1900 is larger than  $2^{\circ}$ C in all scenarios ex-1286 cept the SSP119 where it overshoots 2°C before returning to below 2°C. It should be 1287 noted that the temperature change trajectory is very similar for all scenarios until circa 1288 2040 when it starts to diverge according to the emission trajectory. This highlights the 1289 long timescales associated with the carbon cycle and the climate system (Collins et al., 1290 2013). We also compare on Figure 31 the IPSL-CM5A-LR and IPSL-CM6A-LR mod-1291 els recognizing that the RCP and SSP scenarios are not fully equivalent as the repar-1292 tition of the total net radiative forcing between the different terms has changed (Lurton 1293

et al., 2019). IPSL-CM6A-LR shows more warming than IPSL-CM5A-LR for the high-1294 end scenarios (RCP245 / SSP245, RCP6.0 / SSP460, RCP8.5 / SSP585). This is expected 1295 from the larger TCR and ECS in IPSL-CM6A-LR. More surprising is the larger warm-1296 ing in IPSL-CM5A-LR for the historical period, which we attribute to a number of small 1297 differences in ERF. More specifically, the  $CO_2$  ERF is smaller (1.59 vs 1.83 W m<sup>-2</sup> in 1298 2015), and on the contrary the ERF of the non- $CO_2$  greenhouse gases (CH<sub>4</sub>, CFCs, N<sub>2</sub>O, 1299  $O_3$ ) is larger (1.58 vs 1.03 W m<sup>-2</sup> in 2015) in IPSL-CM5A-LR compared to IPSL-CM6A-1300 LR. The ERF for the anthropogenic aerosols is approximately the same for the two model 1301  $(\approx -0.6 \text{ Wm}^{-2})$ . Assuming that the climate feedback parameter and the ocean heat up-1302 take efficiency are the same in the *historical* and 1pctCO2 experiments, we can indeed 1303 expect more warming in IPSL-CM5A-LR for the historical period compared to IPSL-1304 CM6A-LR (1.61 vs 1.43 K) despite a smaller TCR (2.09 vs 2.45 K). 1305

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## 7.2 Distribution of temperature and precipitation changes

Figure 32 shows the distributions of changes in surface air temperature normalised 1307 by the GMST change, for the SSP126 and SSP585 scenario experiments, both for the 1308 end of the  $21^{\text{st}}$  century (2070–2100 period), and at the end of the  $23^{\text{rd}}$  century (2270– 1309 2300, extended scenario runs). The normalised changes are defined relative to a 100-year 1310 pre-industrial average. The patterns of change are quite similar for SSP126 and SSP585 1311 at the end of the 21<sup>st</sup> century. In contrast, at the end of the 23<sup>rd</sup> century, patterns dif-1312 fer more between the two scenarios: SSP126 shows an Arctic warming pattern quite sim-1313 ilar to that of 2100, whereas the relative warming for this region in the SSP585 scenario 1314 is less severe. However, SSP585 in 2300 shows an overall stronger warming in the South-1315 ern Hemisphere (if we average values on both hemispheres, we have a Northern Hemi-1316 sphere to Southern Hemisphere ratio of 1.13:0.87 for SSP585, versus 1.24:0.76 for SSP126), 1317 and its global patterns are more homogeneous than for the SSP126 scenario. The for-1318 mer shows more warming over the Southern Ocean while the latter exhibits a noticeable 1319 cold spot in the Southern Pacific Ocean. 1320

For precipitation, the patterns at the end of the 21<sup>st</sup> century are similar in both SSP126 and SSP585 scenarios, but they tend to differ more at the end of the 23<sup>rd</sup> century, with a somewhat smoother precipitation signature on some of the equatorial region for the SSP585 experiment (Figure 33). 7.3 Changes in sea ice

We observe a rather large response of sea ice to the  $21^{st}$  century anthropogenic forc-1326 ings, much larger than in IPSL-CM5A-LR. Summertime Arctic sea ice extent (Figure 16) 1327 responds more to changes in global mean temperature (Notz et al., 2020). The simulated 1328 loss rate per °C of global mean temperature change in IPSL-CM6A-LR ( $-3.39 \pm 0.87$ 1329  $\times 10^6$  km<sup>2</sup> K<sup>-1</sup>) has largely increased in comparison to IPSL-CM5A-LR (-1.48  $\pm 0.43$ 1330  $\times 10^{6} \text{ km}^{2} \text{ K}^{-1}$ ) and IPSL-CM5A- MR (-1.67  $\pm 0.87 \times 10^{6} \text{ km}^{2} \text{ K}^{-1}$ ). This is consis-1331 tent with the near-zero summer Arctic sea ice extent for all scenarios in IPSL-CM6A-1332 LR - a feature that is shared with the majority of CMIP6 models. It is also remarkable 1333 that winter sea ice almost disappears by 2100 in the fossil-fuel intensive scenario (SSP585), 1334 which some of the other CMIP6 models also predict. Possible causes for this greater sen-1335 sitivity, which should be further investigated, include the warm winter Arctic atmosphere, 1336 an ocean heat supply, changes in aerosol forcing, and ice drift. The ice volume loss starts 1337 in the early 20<sup>th</sup> century and accelerates in its last three decades of the century. This 1338 is followed by a steady decrease over the 21st century. In the Southern Ocean (Figure 17), 1339 it is mostly winter sea ice that decreases in the 21<sup>st</sup> century. Summer sea ice also decreases, 1340 but less clearly so. 1341

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#### 7.4 Changes in carbon fluxes

The land and oceanic net carbon fluxes for the scenarios are shown on Figure 29. 1343 The oceanic carbon uptake is projected to increase or decrease according to the scenario 1344 being considered, with a clear saturation occurring at large  $CO_2$  concentrations (e.g., SSP585 1345 and SSP460) and a decrease in the sink when the atmospheric  $CO_2$  levels off or decreases. 1346 The net land carbon uptake peaks at about  $3 \text{ PgC yr}^{-1}$  between 2020 and 2060 in all 1347 scenarios before returning to near-zero, or even negative, values. The downregulation of 1348 the maximum photosynthetic capacity that was implemented to account for the impact 1349 of nutrient limitation on the  $CO_2$  fertilization effect (see section 4.2) may overestimate 1350 the limitation effect when atmospheric  $CO_2$  concentration goes over 700 ppm (mainly 1351 after 2050) and thus may explain this extreme behavior at the end of the century. The 1352 formulation was chosen to broadly reproduce the change in gross primary production ob-1353 served at Free Air Enrichment experiment when CO<sub>2</sub> is doubled (FACE, Norby & Zak, 1354 2011) but we overlooked the responses at very high CO<sub>2</sub> concentrations. A new parametriza-1355 tion is being designed and implemented as an option in the model. 1356

#### 1357 8 Conclusions

We have described the main features of the IPSL-CM6A-LR climate model which 1358 has been developed at IPSL for CMIP6. We discuss the implementation of climate forc-1359 ings in the model in Lurton et al. (2019) and will discuss the development philosophy 1360 and methodology in a future paper. In comparison to the previous generation of IPSL 1361 model several improvements have been introduced to the model: more physically-based 1362 parameterizations (e.g., Hourdin, Rio, Grandpeix, et al., 2020), more realistic implemen-1363 tation of some forcings (e.g., stratospheric aerosols), and more systematic tuning of ad-1364 justable parameters with a view to simulate key aspects of the model's climatology (SST, 1365 AMOC, and Arctic sea ice). The IPSL-CM6A-LR model performance is significantly im-1366 proved over IPSL-CM5A-LR and IPSL-CM5A-MR and compares well to other published 1367 CMIP6 models for a number of metrics. However some systematic regional biases and 1368 shortcomings persist (e.g., double ITCZ, frequency of midlatitude wintertime blockings, 1369 and ENSO dynamics). 1370

The effective ECS (computed from a 300-year regression on *abrupt-4xCO2* and divided by a factor 2) increases from 4.1 to 4.8 K between IPSL-CM5A-LR and IPSL-CM6A-LR. The TCR correspondingly increases from 2.1 to 2.4 K. The increased ECS is due to increased contributions from tropospheric rapid adjustments and the combined lapse rate and water vapour feedback, which are only partly compensated by less positive cloud feedbacks.

A grand ensemble of 32 historical members has been performed with IPSL-CM6A-1377 LR. The global mean surface air temperature increase simulated by the model is in the 1378 range 1.1 to 1.6 K in 2014 relative to 1850–1899 (across the ensemble members). While 1379 the ensemble mean warms more than the observations, some members are more consis-1380 tent with observations. The IPSL-CM6A-LR shows a 1.6 to 6.8 K warming in 2100 across 1381 the scenarios relative to the same 1850-1899 period. The IPSL-CM6A-LR model exhibits 1382 a sea-ice response to 21<sup>st</sup> century climate forcings on the high range in comparison to 1383 other CMIP5 and CMIP6 models. 1384

A range of other papers in the Special collection further evaluate particular aspects of the IPSL-CM6A-LR model. A comprehensive assessment of the model will require a lot more work in the coming years. We expect this to take place in the context of the

CMIP6 multi-model ensemble on the basis of the vast amount of data we have published 1388 on the Earth System Grid Federation. 1389

#### Author contribution 1390

Both JS and OB coordinated the writing of this article. JD, CE and GM wrote the 1391 NEMO description section. FH wrote the LMDZ description section. PP, PC, AD, NVu 1392 and FC wrote the ORCHIDEE description section. OB, JD and OM wrote the model 1393 1394 coupling description section. GL, ArC and JG wrote the infrastructure description section. JS and FH performed the evaluation of the model climatology for atmospheric vari-1395 ables and wrote the corresponding sections. FC, JS and FH performed the evaluation 1396 of the model climatology for land surface variables and wrote the corresponding sections. 1397 JM, JS, JD and GG performed the evaluation of the model climatology for the oceanic 1398 variables and wrote the corresponding sections. MV performed the evaluation and wrote 1399 the sea ice section. JerV performed the evaluation and wrote the ENSO section. FDA 1400 and PD performed the analysis of atmospheric blockings. FH performed the performance 1401 analysis across the CMIP5/CMIP6 models. TL performed the  $CO_2$  radiative forcing cal-1402 culations. ANA conducted the ECS and feedback analysis, with tools developed by JesV 1403 and SaB. RB performed the analysis of the historical ensemble. PC performed the anal-1404 ysis of the carbon fluxes and PC, OB and NVu wrote the corresponding section. TL per-1405 formed the analysis and wrote the section on future scenarios. OB coordinated the IPSL-1406 CM6A-LR model development. All other authors contributed to the model development, 1407 the modelling infrastructure, the CMIP6 experiments, data processing and/or data dis-1408 tribution. 1409

- **Competing interests** 1410

1411

## The authors do not declare any competing interests.

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Variable full name	CMIP6 variable short name	Product	Period covered	Reference publication
2-meter air temperature	tas	ERA Interim	1980-2009	Dee et al. (2011)
3D temperature	ta	-	_	_
3D zonal wind	ua	_	_	_
3D specific humidity	hus	-	_	_
3D relative humidity	hur	-	_	_
Sea level pressure	slp or psl	_	_	_
10-meter zonal and meridional wind component	uas, vas	_	_	_
Air temperature at 850 and	ta_850,	_	_	_
200 hPa	ta_200			
Zonal wind component at 850 and	ua_850,	_	_	_
200 hPa	ua_200			
Meridional wind component at 850	va_850,	_	_	_
and 200 hPa	va_200			
Geopotential Height at 500 hPa	zg_500	_	_	_
Precipitation	$\mathbf{pr}$	GPCP	1979–2009	Adler et al. $(2018)$
Precipitable water	prw	REMSS-PRW-	01/1988-	Mears et al. $(2018)$
		v07r01	01/2019	
Longwave cloud radiative effect	rltcre	CERES-EBAF	2000-2012	Loeb et al. (2018)
Shortwave cloud radiative effect	rstcre	_	_	_
Upwelling shortwave at the top-of-	rsut	_	_	_
atmosphere				
Upwelling longwave at the top-of-	rlut	_	_	_
atmosphere				
Sea Surface Salinity	SOS	WOA13-v2	1975 - 2004	Locarnini et al. (2013)
Atlantic Meridional overturning	msftyz	Smeed et al.	2004-2017	Smeed et al. $(2017)$
streamfunction		(2017)		
Northward oceanic heat transport	hfbasin	Ganachaud and	1985 - 1996	Ganachaud and Wunsch
		Wunsch (2003)		(2003)

 Table 1.
 List of evaluated model variables and datasets against which they are evaluated.



Figure 1. Global distribution of the annually-averaged sea surface temperature (SST) bias (in <sup>°</sup>C) for IPSL-CM5A-LR (upper row), IPSL-CM5A-MR (middle row) and IPSL-CM6A-LR (bottom row). Biases are computed against data from the World Ocean Atlas (WOA13-v2, Locarnini et al., 2013). Global-mean biases, root mean square errors (RMSE) and correlation coefficients are provided for each model.



**Figure 2.** Distributions of biases in the surface (2-meter) air temperature (in °C) for the annual mean (ANM, top row), December-January-February (DJF, middle row) and June-July-August (JJA, bottom row) for IPSL-CM5A-LR (left column), IPSL-CM5A-MR (middle column) and IPSL-CM6A-LR (right column). The bias maps are computed against ERA Interim reanalysis (Dee et al., 2011).

Transect	IPSL- CM5A-LR	IPSL- CM5A-MR	IPSL- CM6A-LR	Observations
Barents opening	-0.89	-0.059	4.06	2.0
Bering Strait	1.09	1.13	1.17	0.8
Denmark Strait	-5.47	-5.73	-5.26	$-3.4{\pm}1.4$
Drake Passage	101.46	109.2	150.87	$136.7 {\pm} 6.9$
Fram Strait	0.009	-0.86	-3.59	$-2\pm2.7$
Indonesian Throughflow	-10.72	-11.17	-13.60	-15
Mozambique Channel	-27.96	-27.21	-23.22	$-16.7{\pm}8.9$

Table 2. Mass transports (in Sv) through a selection of key transects of the global ocean as defined in Griffies et al. (2016). In the three model configurations, the transports are computed as time averages over the period 1980–2005 of the historical simulations (r1i1p1f1 member). Mass transports are counted positively eastwards and northwards. Observations are from Griffies et al. (2016) and references therein.



Figure 3. Zonally- and annually-averaged air temperature (top row, in  $^{\circ}$ C) and zonal wind component (bottom row, in ms<sup>-1</sup>) for IPSL-CM5A-LR (left column), IPSL-CM5A-MR (middle column) and IPSL-CM6A-LR (right column). The black contours show the ERA Interim reanalysis (Dee et al., 2011) climatology and the model bias against ERA Interim is depicted by the color scale. The pressure on the vertical axis is expressed in Pa.

	Mean fluxes $(PgCyr^{-1})$							
	1990-199	9	2009-2018					
	IPSL-CM6A-LR	GCP 2019	IPSL-CM6A-LR	GCP 2019				
Emissions								
Fossil fuel $(E_{\rm ff})$	$6.5 \pm 0.15$	$6.4 {\pm} 0.3$	$9.1 {\pm} 0.13$	$9.5 {\pm} 0.5$				
Land cover change $(E_{lcc})$	$0.4{\pm}0.0$	$1.3 \pm 0.7$	$0.7 \pm 0.0$	$1.5 \pm 0.7$				
Total emissions $(E_{\rm ff} + E_{\rm lcc})$	$7.0 \pm 0.15$	$7.7 {\pm} 0.8$	$10.0 {\pm} 0.13$	$11.0{\pm}0.8$				
Partitioning								
Atmospheric growth rate $(G_{\text{atm}})$	3.2	$3.1 {\pm} 0.02$	5.2	$4.9 {\pm} 0.02$				
Oceanic Sink $(S_{ocean})$	$2.1 \pm 0.04$	$2.0 {\pm} 0.6$	$2.7 \pm 0.04$	$2.5 \pm 0.6$				
Terrestrial Sink $(S_{\text{land}})$	$1.7 {\pm} 0.13$	$2.6{\pm}0.9$	$2.2 \pm 0.14$	$3.6{\pm}1.0$				
Total land fluxes $(S_{\text{land}}-E_{\text{lcc}})$	$1.3 {\pm} 0.13$	$1.0 {\pm} 0.8$	$1.5 {\pm} 0.14$	$1.7 {\pm} 0.9$				

Table 3. Decadal mean components of the global CO<sub>2</sub> budget for the 1990–1999 and 2009–2018 periods for IPSL-CM6A-LR and from the Global Carbon Project (Friedlingstein et al., 2019). The GCP carbon budget shows an imbalance of 0.3 to 0.4 PgC yr<sup>-1</sup>. The error bars represent uncertainties for the GCP estimates and standard deviation across the 32 ensemble members for IPSL-CM6A-LR.



**Figure 4.** Zonally- and annually-averaged specific humidity (top row, in kg kg<sup>-1</sup>) and relative humidity (bottom row, in %) for the IPSL-CM5A-LR (left column), IPSL-CM5A-MR (middle column) and IPSL-CM6A-LR (right column). The black contours show the ERA Interim reanalysis (Dee et al., 2011) climatology and the model bias against ERA Interim is depicted by the color scale. The pressure on the vertical axis is expressed in Pa.



**Figure 5.** Global distribution of the annual mean precipitation biases (in mm day<sup>-1</sup>) for the IPSL-CM5A-LR (upper row), IPSL-CM5A-MR (middle row) and IPSL-CM6A-LR (bottom row) models. The bias maps are computed against the Global Precipitation Climatology Project (GPCP, Adler et al., 2018).



Figure 6. Annual mean precipitation rate (in mm day<sup>-1</sup>) for a) GPCP, b) IPSL-CM5A-LR,
c) IPSL-CM5A-MR and d) IPSL-CM6A-LR in the tropical region. The climatology is computed over the 1980–2005 period for the models and 1980–2009 for GPCP (Adler et al., 2018).




Figure 7. Global metrics summarizing the performance of the IPSL-CM6A-LR members of the historical ensemble (blue lines) against IPSL-CM5A-LR (red line) and IPSL-CM5A-MR (green line) and the CMIP5 multi-model ensemble (grey lines). The metrics are for the global monthly spatio-temporal root mean square error (upper panel), global annual mean bias (middle panel) and spatial correlation on the annual mean field (bottom panel) for 17 atmospheric variables (see full name of the variables in Table 1). The statistics are computed for the models on the 1980–2005 climatology against the reference datasets listed in Table 1. Note that the reference period for the observational datasets can be different from the period listed in Table 1, e.g. 1989–2009 for ERA Interim, 01/1979–04/2018 for GPCP, and 2000–2018 for CERES-EBAF. Each model is represented by a line that connects the values of the metric obtained for the different model variables (vertical axes). For readability the columns are sorted so that the line connecting the IPSL-CM5A-MR results goes up from the left to the right. The metrics were computed with the PCMDI Metrics Package (PMP, Gleckler et al., 2016).

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Figure 8. Global distribution of the sea surface salinity (SSS, in ‰) biases in annual mean for IPSL-CM5A-LR (upper row), IPSL-CM5A-MR (middle row) and IPSL-CM6A-LR (bottom row). Biases are computed against data from the World Ocean Atlas (WOA13-v2, Locarnini et al., 2013).

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**Figure 9.** Latitude-depth distribution of the global zonal mean ocean temperature bias (°C) for IPSL-CM5A-LR (upper row), IPSL-CM5A-MR (middle row) and IPSL-CM6A-LR (bottom row). Biases are computed against data from the World Ocean Atlas (WOA13-v2, Locarnini et al., 2013) over the 1955-2015 period.

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Figure 10. Annual monthly maximum of the Mixed Layer Depth (MLD, in m) in the North Atlantic for the reconstruction of de Boyer Montégut et al. (2004, upper left panel), IPSL-CM5A-LR and IPSL-CM5A-MR (upper middle and right panels), and for three IPSL-CM6A-LR historical members: the first member (r1i1p1f1, lower left panel), a member visually identified with a shallow mixed layer (r8i1p1f1, lower middle panel) and a member visually identified with a deep mixed layer (r12i1p1f1, lower right panel).



Figure 11. Same as Figure 10 but for the Southern Ocean. For IPSL-CM6A-LR, the lower left panel is also r1i1p1f1, the lower middle panel is the member visually identified with a shallow mixed layer (r3i1p1f1) and the lower right corner is the member visually identified with a deep mixed layer (r17i1p1f1).



Figure 12. Vertical profile of the meridional overturning streamfunction (in Sv) at 26°N in the Atlantic Ocean for IPSL-CM6A-LR (red line), IPSL-CM5A-LR (dark blue line), IPSL-CM5A-MR (light blue line) and the RAPID-WATCH observations (black line, Smeed et al., 2017).



**Figure 13.** Atlantic meridional overturning streamfunction (in Sv) for IPSL-CM5A-LR (top panel), IPSL-CM5A-MR (middle panel) and IPSL-CM6A-LR (bottom panel) as a function of depth and latitude, on average over the 1980–2005 period.



Figure 14. Barotropic streamfunction (in Sv) for IPSL-CM5A-LR (top panel), IPSL-CM5A-MR (middle panel) and IPSL-CM6A-LR (bottom panel).

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Figure 15. (a) Global and (b) Atlantic Ocean meridional heat transport (in PW) for IPSL-CM5A-LR (dark blue line), IPSL-CM5A-MR (light blue line), and IPSL-CM6A-LR (red line) over the 1980–2005 period and corresponding direct observations (black stars with error bars) from Ganachaud and Wunsch (2003).

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Figure 16. Time series of integrated sea ice diagnostics (area, extent -both in 10<sup>6</sup> km<sup>2</sup>- and volume -in 10<sup>3</sup> km<sup>3</sup>) over the Northern Hemisphere. Sea ice area is the integral of ice fraction within a given region – here the Northern Hemisphere. Ice extent is the area total area enclosed within the 15% sea ice fraction contour. Ice volume is the integral of ice fraction times thickness. IPSL-CM6A-LR simulations feature the *historical* r1i1p1f1 member in black, other *historical* ensemble members in grey (16-85% confidence interval), and selected scenario runs in color. The 16-85% confidence range is also shown for IPSL-CM5A-LR (blue) and IPSL-CM5A-MR (orange), for historical and RCP8.5 runs. Symbols depict passive microwave satellite-based retrievals from three different algorithms: Nasa Team (Cavalieri et al., 1996), Bootstrap (Comiso, 1996), and OSI-SAF (OSI-SAF, 1996). The upper and lower curves correspond to March and September, respectively.



Figure 17. Same as Figure 16 but for the Southern Hemisphere. The upper and lower curves correspond to the September and February months, respectively.



Figure 18. Biases (crosses) and root mean square errors (RMSE, closed circles) against observations of SST averaged over the 65°S-65°N latitudinal region (in °C), outgoing shortwave radiation (OSR, in W m<sup>-2</sup>) and outgoing longwave radiation (OLR, in W m<sup>-2</sup>) for CMIP5 models (in green) including IPSL-CM5A-LR and IPSL-CM5A-MR (in blue) and IPSL-CM5B-LR (in purple) and CMIP6 models (in black) including IPSL-CM6A-LR (in red). The averages of the scores for CMIP5 and CMIP6 models are shown as <CMIP5> and <CMIP6>, respectively. The models are ranked according to their RMSE. <CMIP5> is shown as a range through random sampling of an equivalent number of CMIP6 models. The RMSE is computed by computing the final reserved. seasonal cycle for the period 1979-2005 against SST from the input4MIPs dataset and radia-



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Figure 19. Same as Figure 18 but for the shortwave cloud radiative effect (SW CRE, in  $W m^{-2}$ ), longwave cloud radiative effect (LW CRE, in  $W m^{-2}$ ) and precipitation (PR, in  $mm day^{-1}$ ). The RMSE and biases are computed against the radiative fluxes from CERES-EBAF and precipitation from GPCP.



Figure 20. Same as Figure 18 but for the double ITCZ index (DI Pacific, in mm day<sup>-1</sup>),
Pacific precipitation (in mm day<sup>-1</sup>), and the Amazonian precipitation (in mm day<sup>-1</sup>). The DI
Pacific index is the annually averaged precipitation over the Southeastern Pacific Ocean in the
[150°W-100°W, 20°S-0°S] region (this diagnostics is performed on the native grid). The Pacific
precipitation is annually averaged over the [150°E-100°W, 40°S-40°N] region. The Amazonian
precipitation is annually averaged over the [65°W,50°W,15°S,0°S] region. The map shows the
multi-model averaged CMIP6 precipitation bias, which guided the choice of the regions to compute the indices.



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**Figure 21.** Left and middle panels: mean June-July-August (JJA) near-surface temperature biases (in °C) for *amip* (circles) and *historical* (plus signs) CMIP simulations averaged over a North American  $[105^{\circ}W-80^{\circ}W, 35^{\circ}N-50^{\circ}N]$  region and an Eurasian  $[23^{\circ}E-92^{\circ}E, 45^{\circ}N-58^{\circ}N]$ region. The map shows the multi-model averaged CMIP6 JJA near-surface temperature bias, which guided the choice of the regions to compute the indices. Right panel: same as Figure 18 but for the SH jet latitude. The bias is computed from the annually- and zonally-averaged nearsurface wind  $\overline{u_{10m}}$  as a weighted average of the latitude  $\phi$  over the  $[70^{\circ}S, 10^{\circ}S]$  latitudinal band:  $\int_{70^{\circ}S}^{10^{\circ}S} \max(\overline{u_{10m}} - 1, 0) \phi d\phi / \int_{70^{\circ}S}^{10^{\circ}S} \max(\overline{u_{10m}} - 80^{-1}, 0) d\phi$ . Fine KMSE by CONVERSE by CONVERSE by CONVERSE by CONVERSE by CONVERSE by the SALL of the SALL of the set of the region of the set of



Figure 22. Longitude-time section of the Pacific Ocean  $5^{\circ}S-5^{\circ}N$  average mean seasonal cycle of SST (colors,  $^{\circ}C$ ), zonal wind stress × 1000 (green contours, N m<sup>-2</sup>) and rainfall (black contours, mm day<sup>-1</sup>) for a) observations, b) IPSL-CM5A-LR, c) IPSL-CM5A-MR and d) IPSL-CM6A-LR. Longitude-time section of the  $5^{\circ}S-5^{\circ}N$  typical anomalies during an ENSO event: SST (colors,  $^{\circ}C^{\circ}C^{-1}$ ), zonal wind stress ×1000 (green contours, N m<sup>-2</sup>  $^{\circ}C^{-1}$ ) and rainfall (black contours, mm day<sup>-1</sup>  $^{\circ}C^{-1}$ ) for e) observations, f) IPSL-CM5A-LR, g) IPSL-CM5A-MR and h) IPSL-CM6A-LR. The typical ENSO anomalies are obtained as a lead/lag regression to the November-January averaged Niño3.4 SST anomalies. The average of the 6 available IPSL-CM5A-LR, 3 available IPSL-CM5A-MR and 32 available IPSL-CM6A-LR *historical* members is used. Observations (1980–2018) are from GPCPv2.3 (Adler et al., 2018) for rainfall and TropFlux (Praveen Kumar et al., 2012, 2013) for SST and zonal wind stress.



**Figure 23.** Seasonally-stratified standard deviation of the average Niño3.4 SST anomalies for observations (black), the 6 IPSL-CM5A-LR (dark blue), 3 IPSL-CM5A-MR (light blue) and 32 IPSL-CM6A-LR (red) members. The green shading indicates the range of values from the CMIP5 multi-model database (green curve = median, dark green = 25<sup>th</sup> to 75<sup>th</sup> percentiles, light green = 5<sup>th</sup> to 95<sup>th</sup> percentile). Observations are from GPCP v2.3 (Adler et al., 2018) for rainfall and TropFlux (Praveen Kumar et al., 2012) for SST over the period 1980–2018.

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**Figure 24.** Time evolution of the AMV index (top panel, in K), AMOC maximum (middle panel, in Sv) and Antarctic Circumpolar Current (ACC) measured as the mass transport through the Drake Passage (bottom panel, in Sv) in the IPSL-CM5A-LR, IPSL-CM5A-MR, and IPSL-CM6A-LR model versions. The AMV index is defined as the detrended 10-year low-pass filtered annual mean area-averaged SST anomalies over the North Atlantic basin (0°N-65°N, 80°W-0°E). The AMOC maximum is taken from the meridional streamfunction between 10°N and 60°N and below 500 m. The mass transport at the Drake Passage is integrated from the surface to depth between the Cape Horn and the western Antarctic Peninsula.



**Figure 25.** Pattern of the Atlantic Multidecadal Variability (AMV) defined as the regression (in  ${}^{\circ}C {}^{\circ}C^{-1}$ ) of the global SST on the AMV index defined as the 10-year low-pass filtered annual mean area-averaged SST anomalies over the North Atlantic basin (0°N-60°N, 80°W-0°E) for the HadISST observations (1920–2016 period), and the IPSL-CM5A-LR, IPSL-CM5A-MR and IPSL-CM6A-LR models. The SST weighted average between 60°N and 60°S was subtracted from each grid point prior to any calculation, in order to account for the global warming trend following Trenberth and Shea (2006).



**Figure 26.** Diagnostics of wintertime (DJFM) Norhtern Hemisphere midlatitude dynamics: a) 1980–2005 frequency (in %) of blocked days as a function of longitude in observations (black line: average between JRA-55 Reanalysis, NCEP/NCAR Reanalysis and ECMWF ERA-Interim Reanalysis), IPSL-CM5A-LR (dark blue line), IPSL-CM5A-MR (light blue line) and IPSL-CM6A-LR (red line). CMIP5 (shaded blue) and CMIP6 (shaded orange) multi-model ensemble spread is also shown as the ±1 standard deviation romative ensemble mean; b) **Y**<sub>9</sub>**S**<sub>0</sub>**-2MS**<sub>0</sub>**HeAHrights reserved**. wind (in m s<sup>-1</sup>) at 500 hPa for IPSL-CM6A-LR; c) stormtrack of IPSL-CM6A-LR, computed as



Figure 27. Time evolution of a) the annual global mean near-surface air temperature (GMST, in K) from the 31 *historical* members, prolonged using SSP245 forcings, of the IPSL-CM6A-LR model and b) the anomalies of GMST relative to the 1880-2018 average for the Cowtan and Way (2014, in blue) and the Berkeley (Rohde, Muller, Jacobsen, Muller, et al., 2013; Rohde, Muller, Jacobsen, Perlmutter, et al., 2013, in green) datasets, and for the model ensemble average (in black) of the individual *historical* members (in grey).



Figure 28. Trends in near-surface air temperature  $(K \text{ yr}^{-1})$  from the Cowtan and Way (2014) dataset (upper left panel) and from the *historical* members of the IPSL-CM6A-LR model for the 1978–2018 period. Members are classified (from left to right, and top to bottom) by increasing spatial root-mean-square error (RMSE) relative to the observations. The member number and the RMSE values are indicated on the top-left and bottom-right corners of each panel, respectively. Dotted hatching indicates grid boxes where trends are significant (Mann-Kendall test, p < 0.1).



Figure 29. Prescribed atmospheric  $CO_2$  mixing ratio (ppmv, upper left), and inferred "compatible" emissions (upper right), oceanic (lower left) and terrestrial (lower right) net  $CO_2$ fluxes (in  $PgCyr^{-1}$ ) in IPSL-CM6A-LR for the historical period (*historical* r1i1p1f1 in black and other ensemble members in grey) and the future scenario experiments (colored lines). A five-year running average is applied for the fluxes. Estimates from the Global Carbon Project (Friedlingstein et al., 2019) and their uncertainties are indicated with black circles and grey error

bars.

Quantity / Model	IPSL-CM5A-LR	IPSL-CM6A-LR
ERF $2 \times CO_2 (Wm^{-2})$	_	$3.50\pm0.27$
ERF $4 \times CO_2 \ (W m^{-2})$	$6.65\pm0.18$	$7.64\pm0.22$
$\Delta T 4 \times CO_2$ (900 years, K)	_	$10.02 \ (9.56, \ 10.62)$
$\Delta T 4 \times CO_2$ (300 years, K)	$8.12 \ (7.55, \ 8.74)$	$9.49 \ (8.65, \ 10.40)$
$\Delta T 4 \times CO_2$ (150 years, K)	$8.08\ (7.36,\ 9.90)$	$9.05 \ (8.05, \ 10.20)$
ECS from $4 \times CO_2$ (900 years, factor 2, K)	_	$5.01 \ (4.76, \ 5.28)$
ECS from $4 \times CO_2$ (300 years, factor 2, K)	4.06(3.78, 4.37)	4.75(4.33, 5.21)
ECS from $4 \times CO_2$ (150 years, factor 2, K)	4.04 (3.68, 4.45)	$4.53 \ (4.02, \ 5.10)$
ECS from $4 \times CO_2$ (300 years, scaled by ERF, K)	_	4.35
ECS from $4 \times CO_2$ (150 years, scaled by ERF, K)	_	4.15
ECS from $2 \times CO_2$ (300 years, K)	-	$3.83 \ (3.03, \ 4.88)$
TCR (K)	1.96 (2.09)	2.45

-10

**Table 4.** Effective radiative forcing (ERF, in W m<sup>-2</sup>), equilibrium global mean surface temperature change ( $\Delta T$ , in K), different estimates of the equilibrium climate sensitivity (ECS, in K) as derived from *abrupt2xCO2* and *abrupt-4xCO2* simulations using variants of the Gregory (2004) method, and transient climate response (TCR) for the IPSL-CM5A-LR and IPSL-CM6A-LR. ERF is calculated as in Lurton et al. (2019) by regressing the anomaly of the net radiative flux at the top-of-atmosphere against the anomaly in global-mean surface temperature using the 20 first years of the experiment. The anomalies are computed after substracting the *piControl* values year-by-year. The confidence intervals correspond to  $\pm 2\sigma$ . For IPSL-CM5A-LR, we also provide for reference in parenthesis the TCR value published by (Dufresne et al., 2013).



Figure 30. Diagrams supporting our analysis of the model's equilibrium climate sensitivity (ECS). a) Bar plot showing the relative contributions (in K) to the ECS of the stratospheric adjustment, tropospheric rapid adjustments, combined lapse-rate and water vapour (LR+WV), surface albedo and cloud feedbacks for the IPSL-CM5A-LR and IPSL-CM6A-LR models. The residual term is due to nonlinearities in the feedback terms. b) Anomalies in the tropical relative humidity (%) as a function of atmospheric pressure (hPa) and circulation regime as diagnosed by the vertical pressure velocity,  $\omega_{500}$  in hPa day<sup>-1</sup>. c) Distribution of the net cloud feedback (in W m<sup>-2</sup> K<sup>-1</sup>) for IPSL-CM6A-LR. d) Tropical net cloud feedback (in W m<sup>-2</sup> K<sup>-1</sup>) as a function of the circulation regime as diagnosed by the vertical pressure velocity,  $\omega_{500}$  in hPa day<sup>-1</sup>. The last two diagnostics are computed over the tropical ocean (30°S–30°S)



Figure 31. Upper panel: change in global-mean surface air temperature (GMST, in °C) relative to the 1850-1899 period in the *historical* r1i1p1f1 member (thick black line) and the other members (thin grey lines) and scenario experiments for the r1i1p1f1 member (thick colored lines) and other members (thin colored lines). Anomalies for 0, 1, and 2 °C are indicated for reference. Lower panel: change in GMST relative to the 1850-1899 period for the IPSL-CM5A-LR and IPSL-CM6A-LR models for the historical period and the 21<sup>st</sup> century.



Figure 32. Geographical distributions of the normalised change in near-surface air temperature (in  ${}^{\circ}C {}^{\circ}C^{-1}$ ) for the CMIP6 SSP126 (upper panels) and SSP585 (lower panels) *scenario* experiments, at the end of the 21<sup>st</sup> century (2070–2100 period, left panels) and at the end of the 23<sup>rd</sup> century (2270–2300, right panels) as simulated by the IPSL-CM6A-LR model. The temperature change is defined relative to the pre-industrial value (averaged over 100 years), and the normalisation consists in dividing the local temperature change by the global-mean surface air temperature change.



Figure 33. Same as Figure 32 but for the relative change in surface precipitation normalised by the global-mean surface temperature change (in % °C<sup>-1</sup>).

## IPSL-CM5A-LR r1i1p1 (vs WOA13-v2) 1980-2005

bias=-1.01 ; rmse=1.74 ; corr=0.984



## IPSL-CM5A-MR r1i1p1 (vs WOA13-v2) 1980-2005

tos, ANM

bias=-0.464 ; rmse=1.4 ; corr=0.986










## IPSL-CM5A-LR r1i1p1 (vs GPCP) 1980-2005

bias=0.0384 ; rmse=1.36 ; corr=0.772





## IPSL-CM5A-MR r1i1p1 (vs GPCP) 1980-2005

pr, ANM

bias=0.109 ; rmse=1.4 ; corr=0.769



## IPSL-CM6A-LR r1i1p1f1 (vs GPCP) 1980-2005

bias=0.342 ; rmse=1.45 ; corr=0.81





## PMP - Spatio-temporal RMS on global annual cycle



## PMP - Global annual mean bias



IPSL-CM6A-LR historical ensemble members

## PMP - Spatial correlation on global annual mean





IPSL-CM6A-LR r1i1p1f1 (vs WOA13-v2) 1980-2005

bias=-0.156 ; rmse=0.886 ; corr=0.721



-10



## pted Article

De Boyer Montegut et al. 2004







IPSL-CM6A-LR r1i1p1f1



IPSL-CM6A-LR low MLD member (r8)



IPSL-CM6A-LR deep MLD member (r12)



# pted Article











IPSL-CM6A-LR deep MLD member (r17)



IPSL-CM6A-LR low MLD member (r3)



IPSL-CM6A-LR r1i1p1f1



1350 1250



## **IPSL-CM5A-LR**





Barotropic Stream Function





Sv

IPSL-CM6A-LR

Barotropic Stream Function



## 



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Sea ice in IPSL-CM (Northern Hemisphere)



1960 1980 2000 2020 2040 2060 2080 2100 This article is appretected by copyright. All rights reserved.

### Sea ice in IPSL-CM (Southern Hemisphere)



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# ENSO seasonality





# HadISST

## IPSL-CM6A-LR piControl





1

0.5

0

-1

2

1

0

-1

# **IPSL-CM5A-LR piControl**



# IPSL-CM5A-MR piControl



1

0.5

-0.

-1 -1.1 -2 -2.1 -3 -3.1

3.5 3 2.5 2

1.5

1

0.5



a) Northern Hemisphere blocking frequency DJFM 1980-2005

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# a) Contributions to ECS



# b) Net cloud feedback

Tropical circulation regime ( $\omega_{500}$  hPa/day)







